551.521:551.553.21 (545)

Radiation Climate of New Delhi
Part I : Shortwave Radiation

KATHARINA LETTAU

*Center for ClimaticResearch University of Wisconsin at Madison**

(*Received 2 April 1969*)

ABSTRACT : A budgetary method to appraise the various shortwave attenuation processes in the atmosphere is described. The method is based on observed global and diffuse radiation from sun and sky, given separately for clear days and all days with average cloud cover. New Delhi was selected as a case study because cloudiness, precipitable water, and aerosol content of the air show pronounced annual variations. It is shown that the monsoon period does not produce radical changes in aerosol absorption. More significant is a decrease in the intensity of aerosol scattering. This may be interpreted as a change in particle composition and size distribution. While the monsoon rains remove most of the mineral dust, enough hygroscopic nuclei remain in the air over New Delhi so that attenuation by absorption decreases only slowly after the rains have stopped.

The total heating by all absorbers in the atmosphere varies from 0·54°C/day in January–March to 1·23°C/day in July–September. Aerosol contributes about 30 per cent to the total heating.

1. Introduction

Since 1963 the Rajasthan desert of northwestern India has been investigated jointly by scientists from the University of Wisconsin and the India Meteorological Department to determine whether this desert might possibly be man-made. In the course of these studies Bryson (1963) suggested that radiation attenuation due to dust and haze (aerosol)—so prevalent over India during the pre-monsoon months—could be a contributing factor to the subsidence of air and consequently, to the aridity of the area.

In the following study, the use of existing climatic data for a quantitative determination of aerosol effects on the radiation budget at New Delhi is attempted. A budgetary method to appraise the various attenuation processes quantitatively in the atmosphere above a given locality has been described recently by Lettau and Lettau (1969), under consideration of total energy of the solar spectrum, within the framework of upper and lower boundary conditions. This approach can be applied to a local radiation climate and permits one to discuss how the climate will be affected by natural or man-made changes of environmental factors. This method employs conventional ground-based radiation measurements, but also permits one to include external data from satellites. The physical relationships between attenuation parameters and variables to be considered are summarized in Section 2.

New Delhi was selected as a case study for several reasons. The method requires climatic data on diffuse radiation from the sky, in addition to global radiation. This information is available from New Delhi since 1957, measured with a Moll-Gorczinski solarimeter equipped with a shading ring. First-class climatological observations are also recorded, and the geographical location is not far from the extensive desert area which is of special interest.

Humidity and aerosol content in the air over New Delhi increases from winter to summer. A marked break occurs at the onset of the monsoon rains. Thus, an investigation of the annual trend of attenuation parameters seems to be most revealing with respect to the objective of this study, namely to determine the effects of aerosol on radiative heating in the atmosphere.

8 Summary of Budget Equations

It is convenient to express the mathematical relationships of shortwave attenuation in dimensionless form. This will be achieved by dividing any energy flux-density by the extra-atmospheric irradiance, $I_0 \cos \nu = I$, which is the insolation intercepted horizontally by one cm² area at the top of the atmosphere (ν = zenith angle of the sun). To guarantee a reasonable degree of completeness and detail, no less than nine variables must be considered. These can be divided into three groups:

2·1. *Non-dimensional values of shortwave fluxes*

A = "top albedo", or fraction returned to space at the top of the atmosphere;

G^* = reduced global radiation, or fraction received at ground level from sun and sky;

*This research was supported by Grant GP-5572XI, Section on Atmospheric Sciences, National Science Foundation

