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EFFECT OF STRATOSPHERIC AEROSOL ON LONGWAVE RADIATIVE FLUXES:
VARIATION WITH LATITUDE AND SEASON

Frederick M. Luther

July 19, 1974

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EFFECT OF STRATOSPHERIC AEROSOL ON LONGWAVE RADIATIVE FLUXES:
VARIATION WITH LATITUDE AND SEASON

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Concern about the possible environmental impact of extensive aerospace operations in the stratosphere has led to increased interest in the climatic changes which might be caused by an increase in the concentration of stratospheric aerosol. The effect of a stratospheric aerosol layer on longwave radiative fluxes has been discussed for a tropical atmosphere (Luther, 1974). In this paper, we estimate the effect of increased stratospheric aerosol on longwave fluxes for summer and winter seasons at other latitudes.

The stratospheric aerosol is assumed to be a water solution of sulfuric acid with a concentration of 75% H_2SO_4 by weight (Toon and Pollack, 1973). Diermendjian's (1969) haze H given by a modified gamma function (Eq. 1) is used to represent the size distribution of the aerosol

$$n(r) = a r^2 \exp(-20 r) \quad (1)$$

We consider an increase in the stratospheric aerosol mass loading of $0.1 \mu\text{g cm}^{-2}$, which corresponds to an increase in optical depth of 0.005 at a wavelength of $0.55 \mu\text{m}$. Using Remsberg's (1971) data for the real

and imaginary parts of the index of refraction, the average change (weighted by the longwave spectrum in the region 6 to 50 μm) in optical depth in the infrared, $\Delta\tau_{\text{IR}}$, was found to be 0.00025.

We assume that the particles are completely absorbing in the infrared. The infrared transmissivity, t , and the infrared emissivity, ϵ , of the stratospheric aerosol layer are given by

$$t = e^{-1.66 \tau_{\text{IR}}} \quad (2)$$

$$\epsilon = 1 - t = 1 - e^{-1.66 \tau_{\text{IR}}} \quad (3)$$

To first order in the optical depth, these equations become

$$t = 1 - 1.66 \tau_{\text{IR}} \quad (4)$$

$$\epsilon = 1.66 \tau_{\text{IR}} \quad (5)$$

We assume that the stratospheric aerosol layer radiates as a black-body at the temperature of the surrounding gas, T . Then the infrared flux emitted by the aerosol layer and entering the troposphere is given by

$$F_{\downarrow} = \epsilon \sigma T^4 \quad (6)$$

Emission by the aerosol layer also contributes to the upward flux above the aerosol layer. We further assume that the longwave radiation leaving the troposphere is equivalent to that of a blackbody at the effective temperature T_e . The upward longwave flux above the aerosol layer is

$$F_{\uparrow} = \sigma T_e^4 (1 - \epsilon) + \epsilon \sigma T^4 \quad (7)$$

Considering a change in the infrared optical depth of $\Delta\tau_{IR}$, the changes in the longwave fluxes become

$$\Delta F_{\downarrow} = 1.66 \Delta\tau_{IR} \sigma T^4 \quad (8)$$

$$\Delta F_{\uparrow} = 1.66 \Delta\tau_{IR} \sigma (T^4 - T_e^4) \quad (9)$$

The change in the heating rate of the aerosol layer is

$$\Delta \dot{q} = -\Delta F_{\uparrow} - \Delta F_{\downarrow} \quad (10)$$

so

$$\Delta \dot{q} = 1.66 \Delta\tau_{IR} \sigma (T_e^4 - 2T^4) \quad (11)$$

$\Delta \dot{q}$ will be positive for $T_e > 1.19T$ (approx. 40°K difference in temperature). The various fluxes are shown schematically in Fig. 1.

FIGURE 1. Schematic diagram of longwave fluxes.

Using McClatchey's (1971) temperature data, values for $\Delta F_{\downarrow}$, $\Delta F_{\uparrow}$ and $\Delta \dot{q}$ were computed for tropical, midlatitude and subarctic atmospheres for summer and winter seasons. The temperature at 20 km was assumed for T and the temperature at 6 km (approximately 500 mb) was assumed for T_e . The global mean value for T_e is 253°K, which corresponds closely to the temperature at 500 mb. The results are shown in Table I.

TABLE I. Changes in longwave fluxes above and below a stratospheric aerosol layer for $\Delta\tau_{IR} = 0.00025$.

Quantity	Tropical		Midlatitude		Subarctic	
	This calc.	Ellingson	Summer	Winter	Summer	Winter
$\Delta F_{\downarrow} (W/m^2)$	0.43	0.40	0.53	0.51	0.60	0.49
$\Delta F_{\uparrow} (W/m^2)$	-0.71	-0.75	-0.56	-0.32	-0.36	-0.21
$\Delta \dot{q} (W/m^2)$	0.28	0.35	0.03	-0.19	-0.24	-0.28
T (°k)	207.0	unk.	218.0	215.2	225.0	214.1
T _e (°k)	264.0	unk.	261.0	243.7	253.0	234.1

The results for a tropical atmosphere using Ellingson's model (Luther, 1974) are also shown for comparison. The results for the Ellingson model are based on a parameter study for a tropical atmosphere (Ellingson, 1971) and are not applicable to other latitudes where the temperature profile is different. The temperature profile used by Ellingson for the parameter study is not known.

The results show an increase in the downward longwave flux into the troposphere, $\Delta F_{\downarrow}$, which is greater at higher latitudes and is larger in summer than in winter. The decrease in outgoing radiation, $\Delta F_{\uparrow}$, is smaller at higher latitudes, and the magnitude is larger in summer than in winter. The results for $\Delta \dot{q}$ show a warming of the aerosol layer at low latitudes and a cooling of the layer at high latitudes. Although the changes in longwave fluxes are small when compared to a global mean outgoing longwave flux of 233 watts/m^2 , they may not be negligible. In order to determine whether or not these changes are negligible, they must be compared to the corresponding changes in the solar fluxes which would result as well. The changes in solar and longwave fluxes only

indicate the initial tendency of the perturbation to warm or cool regions of the troposphere and stratosphere. Given the uncertainty as to how the atmosphere will respond to such a perturbation, one cannot accurately determine at this time what the new temperature profile will be.

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