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Variations in Ångström's wavelength exponent at Pune

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सार — सौर प्रकाश प्रकीर्णन और सुविशिष्ट अवशोषण प्रक्रियाओं के कारण वायुमंडल में अपने मार्ग से गुजरते हुए क्षीण हो जाता है। स्पैक्ट्रम के दिखाई पड़ने वाले भाग में भिन्न परमाणु कणों के कारण प्रकीर्णन होता है। स्पैक्ट्रम के विकिरणमापीय अध्ययनों से विलोपकारी कारकों का पता चलता है, जिससे प्रत्येक माप के लिए तरंगदैर्ध्य घातांक α के मानो के निर्धारण में सहायता मिलती है। α मान 1.3 के क्लासिक मान से काफी कम है, जिसका अब आंग्स्ट्राम आविलता गुणांक β के निर्धारण में प्रयोग किया जा रहा है। α कामान सामान्यतः 0.5 के लगभग बराबर है। विस्तृत कणों, विशेष रूप से आर्द्रताग्राही के विद्यमान होने पर मान शून्य के लगभग पहुँच जाते हैं। यहाँ तक कि उनके नकारात्मक मान भी हो सकते हैं, जिससे अवशोषण प्रक्रिया द्वारा अवरक्त तरंग दैर्ध्य में सौर प्रकाश के क्षीण होने की संभावना के संकेत मिलते हैं।

ABSTRACT. Solar irradiance in its passage through the atmosphere gets attenuated by scattering and selective absorption processes. Particulate matter cause scattering in the visible part of the spectrum. Spectral radiometric studies enable the determination of the extinction factors and hence the values of wavelength exponent α for each measurement. α values are much lower than the classic value of 1.3, now being used in the determination of Ångström Turbidity Coefficient β. The value of α is generally found to be of the order of 0.5. In presence of large particles, particularly hygroscopic ones, values reach nearly zero and even have negative values, suggesting the possibility of attenuation of irradiance in red wavelengths by absorption processes.

Key words - Spectral turbidity coefficients, Pyrheliometer measurements, Variations in wavelength exponent values.

1. Introduction

The solar irradiation, in its passage through the atmosphere, gets attenuated by selective absorption and scattering by the atmospheric constituents. The absorption factor is not significant in the visible part of the spectrum, the scattering processes becoming more dominant. The attenuation of the irradiation may be expressed, following Beer's Law, as

$$S = \frac{1}{d} \int_0^\infty S_{0\lambda}\, e^{-A(\lambda)}\, d\lambda \qquad (1)$$

$A(\lambda)$ is expressed as the total attenuation due to Rayleigh Scattering $A_R(\lambda)$, aerosol (Mie) scattering $A_D(\lambda)$ and the water vapour extinction $A_W(\lambda)$. $A_D(\lambda)$ varies inversely as the fourth power of wavelength, λ, $A_W(\lambda)$ depends on the highly variable water vapour content which attenuates the irradiation both by absorption and scattering. Hence it is preferable to study the scattering of the irradiation by limiting it to the visible wavelength ranges. Ångström expressed the aerosol scattering as

$$A_D(\lambda) = m_h\, a_D(\lambda) = m_h\, \beta\, \lambda^{-\alpha} \qquad (2)$$

where, α is the wavelength exponent, m_h is the relative optical airmass and β is the atmospheric turbidity coefficient.

The wavelength exponent α is assigned a value between 0 and 4. α will be equal to 4 under Rayleigh atmospheric conditions (Robinson 1966). Thus when the atmosphere contains a larger quantity of small particles α will tend towards 4 and towards zero when larger particles are predominant. Ångström found that a value of 1.3 for α obtained after extended measurements, will represent the reasonably mean atmospheric conditions. However, on account of industrial development activities, it has been realised that the use of 1.3 as the mean value of α would represent wrong atmospheric conditions (Mani *et al.* 1969 and Rangarajan 1970).

