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THERMAL AND RADIATIVE EFFECTS OF ATMOSPHERIC AEROSOLS IN THE NORTHERN HEMISPHERE CALCULATED USING A RADIATIVE–CONVECTIVE MODEL

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Abstract—A radiative–convective atmospheric model has been used to demonstrate that at all latitudes and at all atmospheric altitudes in the northern hemisphere, a low-lying aerosol (of mean visible refractive index 1.5–0.1 i) with an optical density greater than 0.065 (global background) will cool the surface when the underlying earth's surface has an albedo of 0.1 or 0.3. However, the model indicates that heating will occur at all latitudes over a surface with an albedo of 0.6 for an aerosol with a visible extinction coefficient 10 times that of the present background aerosol.

The reduction in both the incoming solar radiation and the atmospheric radiative cooling was also calculated. It was estimated for the latitudes south of 35°N that the present background level of atmospheric aerosols would produce ~1% decrease in solar radiation reaching the ground (relative to the case for no aerosols). The most significant relative heating of the atmosphere was found at latitudes north of 55°N. An aerosol-related mechanism which could extend the length of the winter season above 55°N is suggested.

1. INTRODUCTION

Recently there has been increased interest in the effects of atmospheric aerosols upon the temperature and the solar radiation near the earth's surface. Specifically: (1) The influence of the aerosol absorption/back-scatter ratio on the mean earth–atmosphere albedo has been considered, for example, by Reck (1974a) using a radiative–convective initial value model, for 35°N latitude, and by Chýlek and Coakley (1974), Ensor *et al.* (1971), Schneider (1971) and Atwater (1970) using mean-radiative-transfer models. The critical ratio which produces surface temperature change was found for each surface albedo, ω_s. Also in the Reck (1974a) paper a comparison was made between the present radiative-transfer model for 35°N latitude, April, and results of the earlier global mean radiative-transfer models of Chýlek and Coakley (1974), Ensor *et al.* (1971) and Atwater (1970). (2) Certain effects of aerosols on the mean surface temperature and climate were considered by Reck (1974, b, c), Rasool and Schneider (1971), Mitchell (1971), Budyko (1969) and McCormick and Ludwig (1967) using average global properties for the atmosphere. (3) Ground-based measurements of the absorption of solar radiation by aerosols have been made by Robinson (1962) in several locations and trends in the amount of atmospheric aerosols measured over the U.S.A. between 1961 and 1966 have been discussed by Flowers *et al.* (1969). (4) Calculations of globally averaged effects of aerosols on the amount of sunlight reaching the earth's surface have been made by Shotkin *et al.* (1973) and Shotkin and Thompson (1973). In addition, trends in mean global temperature changes with changes in solar radiation and turbidity (e.g. from volcanic activity) have been suggested by, among others, Bryson (1974), Reitan (1973) and Newell (1970).

The present work reports calculations of the temperature and solar radiation changes at nine different northern latitudes. It is part of a continuing effort to parameterize the thermal and radiative effects of aerosols in terms of their optical and physical properties using a one-dimensional radiative–convective atmospheric model. Previous calculations (Reck, 1974b) demonstrated the role of a lowlying aerosol layer in the radiative energy balance at 35°N latitude in April. In this paper those calculations are extended to a series of 10° latitudinal bands in the northern hemisphere for the month of April, to determine the manner in which the previous results vary with latitude. The latitudinal variation of temperature when no aerosols are present has already been shown by Manabe and Strickler (1964) for 5°, 45° and 85°N latitude, for the month of April.

Variations of the thermal contribution of a given aerosol layer might be expected due to nonuniformity in solar heating due to zenith angle variations with latitude and to geographic variations in the earth's albedo and absorptivity. For example, both the mean solar zenith angle and the height of the high and middle altitude water clouds decreases as the latitude changes from 35°N toward the north pole (Manabe and Wetherald, 1967). In the present calculations the aerosol properties (height, size distribution, thickness, refractive index) are taken to be the same as those used previously (Reck, 1974a) and are assumed independent of latitude.