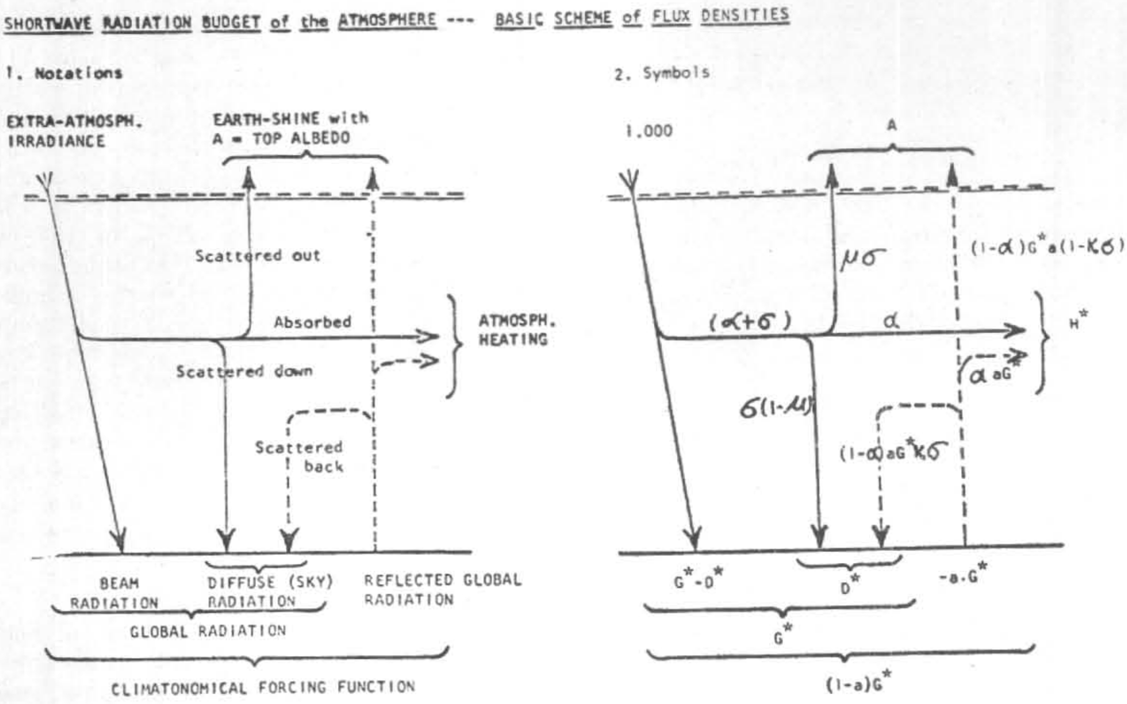

Fig. 1. Shortwave Radiation fluxes : Schematic illustration and notations

D^* = reduced diffuse radiation, or fraction received at ground level from sky alone;

H^* = relative solar heating in the air, or fraction absorbed in the atmosphere; and

a = energy albedo of the lower boundary of the atmosphere; thus, $(1\text{-}a)G^*$ = fraction absorbed by the ground.

2·2. *Contributions to attenuation of non-dimensional direct solar (beam) radiation*

σ=attenuation due to scatterers in the air,

α= attenuation due to absorbers in the air.

2·3. *Coefficients of the scattering process*

μ=fraction describing effective outward scattering to space; or $(1\text{-}\mu)$ = fraction of effective downward scattering to the lower boundary,

K=fraction describing backward scattering of that part of shortwave radiation which has been reflected at least once by the lower boundary.

Fig. 1 illustrates the basic scheme as specified in Section 2. Equation (1) to (5) can be deduced from this scheme. Physical relationships between the nine dimensionless variables produce only five equations, Equation (1) expresses the budget of shortwave radiation energy. The fraction not reflected to space must be equal to the sum of absorption by the ground $(1\text{-}a)\,G^*$ and by the atmosphere (H^*),

$$1-A = G^*(1-a) + H^* \tag{1}$$

The second equation considers non-dimensional direct solar radiation $(G^* - D^*)$. Its attenuation $(1-G^*+D^*)$ must equal the depletion by absorbers and scatterers,

$$1-G^*+D^* = \alpha + \sigma \tag{2}$$

Equation (3) represents the radiation reflected back to space, the "earth-shine," expressed as energy flux-density, or "top albedo" (A) in dimensionless form. The latter is the sum of the outward directed part of primary scattering $(\mu\sigma)$ plus that part of ground-reflected global radiation which is neither abosrbed in the air nor backscattered to the ground,

$$A = \mu\sigma + (1-\alpha)\, a\cdot G^*\,(1-K\sigma) \tag{3}$$

The fourth equation expresses the diffuse radiation arriving at the ground. It consists of the downward directed part of the primary scattering $(1-\mu)\,\sigma$ and of that fraction of ground reflected radiation which is scattered back towards the lower boundary,

$$D^* = (1-\mu)\sigma + (1-\alpha)\ a.G^*K\sigma \qquad (4)$$

Equation (5) represents the overall balance of the preceding equations, and thus is not independent. It results from the necessary condition that the sum of the left-hand sides must equal to that of the right-hand sides of Eqs. (1) to (4). This reduces the heating fraction simply to

$$H^* = \alpha\ (1+\alpha.G^*) \qquad (5)$$

namely, the sum of primary absorption plus that part of ground-reflected radiation which is absorbed in the air.

With four idependent equations and nine variables to be determined, five variables must be independently known. That means they must be either observed, or partly observed and partly prescribed by model assumptions, or externally given. In the following sections it will be assumed that G^* and D^* are observed, while a, μ, K are prescribed, so that α, σ, A, and H^* can be derived in order to permit the discussion of seasonal changes in absorption and heating, and in scattering and planetary or top albedo.

The fundamental equations (1) to (4) can be transformed readily by eliminating certain variables. For example, elimination of D^* in (2) with the aid of (4) produces

$$G^* = (1-\alpha-\mu\sigma)\,/\,[1-aK\sigma\,(1-\alpha)] \qquad (6a)$$

In the special case of $a=0$, the earth's surface reacts like an ideally black body, whereupon (6*a*) reduces to $G^*_{(a=0)} = (1-\alpha-\mu\sigma)$. Substituting d for $K\sigma(1-\alpha)$, Eq. (6*a*) can be restated as

$$G^* = G^*_{(a=0)}\,/\,(1-a.d) \qquad (6b)$$

which is Angstrom's formula as quoted from Möller (1965) where d represents the 'backscatterance" of the sky. Equation (6*a*) immediately shows how backscatterance depends on the efficiency of scatterers and absorbers in the air.

For cloudfree air, it will be assumed that the total absorption (α) can be separated into additive contributions by water vapour (α_w). other absorbing gases, notably ozone and CO_2 (α_g), and aerosol (α_{ae}). The total scatter in cloudfree air (σ) will consist of contributions by neutral molecules (Rayleigh scattering σ_R) and aerosol (α_{ae}).