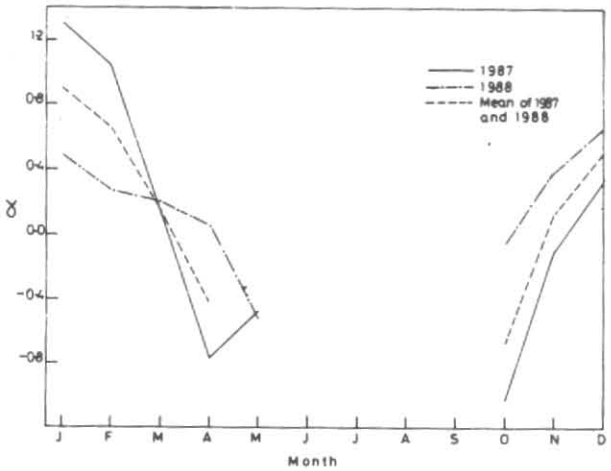


Fig.1. Variation of wavelength exponent during 1987 and 1988

Ångström (1961, 1963) put forward techniques for determining the value of α from radiometric data obtained in different spectral ranges. Spectral measurements of direct solar irradiances can be made using a pyrheliometer and a set of three broad band - pass filters. The three filters used are the OG_1, filter with band pass wavelengths 525-2900 nm, RG_2 filter with 630-2900 nm and RG_8 filter with 700-2800 nm. Measurements without any filter give the irradiances in the range 290-3000 nm. Thus it is possible to work out the spectral irradiances in the narrower bandwidths 300-525 nm and 630-700 nm also. Knowing the extra-terrestrial values and also the effects of Rayleigh scattering and ozone absorption, it is possible to evaluate the integrals over these wavelength ranges using the Beer's law and derive the values of β corresponding to given value of α. The transmission of irradiation through the aerosols alone then at any two given wavelengths λ_1 and λ_2 can be expressed as,

$$\rho_{\lambda_1} = e^{-m_h \beta_1 \lambda_1^{-1.3}} = e^{-m_h \beta_0 \lambda_1^{-\alpha_0}} \quad (3)$$

$$\rho_{\lambda_2} = e^{-m_h \beta_2 \lambda_2^{-1.3}} = e^{-m_h \beta_0 \lambda_2^{-\alpha_0}} \quad (4)$$

Here β_1, and β_2 are turbidity coefficients assuming α to be 1.3. α_0 is the true value of α at the time of measurement and β_0 is the corresponding value of β derived from α_0. In the case of the spectral measurements in the range 300-525 nm and 630-700 nm, λ_1 has a weighted mean value of 454 nm and λ_2 a value of 669 nm β_1 and β_2 are then designated as β_g and β_{rr} respectively for easier identification. Combining the equations (3) and (4), we get,