2. DESCRIPTION OF THE PRESENT MODEL AND CALCULATIONS

The basic model is that developed by Manabe and Wetherald to calculate the temperature profile, assuming radiative–convective steady state with fixed relative humidity. The short wave absorption of the principal gases, CO_2, H_2O, and O_3, is calculated by a method equivalent to that of Manabe and Strickler (1964) and the long wave portion is based upon the computational scheme of Rodgers and Walshaw (1966).

A convective adjustment is made whenever the tropospheric lapse rate exceeds (larger negative value) the measured critical adiabatic rate (approx. -6.5 K km^{-1}). Otherwise local radiative equilibrium is assumed. The convective adjustment approximates the upward heat transfer due to atmospheric motions. With the inclusion of the convective adjustment there is more heat energy in the vicinity of the tropopause, thus obtaining more realistic temperature values there. An assumption of a steady-state requires that, at the top of the atmosphere, the net downward solar radiation flux must equal the net upward terrestrial (infrared) radiation flux. Also, at the earth's surface the imbalance between the solar and terrestrial radiation flux must equal the total radiative cooling of the atmosphere. An assumption of fixed relative humidity permits the outgoing infrared radiation flux to vary with temperature due to changes in the moisture content of the atmosphere. A comparison of the summer and winter data of relative humidity with latitude strongly suggests the atmosphere tends to retain a certain distribution of relative humidity (Manabe and Wetherald, 1967). The net turbulent heat transfer between the surface and the atmosphere also includes the exchange of lateral heat that results from maintenance of fixed relative humidity.

The computational scheme of the Manabe–Wetherald model may be briefly described as follows. An initial temperature profile is assumed and the radiative convective flux divergence at nine vertically aligned points is repeatedly calculated until a steady-state temperature profile is obtained. The nine points are located at pressures and altitudes for the standard global atmosphere as shown in Table 1. To simplify the calculations in this work the pressure at the surface of the earth is assumed to be 1000 mb at all latitudes. The actual mean altitude and mean sea level pressure for each latitude are given in Table 2.

Table 1. Measured average altitude for standard atmosphere

Level No.	Atmospheric pressure (mb)	Altitude (km)
1	9	32
2	78	18
3	188	12
4	336	8.4
5	500	5.6
6	663	3.4
7	811	1.9
8	925	0.75
9	991	0.1
	1000	0

It is into this Manabe–Wetherald-type model that we have introduced Sagan and Pollack* expressions (Sagan and Pollack, 1967) to describe the interaction of the suspended aerosol layer with the radiative field. The aerosol optical properties used are those described previously (Reck, 1974b) and are equivalent to those measured for airborne dust of the northern hemisphere (mean visible refractive index $m = 1.5$–0.1 i) with a $1/r^4$ radius, r size distribution in the range 0.1–10 μm.†

For the present calculations parameters characteristic of April at each latitude were obtained from Wetherald of Princeton University.‡ The time-averaged values of the cosine of the zenith angle, $\cos\theta$, are given in Table 3 (Manabe and Möller, 1961). Also shown are the fractional coverages of long-wave-reflecting clouds (which are taken to be equal to the short-wave cloud coverages). Three layers of clouds have been assumed, the heights of which are described in integers defining the Smagorinsky constant ϵ levels

Table 2. Mean surface elevation and mean sea level pressure*

Lat. (°N)	Elevation (m)	Mean sea level pressure (mb)
5	158	1009.7
15	146	1011.7
25	366	1014.8
35	496	1016.5
45	382	1015.7
55	296	1013.8
65	202	1014.5
75	220	1017.0
85	137	1020.0

* Mean elevation was taken from Sellers (1965); the mean sea level pressures were obtained from linear interpolation of the data assembled by Schutz and Gates (1973).

