Relative Influence of Stratospheric Aerosols on Solar and Longwave Radiative Fluxes for a Tropical Atmosphere[1]

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ABSTRACT

The solar and longwave radiative effects of a stratospheric aerosol layer between 18 and 22 km are compared for a tropical atmosphere. The changes in the daily mean solar and longwave radiative fluxes above and below the aerosol layer are computed for two particle size distributions and as a function of the albedo of the earth's surface. The changes in the solar and longwave fluxes above the aerosol layer are found to be comparable in magnitude. In the troposphere, the reduction in the incoming solar radiation (cooling) is several times greater than the increase in the downward longwave radiation (warming), the difference decreasing with increasing surface albedo.

1. Introduction

Concern about a potential large-scale increase in the concentration of aerosols due to man's activity has led to several studies of the radiative effects of aerosols. Several authors (e.g., Atwater, 1970; Rasool and Schneider, 1971; Mitchell, 1971; Yamamoto and Tanaka, 1972; Braslau and Dave, 1973; Chýlek and Coakley, 1974; Wang and Domoto, 1974; Reck, 1974, 1975) have shown that an increase of aerosols may cause either a cooling or a warming of the earth-atmosphere system depending upon the complex index of refraction of the aerosol and the albedo of the earth's surface. Other studies (e.g., Pollack and Toon, 1974; Herman, 1974; Coakley and Grams, 1974) have dealt specifically with an increase of stratospheric aerosols which may result from extensive aerospace operations in the stratosphere. Most studies have emphasized the effect of aerosols on the transfer of solar radiation and have neglected their effect on the transfer of longwave (infrared) radiation. Consequently, the relative influence of a stratospheric aerosol layer on solar and longwave fluxes is not adequately resolved.

The purpose of this research is to investigate the relative magnitudes of the solar and longwave effects of a stratospheric aerosol layer in order to determine the importance of the frequently neglected longwave effects. In this paper the diurnally averaged solar and longwave effects of a stratospheric aerosol layer are compared for different particle size distributions as a function of surface albedo. The results of a parameter study reported by Ellingson (1972) are used to estimate the longwave radiative effects, and a solar radiative transfer model similar to that of Braslau and Dave (1973) is used to compute the solar radiative effects. Since Ellingson's parameter study applies to a cloud-free tropical atmosphere with an aerosol layer uniformly distributed between 18 and 22 km, our results are limited to such a model atmosphere.

The Ellingson model, in addition to the water vapor continuum, includes the pure rotational water vapor band, the 6.3 μm water vapor band, the 15 μm carbon dioxide band, the 14 μm ozone band, the 0.6 μm ozone band, the 7.66 μm methane band, and the 7.78 μm nitrous oxide band. The spectrum between 0 and 2814 cm^{-1} (3.55 μm to ∞) is divided into 100 spectral intervals, and the multiple Mie scattering is included in the calculation.

The solar radiation model includes multiple Rayleigh and Mie scattering along with absorption by aerosol, ozone, water vapor, carbon dioxide and oxygen. The solar spectrum between 0.285 and 2.5 μm is divided into 83 spectral intervals, and the vertical column is divided into as many as 500 layers depending upon the optical thickness of the atmosphere.

2. Description of the aerosols

The stratospheric aerosols were assumed to be a water solution of sulfuric acid with a concentration of 75% H_2SO_4 by weight (Toon and Pollack, 1973). Two distinctly different size distributions were used to test the sensitivity of the results to the size distribution. Deirmendjian's (1969) haze H, given by a modified gamma function, is used to represent the size distribution of existing stratospheric aerosols, and Deirmendjian's haze M is used to represent a size distribution

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TABLE 1. Summary of input parameters for Ellingson's (1972) longwave radiation calculation.

Distribution	ω	τ	β
Haze H	0.05	0.005	0.5
Haze M	0.2	0.006	0.1

with larger particles than existing stratospheric aerosols. It is generally assumed that the aerosols which are formed from aircraft effluent in the stratosphere have a size distribution similar to that of the existing Junge layer.

The same stratospheric aerosol mass loading of 2.56 μg cm^{-2} was used for both size distributions. This yields a total of 5.00×10^7 particles in a centimeter square column of the aerosol layer for haze H and 3.175×10^6 particles in a centimeter square column of the aerosol layer for haze M. No tropospheric aerosols were included in the model.

