A Study of Ångström's turbidity parameters from solar radiation measurements in India

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ABSTRACT

Values of Ångström's turbidity coefficient β and the wavelength exponent α have been computed for a number of stations in India from pyrheliometric measurements of direct solar radiation, for the whole spectrum and for specified spectral regions using filters OG_1, RG_2 and RG_8. Large seasonal variations are noticed in β, with high values in summer and low values in winter at all stations. Rainout and washout are effective in the removal of aerosols from the atmosphere and a marked fall is noticed in β after thunderstorms and after the onset of the monsoon at all stations.

Values of α at stations in the southern half of the subcontinent remain more or less constant throughout the year with a mean value of about 1.0, indicating that smaller haze particles predominate and the size distribution remains the same despite the large increase of turbidity in summer. In the northern half of the sub-continent, α shows seasonal variations with low values, sometimes becoming zero or negative in summer and the normally accepted values in winter. Over north and central India therefore, while smaller particles are more numerous in winter, large particles predominate in summer.

1. Introduction

It is well known that the atmosphere over the tropics is generally more turbid than that over the temperate latitudes. Ångström (1961) pointed out that his turbidity coefficient β varies with latitude and altitude according to an empirical relationship

$$\beta = \{0.025 + 0.100 \operatorname{Cos}^2 \phi\} e^{-0.7h}, \tag{1}$$

where ϕ is the latitude and h the height in km above sea level. He found a mean value for the tropics of about 0.12, which rapidly decreases with increasing latitude to about 0.06 in the temperate zones and to 0.02 or less over the Arctic and Antarctic. Ångström's empirical relationship was confirmed by Yamamoto, Tanaka & Arao (1968), who studied the hemispherical distribution of the turbidity coefficient β and suggested that the relation found by Ångström might be an alternate way of expressing the relation between β and the annual mean value of precipitable water ω, as the amount of water vapour in the atmosphere generally decreases with an increase of latitude.

Values of Ångström's turbidity coefficient β have been determined at a number of stations in India since the IGY and the results for Poona (18° N) and New Delhi (28° N) published (Mani & Chacko, 1963; Chacko & Desikan, 1967). While individual values of β were found to vary from 0.01 to 0.13, the average values of β for Poona and Delhi were only about 0.02 to 0.08. Even during the hot, dry summer months, when large and giant particles are known to be introduced on a massive scale into the atmosphere by dust raising winds and dust storms over central and north India, reducing visibility to a few km or less, the maximum values of β observed over Delhi were only about 0.13. There were either serious instrumental or observational errors leading to such low values or major errors in the assumptions made in calculating β. The computations of β had been made on the assumption that the wavelength exponent α in Ångström's empirical formula for turbidity is a constant and has a value of 1.3. A re-examination of the observational data showed that this assumption was not applicable at all stations in all seasons, particularly at Delhi during the summer months; and that there were errors in the filters used in the Poona measurements till 1965. Unreliable data have now been omitted. Values of both α and β have been re-computed using Ångström's simplified method (1961) and the results of the study of atmospheric turbidity parameters at 12 stations in India are presented in this paper.

2. Atmospheric turbidity parameters and their evaluation

In Ångström's well known empirical formula for the wavelength dependence of total haze scattering, the coefficient of extinction of solar radiation by haze is given by $m_h \cdot \beta\lambda^{-\alpha}$ where β is Ångström's turbidity coefficient, m_h is the relative optical airmass and α a constant depending on the size distribution of the haze particles. α can vary from 0 to 4, but in practice is not generally known to fall outside the limits 0.5 and 3.0. Both Ångström (1961) and Volz (1956) found $\alpha = 1.3 \pm 0.2$ to be a reasonable average value, although Ångström has shown that in a polluted atmosphere, for instance after volcanic outbreaks or forest fires, α may be as low as 0.5 or less and he emphasized that α should be determined as far as possible with β in every case.

The method of evaluating both atmospheric turbidity parameters α and β, from measurements of direct solar radiation using optical filters OG_1, RG_2 and RG_8, was developed by Schüepp (1949) and Ångström (1961). If the observed intensity of solar radiation is I_{obs}, and that reduced to the mean sun-earth distance R_0,

$$I = \left(\frac{R}{R_0}\right)^2 I_{obs} = \int_0^\infty I_0(\lambda)\,\tau_R^m(\lambda)\,\tau_{03}^m(\lambda)\,\tau_M^m(\lambda)$$

$$\times\, \bar{\tau}_{H_2O}(\lambda, m\omega, P_e, H_2O)\,\bar{\tau}_{CO_2}(\lambda \cdot mu\, P_e\, CO_2)\, d\lambda, \quad (2)$$

where

R = the instantaneous distance between the sun and earth,

$I_0 = \int_0^\infty I_0(\lambda)\, d\lambda$ = solar constant,

$\tau_R(\lambda)$ = transmission due to the Rayleigh atmosphere of unit air mass,

$\tau_{03}(\lambda)$ = transmission due to ozone included in unit air mass,

$\tau_M(\lambda)$ = transmission due to aerosols included in unit air mass,

$\bar{\tau}_{H_2O}(\lambda, mu, Pe, H_2O)$ = mean transmission by each of the near infrared water vapour bands,

$\bar{\tau}_{CO_2}(\lambda, mu, P_e, CO_2)$ = mean transmission by each of the near infrared carbondioxide bands,

m = sec z, where z is the zenith angle of the sun,

ω, u = water vapour and carbondioxide amounts respectively in the vertical air column,

P_e = the effective pressure parameter.

Assuming Junge's power law for the distribution for aerosol particles which states

$$n(r) = C \cdot r^{-\nu}, \quad (3)$$

where $n(r)dr$ is the number of aerosol particles in unit airmass with radius between r and $r + dr$, ν is a number generally about 4, and C is a constant, which is related to the total number of aerosol particles, the transmission due to aerosol particles $\tau_M(\lambda)$ is given by,

$$\tau_M(\lambda) = \exp - \int_0^\infty \pi r^2 K\left(\frac{2\pi r}{\lambda}, m^*\right) n(r)\, dr = \frac{e^{-\beta}}{\lambda^\alpha}, \quad (4)$$

where $\beta = 2\pi^2 C \int_0^\infty K(\omega, m^*)\,\omega^{-3} d\omega$, and $\omega = 2\pi r$ and $K(\omega, m^*)$ of Mie scattering and m^* is the refractive index of the particles. Junge's power law thus gives a simple and satisfactory explanation of Ångström's empirical formula for turbidity.

In Ångström's original method recommended in the IGY Instruction Manual on Radiation Instruments and Measurements, and used in India for routine determination of atmospheric turbidity, the evaluation is based on the assumption $\alpha = 1.3$ and on the measurement, using a single filter RG_2, of the solar radiation intensity in the short wavelength region $I_k = \int_{0.300}^{0.630} \times F(\lambda)\, d\lambda$ to obtain β_r. It was pointed out that although the above simplifying assumption can give somewhat incorrect values for a single measurement, the definition of β_r is based on a physically correct conception of the extinction by haze and the method has the great advantage that β_r can be easily determined, yielding usable numerical values for comparisons over time and space.