*Since the completion of this work, Coakley and Chýlek (1975) have presented a zenith-angle refinement of the Sagan and Pollack scattering approach used here. Using the Coakley-Chýlek solutions at 55°N latitude, the largest relative contribution is found for optical density (τ) = 0.065 and mean surface albedo (ω_s) = 0.1, for which the additional temperature contribution is -0.1 K.

† To determine the sensitivity of these results (Reck, 1974a) to an order of magnitude smaller imaginary refractive index (as suggested by the referee), additional calculations have been performed for 55°N latitude. Using the single-particle multiple scattering results of Deirmendjian (1969) for a mean visible refractive index, 1.55–0.0155 i (infrared portion unchanged), at a wavelength 0.589 μm, and for a size distribution, haze M, the largest relative contribution at this latitude was an additional -0.2 K for $\tau = 0.065$ at $\omega_s = 0.1$. The parameterization with respect to a variety of aerosol optical properties will be the topic of a later publication.

‡ Private communication.

Table 3. Cloud input data

Lat.	Cosine of zentih angle	Fraction of cloud cover Upper cloud	Middle cloud	Lower cloud	Cloud levels* Upper cloud	Middle cloud	Lower cloud
5	0.61	0.225	0.075	0.317	4–4	6–6	7–8
15	0.593	0.181	0.064	0.264	4–4	6–6	7–8
25	0.560	0.160	0.063	0.248	4–4	6–6	7–8
35	0.512	0.181	0.079	0.302	4–4	7–7	7–8
45	0.450	0.210	0.110	0.388	5–5	7–7	7–8
55	0.381	0.242	0.131	0.438	5–5	7–7	7–8
65	0.309	0.254	0.119	0.444	5–5	7–7	7–8

* Cloud levels indicate the location of the cloud. 4–4 is an infinitely thin cloud located at 336 mb; 5–5 is an infinitely thin cloud located at 500 mb; 6–6 is an infinitely thin cloud located at 663 mb; 7–7 is an infinitely thin cloud located at 811 mb; 7–8 is a thick cloud extending between 811 and 925 mb.

Table 4. Water cloud optical properties

	Upper cloud	Middle cloud	Lower cloud
Reflectivity of solar radiation by cloud layer	0.21	0.48	0.69
Absorptivity of solar radiation by cloud layer	0.005	0.02	0.035
Absorptivity of infrared radiation by cloud layer	0.50	1.00	1.00

as discussed previously (Reck, 1974b). Table 4 contains the assumed water cloud optical properties.

Starting at 5°N latitude in April, calculations have been performed for each 10° latitudinal band for a low-lying aerosol layer located between 925 and 991 mb. The particle density in the aerosol is assumed to vary in proportion to the pressure in this layer. Three values of mean aerosol optical density τ (extinction coefficient σ times path length l) in the visible have been considered (for 35°N the average path length is 0.65 km and extinction coefficient σ values are 0.1, 0.4 and 1.0 km^{-1}). These roughly correspond to the present background level and 4 and 10 times that amount (Porch *et al.*, 1970; Herman *et al.*, 1971). It is assumed that σ increases as the mean aerosol density increases. As the aerosols fill the space between levels eight and nine, and as the distance between levels eight and nine changes with latitude, the three optical density values considered have slightly different extinction coefficient values at each latitude. Values of the calculated aerosol optical properties are shown in Table 5. The radiative–convective calculation was performed for surface short-wave albedo values (ω_s) of 0.1, 0.3 and 0.6. In certain cases an unusually large number of computer iterations would have been required to obtain a steady-state solution by the forward time-step integration method, and hence no results are reported for those few cases. The same calculations without aerosols have also been made for each value of ω_s so that temperature and radiation differences could be obtained.