If a fraction (c) of the sky is covered by clouds, the relative flux densities of the cloudless area must be multplied by the weight factor $(1-c)$. Let σ_c denote the effective coefficient of scattering and α_c the absorptivitity of air in the cloud, and let μ' be a fraction so that $\mu'\sigma_c = A'$ will be scattered back to space by the upper cloud surface and $(1-\mu')\ \sigma_c$ will reach the ground. It will be assumed that the effective albedo of the cloud base will not be different from that of the cloud top (A'). The employment of "effective" or bulk coefficients represents a considerable model simplification of the scattering processes, which are known to possess pronounced dependencies on particles sizes (and their spectral distribution), liquid water content of the cloud, and solar elevation angle. Compressing these characteristics into a few single parameters can be tolerable only for the study of climatic gross-effects and overall energey partitions.

In the following transformations of energy budget equations, let the subscripts o and c indicate values for cloudless sky (with prescribed conditions of clear-air attenuation and ground albedo), and for cloudy sky, respectively. As an important model characteristic for the calculation of effective coefficients in partly cloudy to overcast air, it is assumed that the contributions by absorbers are additive while contributions by scatterers are distributive (or must be prorated). Thus, in air with cloudiness c, effective absorption and ascattering are given by

$$\alpha = (1-c)\ \alpha_o + c\,(\alpha_o + \alpha_c) = \alpha_o + c\,\alpha_c \qquad (7)$$

$$\sigma = (1-c)\ \sigma_o + c\sigma_c = \sigma_o + c\,(\sigma_c - \sigma_o) \qquad (8)$$

In a logical extension of equation (3) it follows from continuity principles that the effective top albedo is composed of a prorated contribution from the clear area and direct reflection from the cloud surface plus diffuse radiation reflected from the ground, and reduced by what is either reflected back downward from the cloud base, or absorbed in the air. Hence,

$$A = (1-c)A_o + cA' + a.G^*(1-\alpha)(1-A') \qquad (9)$$

Corresponding developments for the effective heating function H^* yield

$$H^* = (1-c)\ H_o^* + c.\alpha\ (1 + a.G^*) \qquad (10)$$

The sum of equations (9) and (10) produces $(A+H^*)$ which is then used to eliminate this term in equation (1). The resulting relation can be solved for G^*,

$$G^* = [1 - A_o - H_o^* + c(A_o + H_o^* - A' - \alpha)] / [1 - a + c \cdot a\ (1 - A' + \alpha A')] \qquad (11)$$

After G^* has been determined for a given cloudiness, A follows from Eq. (9), and H^* either from Eq. (5) or (10).

If supporting data on planetary albedo (A), as measured from orbiting satellites, are available, a transformation of equations (2), (3), (4), and (9) yields the value of the surface albedo (a).

The resulting equation can be written as

$$a = [A - (1-c)\ (1-\alpha_o - G_o^*) - cA'] / [(1-c)\ (1-\sigma_o)G_o^* + c\,(1-A')\ (1-\alpha)G^*] \qquad (12)$$

Albedo values derived from satellite observations cannot be considered "point values". They represent integrated values over a rather large area depending on the resolution of the sensors. They vary greatly with the amount of clouds present at the time of the satellite's overpass. If surface albedo (a) is known in addition to top albedo (A), equation (12) could be used to get an indication of the amount of clouds present at the time in regions where no climatic observations are available.

3. Method for calculating Attenuation Parameters

Let us consider cloudless conditions first. With the given set of observed values (G_o^*, D_o^* and a) and prescribed values (K and μ), the total attenuation of direct solar radiation by absorption α_o and by scattering σ_o can be computed. The sum $(\alpha_o + \sigma_o)$ is immediately deducible from equation (2). It was assumed in Section 2 that $\alpha_o = \alpha_w + \alpha_g + \alpha_{ae}$ and $\sigma_o = \sigma_R + \sigma_{ae}$.

The absorptivity of water vapour can be calculated by using a semi-empirical formula derived from data reported by Yamamoto (1962)

$$\alpha_w = 0.102\ (w.M)^{0 \cdot 276} \qquad (13)$$

where w=precipitable water, and M=optical air mass. Total precipitable water can be determined from readiosoundings, or estimated from the dewpoint temperature at the surface. It was shown by Smith (1966) that the following empirical relationship produces fairly representative values,

$$\ln w = 0.1133 - \ln\ (\lambda+1)\ 0.0393\, t_d \qquad (14)$$

where t_d is the dewpoint temperature and λ an empirical number depending on the season, ranging from 2·8 in summer to 3·3 in winter. Although aerological observations are available from New Delhi, the use of equation (14) was preferred because of its wider application possibilities.

Ozone absorption is a complex process which is known to depend on total amount as well as vertical distribution of this constituent gas. Dave and Sekera (1959) have calculated and parameterized it with the aid of a simplified model of vertical O_3–distribution. Because ozone absorption is normally small in comparison with water vapour effects, it is suggested here to carry the simplification one step farther by assuming that this absorption is directly related to the total O_3–amount as well as proportional to the readily calculated intensity of Rayleigh scattering. If the tabulations by Dave and Sekera (1959) are taken as the basis for interpolation, a crude estimate suggests for example, that for a total of 0·23 cm, the ozone absorption will approximately amount to 0·16 of the calculated Rayleigh scattering. Even though this estimate may be wrong by 40 per cent the effect on the relative error of total attenuation will remain less than 5 per cent.