But it soon became clear that values of βr thus computed, during and after the IGY, at a number of radiation stations in India for periods varying from 2 to 12 years were not representative of actual conditions at all stations in all seasons. Fortunately regular measurements of direct solar radiation had been made at all principal radiation stations using all three filters OG_1, RG_2 and RG_8. So using the simplified method described by Ångström (1961), values of both α and β for various spectral regions have now been evaluated.

From the measurement of the intensity of solar radiation for the whole spectrum and for specified spectral regions using filters OG_1, RG_2 and RG_8 with cut-off values at 0.525 μ, 0.630 μ and 0.710 μ, solar radiation intensities for two narrow wavelength regions $\int_{0.300}^{0.525} F(\lambda)\, d\lambda$ and $\int_{0.630}^{0.710} F(\lambda)\, dy$ are first obtained. Values of β_g and β_{rr} for these two wavelengths are then evaluated on the assumption that $\alpha = 1.3$. The two values of turbidity β_{rr} and β_g for the two wavelength regions will be equal as long as $\alpha = 1.3$.

In general they are not equal, indicating that $\alpha \neq 1.3$, β_{rr} being sometimes less and sometimes more than β_g. Ångström has shown that in such cases the real value of $\alpha(\alpha_0)$ and the real value of $\beta(\beta_0)$ can be obtained from

$$\alpha_0 = 1.3 - 5.94 \log \beta_{rr}/\beta_g \tag{5}$$

assuming the weighted mean wavelengths of the two spectral regions to be 0.669 μ and 0.454 μ respectively. β_0 is given by

$$\log \frac{\beta_0}{\beta_{rr}} = (\alpha_0 - \alpha) \log \lambda_{rr}. \tag{6}$$

3. Evaluation of data

3.1. *Errors in the evaluation of α_0 and β_0*

Errors in the evaluation of α_0 and β_0 arising from the many simplifying assumptions made, have been discussed at length by Schüepp (1949) and Ångström (1963). In actual measurement errors also arise from lack of an exact knowledge of the filter constants and from the non-uniformity of turbidity conditions during the long interval of time, of the order of 15 to 20 minutes, required for a complete series of measurements with three different filters. But these are small compared with the errors involved in measuring radiation integrals for specified spectral regions using broad band pass filters.

The use of differences between several measured values of I, to obtain α_0 and β_0 infers a very high degree of precision in measurement not possible in actual practice. Even the use of the single differences $I_t - I_{R_2}$ or $I_t - I_G$ to obtain β_r or β_g, can lead to appreciable errors in their determination. And an error of $\pm 1\%$ in the measurement of each of the integrals

$$\int_{0.630}^{\infty} F(\lambda)\, d\lambda, \quad \int_{0.525}^{\infty} F(\lambda)\, d\lambda \quad \text{and} \quad \int_{0.710}^{\infty} F(\lambda)\, d\lambda$$

can introduce an error of 10 per cent in the difference $IR_2 - IR_8$, 50 per cent in β_{rr}, 100 per cent in α_0 and 300 per cent in β_0, giving absurd values of α_0 and β_0. Unless solar radiation intensities are measured to an accuracy of ± 1 per cent too much importance cannot be attached to values of β_0 and α_0 derived from pyrheliometric measurements of solar radiation. A study of β_r can, however, yield numerical values which can be compared over time and space and which are of practical value in synoptic or climatological studies of scattering by haze. And the results can be considered to provide a broad view of the transmission conditions of the atmosphere and to give information on the dependence of the atmospheric turbidity on air masses, seasons, latitude and local conditions.

3.2. *Evaluation of Ångström's turbidity coefficient β*

Values of the intensity of direct solar radiation at normal incidence, for the whole spectrum and with the three filters RG_2, OG_1 and RG^8, are obtained using Ångström's pyrheliometers at all 12 principal stations in India, whenever weather conditions permit, at all main synoptic hours of observations from sunrise to sunset. Values of β_r are as a routine computed at the stations themselves. Values of β_g, β_{rr}, α_0 and β_0 were computed from eqs. (5) and (6).

In order to make a quantitative assessment of the accuracy possible in routine observations of α and β at radiation stations, a programme of careful measurements of β was initiated at the Central Radiation Laboratory at Poona in October 1968. Thirty individual observations selected at random from the series of 150 ob-

Table 1. *Atmospheric turbidity parameters at Poona*

S. No.	Date	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B_{cal}	B_{obs}	Weather
1	9.10.68	.055	.083	.073	0.879	1.63	.064	.086	.075	3/8 cu, sc, ci
2	11.10.68	.082	.157	.230	1.465	0.34	.337	.182	.095	3/8 cu, sc, ci
3	14.10.68	.045	.021	.030	1.429	0.40	.042	.024	...	1/8 cu
4	15.10.68	.037	.043	.061	1.419	0.42	.087	.050	.055	1/8 ci
5	19.10.68	.065	.089	.038	0.427	3.45	0.16	.076	.080	2/8 ci, sc
6	24.10.68	.024	.043	.189	0.439	3.38	.083	.386	.085	1/8 ci
7	1.11.68	.046	.061	.068	1.115	1.03	.076	.067	.085	clear
8	4.11.68	.063	.091	.141	1.549	0.20	.218	.108	.078	1/8 ci
9	13.11.68	.050	.077	.065	0.844	1.73	.055	.080	.070	ci trace
10	13.11.68	.023	.039	.025	0.641	2.42	.016	0.37	.040	2/8 ci
11	14.11.68	.016	.021	.020	0.952	1.42	.019	.022	.050	clear
12	25.11.68	.072	.092	.121	1.315	0.61	.159	.105	.090	3/8 ci
13	27.11.68	.087	.113	.050	0.442	3.36	.022	.100	.087	clear
14	2.12.68	.088	.140	.143	1.021	1.25	.146	.150	.100	5/8 sc
15	3.12.68	.109	.151	.240	1.589	0.13	.381	.179	.090	2/8 ac
16	4.12.68	.040	.061	.040	0.656	2.36	.026	.059	.092	ac trace
17	9.1.69	.043	.045	.101	2.244	−0.74	.227	...	.083	3/8 ci
18	10.1.69	.102	.126	.073	0.579	2.68	.042	.118	.112	cu trace
19	18.1.69	.082	.097	.150	1.546	0.20	.232	.115	.127	1/8 ci
20	3.2.69	.047	.063	.110	1.746	−0.11	.192	...	.110	2/8 ci
21	5.2.69	.119	.165	.105	0.636	2.34	.068	.149	.130	haze
22	5.2.69	.093	.124	.130	1.048	1.18	.136	.133	.117	clear
23	6.3.69	.063	.092	.105	1.141	0.97	.120	.102	...	2/8 ac
24	18.3.69	.093	.142	.108	0.761	1.99	.082	.142	...	clear
25	14.4.69	.112	.156	.192	1.231	0.78	.236	.175	.173	clear
26	26.4.69	.137	.251	.238	0.948	1.43	.226	.264	.140	clear
27	27	.163	.254	.204	0.803	1.85	.164	.256	.195	haze
28	7.5.69	.157	.199	.202	1.015	1.26	.205	.215	.167	white sky
29	30.5.69	.083	.113	.165	1.460	0.34	.240	...	.135	2/8 sc, white sky
30	31.6.59	.066	.113	.115	1.020	1.25	.117	...	.120	cu trace, white sky

Table 2. *Mean-monthly values of atmospheric turbidity parameters at Poona*

Special measurements

Months		n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B_{cal}	B_{obs}
Oct.	1968	26	.044	.064	.075	1.17	0.9	.088	.070	.070
Nov.	1968	36	.050	.076	.079	1.04	1.2	.082	.081	.075
Dec.	1968	28	.046	.076	.090	1.18	0.9	.107	.086	.070
Jan.	1969	15	.070	.095	.116	1.22	0.8	.142	.106	.103
Feb.	1969	11	.072	.095	.109	1.15	0.9	.125	.100	.108
March	1969	9	.058	.095	.110	1.16	0.9	.127	.102	.118
April	1969	3	.107	.172	.189	1.10	1.1	.208	.192	.137
May	1969	5	.122	.169	.191	1.19	1.2	.233	.231	.162
Mean			.071	.105	.120	1.15	1.0	.139	.121	.105

servations made during October 1968 to May 1969 at Poona are presented in Table 1. The mean-monthly values of β_r, β_g, β_{rr}, β_{rr}/β_g α_0 and β_0 based on all 150 observations are given in Table 2.