Figures 1–6 show the results of the calculations of changes (due to the addition of the aerosol to the atmosphere) which occur in the surface temperature, T_S (Fig. 1), the atmospheric temperature T (Figs. 2–3), the net solar downward radiation at the top of the atmosphere, SD_T (Fig. 4), the net solar downward radiation at the bottom of the atmosphere, SD_B (Fig. 5), and the radiative cooling of the atmosphere, RC (Fig. 6). The changes rather than the magnitudes of the quantities are reported since the same pressure (1000 mb) at the earth's surface was used for each latitudinal band (see Table 2). Nevertheless, the differences, ΔT_S, ΔT, ΔSD_T, ΔSD_B, and ΔRC as a function of latitude as reported in this paper should be reasonably accurate for actual pressures. Because of the finite difference method of this model, the combined effect of the cirrus (upper) clouds and the middle clouds may cause some of the smaller variations in the curves.

3. LATITUDINAL DEPENDENCE OF AEROSOL EFFECTS ON SURFACE-TEMPERATURE T_S

The difference in surface temperature ΔT_S due to the presence of aerosols is taken to be T_S (with aero-

Table 5. Aerosol optical properties*

τ	Short-wave α	Short-wave t	Short-wave r	Infrared α	Infrared t	Infrared r
0.065	0.01717	0.9560	0.02683	0.01338	0.9857	0.0009292
0.26	0.06686	0.8387	0.09445	0.05245	0.9440	0.003561
0.65	0.1579	0.6546	0.1876	0.1260	0.8658	0.00819

* α = Portion of flux absorbed, t = Portion of flux transmitted, r = Portion of flux back reflected.

sols)—T_S (without aerosols). The variation of ΔT_S with latitude is shown in Fig. 1. It is apparent that the sign of ΔT_S is independent of latitude. Cooling is observed over surfaces for which $\omega_s \leqslant 0.3$ (b, c), while minor heating is obtained for aerosol levels 10 times the present background value when $\omega_s = 0.6$ (a).

The calculations indicate that the magnitude of ΔT_S varies appreciably with latitude only for the intermediate value of $\omega_s = 0.3$ (b). For this surface albedo the behavior of ΔT_S for the latitudes 5–45°N differs from that for 45°N and above. For latitudes up to 45°N, the curves are relatively flat while above this latitude, there is an appreciable increase in the cooling effect of the aerosol for $\omega_s = 0.3$.

This dependence on ω_s could affect the seasonal changeover in the northern latitudes in the following way. Table 6 gives experimental values of the earth's surface albedo (Schutz and Gates, 1972, 1973, 1974). From the data in this table, it appears that ω_s approaches, or exceeds, 0.3 between the months of April and October only for the latitudes above 55°N. This seasonal data, coupled with the ω_s dependence shown in Fig. 1, suggests that an aerosol layer can produce increased cooling as $\omega_s \rightarrow 0.3$, from either lower or higher values. Hence in the more northern latitudes (as for, e.g. 65°N), aerosols could *delay* the seasonal changeover from spring to summer (ice-in to ice-out, ω_s decreasing $\rightarrow$ 0.3) and *accelerate* the changeover from summer to fall (ice-out to ice-in, ω_s increasing $\rightarrow$ 0.3). Such an effect could possibly account for the 1971–1972 season 10% increase in snow coverage in the north polar region (Kukla and Kukla, 1974; Reck, 1975), if aerosols in the northern latitudes could be shown to have increased. In this connection Hofmann and Rosen (1975) have stated that their *in situ* light scattering measurements at 85°N latitude suggest an apparent volcanic injection of aerosols in 1971 followed by slow decay during 1972 and 1973. We have noted that four volcanoes (Macdonald, 1972) are known to have erupted at the upper northern latitudes during 1970–1971, compared with the previous 10-y period where the average was one eruption per year. Also the refractive index in the present calculation lies within the range of values considered by Rosen and Hofmann (1974) for the interpretation of their data which indicates an increase in the aerosol abundance for 1971.