$$\alpha_0 = 1.3 - 6.02 \log \frac{\beta_{rr}}{\beta_g} \quad (5)$$

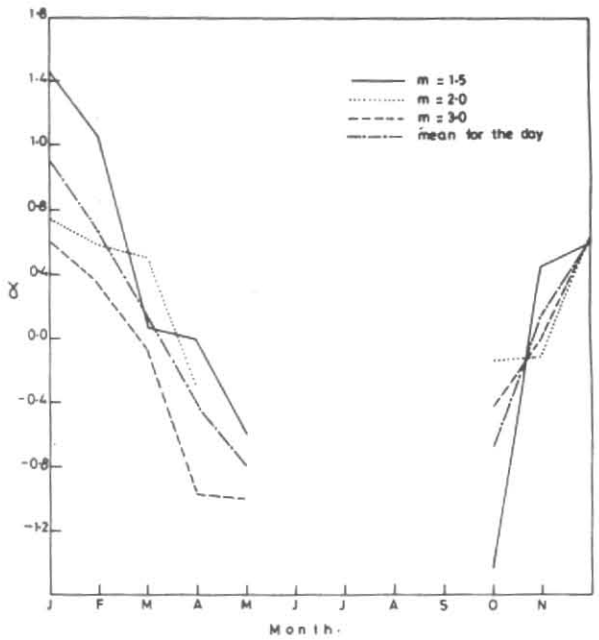
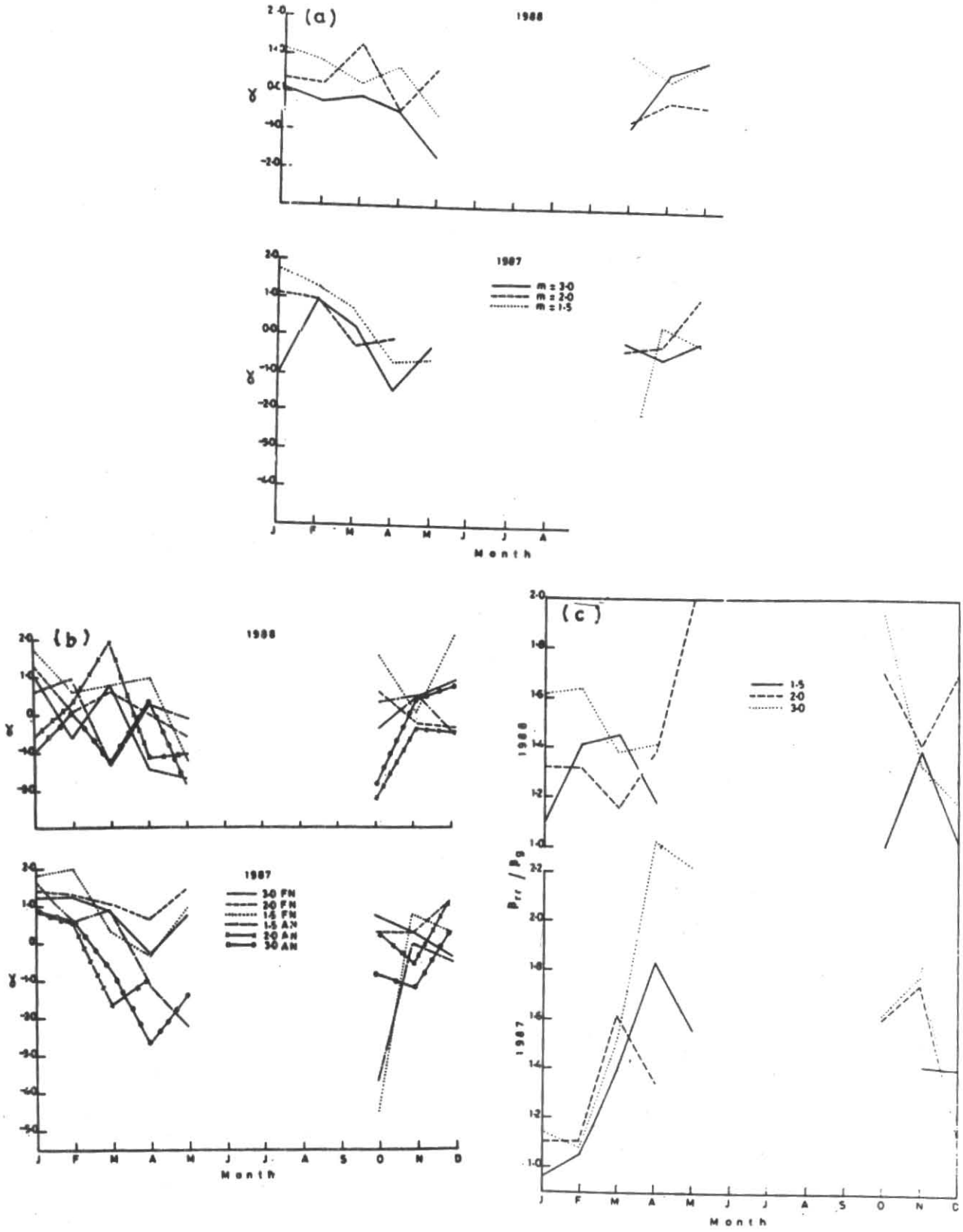


Fig.2. Mean variation of α for different air masses

From the knowledge of and it is possible to evaluate α_0. Nomograms prepared by Rangarajan (1970) were used for a quick evaluation of β_g and β_{rr}. It is found that so long as $\beta_g \approx \beta_{rr}$ value of α_0 is around 1.3. When $\beta_g > \beta_{rr}$, the value of α_0 becomes less than 1.3. If the ratio $\beta rr / \beta g$ is greater than 1.7, α_0 values become negative.