It will be observed that β_g and β_{rr} are roughly equal, although individual values of β_{rr}/β_g vary from 0.19 to 5.7. The frequency distribution of β_{rr}/β_g at Poona is shown in Fig. 1, where values of β_{rr}/β_g are arranged in convenient groups of

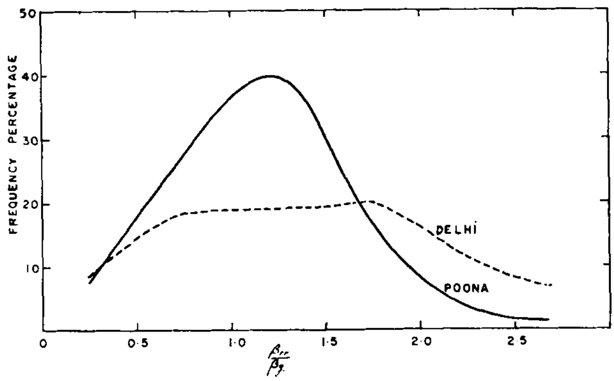

Fig. 1. Frequency distribution of β_{rr}/β_g at Poona and Delhi.

$\leqslant 0.5$, 0.6–1.0, 1.1–1.5, 1.6–2.0 and 2.0–2.5. The maximum frequency occurs in the group 1.1–1.5. The value of α corresponding to this is 1.0, not very different from the mean value assumed in the computation of β_r. It will also be seen that, in spite of the large errors which can occur in the evaluation of α_0 and β_0, careful measurements taken by skilled and experienced observers, can lead to consistent and reliable results.

This cannot be said of the routine measurements made at stations in the radiation network, as will be seen from Table 3, where values of β_r, β_g, β_{rr}, β_0 and α_0 obtained for the 12 principal radiation statins in India are presented. There are considerable observational errors leading to large values of $\beta_{rr}/\beta_g \geqslant 2.2$ and large negative values of α. Even after such values are arbitrarily omitted, the scatter is considerable. The routine observations at Poona listed in Table 3 (*a*) were made at the Central Agrimet Observatory, Poona, during 1966–67. There are significant differences in the values for Poona given in Tables 2 and 3 (*a*), some genuine, as the variations in β_r and β_g measured for different measured for different years, and some spurious, as for β_{rr} and β_0. The differences give an idea of the magnitude of the observational errors likely to occur in routine radiation measurements in α_0 and β_0 in a network of stations.

3.3. *Measurement of Schüepp's coefficient B*

Ångström's turbidity coefficient β refers to a wavelength 1.0 μ outside the visible spectrum and therefore its actual magnitude is of only abstract interest. The main advantage is that β gives a direct idea of the fraction of incident solar radiation scattered and absorbed by aerosols and when multiplied by 2 gives a direct measure of the amount of radiation scattered in the visible region.

Schüepp's turbidity coefficient B on the other hand refers to the wavelength 0.5 μ in the central part of the visible spectrum. The two coefficients are related by the formula:

$$B = \beta \cdot 2^{\alpha} \cdot \log e. \quad (7)$$

Ångström found that values of β_r, derived by means of the filter RG_2 are related to B by a constant factor which is practically independent of α and only slightly differs from unit, i.e. $B = 1.05\,\beta_r$ when $\alpha = 1.3$. For filters for other wavelength regions this will not be valid.

Measurements of the Schüepp's turbidity coefficient B using the Volz sunphotometer were made during 1966–1969 at 3 stations, Poona, Madras, and Trivandrum. The mean-monthly values of B are given in Tables 1, 2 and 3 (*a*, *h*, *k*, *m*). Values of B taken at Delhi during 1963 are included in Table 3 (*b*).

For stations where sunphotometer observations were not available B was computed from eq. (7) and Fig. 2 where values of $C = 0.434 \times 2^{\alpha}$ are plotted against α. Values of B calc. for the various stations are given in Tables 1, 2 and 3. Considering the inaccuracies involved in the measurement of α at routine radiation stations the computed values of B at all stations have to be accepted with a certain amount of reservation. That fairly reliable values of B calc. can be obtained from careful observations of B

Table 3. *Mean-monthly values of atmospheric turbidity parameters*

Routine measurements

(*a*) *Poona* (18°32′ N 73°51′ E, 555 m a.s.l.)

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B
Jan.	22	.055	.055	.055	1.00	1.3	.055	.107
Feb.	23	.051	.080	.092	1.15	0.9	.106	.115
March	30	.065	.101	.137	1.36	0.5	.186	.124
April	14	.069	.107	.121	1.13	1.0	.137	.137
May	7	.063	.098	.162	1.65	0.0	.267	.129
June	3	.053	.070	.076	1.09	1.1	.083	.120
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	...	...	...	...	...	...	...	...
Oct.	13	.057	.079	.094	1.19	0.9	.112	.127
Nov.	8	.038	.062	.097	1.56	0.2	.151	.120
Dec.	11	.055	.079	.144	1.82	−0.2	.262	.130
Mean		.056	.081	.109	1.33	0.7	.151	.123

(*b*) *New Delhi* (28°55′ N 77°12′ E, 212 m a.s.l.)

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B
Jan.	18	.085	.079	.111	1.41	0.4	.157	.157
Feb.	6	.067	.076	.093	1.22	0.8	.113	.155
March	13	.073	.082	.103	1.26	0.7	.130	.195
April	12	.087	.120	.184	1.53	0.2	.282	.250
May	25	.087	.106	.136	1.28	0.7	.174	.262
June	6	.110	.200	.272	1.36	0.8	.370	.289
July	6	.093	.169	.219	1.29	0.7	.283	...
Aug.	...	...	...	...	...	...	...	...
Sept.	18	.072	.076	.103	1.35	0.5	.139	...
Oct.	30	.073	.083	.095	1.14	1.0	.108	...
Nov.	28	.061	.079	.106	1.34	0.6	.142	...
Dec.	17	.073	.098	.100	1.02	1.3	.102	...
Mean		.081	.106	.138	1.30	0.7	.182	.218

(*c*) *Jodhpur* (26°18′ N 73°01′ E, 217 m a.s.l)

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B_{calc}
Jan.	6	.040	.049	.075	1.53	0.2	.115	.057
Feb.	4	.055	.100	.138	1.38	0.5	.190	.116
March	...	...	...	...	...	...	...	...
April	6	.064	.086	.120	1.39	0.5	.167	.102
May	13	.083	.117	.142	1.21	0.8	.172	.129
June	9	.103	.085	.125	1.47	0.3	.184	.098
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	5	.048	.111	.093	0.84	1.7	.078	.111
Oct.	10	.050	.074	.071	0.96	1.4	.068	.078
Nov.	14	.034	.062	.090	1.45	0.4	.131	.075
Dec.	17	.041	.058	.065	1.12	1.0	.073	.064
Mean		.057	.082	.102	1.26	0.7	.131	.092

(*d*) *Ahmedabad* (23°04′ N 72°38′ E, 55 m a.s.l.)