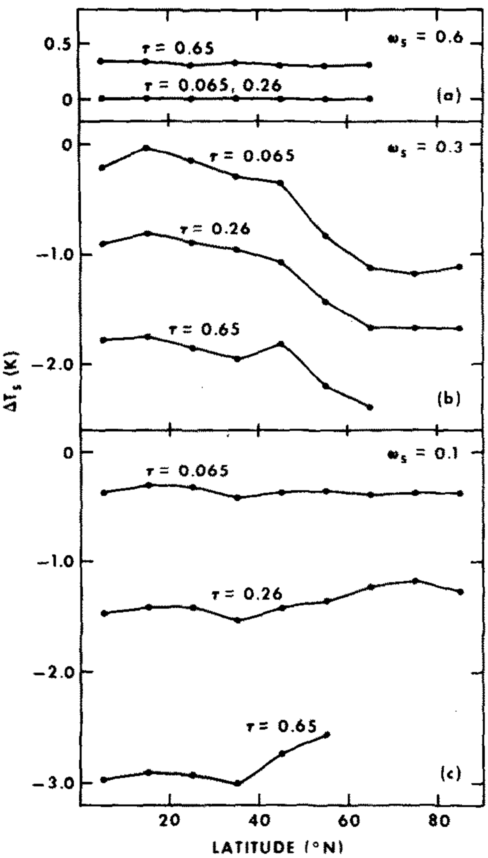

Fig. 1. (a) Difference in surface temperature [$\Delta T_S = T_S$ (with aerosols)—T_S (without aerosols)] over a surface having an albedo of $\omega_S = 0.6$, as a function of latitude, for three values of particulate visible optical densities τ, 0.65, 0.26, and 0.65; (b) Same as (a) but over a surface having an albedo of 0.3; (c) Same as (a) but over a surface having an albedo of 0.1.

4. LATITUDINAL DEPENDENCE OF AEROSOL EFFECTS ON THE VERTICAL TEMPERATURE PROFILE

We shall now turn our attention to the influence of the low-lying aerosol on the temperature profile of the earth's atmosphere. The effect is seen most readily by plotting ΔT (T(with aerosols)—T(without aerosols)) as a function of latitude for $\tau = 0.65$. In Figs. 2 and 3 are a series of curves showing this lati-

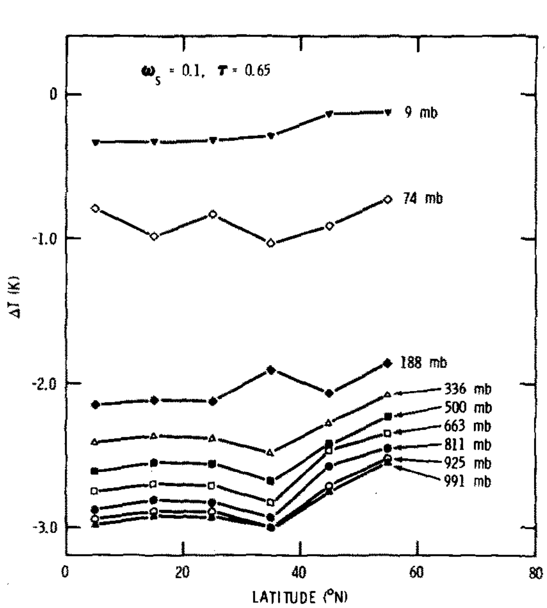

Fig. 2. Temperature difference [$\Delta T = T$ (with aerosols) − T (without aerosols)] for $\tau = 0.65$ as a function of latitude for atmospheric levels corresponding to pressures of 991, 925, 811, 663, 500, 336, 188, 74 and 9 mb, over a surface having mean short-wave albedo $\omega_s = 0.1$.