The choice of aerosol mass loading was arbitrary, but it was based on two fundamental considerations. First, larger aerosol optical depths lead to greater precision for the longwave calculation. This occurs because the longwave effects are determined by interpolating from Ellingson's results. Second, aerosol optical depths much greater than 0.1 in the visible may lead to excessive multiple scattering in the aerosol layer. By computing the solar radiative effects for various aerosol optical depths, the change in the solar flux was found to vary almost linearly with aerosol optical depth for optical depths up to about 0.1 at a wavelength of 0.55 μm.

A complex index of refraction of $m=1.45-0.005i$ was assumed for the aerosol for the solar spectrum, and Remsberg's data (1973) for 75% H_2SO_4 were used for the longwave spectrum. The aerosol mass loading of 2.56 μg cm^{-2} corresponds to an optical depth of 0.109 for haze H and 0.034 for haze M at a wavelength of 0.55 μm. For comparison, other authors (CIAP Report of Findings, 1974, p. F-116) obtain aerosol optical depths between 0.084 and 0.097 for this mass loading while assuming different size distributions and radiative properties of the aerosols. These optical depths are approximately five times the estimate of Elterman *et al.* (1973) of the normal optical depth of the stratosphere or about half the increase of the optical depth in the Southern Hemisphere stratosphere due to Agung. Palmer and Williams (1975) reported values of the real and imaginary components of the index of refraction for sulfuric acid as a function of wavelength in the visible and near infrared, but their data appeared subsequent to these calculations. According to their measurements, the imaginary component of the index of refraction is on the order of 10^{-7} in the visible for a 75% solution of H_2SO_4, which suggests that the present calculations tend to overestimate the solar absorption by the aerosol particles.

3. Longwave and solar results

The input parameters for Ellingson's (1972) longwave calculation are wavelength-averaged values of the single scattering albedo of the particles (ω), the vertical optical depth of the aerosol layer (τ), and the fraction of incident radiation singly scattered into the backward hemisphere (β), as defined by Sagan and Pollack (1967). For a given set of parameters (ω,τ,β), one may calculate the upward and downward longwave fluxes at the top and bottom of the aerosol layer. Using Mie theory, values of ω and τ were computed at 1 μm wavelength intervals from 6 to 14 μm, and the values at 14 μm were assumed valid for wavelengths >14 μm. Mean values of ω and τ were then obtained by weighting the results by the longwave spectral flux. The mean values are estimated to be accurate to no more than one significant figure because of the assumed radiative properties beyond 14 μm. The values of β used were determined from the scattering phase functions for a wavelength of 10 μm. A summary of the input parameters is shown in Table 1.

The aerosol layer affects the upward longwave flux above the layer and the downward longwave flux below the layer. The values of the upward and downward fluxes and the change in the *net* downward longwave flux ΔF_{LW} (F_{LW}= downward flux − upward flux) at 17 and 23 km induced by the aerosol layer are shown in Table 2. Since F_{LW} is positive downward, an increase in F_{LW} implies a tendency to warm the atmosphere below that height. Thus, the results at 17 km show that a stratospheric aerosol layer tends to warm the troposphere. The results indicate that the longwave radiative effects of the stratospheric aerosol layer are not very sensitive to the size distribution of particles. The mean particle size is small compared to the wavelength. Hence the absorption cross section is dominant over the scattering cross section, and the absorption optical depth of the aerosols is proportional to the volume (or mass loading) of the particles.

Since Ellingson (1972) did not report the vertical profiles of pressure, water vapor density and ozone density which he used, vertical profiles given by McClatchey *et al.* (1971) for a tropical atmosphere were used for the solar radiation calculation. This calcula-

TABLE 2. The upward, downward and net longwave fluxes at 17 and 23 km (positive downward). All values in W m^{-2}.

		Upward flux	Downward flux	Net flux	Change in net flux
23 km	No aerosol	−288.9	9.1	−279.8	
	Haze H	−287.4	9.1	−278.3	+1.5
	Haze M	−287.4	9.1	−278.3	+1.5
17 km	No aerosol	−290.9	10.8	−280.1	
	Haze H	−290.9	11.6	−279.3	+0.8
	Haze M	−290.9	11.6	−279.3	+0.8

tion also used a solar constant of 1340 W m^{-2}. Daily average (24 h) solar fluxes were obtained by combining results for solar zenith angles of 30°, 60° and 80°. Computational limitations precluded using more increments in the solar zenith angle.