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B_{calc}
Jan.	15	.039	.066	.091	1.38	0.5	.126	.078
Feb.	5	.032	.047	.064	1.36	0.5	.087	.054
March	5	.059	.088	.144	1.64	0.1	.236	.108
April	5	.055	.082	.130	1.93	−0.4	.251	...
May	7	.057	.078	.080	1.02	1.3	.082	.088
June	1	.064	.080	.118	1.47	0.3	.173	.092
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	1	.079	.080	.113	1.41	0.4	.159	.091
Oct.	4	.050	.077	.074	0.96	1.4	.071	.081
Nov.	5	.033	.037	.053	1.43	0.4	.076	.043
Dec.	9	.043	.071	.077	1.08	1.1	.083	.077
Mean		.051	.071	.094	1.37	0.6	.134	.079

(*e*) *Bhavnagar* (21°45′ N 72°11′ E, 5 m a.s.l.)

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B_{calc}
Jan.	20	.037	.055	.077	1.46	0.3	.112	.159
Feb.	19	.025	.041	.060	1.51	0.3	.116	.061
March	18	.045	.062	.092	1.66	0.0	.128	.055
April	14	.050	.077	.124	1.61	0.1	.200	.092
May	...	...	...	...	...	...	...	...
June	5	.077	.097	.181	1.90	−0.3	.344	...
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	5	.056	.075	.115	1.45	0.4	.167	.095
Oct.	10	.051	.065	.103	1.61	0.1	.166	.076
Nov.	18	.036	.052	.083	1.61	0.1	.134	.062
Dec.	12	.033	.047	.076	1.32	0.0	.128	...
Mean		.046	.063	.101	1.57	0.1	.166	.071

(*f*) *Nagpur* (21°06′ N) 79°03′ E, 308 m a.s.l.)

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B_{calc}
Jan.	1	.063	.035	.032	0.91	1.5	.029	.036
Feb.	21	.067	.042	.113	2.27	−0.8	.257	...
March	12	.057	.122	.091	0.75	2.0	.068	.118
April	11	.064	.139	.085	0.61	2.5	.052	.129
May	15	.077	.157	.174	1.11	1.0	.193	.166
June	6	.084	.116	.129	1.11	1.0	.143	.123
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	...	...	...	...	...	...	...	...
Oct.	6	.044	.125	.113	0.90	1.6	.102	.135
Nov.	15	.047	.066	.101	1.53	0.2	.155	.077
Dec.	3	.056	.096	.042	0.44	3.4	0.18	.085
Mean		.062	.100	.098	0.98	1.5	.113	.109

Table 3 (*cont.*)

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B_{calc}
(*g*) *Calcutta* (22°39′ N 88°27′ E, 4 m a.s.l.)								
Jan.	...	...	...	...	...	...	...	...
Feb.	14	.106	.130	.287	2.21	−0.7	.634	...
March	25	.110	.131	.202	1.54	0.2	.311	.155
April	20	.099	.159	.225	1.41	0.4	.317	.181
May	12	.118	.134	.233	1.74	−0.1	.405	...
June	1	.113	.246	.208	0.85	1.7	.177	.251
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	...	...	...	...	...	...	...	...
Oct.	5	.075	.106	.162	1.53	0.2	.248	.124
Nov.	11	.085	.114	.116	1.02	1.3	.118	.127
Dec.	10	.075	.137	.095	0.69	2.2	.066	.132
Mean		.098	.145	.191	1.32	0.7	.285	.162
(*h*) *Shillong* (25°34′ N 91°53′ E, 1500 m a.s.l.)								
Jan.	4	.031	.048	.076	1.62	0.1	.123	.057
Feb.	8	.039	.062	.101	1.63	0.1	.165	.076
March	1	.044	.046	.082	1.78	0.1	.146	...
April	3	.029	.076	.100	1.32	0.6	.132	.086
May	3	.055	.077	.052	0.67	2.3	.035	.075
June	...	...	...	...	...	...	...	...
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	...	...	...	...	...	...	...	...
Oct.	1	.013	.037	.062	1.67	0.0	.104	.045
Nov.	1	.007	.032	.014	0.44	3.4	.006	.028
Dec.	3	.109	.092	.105	1.13	1.0	.119	.102
Mean		.030	.061	.074	1.28	0.9	.104	.067
(*j*) *Visakhapatnam* (17°48′ N 83°14′ E, 41 m a.s.l.)								
Jan.	1	.066	.061	.130	2.13	−0.6	.277	...
Feb.	...	...	...	...	...	...	...	...
March	12	.097	.208	.311	1.49	0.3	.463	.245
April	3	.061	.143	.173	1.21	−0.2	.209	...
May	1	.061	.250	.163	0.65	0.2	.106	.053
June	...	...	...	...	...	...	...	...
July	...	...	...	...	...	...	...	...
August	2	...	.045	.055	1.22	0.8	.067	.050
Sept.	...	...	...	...	...	...	...	...
Oct.	1	.071	.032	.030	0.94	1.5	.028	.034
Nov.	3	.066	.104	.184	1.77	−0.1	.326	...
Dec.	3	.065	.153	.187	1.22	0.8	.228	.171
Mean		.070	.125	.154	1.33	0.5	.213	.111

	n	β_r	β_g	β_{rr}	$\frac{\beta_{rr}}{\beta_g}$	α_0	β_0	B
(*k*) *Madras* (13.00′ N 80°11′ E, 10 m a.s.l.)								
Jan.	3	.047	.088	.091	1.03	1.2	.094	.056
Feb.	13	.063	.127	.102	0.80	1.9	.082	.050
March	11	.088	.139	.153	1.10	1.1	.168	.050
April	8	.062	.117	.108	0.92	1.5	.099	.062
May	8	.087	.159	.217	1.36	0.5	.295	.090
June	3	.060	.117	.180	1.54	0.2	.277	.105
July	1	.052	.207	.109	0.53	2.9	.058	.035
August		...	...	...	...	...	...	...
Sept.	...	...	...	...	...	...	...	...
Oct.	2	.059	.196	.257	1.31	0.6	.337	.046
Nov.	1	.026	.048	.107	2.23	−0.7	.239	.051
Dec.	2	.047	.061	.116	1.90	−0.3	.220	.048
Mean		.059	.126	.144	1.27	1.0	.187	.055
(*l*) *Goa* (15°29′ N 73°49′ E, 55 m a.s.l.)								B_{calc}
Jan.	11	.052	.051	.052	1.02	1.3	.053	.057
Feb.	...	...	...	...	...	...	...	...
March	...	...	...	...	...	...	...	...
April	...	...	...	...	...	...	...	...
May	3	.069	.087	.141	1.62	0.1	.228	.105
June	...	...	...	...	...	...	...	...
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	1	.016	.046	.068	1.45	0.4	.099	.056
Oct.	6	.053	.070	.055	0.79	1.9	.043	.070
Nov.	8	.033	.069	.052	0.75	2.0	.039	.068
Dec.	9	...	.108	.078	0.72	2.1	.056	.105
Mean		.044	.072	.074	1.06	1.3	.086	.077
(*m*) *Trivandrum* (08°29′N 76°57′ E, 60 m a.s.l.)								
Jan.	6	.079	.079	.039	0.49	3.1	.019	.090
Feb.	8	.071	.097	.082	0.85	1.7	.070	.103
March	3	.051	.125	.131	1.05	1.2	.138	.119
April	2	.083	.127	.073	0.57	2.7	.042	.093
May	...	...	...	...	...	...	...	...
June	...	...	...	...	...	...	...	...
July	...	...	...	...	...	...	...	...
Aug.	...	...	...	...	...	...	...	...
Sept.	...	...	...	...	...	...	...	...
Oct.	1	.057	.078	.037	0.47	3.2	.017	.048
Nov.	3	.047	.077	.028	0.36	3.9	.010	.065
Dec.	1	.056	.088	.009	0.10	...	.001	.072
Mean		.063	.098	.071	0.67	2.6	.042	.080