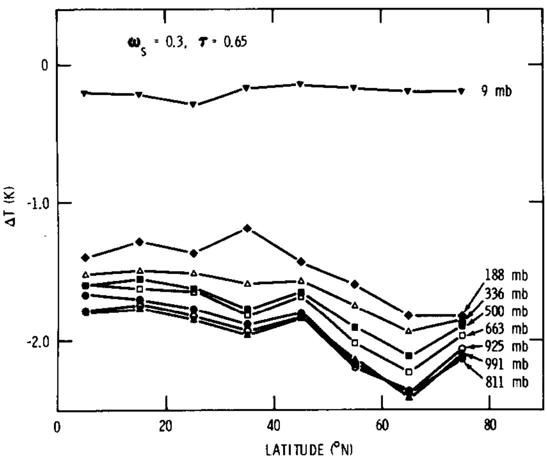

Fig. 3. Same as Fig. 2 except over a surface having an albedo of $\omega_s = 0.3$.

tudinal dependence at each of nine vertically aligned points in the atmosphere over a surface albedo of 0.1 and 0.3. The height of each point is defined in terms of pressure (mb).

In general, the magnitude of the thermal effect decreases with altitude, i.e. with decreasing pressure. Between the 991 and 336 mb levels, the curves are relatively flat from 5 to 35°N latitude. From 35 to 65°N the thermal effect decreases over $\omega_s = 0.1$ and increases over $\omega_s = 0.3$. The ΔT magnitudes over surfaces with $\omega_s = 0.1$ are always larger than those calculated for $\omega_s = 0.3$, except at the 9 mb level, where thermal effects are comparable. As was the case for ΔT_S, the larger latitudinal effect is found for $\omega_s = 0.3$. For $\omega_s = 0.6$, all values are between $\Delta T = 0$ and $\Delta T = 0.4$°C with no specific variations with latitude (Reck, 1974c).

5. LATITUDINAL DEPENDENCE OF AEROSOL EFFECTS ON INCOMING SOLAR RADIATION

In earlier calculations for 35°N latitude in April (Reck, 1974c) it was found that the presence of an aerosol layer increases the total earth atmosphere short-wave albedo slightly (i.e. causes a reduction ΔSD_T in the net downward solar flux at the top of the atmosphere). The present work calculates the effect of aerosols on the latitudinal dependence of the change in the net downward solar flux at the top of the atmosphere (ΔSD_T) and at the earth's surface (ΔSD_B) (Figs. 4 and 5).

The net reduction in solar radiation at the bottom of the atmosphere is always greater than at the top. Furthermore, the largest reduction is seen at the latitudes nearest the equator. The largest values of solar flux decreases are found for $\tau = 0.65$ (10 times that for present background) at 5–25°N latitude with $\omega_s = 0.1$. In that case the reduction corresponds to $\sim 10\%$ of the incoming solar radiation (see Table 7). The latitudinal dependence of the flux change caused by particles is most dramatic at latitudes ≥ 35°N;

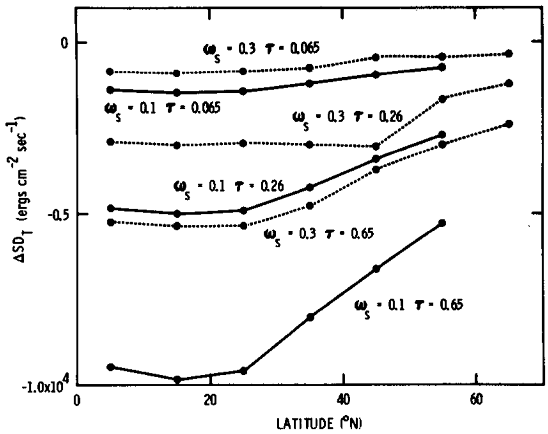

Fig. 4. Difference in net incoming solar radiative flux at the top of the atmosphere $\Delta SD_T[SD_T$ (with aerosols) $- SD_T$ (without aerosols)] as a function of latitude over a surface having $\omega_s = 0.1$, for $\tau = 0.065$, 0.26 and 0.65, and over a surface having $\omega_s = 0.3$ for the same values of τ.

however, on a percentage basis it is essentially constant.

