Radiative effects of desert aerosols

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The heating rates for background, wind-carrying dust and sandstorm aerosol concentrations in the atmosphere have been calculated using a simple realistic model of earth–atmosphere system. Based on the data for Saharan desert aerosols the paper analyses heating rates in troposphere in shortwave and infrared region. It is found that net warming occurs for all three types of aerosol concentrations in winter and summer months. The maximum heating is associated with wind-carrying dust aerosol concentration and the least with background aerosol concentration.

AEROSOL particles play an important role in the earth–atmosphere system because of their direct interaction (absorption and scattering) with solar and terrestrial radiation. They also affect indirectly through their influence on cloud processes. Thus the presence of aerosols leads to change in the earth's radiation budget and hence its climate. The aerosols have potential to warm or cool the earth–atmosphere system depending upon their absorptivity and surface albedo[1]. The scattering and absorption properties of individual particles depend upon the particle shape, size, refractive index and wavelength of incident radiation. The radiative properties of an aerosol layer depend on the above factors as well as on the spatial distribution of the particles, angle of incidence of radiation, and for the nonspheric particles their orientation[2]. However, the impact of aerosol layer is not only determined by its particle size distribution and composition, but also by its location and thickness in the atmosphere, the nature of the underlying surface and the presence of the clouds.

In the present study, a simple model has been used to analyse the effect of an aerosol layer overlying the earth surface, on the climate for three aerosol concentrations, viz. background concentration, wind-carrying dust concentration and sandstorm concentration. Theoretical framework of the model follows along with discussion of our results.

Theory

A two-layered aerosol model as shown in Figure 1 has been considered. The aerosols are assumed to be monodispersive, i.e. same size and same chemical composition, and homogeneously distributed throughout the layer. A further assumption is that only forward scattering takes place. A vertical column of unit cross-section from ground surface $z = 0$ to height $z = 10$ km in hydrostatic equilibrium is assumed. Thickness of aerosol layer suspended at the height of 10 km is considered to be 2 km for the study. The internal energy (U) of each infinitesimal layer of thickness Δz is $U = C_v \rho T \Delta Z$, where C is the specific heat ratio at a constant volume and ρ the density of air. Therefore,

$$\Delta U = C_v \rho \Delta T \Delta Z. \tag{1}$$

The absorbed radiant energy in the earth–atmosphere system is $(1 - \alpha_s)F\downarrow$ while the absorbed radiant energy in the earth–atmosphere aerosol system is $(1 - \alpha_{\text{mod}})F\downarrow$. The change of absorbed radiant energy due to the presence of aerosol is

$$\Delta\alpha_p F\downarrow, \tag{2}$$

where α_s is the albedo of earth–atmosphere system and α_{mod} is the albedo of modified earth–atmosphere system due to the presence of aerosols. From equations (1) and (2), $C_v \rho \Delta T \Delta z = \Delta\alpha_p\, F\downarrow$.

Thus the variation in temperature due to aerosol presence in shortwave (SW) region is

$$\Delta T = \Delta\alpha_p F\downarrow / C_v \rho \Delta z, \tag{3}$$

where, $\Delta\alpha_p$ is the change in albedo due to the presence of aerosol layer and $F\downarrow$ the total daily solar radiation at the top of the atmosphere at 30°N.

It may be noted from Figure 1 that the infrared (IR) region, radiation $F_s\uparrow$ emitted from earth's surface interacts with the aerosol layer and gives rise to a reflected component of the radiation $\text{Re}F_s\uparrow$ towards the ground from the layer, an absorbed component $\text{A}F_s\uparrow$ and a transmitted component $\text{Tr}F_s\uparrow$, where Re, A and Tr are reflectance, absorptance and transmittance coefficients respectively. The reflected component, after reflection again from earth's surface, produces a component $\alpha_s \text{Re}F_s\uparrow$, which further interacts with the aerosol layer to produce a reflected component $\alpha_s \text{Re}^2 F_s\uparrow$ and an absorbed component $\text{A}\alpha_s \text{Re}F_s\uparrow$ and so on. The absorption of $\text{A}F_s\uparrow$ in the aerosol layer heats it and emits a flux $\varepsilon \text{A}F_s\uparrow$