With the irradiance measurements at Pune and elsewhere, some authors have discussed the changes in the values of α. Rangarajan (1970) studied the values of α from pyrheliometric measurements and again (Rangarajan 1972) made with a two filter sunphotometer. Mani *et al.* (1969) made a very interesting study of values of α derived from pyrheliometric measurements made in the Indian radiation network. Again Mani *et al.* (1977) made special measurements at various high altitude locations in India and discussed the different ranges of α values. Prabhu (1989) made a detailed study of atmospheric turbidity measurements made at Pune. The present paper makes an attempt to present the values of α obtained from spectral irradiance measurements carried out at the Pune network station during 1987 and 1988. It is pointed out that the annual rainfall during 1987 has been near normal, though it was below normal during the monsoons. It was excess in 1988 but there had been no significant rains outside the four monsoon months (Table 2). The paper tries to bring out the effective roles of rains in the dispersal of the pollutants and of the water vapour content in the atmosphere.



Figs.3(a-c). (a) Variations in for different airmasses during 1987-1988 (b) Diurnal variations in α during 1987 and 1988 and (c) Mean variation of β_m/β_g for different air masses during 1987 and 1988

TABLE 1
Monthly rainfall over Pune (mm)

Year	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
1987	5.2	2.5	Tr	4.0	116.8	72.7	61.0	223.6	23.1	166.4	8.3	35.3
1988	0.0	Tr	0.0	4.8	15.6	67.0	299.9	84.1	431.1	0.3	0.0	0.0

TABLE 2
Mean water vapour pressure (hPa) during February and November 1987 and 1988

	February		November	
Airmass	1987	1988	1987	1988
3.0 FN	12.7	13.3	17.0	13.5
2.0 FN	12.7	13.8	16.6	12.9
1.5 FN	11.2	12.4	17.3	12.8
1.5 AN	10.5	8.4	13.7	10.2
2.0 AN	10.7	9.4	13.4	10.4
3.0 AN	11.1	10.1	15.7	10.4

2. Results

Fig.1 shows the annual variation of α. These values are means of the measurements made in the forenoons and the afternoons. The value of α is close to the classical value of 1.3 in winter but steadily decreases from January to May, becomes nearly zero or even negative during April-May. This must be due to dust particles of larger sizes during this period. In the mean α is nearly equal to 0.9 in January and it gradually decreases to 0.7 in February. The March values drop down to 0.2 and the mean value for April even becomes negative. The October value remains negative many times despite the efficient scavenging by the rains during the monsoon. This then should be attributable to small water droplets or large hygroscopic particles that may remain suspended during October. The gradual removal of moisture by dry air causes a sharp increase of more than 120 per cent to reach a mean value of 0.14 in November. α continues to increase from November to January by 82 per cent to reach nearly the unity value.

The pyrheliometer measurements are made at three optical air masses 1.5, 2.0 and 3.0, both in the forenoons and afternoons. Fig.2 gives variations in α for different air masses. α values show the general tendency of steady increase in the aerosol sizes from January, though to different degrees for different air masses. The restoration in the α values in the post monsoon season shows the tendency in the reverse direction. The sudden drop in α for m=1.5 from February to March by more than 200 per cent is noteworthy. The generally increasing sizes of aerosols register their effect on α for all airmasses during the premonsoon period. The increasing tendency in α values are more sharp for m=1.5 during October-January period. The noteworthy point is the convergence of α values to 0.5 in December for different airmasses. This should then show the stable nature being reached in December.