is illustrated in Tables 1 and 2 where the last two columns give the observed and calculated values of B.

4. Results

Direct solar radiation measurements can be made only when the skies are relatively clear and the turbidity data therefore relate to a particular meteorological condition. The data also give only daytime turbidity values and do not represent all conditions, particularly those in winter when strong inversions exist at night. No observations are generally possible in India during the four monsoon months June to September when the skies are overcast with low clouds and the data are limited to the remaining eight months October to May.

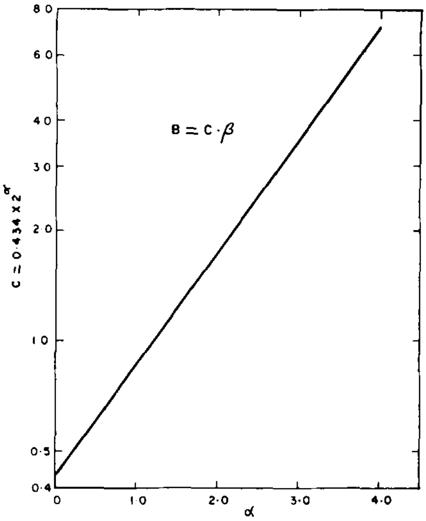

Fig. 2

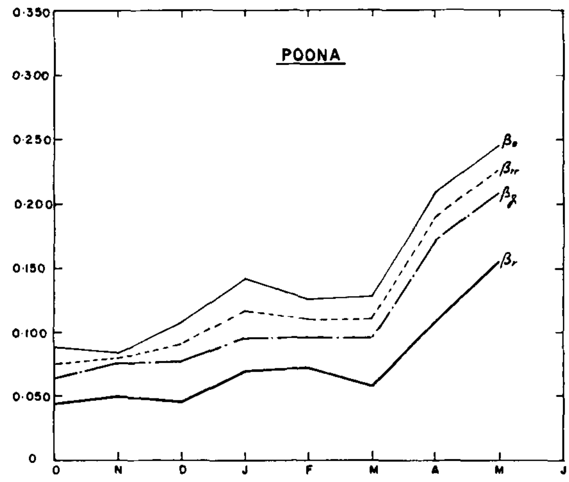

Fig. 3

Fig. 3. Annual variation of turbidity coefficient β_0, β_{rr}, β_g and β_r at Poona 1968–1969.

4.1. *Turbidity measurements at Poona*

The individual values of β_r, β_g and β_{rr} listed in Table 1 show marked diurnal, day-to-day and seasonal variations but the various β values themselves are roughly equal and practically independent of the wavelength interval over which they are determined. The ratio β_{rr}/β_g which determines the value of α_0 and the wavelength exponent α_0 itself shows no seasonal variations. The mean-monthly value of α_0 is remarkably consistent, with a mean value of about 1.0 ± 0.2, although individual values range from -0.7 to $+3.5$. This is not very different from the value generally accepted as valid over a wide range of conditions and indicates that the wavelength dependence of the scattering of the atmosphere over Poona has approximately the classical $\lambda^{-1.3}$ proportionality and the size of the scattering particles consequently is roughly about the same as that found generally for high altitude stations where Junge's power law holds good.

While the wavelength exponent α_0 is constant during the year turbidity shows a striking seasonal variation, with maximum values in the hot, dry premonsoon "summer" months April–May and minimum values during the "winter" months October–November. There is a steady increase in β_r from October (0.04) to May (0.15) and in β_0 from 0.08 to 0.23. So while the particle concentration increases in summer to three to four times the winter value, the particle size distribution remains basically unchanged throughout the year and is independent of β.

The pronounced maximum in β in the hot dry summer months indicates that the source of the aerosols is of entirely local origin, and turbidity depends on conditions in the first few km of the atmosphere and the dispersion of dust from the ground to higher levels. Even during these highly turbid months the turbidity can show remarkable sudden falls in value, for example immediately after a thunderstorm when the sky which was a drab white turns a startling blue with values of β_r as low as 0.005. Rainout within clouds and washout below clouds are known to be especially effective in the wet removal of aerosols from the atmosphere. Turbidity also shows a sharp fall as soon as the monsoon sets in, although it rises again whenever the monsoon is weak and the effectiveness of rainout or washout processes is reduced. The values remain low throughout winter.

4.2. *Turbidity measurements at Delhi*

Conditions over Delhi (28° N), in close proximity to the Rajasthan desert, are completely different from those over Poona (18° N), a hill station 555 metres above sea level. The values of β_r, β_g, β_{rr} and β_0 at Delhi are about two to three times those over Poona and the seasonal variations are also more pronounced, with maximum values in summer and minimum values in winter (Fig. 4). As a result of the late arrival of the monsoon over Delhi, the fall in turbidity occurs in July at Delhi and not in June as at Poona. The passage of western disturbances during winter also contributes to the low values in winter over Delhi.

The ratio β_{rr}/β_g is about 1.3 and varies with the seasons, with generally higher values in summer and lower values in winter. The frequency distribution of β_{rr}/β_g for Delhi is shown in Fig. 1. The maximum frequency occurs in the group 1.5 to 2.0, with corresponding values of α of the order of 0.2 or less. If we take the seasons separately, it will be seen that in winter the values of α are not far from the classical value, while in summer, α_0 can be often zero or negative. The marked seasonal variation in both β and α shows that both the particle concentration and the particle size distribution vary with the seasons at Delhi and that smaller particles are more numerously represented in winter when the turbidity is low and large numbers of particles of large size predominate in summer when the turbidity is high.

Studying two years' data at Delhi and a few months' data at Poona, Rangarajan (1966) came to the conclusion that aerosol scattering over India in summer is independent of wavelength and aerosol particles at New Delhi and possibly at many locations in India have a particle size distribution different from that in places in the temperate latitudes. He concluded that the increase of turbidity in summer, being caused by the generation of dust in the lower atmosphere by surface winds and the transport of this dust by convection and turbulance into the higher levels, the dust particles with sizes larger than normal aerosols found in the atmosphere was largely responsible for bringing down the values of α to near zero.