According to Roach & Yamamoto — as quoted from Moeller (1964)—water vapour absorbs ten to twenty times more radiation than carbon dioxide in the lower atmosphere. Only above 12 km absorption by CO_2 exceeds that by H_2O by approimately a factor of five. The absolute value of both, however, is very small. Thus, in the following calculation of shortwave radiation absorption the contribution of carbon dioxide is neglected (*see* also Robinson 1963).

Molecular, or Rayleigh scattering can be computed from known geographical and astronomical data with the help of convenient tabulations such as included in the IGY Tables (1958). The remaining problem is to separate aerosol attenuation into a scattering and absorbing fraction under cloudless sky conditions. If diffuse radiation D^* is known, this amounts to prescribing the factors μ and K, as defined in Section 2·3. Tentative values of μ=1/3, and K=1·0 were used following conclusions by Lettau and Lettau (1969), Then, α_o and σ_o result from equations (2) and (4), and consequently, α_{ae} and σ_{ae} are established as remainders.

If part of the sky is covered by clouds, equations (7) to (9) can be employed to determine the parameters α_c, σ_c, and $\mu' = A'/\sigma_c$. Even if all three parameters would be independent of the degree of cloudiness, equation (11) shows that the variation of G_c^*/G_o^* with cloudiness will still depend on the turbidity in the cloudfree atmosphere at the locality under investigation. Studies by Kimball, Angström, Mosby and Haurwitz—as quoted on page 440 of *Smithsonian Tables* (1963)—suggest simplified universal relationship between insolation and cloudiness. Practically useful statistical relationships are extensively discussed in the summary presented by Budyko (1958).

A new approach to the general problem of calculating cloudcover effects on global radiation became necessary, under due consideration of the shortwave radiation budget-relationships discussed in Section 2. For this purpose results of a study by Haurwitz (1934) are useful, which relate global radiation to cloud types during overcast as a fraction of global radiation under cloudless conditions. Haurwitz employed data from Blue Hill Observatory in Massachusetts. A recent study by Mooley and Raghavan (1963) resulted in similar values for India (Madras Airport). For example, the ratio G_1/G_o, that is the percentage of clear-sky radiation transmitted through overcast at an optical airmass of $1 \cdot 2$ for altostratus clouds, was 41 per cent at Blue Hill and 39 per cent at Madras. Both studies show that the dependency on airmass is slight.

Even in the present era of world-wide measurements of solar energy fluxes including data from orbiting satellite stations, there will exists a paucity of data on the attenuation effects of clearly defined cloud masses, or cloud thicknesses. One of the few systematic evaluations of fractions absorbed, transmitted, and reflected, is the study by Neiburger (1949), which is based on a series of aerological measurements of shortwave radiation fluxes above and below a coastal stratus in California. He finds that absorption is generally small rarely exceeding 10 per cent so that the amount of energy transmitted is mainly dertermined by reflection at the upper cloud surface (A') which increases significantly with cloud thickness.

According to the very comprehensive study by Budyko and Kondratiev (1961) on the energy balance of the northern hemisphere, the average α_c equals $0 \cdot 04$. and the average A' about $0 \cdot 6$. At New Delhi, it was tentatively assumed that α_c equals $0 \cdot 04$ during the "monsoon months" June through September, while for the remaining eight months α_c was taken to be $0 \cdot 02$. On this basis, an estimate of monthly mean σ_c is possible by employing the observed c values and the difference between direct solar radiation on clear *versus* cloudy days, that is, $(G_o{}^*—D_o{}^*)$ *minus* $(G^*—D^*)$. The result for 1959 was $\sigma_c = 0 \cdot 60 \pm 0 \cdot 06$. Since the departures from the mean σ_c did not appear to be systematically coupled with the season, it was concluded that a constant value of $\sigma_c = 0.60$ could be used at this location.

In order to obtain the albedo A' of the average cloud surface at New Delhi, equation (9) was considered in combination with

$$A + D^* = \sigma + (1 - \alpha)\ a.G^* \qquad (15)$$

which follows from euqations (1), (2), and (5). Solving Eqns. (9) and (15) for A' yields—

$$A' = [\sigma - D^* + (1-c)\ (a.G^* - \alpha\ a.G^* - A_o)] \,/\, [c - c.a.G^*\ (1-\alpha)] \qquad (16)$$

After inserting the repspective monthly means on the right-hand side of (16), it was found that $A' = 0 \cdot 30 \pm 0 \cdot 03$. The relatively low value of the cloud albedo at New Delhi may be due to the fact that the bulk of the monsoon clouds does not reach great heights, and dust or haze may still be above them (reference is made to Section 5).

It follows from the average of $\sigma_c = 0 \cdot 60$ and $A' = 0 \cdot 30$, that the fraction μ' equals $\frac{1}{2}$. Together with $\alpha_c = 0 \cdot 02$ to $0 \cdot 04$, and the previously determined local clear-air attenuation coefficients α_w, α_g, α_{ae}, σ_R, σ_{ae}, the basic information is now assembled for a comprehensive evaluation of the radiation climate of New Delhi.