Daily average values of the change in the *net* downward solar flux $\Delta F_{\mathrm{solar}}$ (F_{solar}= downward direct+downward diffuse−upward diffuse) caused by the stratospheric aerosol layer as computed at 23 and 17 km are shown in Figs. 1 and 2, respectively. The results show a strong dependence on surface albedo and size distribution. Since F_{solar} is positive downward, a decrease in F_{solar} implies a tendency to cool the atmosphere below that height. The results shown in Fig. 1 indicate that in the solar spectrum, for sufficiently high values of surface albedo, a stratospheric aerosol layer tends to warm the earth-atmosphere system (i.e., the upward diffuse solar flux above the aerosol layer is decreased). For lower values of surface albedo there is cooling. Wang and Domoto (1974) obtained similar results for a global average calculation. They found that an increase in aerosols distributed within the troposphere and stratosphere decreases the global albedo (i.e., increases solar heating of the earth-atmosphere system) for a surface albedo >0.3.

The value of ΔF_{LW} at 23 km is also shown in Fig. 1. The longwave heating equals the solar cooling at a surface albedo of 0.11 for haze M and 0.35 for haze H. This indicates that there would be a net warming for surface albedoes greater than these values and a net cooling for smaller values. For comparable mass loadings, the larger particle haze M distribution is not as effective as the haze H distribution in rejecting solar radiation and cooling the atmosphere for small values of surface albedo. For a surface albedo of 0.3, which closely approximates the mean global albedo, the changes in the net solar and net longwave fluxes are comparable in magnitude and opposite in sign.

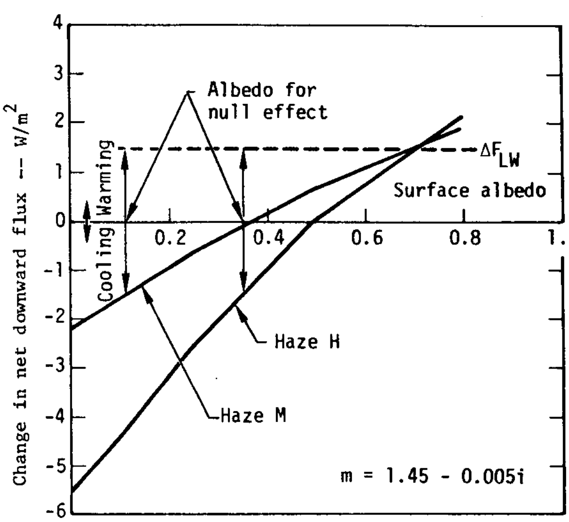

FIG. 1. Change in net downward fluxes at 23 km (24 h average) for aerosol mass loading of 2.56 μg cm^{-2}. ΔF_{LW} indicates the change in the net longwave flux.

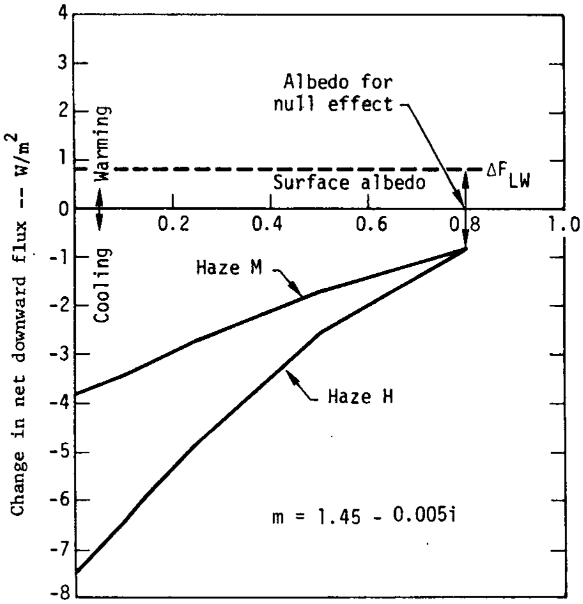

FIG. 2. As in Fig. 1 except at 17 km.