Ramanathan & Karandikar (1949), who carried out a comprehensive study of the effect of haze scattering on the measurement of atmospheric ozone with a Dobson spectrophotometer, found that α is either zero or even negative in summer over Delhi. They measured δ and δ' the extinction coefficients due to large particles scattering at two wavelengths $0.330\,\mu$ and $0.448\,\mu$ and found that while $\delta - \delta'$ is positive in winter, the effect of haze being similar to that over Europe, in summer it is negative, i.e. the depletion of solar radiation by haze scattering was larger for $0.448\,\mu$ than at $0.330\,\mu$. They also found that while $\delta - \delta'$ was negative in summer, its actual value was never very large, even with the haziest of skies and so they concluded that the scattering by haze is for the most part independent of wavelength. An increase in haze scattering at shorter wavelengths and the negative value of α was ascribed to an increase in mean particle size in summer over north India, when light of higher wavelength is scattered more than that of shorter wavelength.

That a deep, dense layer of dust and haze lies over central and north India during the hot premonsoon months, with a maximum concentration from March to June, is well known. Bryson *et al.* (1963), who revived interest in the phenomenon and made an extensive study of it, called it "a great brown sea lapping against the Himalayas", extending as it does across the whole of north India, with its top at about 10 km over Rajasthan desert, lowering to about 5 km between Delhi and Calcutta. A double horizon is a common phenomenon clearly visible from aircraft during this season over north India with slant visibility down to almost zero. IAF aircraft observations made in 1963 showed that the haze varies from day to day in density, depth and spatial distribution, being sometimes layered and sometimes uniform up to 10 km. These visual observations were confirmed by Bryson and his colleagues during their dust patrol over India in 1966. They found the thickest layer to be confined to 3 km with less dense layers upto 5 km over north and central India in April–May.

The description of the atmospheric haze over north India by Ramanathan & Karandikar (1949) cannot be bettered and is quoted verbatim below:

> The character of the haze over North and Central India during the dry months of the year undergoes a marked change from winter. The winter haze is dry ground haze composed of inorganic and organic particles, mainly from places of human habitation. It is light brown in colour, generally settles during the night and rises up in the forenoon and gets dissipated by thermal turbulence and mixing with upper winds by noon. In the hot days of

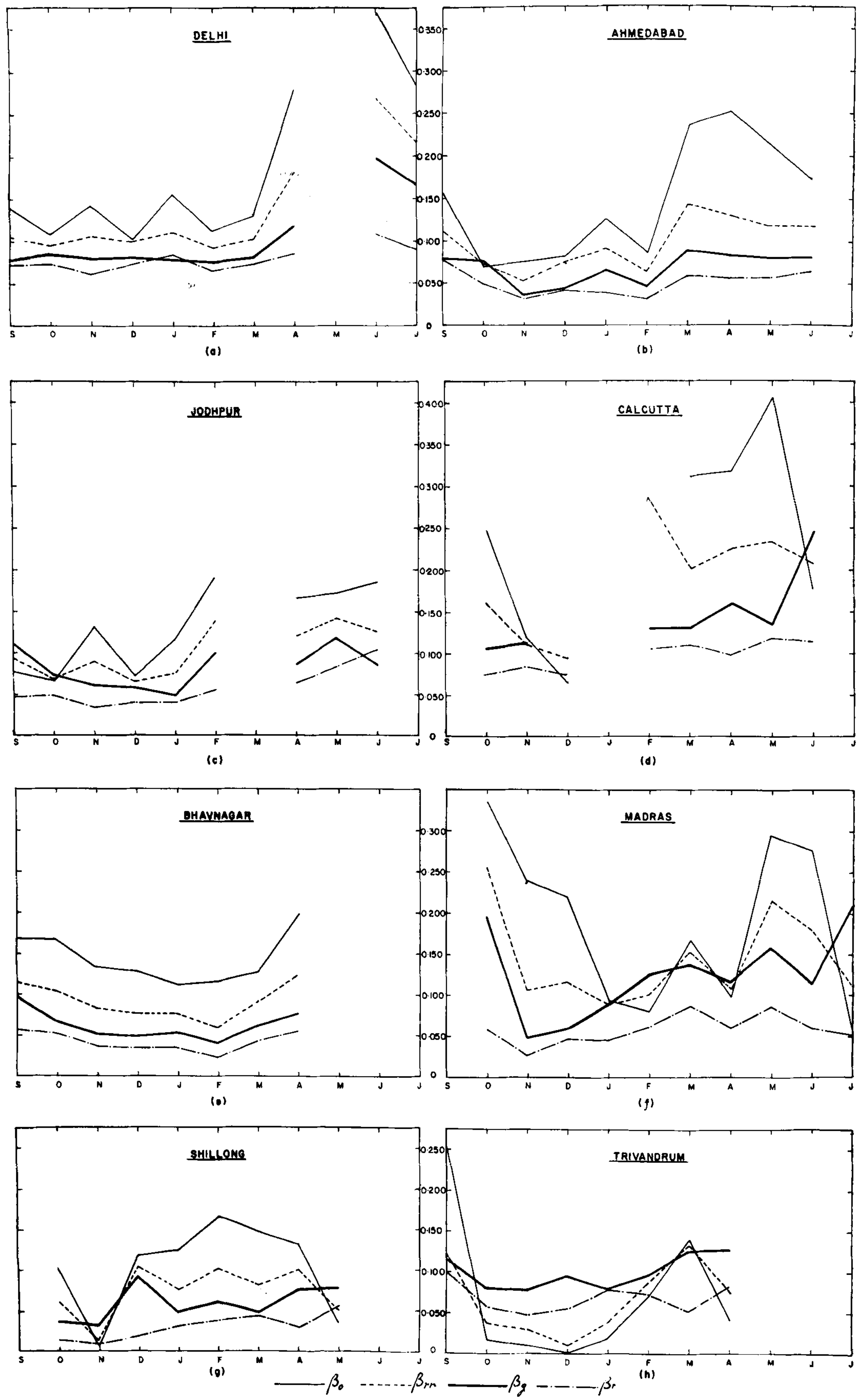

Fig. 4. Annual variation of turbidity coefficient β_0, β_g, β_{rr} and β_r.