4. Climatology of Shortwave Attenuation for New Delhi

The information summarized in Tables 1 and 2 came from various sources. The extra-atmospheric irradiances are values tabulated by Bolsenga (1964); global and sky radiation are observed data as published separately for clear days and all days by Mani and Chacko (1963). Surface albedo was measured at the grounds of the New Delhi Observatory by the University of Wisconsin study group only in January of 1967, and tentatively estimated for the other months. Climatic data for New Delhi were taken from *Monthly Climatic Data for the World*, published by US Department of Commerce.

The calculations of attenuation coefficients was based on the formulæ summarized in Section 3. Results are listed in Table 3, separately for absorption and scattering constituents. All coefficients show a pronounced annual trend. Absorptivities are highest during July and August at the peak of the monsoon season. They decrease only slowly after the rains have stopped because of the still high precipitable water content in late summer. Aerosol scattering increases steadily from January to June with a significant break when the rains begin.

Totally cloudless days are rare during the rainy season. At New Delhi in 1959, none was recorded in July, one in August, and two in September. Consequently, the coefficients listed in Table 3 for these months are rather uncertain and may not be truly representative. However, the steady increase in

TABLE 1

Monthly means of Shortwave Radiation Fluxes at New Delhi

Recorded mean daily totals are given separately for clear days (G_0^*, D_0^*) and all days (G^*, D^*) in 1959.

$\triangle G^* = G^*_{obs} - G^*_{adj}$, and $\triangle D^* = D^*_{obs} - D^*_{adj}$, are adjustments for constancy of cloud albedo and scattering.

Month	Reduced global and sky radiation				Adjustment	
	G_0^*	G^*	D_0^*	D^*	$\triangle G^*$	$\triangle D^*$
January	0·792	0·694	0·119	0·157	—0·023	—0·012
February	0·812	0·769	0·132	0·166	—0·002	0·008
March	0·765	0·753	0·140	0·180	0·070	—0·011
April	0·739	0·672	0·134	0·184	0·005	0·003
May	0·772	0·694	0·173	0·210	0·026	0·008
June	0·664	0·604	0·213	0·255	0·046	0·014
July	0·622[1]	0·458	0·182[1]	0·290	—0·010	0·014
August	0·674[2]	0·485	0·079	0·219	—0·026	0·011
September	0·687[3]	0·522	0·121	0·184	—0·036	—0·033
October	0·690	0·641	0·125	0·146	—0·003	—0·016
November	0·709	0·690	0·099	0·108	—0·014	—0·016
December	0·730	0·676	0·121	0·167	0·002	0·012
Annual mean	0·717	0·638	0·137	0·189	0·003	—0·001

(1) Values are estimated baecause there was no clear day in July 1959

(2) Only one clear day

(3) Only two clear days

aerosol scattering from January to June appears significant. Aerosol absorption increase at a slower rate and more erratically than scattering. Thus the ratio α_{ae}/σ_{ae} is smaller than unity during most of the year except in the months after the monsoon. This may indicate that the lower atmosphere is cleared of mineral dust but not of hygroscopic nuclei which still form haze when the water vapour content is high.

Finally, Table 4 shows the monthly values of the shortwave radiation budget at New Delhi. It was necessary to adjust the observed G^*-and D^*-values in order to conform with the budget requirements because average values of $A'=0·30$ and $\sigma_c=0·60$ were used to calculate the date of Table 3*. The fraction of solar energy absorbed by the ground (and used for heat conduction and evaporation) is highest in February and lowest in July whereas the fraction absorbed in the atmosphere shows the opposite trend, a maximum in July and a minimum in February. The fraction reflected to space, established as the remainder of the budget equation (1), is lowest in December and highest in July. The magnitude of the top albedo appears to be somewhat low (annual mean of 0·21), since the northern hemispheric mean is 0·36 according to Budyko and Kondratiev (1961), but only 0·295 according to Vonder Haar (1968). At New Delhi,

*The adjustments made are indicated in Table 1

TABLE 2

Monthly means of Climatic Station Data at New Delhi

Sea level pressure P (mb), vapour pressure e (mb), daily mean air temperature T (°C), precipitation p (mm/month), Cloudiness c (fraction), total ozone O_3 (cm), and estimated values of surface albedo (a)

Month	P	e	T	p	$*c'$	$*c''$	O_3	a
January	1017	11·49	14·2	41	0·27	0·36	0·27	0·18
February	1013	9·81	16·3	1	0·15	0·27	0·29	0·18
March	1010	11·00	23·5	3	0·30	0·15	0·30	0·19
April	1007	10·75	28·9	0	0·27	0·24	0·30	0·19
May	1000	13·62	32·4	22	0·21	0·16	0·29	0·20
June	997	20·69	34·9	26	0·38	0·27	0·29	0·20
July	997	29·71	31·3	128	0·70	0·63	0·27	0·18
August	998	30·65	30·3	218	0·55	0·65	0·26	0·15
September	1002	27·94	29·7	113	0·49	0·48	0·26	0·15
October	1009	19·36	27·1	0	0·20	0·13	0·25	0·15
November	1015	12·61	20·6	3	0·12	0·12	0·25	0·15
December	1017	9·63	16·3	0	0·19	0·23	0·26	0·15
Annual mean	1007	17·27	25·5	—	0·32	0·31	0·27	0·17
Annual total				554				