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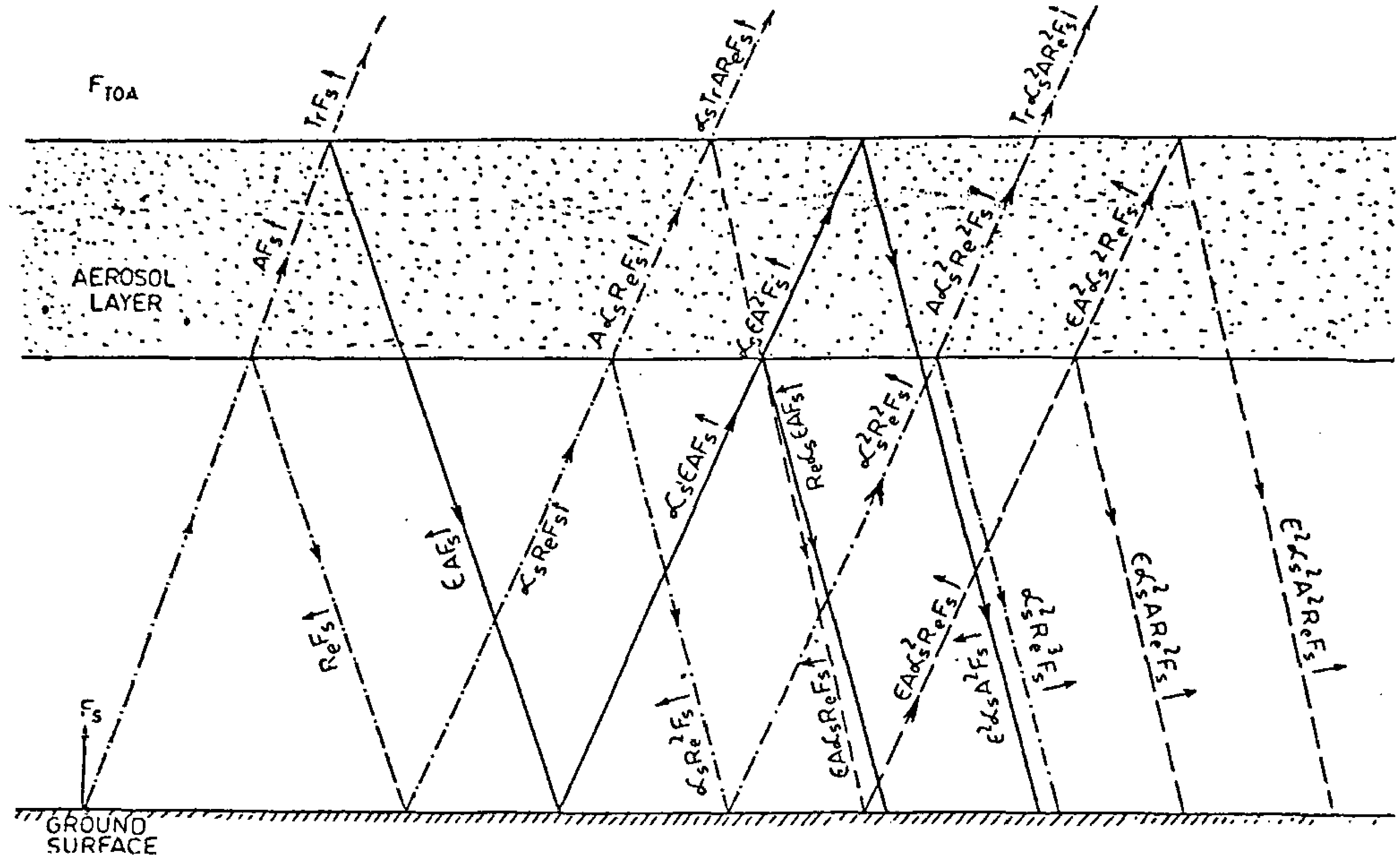