Fig.3(a) shows the mean variations in the value of α for different airmasses during the two years 1987 and 1988. A notable feature in January 1987 is the nearly classical value of 1.3 for α. It is as high as 1.7 for m = 1.5. The decrease in the values starts from February onwards itself. With the steady increase in the solar elevation as the days advance, the dust load in the sky also increases from March onwards. The values of α become highly variable, obviously due to the changing clouding activities during the period and the consequent changes in the aerosol concentrations. The heavy rains (113 per cent more than the normal) in October 1987 have restricted the number of observations during the month possibly giving rise to a very high negative value of α (more than -4.0) for an airmass of 1.5 due to the prevailing conditions obtainable at that time. An interesting feature seen in 1987 is the general increasing trend in the value of α during November and December.

December 1987 had a similar precipitation regime as December 1986 (both had about 35 mm rain during the month). January-February 1988 did not register any rainfall and perhaps this has caused the α values during these months in 1988 to be lower than those of the corresponding period in 1987. The atmosphere also remained drier in 1988 during February (Table 2) in the afternoons perhaps inhibiting the growth and consequent removal of the small sized hygroscopic particles.

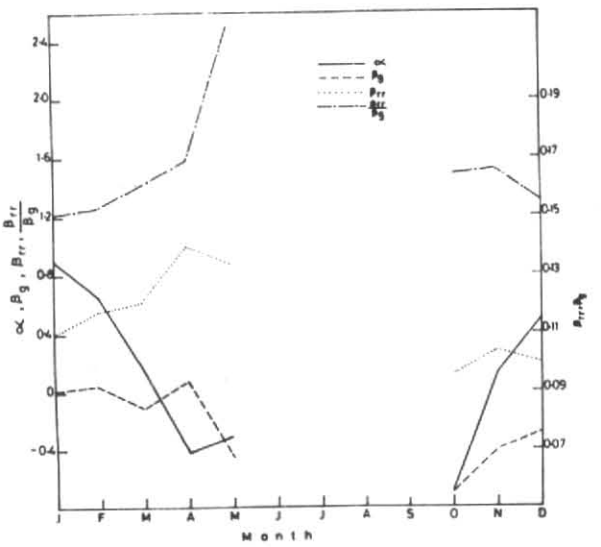


Fig.4. Mean variation of α, β_{rr}, β_g and β_{rr}/β_g

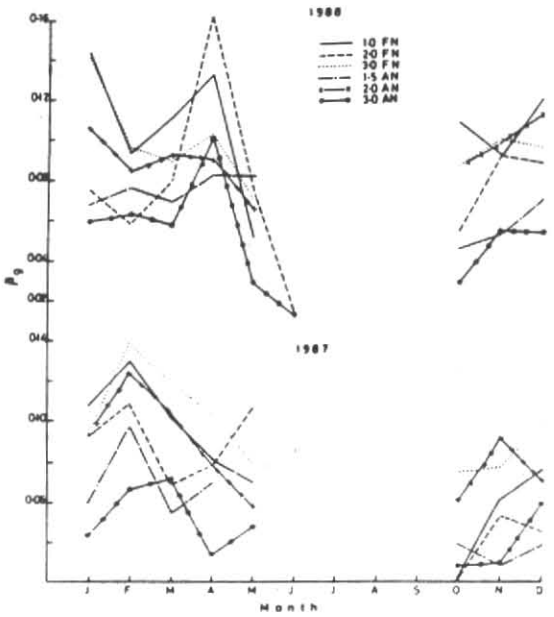


Fig.5. Mean variation of β_g for different airmasses during 1987 and 1988 (FN & AN)

A perusal of Fig.3(b) gives a better insight into such values of α as seen in Fig.3(a). The general tendency seen is the steady increase in the value of α from sunrise and a somewhat rapid decrease in the values becoming more negative in many cases by the evening time. The obvious inference would be that the larger particles in the early hours of the day get dispersed and the smaller particles predominate the aerosol particles as the noon time approaches. The values decrease in the afternoon indicating the quick removal of the predominant smaller particles, perhaps by growth into larger particles. This is particularly so during the cloudy months, indicating the possible role of the moisture in the removal of smaller particles which probably grow by absorption of the water vapour. 1987 had rains during almost all months of the year. This perhaps was one of the reasons for the low values (mostly negative) of α especially in the afternoons. April 1987 shows that the initial build up of the small particles got arrested in the forenoon itself. May 1988 on the other hand shows the α values to be negative through out the day. With nearly no rains during the month and with the prevailing higher humidity conditions, the wet haze particles which will be larger in size must be dominating the particulate concentration in the atmosphere.