* c'-values for cloudiness were calculated from the figures of mean hours of cloudiness and mean hours of sunshine given by Mani and Chacko (1963). c''-values were kindly supplied by the reviewer of this study. They represent the correct value for mean cloud amount at New Delhi in 1959 based on six observations per day. The calculated attenuation coefficients as shown in Table 3 are based on c'-values. Since the contribution of $c\ a_c$ is so small, the difference between c' and c'' becomes insignificant. $c\sigma_c$ was determined as the remainder term.

the annual mean of cloudiness was only 0·3,, and that of the estimated surface albedo only 0·17 for 1959. In addition, the top albedo values for July and August could be too low due to the uncertain G_o*- and D_o*-values, as indicated in Table 1. In spite of the opposite annual trend of $(1-a)G^*$ and H^*, multiplicatiion by extra-terrestrial irradiance shows that $(1-a)G$ as well as H have a summer maximum and a winter minimum.

For a comparison of the conditions before and after the monsoon the budget terms are presented schematically in Fig. 2 for June and October (*see* also Fig. 1 for notations). The significant result seems to be that the energy absorbed by the soil is increased by about 30 per cent after the monsoon while the heat absorbed by the air is decreased by only 10 per cent, which means that the top albedo A in October is 40 per cent less as compared to June.

The calculated top albedos were compared with independent measurements from orbiting satellites. On special request, Dr. Vonder Haar of the Meteorology Department of the University of Wisconsin was kind enough to estimate planetary albedos for the area with 79°E and 28°N as the centre which corresponds approximately to the location of New Delhi. The sensor resolution of the satellite is such that the albedo estimates are representative for an area of no less than 250,000 square miles. The bi-monthly values (Table 5) of the satellite albedos were derived for the period of July 1964 to June 1965; none were available for 1959.

The obvious discrepancies may be caused by the albedo structure of the large area scanned by the

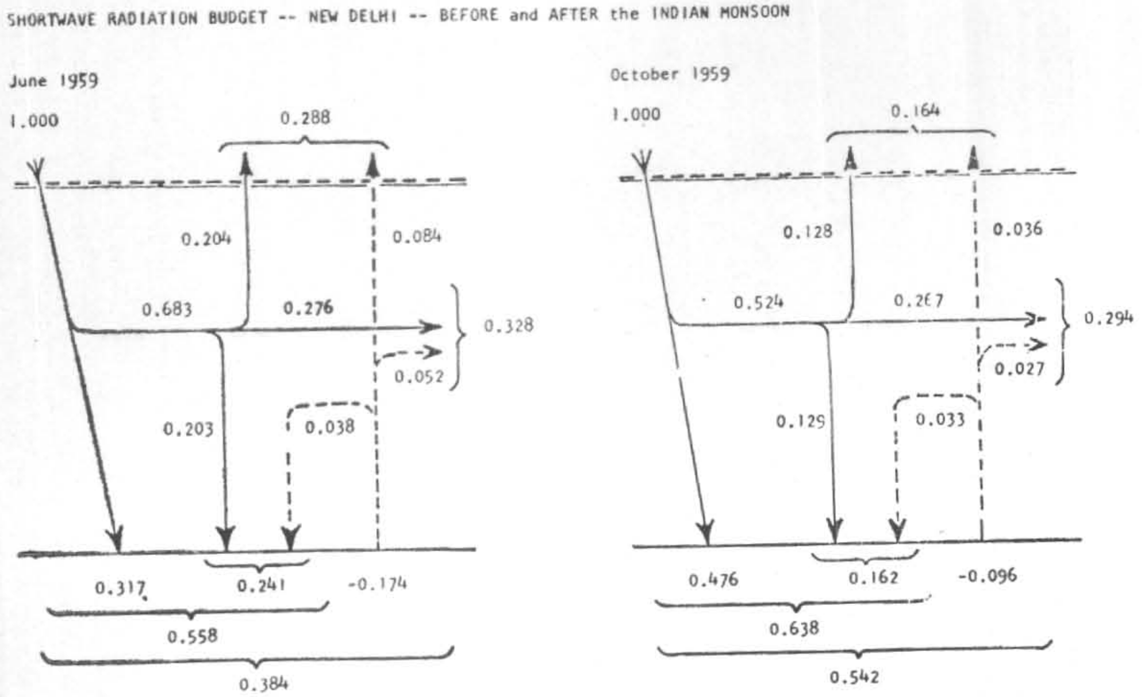

Fig. 2. Shortwave Radiation budgets for June and October 1959 at New Delhi as example of changes caused by the Indian monsoon

TABLE 3

Calculated attenuation coefficients of direct solar radiation in the atmosphere at New Delhi (1959) due to total absorption (α) and scattering (σ), showing individual contributions by neutral molecules (σ_R), water vapour (α_w), ozone (α_g), aerosol (α_{ae}, σ_{ae}) and average cloud cover ($c\alpha_c$, $c\sigma_c$).