Figure 1. A schematic diagram of a two-layered model.

towards the ground, where ε is the emissivity of the aerosol layer. The emitted flux is reflected from earth's surface to produce $\alpha_s \varepsilon A F_s \uparrow$, which further interacts with aerosol layer to produce higher order reflected and emitted fluxes. Combining various fluxes directed towards the ground, as shown in Figure 1, the net flux ΔF producing the IR radiative forcing can be expressed by the following equation

$$\begin{aligned}\Delta F = &[\mathrm{Re}\, F_s \uparrow + \alpha_s \mathrm{Re}^2 F_s \uparrow + \alpha_s^2 \mathrm{Re}^3 F_s \uparrow + \ldots] \\ &+ [\varepsilon A F_s \uparrow + \varepsilon A \alpha_s \mathrm{Re}\, F_s \uparrow \\ &+ \varepsilon A \alpha_s^2 \mathrm{Re}^2 F_s \uparrow + \ldots] + [\varepsilon \alpha_s \mathrm{Re}\, A F_s \uparrow \\ &+ \varepsilon \alpha_s^2 \mathrm{Re}^2 A F_s \uparrow + \varepsilon \alpha_s^3 \mathrm{Re}^3 A F_s \uparrow + \ldots] \\ &+ [\varepsilon^2 \alpha_s A^2 F_s \uparrow + \varepsilon^2 \alpha_s^2 A^2 \mathrm{Re}\, F_s \uparrow \\ &+ \varepsilon^2 A^2 \alpha_s^3 \mathrm{Re}^2 F_s \uparrow + \ldots]. \end{aligned} \quad (4)$$

The terms in the first bracket represent the first reflected component from aerosol layer $\mathrm{Re}F_s\uparrow$, and its successive reflections from the ground and aerosol layer. The terms in the second bracket represent the downward flux emitted by the aerosol layer $\varepsilon A F_s$, and its successive reflections from the aerosol layer. The terms in the third bracket arise out of the interaction of the flux $\alpha_s \mathrm{Re} F_s \uparrow$ from the ground with the aerosol layer. After absorption in aerosol layer the emitted component is $\varepsilon \alpha_s \mathrm{Re} A F_s \uparrow$, which is the first term in the third bracket. The later terms arise due to successive reflections of this component from the aerosol layer. The terms in the fourth bracket arise out of absorption of the flux $\alpha_s \varepsilon A F_s \uparrow$ in the aerosol layer and subsequent emitted component and its su reflections. It is evident that the terms in the sec third brackets correspond to first-order absorp emission process, whereas the terms in the fourtl result from second-order absorption and emissior Further higher order absorption, reflection an mission terms have been neglected.

After rearranging the terms and simplification, (4) reduces to

$$\Delta F = F_s \uparrow [1 + \varepsilon A \alpha_s][\varepsilon A + \mathrm{Re}]\left[\frac{1}{1-\alpha_s \mathrm{Re}}\right].$$

Since $\varepsilon A \alpha_s$ is very small, we can write equation (5

$$\Delta F = F_s \uparrow (\varepsilon A + \mathrm{Re})\left[\frac{1}{1-\alpha_s \mathrm{Re}}\right].$$

Assuming that ΔF due to the presence of aeros causes a temperature change ΔT, equations (1) yield,

$$\Delta T = \frac{1}{C_v \rho \Delta z}[\mathrm{Re}\, F_s \uparrow + \varepsilon A F_s \uparrow]\left[\frac{1}{1-\alpha_s \mathrm{Re}}\right],$$

where, $F_s \uparrow = \sigma T_s^4; T_s$ is the surface temperature taken as 17.8°C for winter months and 24.8°C for months. σ is the Stefan Boltzman constant.

Reflectance (Re) and absorptance (A) of the layer can be written as

$$\mathrm{Re} = \omega(1 - g)(1 - e^{-\tau e}), \tag{8}$$

$$A = (1 - \omega)(1 - e^{-\tau e}), \tag{9}$$

where ω is the single scattering albedo, g the asymmetry factor and τ_e is the aerosol optical depth.