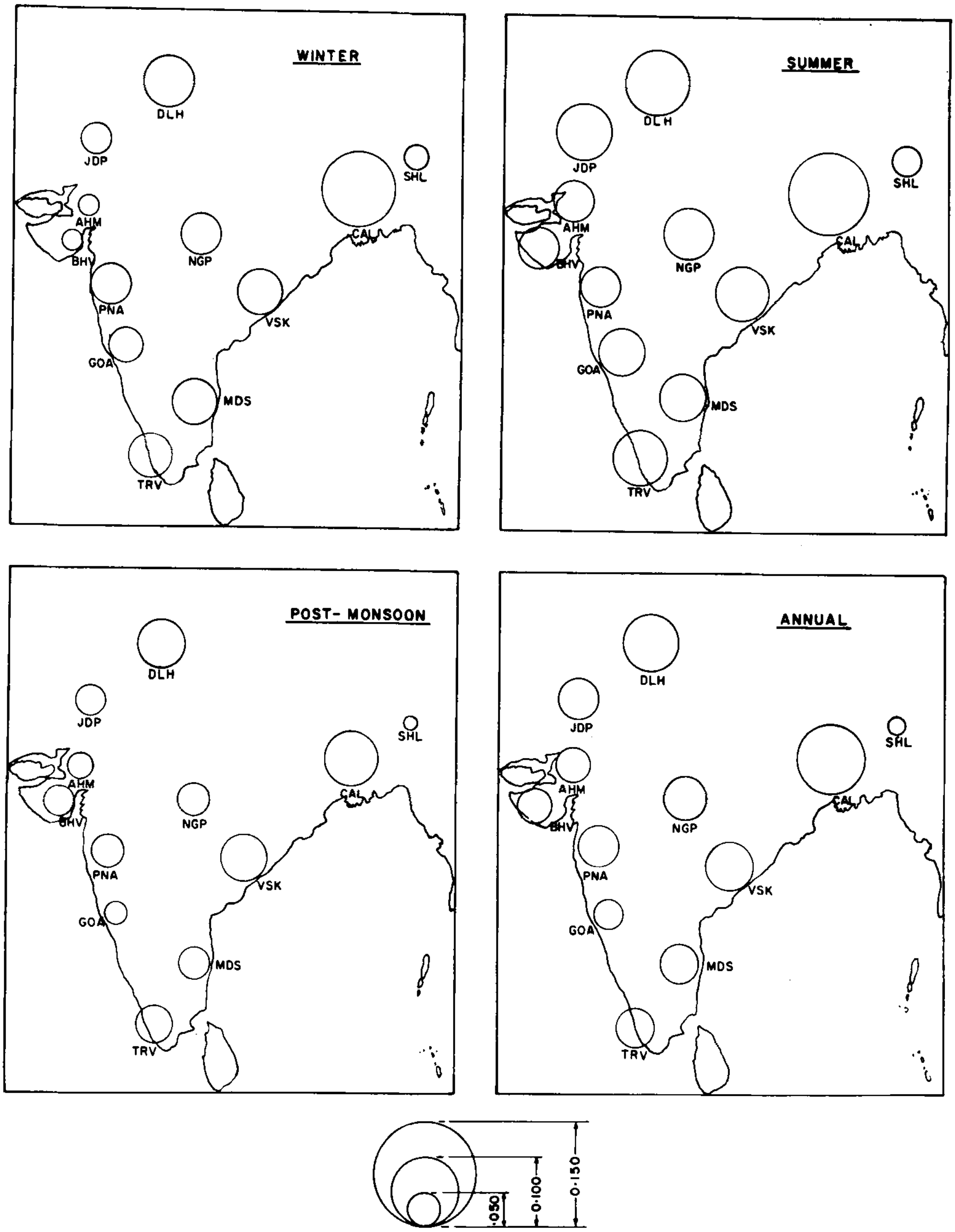

Fig. 5. Distribution of turbidity coeffient β_r.

the premonsoon season, the haze is generally thicker, of a more permanent character and whitish in colour. Lapse rates are high in the first 4 or 5 km and with the rather frequent occurrence of dust devils and dust storms and sometimes thunderstorms, the dust extends upwards into the atmosphere and when the upper winds weaken, the haze persists over a large part of north India as a milky canopy. At small angles from the sun, the scattered light from the sky is of the same order of intensity as the direct transmitted light from the sun.

The whitish appearance of the scattered light itself would indicate the neutral nature of haze scattering in summer.

A series of radiometer soundings were made over central and north India in April 1963 at 4 stations Srinagar (34° N), Delhi (28° N), Ahmedabad (24° N) and Poona (18° N) to study the radiative properties of the haze layer (Bryson *et al.*, 1963; Mani & Sreedharan, 1965). The results suggested that the emission from the dust is comparable in importance to that of carbondioxide in the atmosphere and a significant fraction of that due to water vapour. Bryson *et al.* assuming that the dust particles were in the diameter range 0.25 μ to 1.0 μ found the particle densities over Poona, Delhi and Ahmedabad to be of the order 30–180 per cm^3 for particles having a diameter of 1 μ and 500–3000 per cm^3 for particles having a diameter of 0.25 μ in the surface layer 3 km thick. In the lowest 100 mb layer they found the particle concentration to be as high as 200–450 per cm^3 for the 1 μ particles and 2000–7000 for the 0.25 μ particles. These densities are in good accord with the visibility ranges encountered and in the range of particle numbers measured by Junge (1962) over Hydebad (17° N) in 1961.

4.3. *Turbidity at coastal stations*

Turbidity values at 5 coastal stations, Bhavnagar (22° N), Goa (15° N), Visakhapatnam (18° N), Madras (13° N) and Trivandrum (8° N), are given in Tables 3 (*e, l, j, k, m*). The annual march of β at Bhavnagar, Madras and Trivandrum is shown in Fig. 4. All stations show fairly high values of β_r ranging from 0.03 to 0.09 although the atmosphere is comparatively dust-free, in contrast to the continental stations to the north. As pointed out by Yamamoto *et al.* (1968) β increases with an increase in precipitable water content and should be high at humid coastal stations. Bhavnagar and Goa are much less turbid than Trivandrum, Madras and Visakhapatnam to the south. Apart from different local factors which influence turbidity at these stations the water vapour content is also higher than at the northern stations.

4.4. *Turbidity at hill stations*

As may be expected, turbidity is low at high altitude stations. At Shillong (26° N), 1500 metres above sea level, β_r has a value of only 0.01 in winter and 0.06 in summer. The corresponding values for Poona (555 m.a.s.) are 0.04 and 0.07 respectively.

Values of β obtained at another high altitude station, Kodaikanal (10° N), 2340 m above mean sea level, on 1 February 1965, were as follows: $\beta_r = 0.015$, $\beta_g = 0.047$, $\beta_{rr} = 0.050$, $\alpha_0 = 1.1$, $\beta_0 = 0.05$. β is as low as at Shillong and α has the classical value characteristic of high altitude stations. At Kodaikanal, in the mornings the skies are a deep blue, with the haze lying like a brown blanket over the plains below. In the afternoons the haze rises as a result of increased turbulent convection and the dispersal of dust to higher levels, to the mountain top and skies become hazy.

Turbidity measurements in the Everest region Silver Hut Glacier, 28° N, 6200 m.a.s.l, on the other hand, gave unexpectedly high values of the order of 0.05 (Ångström & Drummond, 1966). But Bishop's observations were made in April–May when a dense haze layer lies over central and north India and extends 6–10 km into the atmosphere. The values obtained at Silver Hut Glacier are also not different from values obtained at Shillong in the same months.

4.5. *Schüepp's coefficient B*

The annual variation of B at 4 stations, Delhi (28° N), Poona (18° N), Madras (13° N) and Trivandrum (8°N), is shown in Fig. 6. The homogenized haze layer corresponding to these observed values of B extends to 2600 m. Considering all factors, particularly the extreme simplicity of measuring B with the Volz sunphotometer and the large errors likely in routine measurements of β_0 from pyrheliometric measurements, the sunphotometer is to be preferred for routine use at radiation stations in a national network.

4.6. *Variation of α*

At coastal stations like Trivandrum and Goa and high altitude stations like Shillong, Poona and Kodaikanal α is fairly uniform throughout the year, and has a value about 1.0, indicating a predominance of small particles at these stations. At continental stations like Delhi, Jodhpur and Ahmedabad α is generally of this order in winter. In summer, α is 0.5 or less, falling to zero or becoming negative on many occasions.