October 1987 shows a rather very high negative value for α both in the forenoon and the afternoon for $m = 1.5$. This particular month had excess rain (113 per cent more) which probably provided a larger concentration of large size hygroscopic particles. Large size hygroscopic particles obviously attain the size by the absorption of water vapour and they become almost water droplets as the hygroscopic substance may get completely dissolved. In such circumstances absorption of the red radiation in the wings will come into play, causing a very low transmission in the wavelength range 630-700 nm. The low irradiances in the red wavelengths give rise to high values of β_{rr} and low values of β_g as can be seen from Figs. 5 & 6. These striking features are not seen in October 1988 which had recorded no rains. The sudden increase in α for $m = 2.0$ in the afternoon should be due to the removal of the large particles by the cloud formation activities leaving smaller particles that still remain to play their roll in the attenuation processes. This can be easily inferred from Figs 5 and 6, which had a low 0.02 value at $m = 1.5$ in the afternoon showed sharp increase to 0.04 for $m = 2.0$ and then reduced to 0.027 in less than an hour by the time the airmass became 3.0. β_{rr} on the other hand dropped from 0.08 at $m = 1.5$ AN to 0.045 at $m = 2.0$ AN. β_{rr} further got reduced to about 0.027 for $m = 3.0$ AN.

October 1988 in contrast shows negative values only towards the evenings. A look into the variations in βg and βrr show that the sharp increase seen in α for $m = 1.5$ FN should be due to sharp increase in the induction of dust particles of small size which however start growing immediately, by inference, by accretion processes as the values of βrr show sharp decrease during this period again obviously due to coagulation processes. The growth of the particle beyond this stage is probably taken over by the adhesion of moisture to the particulate nuclei. βrr almost doubles while βg remains nearly stable.

3. Discussions

The values of α are found to be nearer the classical value of 1.3 during winter while in summer they become lower and lower towards zero. This can only be due to the variations in the particle size distribution with seasons. The