Result and conclusions

The study of radiative characteristics of a dust-laden atmosphere requires an understanding of various parameters such as extinction coefficient, absorption coefficient, single scattering albedo and asymmetry factor. For background, wind-carrying and sandstorm aerosol concentrations, these parameters have been evaluated in the SW and IR regions[3]. The average values for various parameters have been calculated using the relation,

$$\overline{A} = [\Sigma A_i(\lambda)\lambda_i]/\Sigma\lambda_i\,, \tag{10}$$

where $A_i(\lambda)$ is a particular parameter and λ_i the wavelength associated with that parameter.

Optical depths in the SW region for background, wind-carrying and sandstorm concentrations are taken as 0.457/km, 1.902/km and 4.465/km respectively. For the IR region the optical depths for the above-mentioned concentrations of aerosol are taken as 0.179/km, 1.561/km and 1.598/km respectively.

First we consider the radiative forcing due to aerosols in the SW region and calculate the changes in temperature for different types of aerosol concentrations using equation (3). Figure 2 shows the heating rates for background, wind-carrying dust and sandstorm concentrations in SW for all the months (Jan–Dec). It may be seen that maximum heating occurs for wind-carrying dust concentration in the months of JJA, whereas cooling

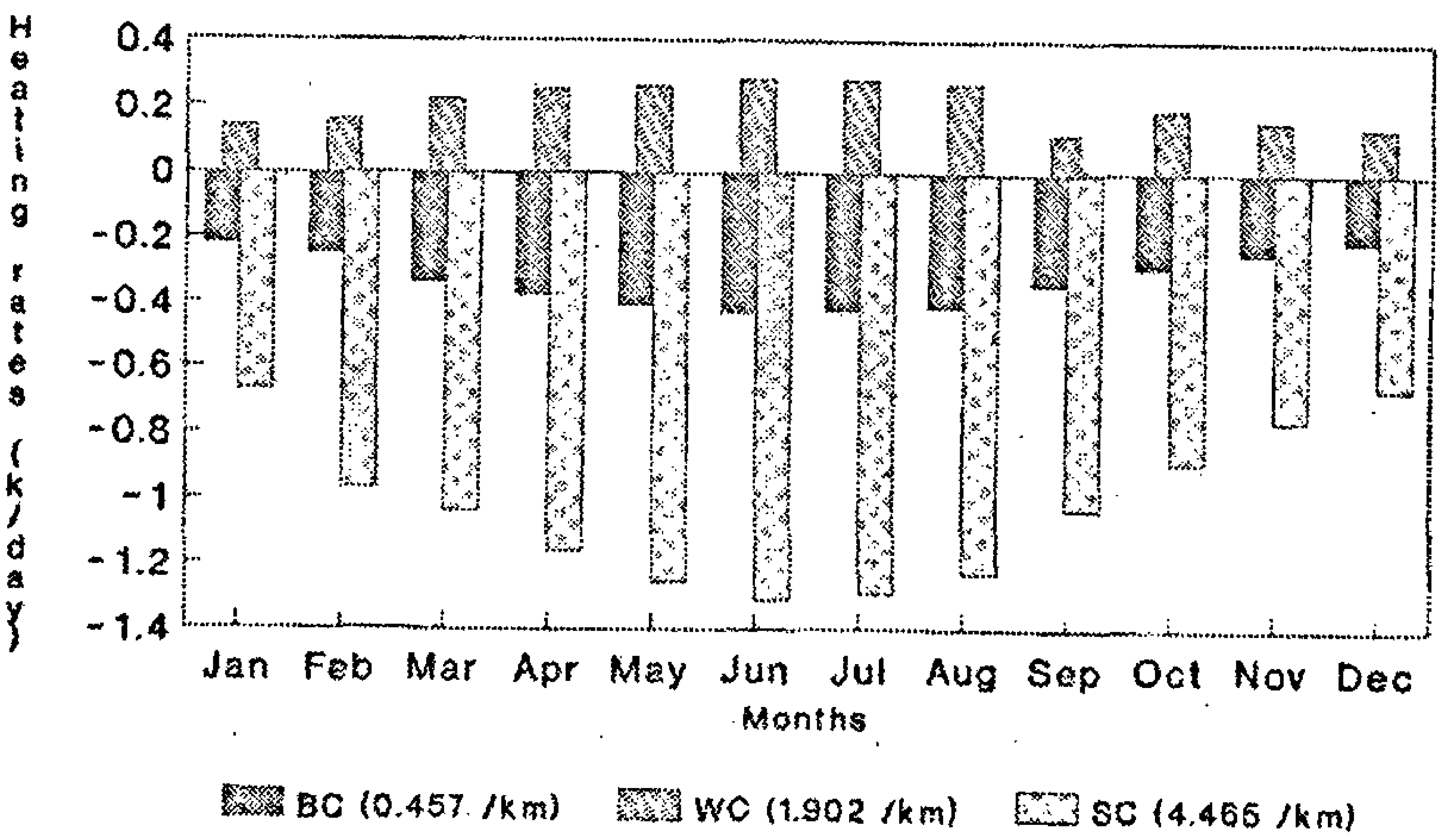

Figure 2. Heating rates for background, wind-carrying dust and sandstorm concentration in solar region.

Table 1. Heating rates in the SW region

Conc.	Optical depth (/km)	Heating rates (k/day) Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
BC	0.457	−0.22	−0.25	−0.34	−0.38	−0.41	−0.43	−0.42	−0.41	−0.34	−0.29	−0.25	−0.21
WC	1.902	0.14	0.16	0.22	0.25	0.26	0.28	0.28	0.27	0.12	0.19	0.16	0.14
SC	4.465	−0.67	−0.9	−1.04	−1.16	−1.25	−1.31	−1.29	−1.23	−1.03	−0.89	−0.76	−0.66

Table 2. Heating rates in SW, IR and net warming

Conc.	Solar region Optical depth (/km)	Solar region Heating rate (k/day) Winter	Summer	Infrared region Optical depth (/km)	Infrared region Heating rate (k/day) Winter	Summer	Net warming (k/day) Winter	Summer
BC	0.457	−0.23	−0.42	0.179	0.43	0.47	0.20	0.05
WC	1.902	0.15	0.27	1.561	2.05	2.25	2.20	2.53
SC	4.465	−0.74	−1.28	1.958	2.29	2.51	1.55	1.23