Estimates from scattering measurements show that the natural background level of aerosols on a worldwide basis has a scattering optical density of $\tau = 0.065$ (Porch *et al.*, 1970). Such an optical density is shown in this work to cause a $\sim 1\%$ decrease in solar radiation reaching the ground at all latitudes below 55°N (see Table 7 and Fig. 5) compared with an atmosphere with no aerosols present. (Of course, the zero level is physically unrealistic since volcanoes sporadically add aerosols to the atmosphere).

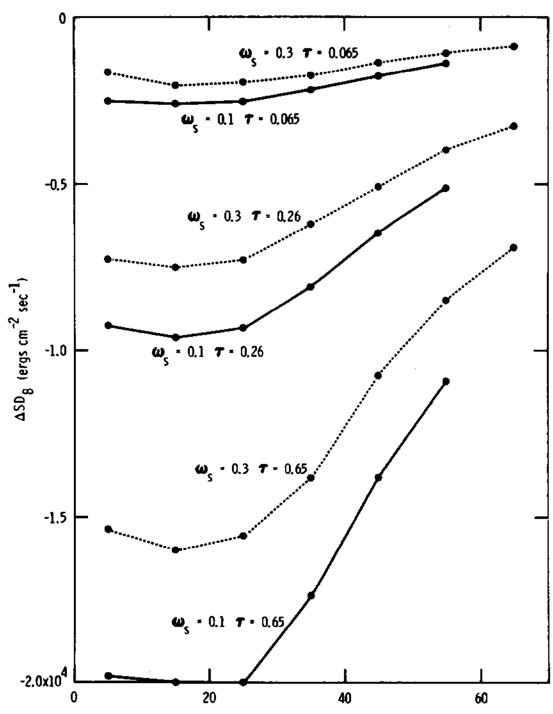

Fig. 5. Difference in net incoming solar radiative flux of the bottom of the atmosphere ΔSD_B for the same set of conditions as in Fig. 4.

Table 6. Values of short-wave surface albedo (Schutz and Gates, 1972, 1973, 1974)

Latitude (°N)	April	July	October	January
5	0.068	0.066	0.066	0.068
15	0.085	0.081	0.076	0.093
25	0.101	0.104	0.107	0.100
35	0.111	0.096	0.110	0.119
45	0.117	0.111	0.125	0.237
55	0.243	0.129	0.176	0.313
65	0.539	0.187	0.384	0.501
75	0.722	0.310	0.612	0.696

Table 7. Measured average net downward solar radiation flux in April at top (SD_T) and bottom (SD_B) of the atmosphere*

Latitude (°N)	SD_T (erg cm^{-2} s^{-1})	SD_B (erg cm^{-2} s^{-1})
5	10.19×10^5	2.21×10^5
15	9.81	2.57
25	9.72	2.59
35	8.26	2.19
45	6.85	1.75
55	5.45	1.56
65	3.81	1.57
75		1.46
85		

* SD_B—Values obtained from linear interpolation of data assembled by Schutz and Gates (1973).

SD_T—Values obtained from linear interpolation of data assembled by Schutz and Gates (1973) on mean planetary albedo and zenith angle.

The present results may, e.g. be compared with the empirical estimates of Budyko (1974). For middle latitudes at constant "planetary" conditions he estimates that a change in radiation of one per cent corresponds to a surface temperature change of 0.5 K (assuming 50% cloudiness). Our results show that a 1% decrease in direct solar radiation occurs for $\omega_s = 0.1$ and $\tau = 0.065$. From Fig. 1 these values of ω_s and τ give a temperature reduction of $\Delta T_S = 0.3$–0.4 K, in good agreement with Budyko's value.