The change in the net solar flux below the aerosol layer (17 km) is shown in Fig. 2. The stratospheric aerosols cause a reduction in the net solar flux into the troposphere for all values of surface albedo considered. Consequently, the aerosol layer causes tropospheric cooling in the solar spectrum, even though the net solar flux above the aerosol layer may either increase or decrease. The aerosols scatter incoming solar radiation back to space, and there is increased absorption by gases and particles within the aerosol layer. This tends to increase the upward diffuse flux above the aerosol layer and tends to reduce the solar transmission into the troposphere. The aerosol layer also attenuates radiation scattered upward from the troposphere, thus allowing less of this radiation to escape to space. When the surface albedo is small, the effect of backscattering to space dominates. When the flux scattered upward from the troposphere is large, which is the case for a large surface albedo, attenuation of the upward flux dominates, thereby leading to a change in the sign of $\Delta F_{\mathrm{solar}}$ above the aerosol layer as noted above.

The aerosol-induced change in the net solar flux below the aerosol layer is greater than the change above the layer. Consequently, a stratospheric aerosol layer may have a significant effect on the tropospheric energy balance even though it has a small effect on the planetary albedo.

The value of ΔF_{LW} at 17 km is also shown in Fig. 2. In this case the longwave and solar effects balance at a surface albedo of about 0.8 for both haze H and haze M. The change in the net solar flux is several times greater than the change in the net longwave flux for small values of the surface albedo.

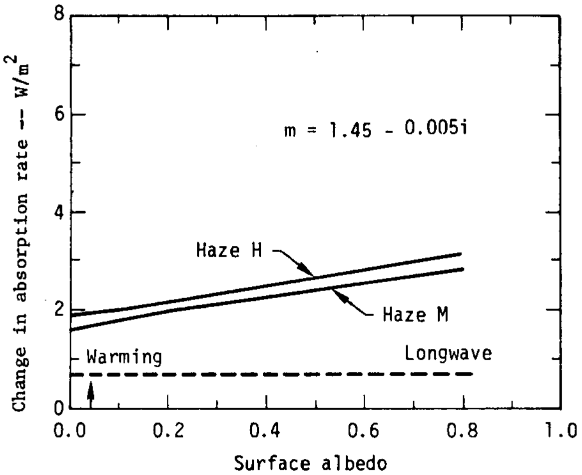

FIG. 3. Change in the absorption rate of the 17–23 km layer (24 h average) for aerosol mass loading of 2.56 μg cm^{-2}.

The change in absorption of solar energy within the aerosol layer is computed by taking the difference ΔF_{solar} (23 km) $-\Delta F_{\text{solar}}$ (17 km). This is shown in Fig. 3, a positive sign indicating a warming of the layer. The results are not as sensitive to size distribution or surface albedo as those shown in Figs. 1 and 2. The rate at which the aerosol layer warms or cools due to changes in the longwave fluxes is similarly computed by taking the difference ΔF_{LW} (23 km) $-\Delta F_{\text{LW}}$ (17 km). Here again, a positive sign indicates a warming of the layer. The change in the rate of absorption of longwave energy is also plotted in Fig. 3. This shows that a stratospheric aerosol layer tends to produce flux convergence and heating of the layer by both the solar and longwave radiation for a tropical atmosphere.

4. Discussion

The results indicate that the longwave radiative effects of the stratospheric aerosol layer are not very sensitive to the size distribution of the particles, assuming a constant mass loading. On the other hand, the solar radiative effects are sensitive to both aerosol size distribution and surface albedo.

For a tropical atmosphere, the presence of a stratospheric aerosol layer leads to a decrease in the upward longwave flux above the aerosol layer, thus tending to warm the earth-atmosphere system by reducing the longwave emission. The aerosol-induced change in the net solar flux above the aerosol layer may tend to either cool or warm the earth-atmosphere system depending upon the surface albedo. For small values of surface albedo, the changes in the net solar and longwave fluxes are comparable in magnitude and opposite in sign.

Longwave emission by the aerosol particles tends to warm the troposphere, while reduced solar transmission due to increased scattering and absorption within the aerosol layer tends to cool the troposphere. In the troposphere, the reduction in the incoming solar flux is several times greater than the increase in the downward longwave flux, the difference decreasing with increasing surface albedo. For a tropical atmosphere, the solar and longwave effects both tend to warm the aerosol layer, the solar effect being several times greater.

These calculations show that the longwave effects of aerosols cannot be neglected. Additional calculations are needed to quantify the solar and longwave effects of aerosols by season at other latitudes.

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