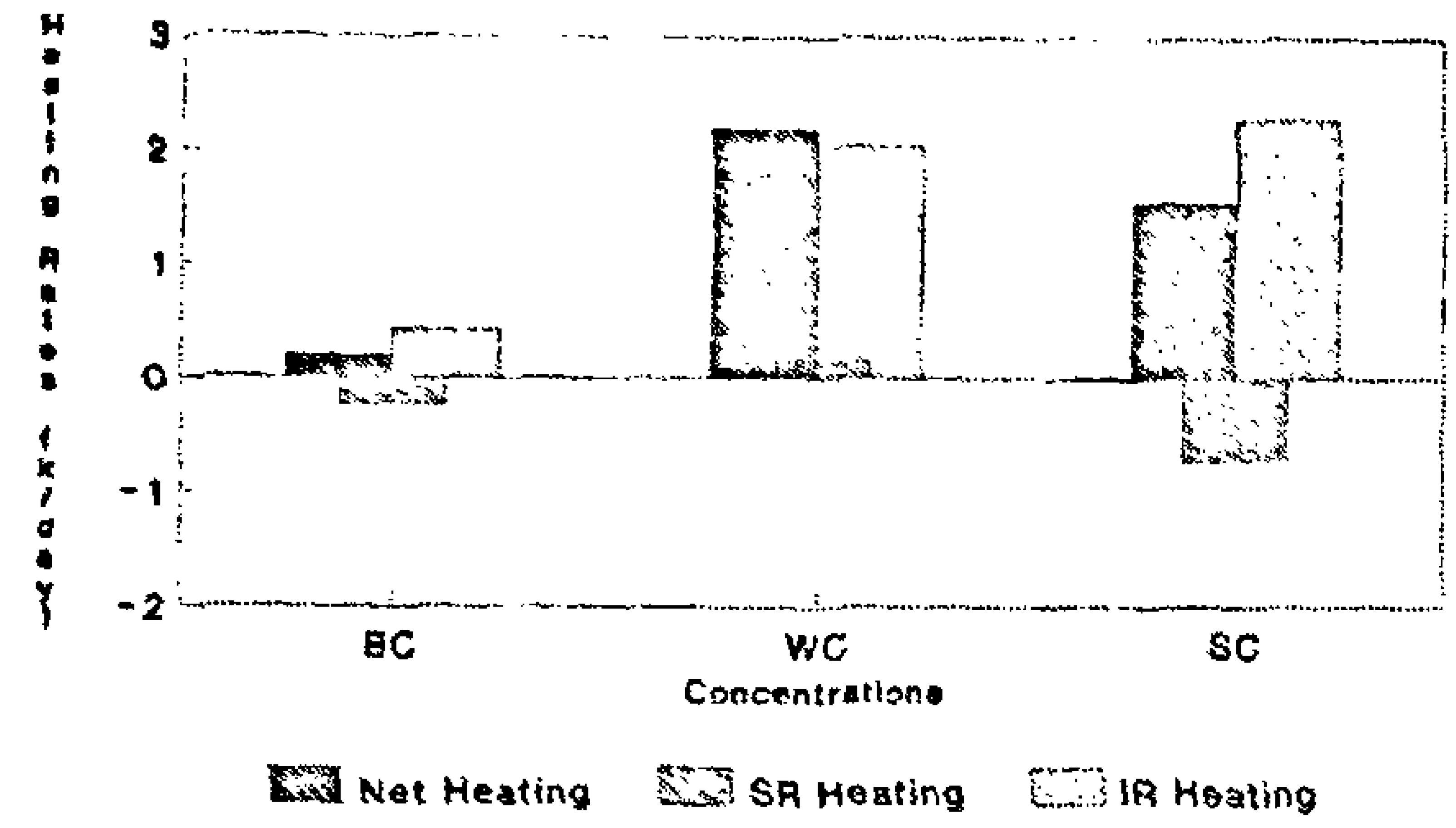

Figure 3. Average heating rates for different concentrations during winter months (DJF).