6. LATITUDINAL DEPENDENCE OF AEROSOL EFFECT ON THE TOTAL RADIATIVE COOLING OF THE ATMOSPHERE

The calculated change in the total radiative cooling of the atmosphere when aerosols are added is shown as a function of latitude for $\omega_s = 0.1$ and 0.3 (Fig. 6). In all cases the radiative cooling of the atmosphere is reduced as the aerosol optical density increases. This is consistent with the previous results and conclusions for 35°N latitude in April (Reck, 1974c). The largest reduction in radiative cooling occurs at latitudes nearest the equator for $\omega_s = 0.1$ and $\tau = 0.65$; for this case, the reduction in radiative cooling by aerosols is on the order of 10% of the value without aerosols (see Table 8 and Fig. 6). In general, the percent reduction in radiating cooling is most significant

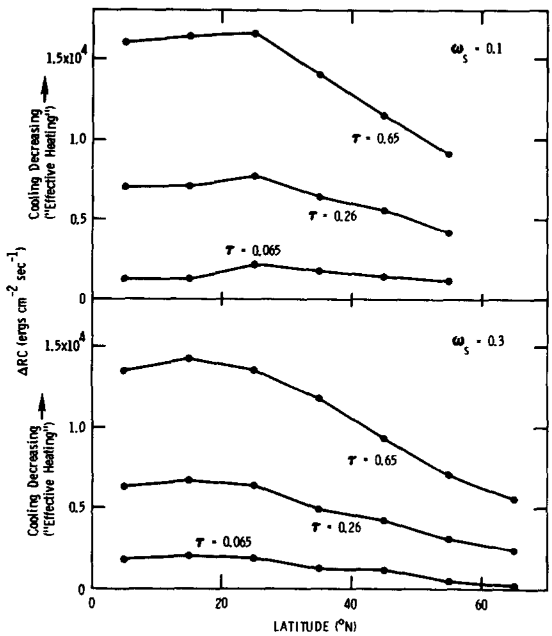

Fig. 6. Difference in the radiative cooling of the atmosphere ΔRC (RC (with aerosols)—RC (without aerosols)) as a function of latitude (°N) for three values of aerosol optical density, $\tau = 0.065$, 0.26, and 0.65 and for two values of surface albedo, (a) $\omega_s = 0.1$ and (b) 0.3.

at the most northern latitudes where it is the order of 20% (see Table 8 and Fig. 6).

SUMMARY

1. This work suggests a mechanism by which aerosols may cause an increase in the length of the winter season.

2. At all latitudes, and at all altitudes, our calculations show that a low-lying aerosol layer of mean visible refractive index, 1.5–0.1 i cools the atmosphere over surfaces with an albedo of 0.3 or less and produces no effect, or slight heating, over surfaces with an albedo equal to 0.6. The magnitude of the temperature change is found to decrease with increasing altitude.

Table 8. Measured average April temperature, T_S, and atmospheric radiative cooling, RC, of the surface of the earth*

Latitude (°N)	T_S (K)	RC (erg cm^{-2} s^{-1})†
5	300.16	1.482×10^5
15	299.61	1.675
25	295.57	1.586
35	287.67	1.338
45	280.64	1.011
55	274.58	0.711
65	265.90	0.309
75	254.73	
85	248.82	

* T_S and RC were obtained from a linear interpolation of the data assembled by Schutz and Gates (1973).

† 1 erg cm^{-2} s^{-1} = 10^{-3} Wm^{-2}.

3. The effect of the aerosol layer is always to decrease the incoming solar radiation at all latitudes. The reduction in radiative flux is larger in magnitude at the surface of the earth than at the top of the atmosphere as would be expected.

4. Aerosols in the atmosphere decrease the radiative cooling of the atmosphere. The largest reduction occurs at latitudes nearest the equator and over surfaces having the smallest ω_s values. However, the most significant *relative* reduction occurs at the most northern latitudes.

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