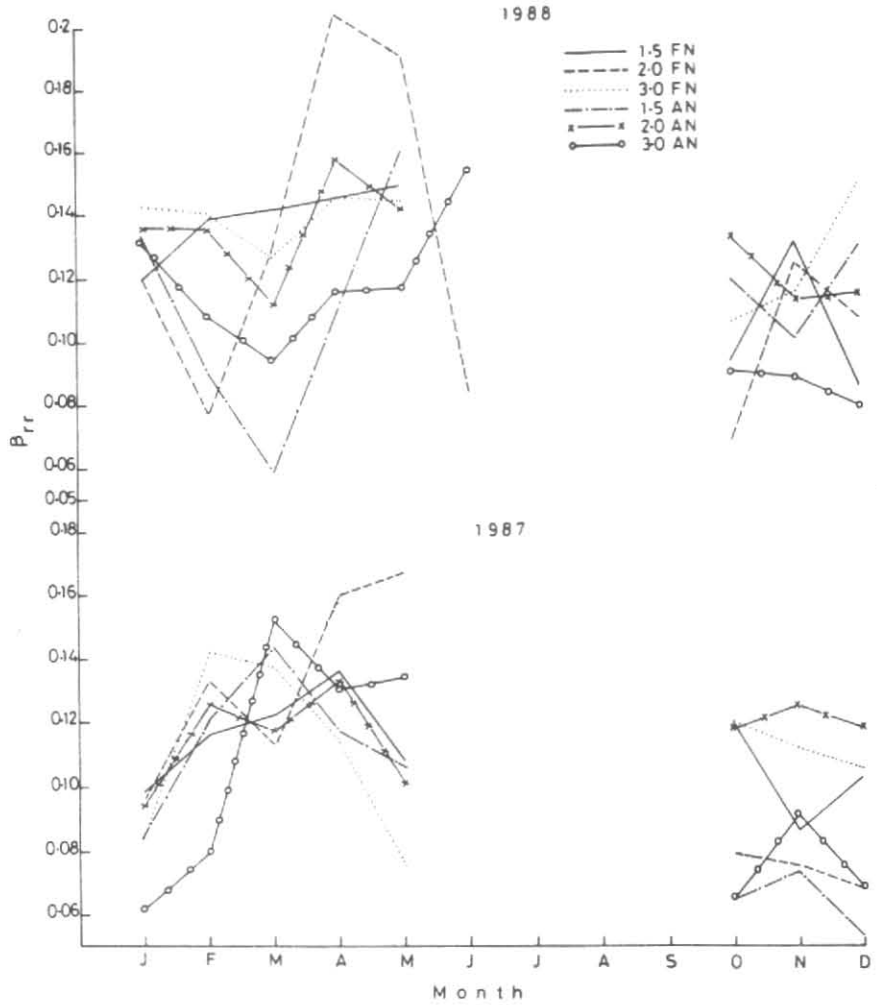


Fig.6. Mean variation of β_{rr} for different air masses during 1987 and 1988 (FN & AN)

larger particles get washed down by the monsoon and post monsoon rains and smaller particles predominate in winter when the turbidity is low. As the winter gives place to summer, the build up of aerosol particles begins; leading to higher concentration of larger particles in the lower layers and resulting in high values of turbidity.

Rangarajan (1970) in his delightful discussions concludes that the aerosol scattering over India in summer is independent of wavelength and aerosol particles. He argues that the increase of the turbidity in summer, being caused by the generation of dust in the lower atmosphere by surface wind and the transport of this dust by convection and turbulence into the higher levels, the dust particles with larger sizes were responsible for bringing down the values of α to near zero. Rangarajan states: "when the wavelength exponent α attains a value of zero or nearly zero, it means that the scattering of light by the aerosol particles is independent of wavelengths. In other words, the haze extinction becomes practically neutral", Ångström (1963) suggested that the size of the particles emanating from the ground and from combustion in the plains is, in general, smaller than the particles of the more pure atmosphere above the mountain stations. He had studied the measurements made at Potsdam and Davos. In marked contrast, the α values over India reveal a minimum during summer. The origin of the large sized aerosols, Rangarajan reasons, could be: (*i*) fine dust raised from ground through convection and thermal instability (during the premonsoon period) and (*ii*) the condensation of water vapour on the hygroscopic nuclei, which could lead to an increase in the number of the larger particles.

Mani *et al.* (1969) got a mean value of 1.0 for α after rejecting some of the measurements suspected to be due to observational errors. This would seem to be rather contentious rejection as the values presented in Table 1 of the work under reference do contain negative values, as high as –0.74