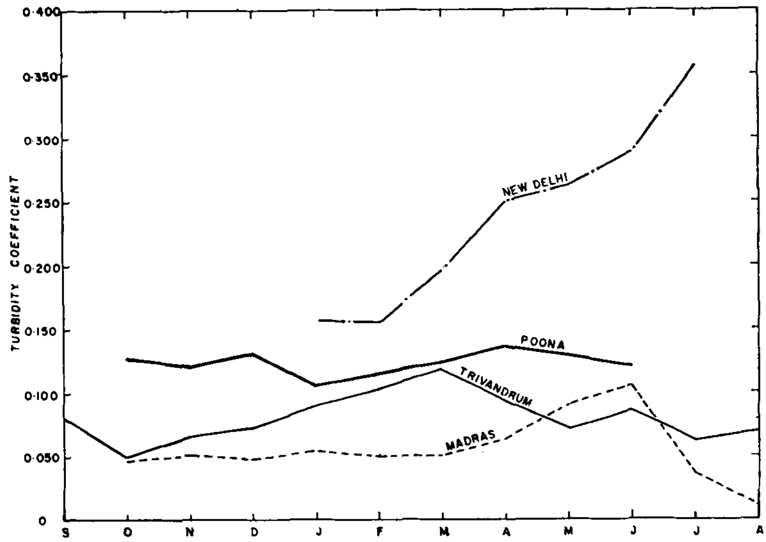

Fig. 6. Annual variation of turbidity coefficient B at Delhi, Poona, Trivandrum and Madras 1966–67.

It may generally be concluded that in the southern half of the subcontinent, the classical particle size distribution is maintained on the whole throughout the year with values of α about 1.0. In north and central India, such conditions exist only during winter. In summer α falls to zero or becomes negative indicating neutral haze scattering and the presence of predominantly large particles.

4.7. *Spatial distribution of β over India*

The geographical distribution of β_r during the winter, summer and postmonsoon seasons and for the whole year is shown in Fig. 5 (*a–d*). Yamomoto's method (1968) of graphic representation has been followed since atmospheric turbidity changes occur generally in mesoscale and the drawing of isopleths of β_r is not quite justified, particularly when one considers the sparseness of both stations and data. The diameter of each circle indicates the seasonal average value of β_r for that station.

It will generally be seen that for the whole year β_r values are higher over central and north India than over the peninsula and over the subcontinent, as a whole, they are highest during the hot, dry summer months and least during winter. Calcutta is an exception with high turbidity values, obviously the result of high local industrial pollution.

4.8. *Turbidity in air masses*

The air over western, central and north India in the "summer" months March to May is tropical continental air, *Tc*, with its source region in southwest Asia. It is the driest and hottest air over the Indian sub-continent, with marked instability, intense insolation and turbulence, leading to the development of dust-raising winds and duststorms which persist for days reducing visibility.

With the arrival of premonsoon thunderstorms and the establishment of the monsoon over the country, a highly humid, cool and convectively indifferent equatorial maritime air, *Em*, lies over south and central India during the monsoon months June to September, characterized by mostly covered to overcast skies, frequent rain and drizzle and visibility fair to good, except in heavy rain. In the north this air mass becomes modified as it turns west round the seasonal trough to the north.

During the winter months December to February, the modified cold dry air over north and central India is of continental polar origin, *Pc*, characterized by convective stability, clear skies and good visibility, except for local mist in the morning and occasional haze in the afternoon. This air mass over the Deccan acquires in its travel across the country to the south tropical characteristics, especially after stagnating in the anticyclonic field over tropical latitudes for some time.

Representative values of the atmospheric turbidity coefficient β for the three air masses, *Tc* over western, north India in summer, *Pc* over north and central India in winter, and *Em* over south and central India during the monsoon and postmonsoon months have been calculated from values of β for typical stations in the north, centre and the south of the subcontinent. Delhi, Jodhpur, Ahmedabad and Nagpur have been taken as representative of the north, western and central parts of the country and Goa and Trivandrum of the south. The air mass over Madras and Vizag on the east coast of the peninsula is modified *Em* and data for these southern stations have not been considered in calculating β for *Em*.

During the non-summer months, when α has the generally accepted value of 1.3, turbidity coefficient β_r gives realistic values of the turbidity and mean values of β_r for *Pc* and *Em* are given below. β_r values for summer over north and central India are too low (0.08) and therefore the mean value of β_0 for the four northern and central stations is given below:

	β_r	β_0
Tc	0.08	0.20
Em	0.05	—
Pc	0.05	—

5. Conclusion

A study of the atmospheric turbidity parameters from pyrheliometric measurements of solar radiation at a number of stations in India has yielded the following results.

1. Reliable values of both the wavelength exponent α and the turbidity coefficient β can be obtained from pyrheliometric measurements of solar radiation for the whole spectrum and for specified spectral regions using broad band-pass filters, if the observations are carefully made by skilled observers. At routine radiation stations such measurements cannot be considered to be reliable but measurements of β_r serve to give a broad picture of the aerosol climatology of the subcontinent, except in the north, where careful measurements of both α and β are necessary to obtain realistic values of β.

2. Turbidity over India shows a marked annual variation, with very high values in summer about 2 to 3 times those of the values in the post-monsoon and winter seasons.

3. At all stations during winter the wavelength exponent α has approximately the classical value of 1.3, indicating a particle size distribution with predominantly small particles.

4. In the southern half of the subcontinent, not affected by industrial pollution, α has roughly the same value during the summer months as well, indicating that no large change occurs in particle size distribution, despite the increased atmospheric turbidity in summer.

5. In north and central India during summer, the atmosphere is 3 to 4 times turbid as in the rest of the country. Values of α are very low, often becoming zero or negative, indicating a predominance of large particles. Under these conditions, haze scattering is independent of wavelength and neutral.

6. Reduction of turbidity as a result of rainout and washout in precipitation is observed at all stations, both in the premonsoon thunderstorm season and during the monsoon months. The values of β corresponding to the tropical maritime, tropical continental and polar continental air masses that lie over India during the monsoon, summer and winter seasons are about 0.05, 0.20 and 0.05 respectively.

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ИЗУЧЕНИЕ ПАРАМЕТРОВ МУТНОСТИ АНГСТРЕМА С ПОМОЩЬЮ ИЗМЕРЕНИЙ СОЛНЕЧНОЙ РАДИАЦИИ В ИНДИИ

Для ряда станций в Индии из пиргелиометрических измерений прямой солнечной радиации были вычислены величины коэффициента мутности Ангстрема β и показатель спепени α в зависимости от длины волны. Измерения проводились как для всего спектра, так и для отдельных спектральных областей при использовании фильтров OG, RG_2 и RG_8. На всех станциях отмечены большие сезонные вариации величины β с максимумом летом и минимумом зимой. Величины α для всех станций южной половины страны остаются более или менее постоянными в течение года, имея среднюю величину 1,0, что указывает на преобладание более мелких частиц дымки и на неизменность распределения частиц по размерам несмотря на большое увеличение замутненности летом. В северной части страны величина α проявляет сезонные вариации с минимумом, достигающим иногда нуля, летом и обычными значениями зимой. Поэтому над северной и центральной частями Индии крупные частицы превалируют летом, в то время как более мелкие частицы более многочисленны зимой.