Month	α	α_w	α_g	α_{ae}	$c\alpha_c$	σ	σ_R	σ_{ae}	$c\sigma_c$
January	0·176	0·134	0·025	0·012	0·005	0·276	0·134	0·022	0·162
February	·152	·123	·025	·001	·003	·235	·118	·052	·090
March	·206	·125	·026	·049	·006	·302	·105	·070	·180
April	·222	·120	·025	·072	·005	·292	·098	·080	·162
May	·226	·128	·026	·068	·004	·308	·093	·136	·126
June	·276	·147	·029	·086	·014	·406	·093	·194	·228
July	(·303)	·151	·028	(·096)	·028	(·505)	·093	(·192)	·420
August	·317	·168	·030	·097	·022	·380	·095	·015	·330
September	·273	·165	·030	·058	·020	·386	·101	·080	·294
October	·267	·152	·026	·085	·004	·257	·113	·059	·120
November	·256	·140	·024	·090	·002	·192	·129	·007	·072
December	·230	·133	·024	·069	·004	·247	·140	·025	·114
Annual mean	·242	·141	·027	·065	·010	·316	·109	·078	·192

Figures in paranthesis are estimated values

TABLE 4

Shortwave Radiation Budget

Extra-Atmospheric Irradiance (I), Adjusted Global (G) and Sky (D) Radiation, Portion absorbed by the Ground $(1-a)G$, absorbed in the air (H), and Reflected at the upper boundary of the atmosphere (AI). All values in ly/day.

Month	I	G	D	$(1-a)G$	H	AI
January	530	380	90	312	104	114
February	646	498	102	408	112	126
March	774	529	149	428	180	166
April	886	591	160	479	222	185
May	962	643	194	514	246	202
June	988	551	238	441	304	243
July	972	455	268	373	319	280
August	925	473	192	402	316	207
September	825	460	179	391	244	190
October	694	443	112	377	204	113
November	564	381	70	324	159	81
December	497	337	77	286	126	85
Annual mean	772	478	153	395	211	166

Solar constant = 1·980 ly/min

TA5LE 5

Bi-monthly means of Top Albedo

A_{clim} are calculated for January 1959 through December 1959 at New Delhi; A_{sat} are averaged from satellite measurements and interpolated for 250,000 square miles centered at New Delhi for July 1964 through June 1965.

	Jan/Feb	Mar/Apr	May/Jun	Jul/Aug	Sep/Oct	Nov/Dec	Annual Mean
A_{clim}	0·21	0·21	0·23	0·26	0·22	0·16	0·21
A_{sat}	0·38	0·31	0.34	0·36	0·40	0·39	0·36

satellite and also by cloud-cover differences from year to year. The values suggest that the snow-covered mountains to the north of New Delhi lie within the region scanned by the satellite. It should also be remembered that the climatic A-values were derived from budget requirements with estimated, not measured ground albedo values (a).

5. Conclusions—Contribution of aerosol to atmospheric heating

It may be asked how significantly the various absorbers of shortwave radiation contribute individually to atmospheric heating. An over-all temperature change in an atmospheric layer can be computed which is proportional to the absorbed

TABLE 6

Seasonal heating rates (°C/day) due to total shortwave radiation absorption with a break-down into constituent contributions, at New Delhi during 1959

Season	Water vapour	Gases	Aerosol	Clouds	Total
January—March	0·39	0·08	0·06	0·01	0·54
April—June	0·58	0·12	0·33	0·04	1·07
July—September	0·67	0·12	0·35	0·09	1·23
October—December	0·36	0·06	0·21	0·01	0·64
Annual Mean	0·50	0·09	0·24	0·04	0·87

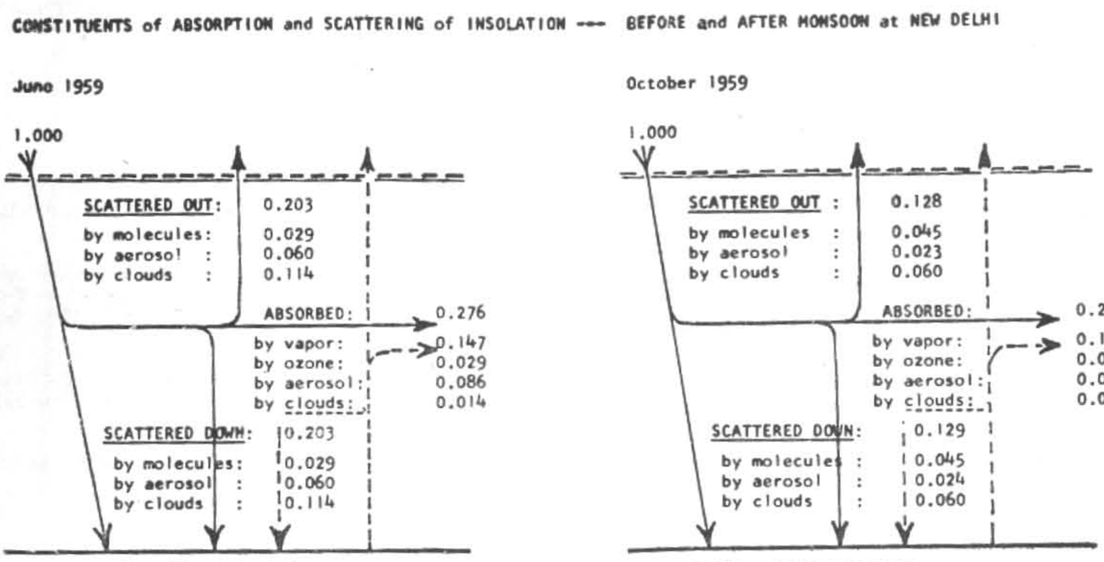

Fig. 3. Attenuation coefficient for June and October 1959 at New Delhi

Constituents of absorption and scattering are listed separately, also direct solar radiation (beam radiation)

radiation flux and inversely proportional to air density and specific heat. Changing from height to pressure by using the hydrostatic equation and taking the total pressure difference the heating rate is $g \triangle H/c_p p_o$, when p_o is the surface pressure. This relation was used to calculate the heating rates summarized in Table 6.