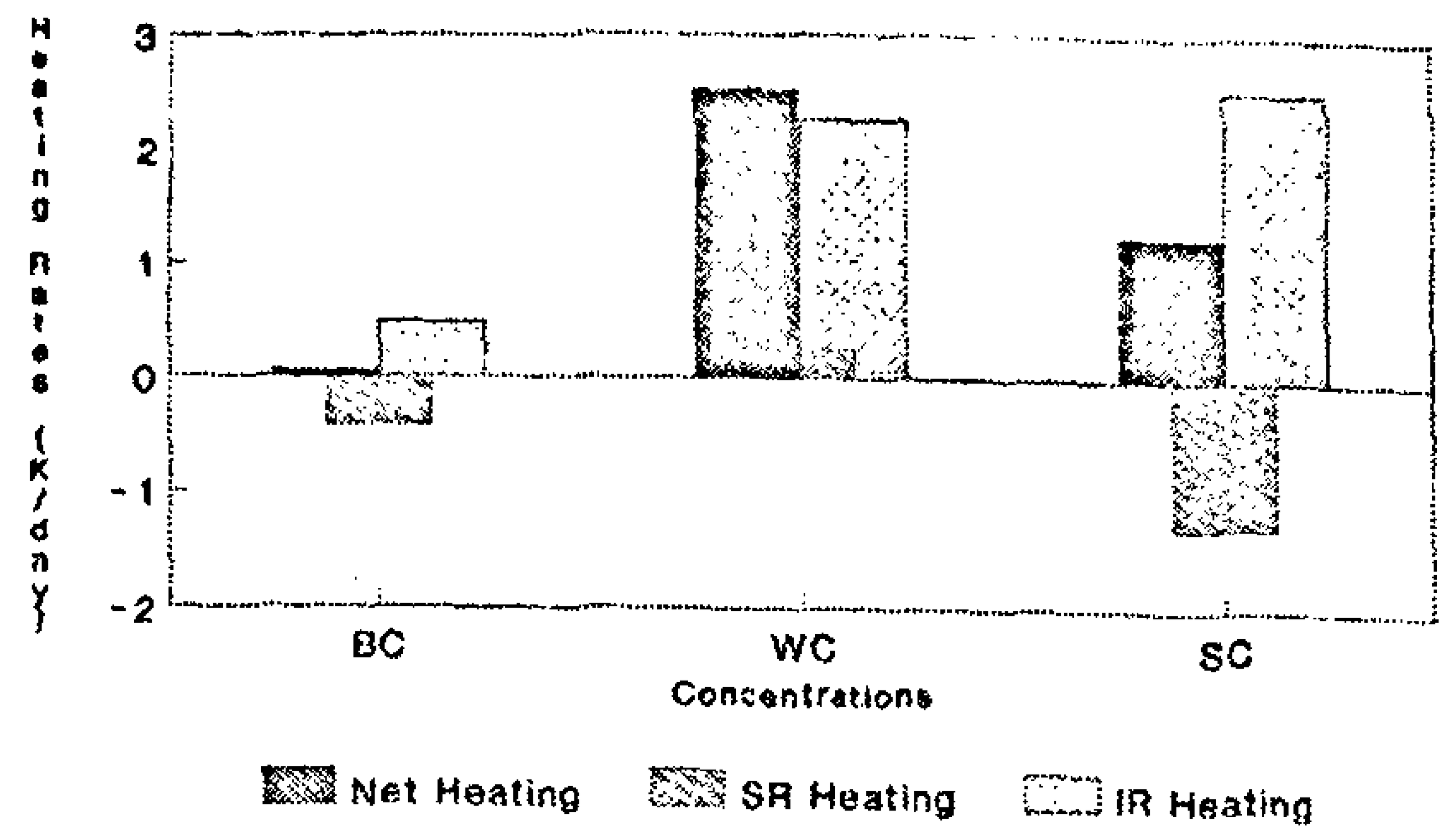

Figure 4. Average heating rates for different concentrations during summer months (MJJ).

occurs for background and sandstorm concentrations. Maximum cooling for both background and sandstorm aerosol concentrations is observed during JJ. It may be noticed that absolute temperature change in sandstorm concentration is significantly higher as compared to background and wind-carrying dust concentrations. Heating rates for SW region are given in Table 1. The highest heating rates in SW region are observed for the month of June, i.e. – 0.43 k/day for background, – 1.31 k/day for sandstorm and 2.8 k/day for wind-carrying dust concentration.

In the IR region the heating rates are calculated using equation (7) for winter and summer months and are presented in Table 2 and compared with the heating rates of solar region. It is found that heating rates in IR region are slightly higher in summer months than winter months for all three types of aerosols. The net result (Table 2) of the radiative forcing in SW and IR region is warming, which is greater in winter months compared to summer for background and sandstorm concentration. For wind-carrying dust concentration, however, the net warming is greater in summer. For winter and summer months the SW, the IR and net heating rates for different types of aerosol concentrations are shown in Figures 3 and 4 respectively.

To summarize, the present study shows that net heating would result in the case of all three types of aerosol concentration in both winter and summer months. Maximum heating occurs for wind-carrying dust concentration followed by sandstorm concentration. Heating is the least for background concentration. For wind-carrying dust concentration, heating rate is maximum for both winter (2.2 k/day) and summer (2.53 k/day). For sandstorm concentration the heating rates for both winter and summer months are 1.55 k/day and 1.23 k/day respectively. For background concentration very small heating occurs for summer months (0.05 k/day) and winter months (0.2 k/day).

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MEETINGS/SYMPOSIA/SEMINARS

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