As could be expected, water vapour is most important. It contributes in each season more than half of the total to atmospheric heating. This fraction remains high even in the dry season. Absorption by clouds is least important. It amounts to less than 10 per cent even in the cloudy season. The annual mean of absorption by gases is also relatively low compared to water vapour and aerosol absorption.

With the exception of the winter months, the contribution of aerosol to atmospheric heating amounts to approximately 30 per cent of the total. It remains rather high during and right after the monsoon season. This could indicate that haze still prevails in the upper troposphere after some dust has been washed out of its lower portion by the monsoon rains. While cumulonimbus tops can reach great heights during the monsoon season, the average monsoon clouds lie below 10,000 feet. Interesting in this connection is an observation by Srivastava and Ronne (1966) who describe cumulus clouds imbedded in dense haze during the south-west monsoon over the Arabian Sea at 12,000 feet.

Results of calculating intensity of shortwave radiation heating in the air over New Delhi are included in the work by Gupta (1966), in a study restricted to the pre-monsoon season. His values are approximately one half of those computed with the set of equations described in this study. The discrepancy can be explained by Gupta's omission of the processes of back-scattering and absorption of energy reflected from the ground, while, simultaneously, he employs an extremely large surface albedo of 0·42 for the New Delhi region.

Fig. 3 illustrates differences before and after the monsoon in relative fractions. The absolute intensities follow from multiplication by the extra terrestrial irradiance and can be seen in Table 6, or can be computed with the aid of data in Tables 3 and 4. The major conclusion is that the relative absorption does not change significantly, while scattering effects are noticeably different. Peterson and Bryson (1968) reported that the analysis of samples from airborne dust collectors during flights over the periphery of the NW-Indian desert (in April of 1966) has shown it to be 36 per cent quartz, 22 per cent feldspar, 22 per cent carbonates and 20 per cent mica. It had been shown previously that desert aerosol tends to increase scattering, and city aerosol tends to increase absorption. On such premises, Fig. 3 suggests the conclusion that, after the monsoon rains, mineral aerosol is partly washed out, but hygroscopic haze is increased resulting in little change in the total absorption fraction.

REFERENCES

Bolsenga, S. J. 1964 Daily Sums of Global radiation for Clondless Skies. U.S. Army Material Command, Cold Regions Res. & Eng. Lab, Hanover, N. H.

Bryson, R. A., Wilson, C. and Kuhn, P. M. 1963 Some Preliminary Results from Radiation Sonde Ascents over India. *Proc. WMO IUGG Symp. on Tropical Meteorology*, Rotorua, New Zealand, N.Z. Met. Service, pp. 507-516.

Budyko, M. I. 1958 The Heat Balance of the Earth. (Transalation by N. A. Stepanova) U.S. Weather Bureau, PB 131692, Washington, D. C.

Budyko, M. I. and Kondratiev, K. Y. 1961 *Res. Geophys.*, 2, pp. 529-544, M.I.T. Press, Cambridge, Mass.

Dave, J. V. and Sekera, Z. 1959 *J. Met.*, **16**, pp. 211-212.

Gupta, M. G. 1966 *Indian J. Met. Geophys.*, **17**, pp. 101-108.

Haurwitz, B. 1934 *Harv. met. Stud.*, 1.

IGY Manual 1958 *Annals of the International Geophysical Year*, **5**, Pt. IV, Pergamon Press.

Lettau, H. and Lettau, K. 1969 *Tellus*, 21, p. 208.

Mani, A. and Chacko, O. 1963 *Indian J. Met. Geophys.*, 14, pp. 416-432.

Moller, F. 1964 *Appl. Optics*, 3, 2, pp. 157-165.

Mooley, D. A. and Raghavan, S. 1963 *Indian J. Met. Geophys.*, 14, pp. 482-485.

Neiburger, M. 1949 *J. Met.*, 6, pp. 98-104.

Peterson, J. T. and Bryson, R. A. 1968 Influence of Atmospheric Particulates on the Infra-Red Radiation Balance of Northwest India. *Proc. First National Conf. on Weather Modification*, Albany, N.Y., pp. 153-162.

Robinson, G. D. 1963 *Arch. Met. Wien*, Series B, 12, pp. 19-40.

REFERENCES (contd)

Smith, W. L.	1966	*J. Appl. Met.*, **5**, pp. 726-727
—	1963	*Smithsonian Meteorological Tables.* Washington, D.C., p. 440
Srivastava, R. C. and Ronne, C.	1966	*Indian J. Met. Geophys.*, **17**, pp. 587-590.
Vonder Haar, T. H.	1968	Ph. D. Thesis, Department of Meteorology, University of Wisconsin, Madison.
Yamamoto, G.	1962	*J. atmos. Sci.*, **19**, pp. 182-188.