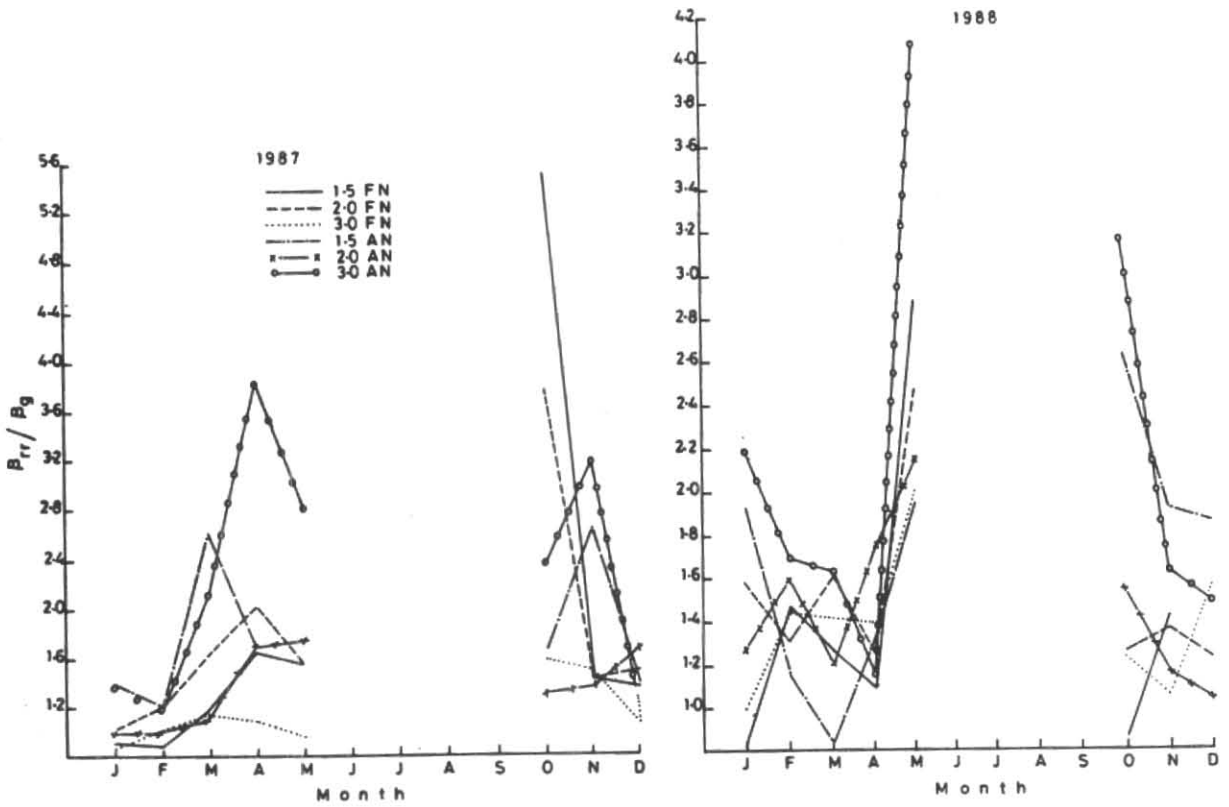


Fig.7. Mean variation of β_{rr}/β_g for different air masses during 1987 and 1988 (FN & AN)

in one occasion. These measurements were stated to have been taken by skilled and experienced observers and with great care. Elsewhere in measurements made at Gulmarg, a very high altitude station in India. Mani *et al.* (1977) do report obtaining negative values, though very close to zero. Thus the negative values of α seem to be possible and the measurements leading to the negative values can not be rejected to be due to observational errors.

It has already been stated that the values of β_g and β_{rr} have been derived using a value of 1.3 for α. When $\beta_g \approx \beta_{rr}\alpha$ remains nearly 1.3. When this ratio β_{rr}/β_g becomes larger than 1.7, α values become negative. The treatment of extinction has been centred around the assumption that only scattering is the process that can cause extinction in the wavelengths shorter than 700 nm. Besides the large particles becoming neutral, the attenuation may be due to a small contribution by absorption in the wings of red regions of the spectrum. The hygroscopic particles when they grow by accretion and coalescence do get diluted and become water droplets, sufficient to cause scattering as well as absorption. An examination of the values of β_g, β_{rr} and β_{rr}/β_g do support such a possibility.

Fig.3(c) shows that the mean value of β_{rr}/β_g to be quite high in October - of the order of 1.7 for optical airmasses 2.0 and 3.0 and this drops by 5 per cent for m = 2.0 in November (Prabhu 1989). The corresponding drop for m = 3.0 is 12 per cent. Interestingly the value for m = 1.5 shows an increase of 39 per cent from 1.012 to 1.411. This makes one to infer that the hygroscopically large particles coagulate to become smaller in size as the solar elevation angles increase and then again grow in size in the late afternoons by coalescence and by accretion. When the moisture content in the air decreases, the ratios β_{rr}/β_g are nearly the same for all the airmasses.

The individual study of β_g and β_{rr} gives a lending hand to the above supposition. Fig.4 shows an increasing trend in the value of β_{rr} from January to May while β_g remains low and has a tendency to a lower value, however small it may appear to be. The values of β_{rr} in the most monsoon period are nearly even. The water vapour content in October is higher as moisture field persists then. The much lower atmospheric temperatures in November and December cause the humidity to increase and thus the values of β_{rr} remains even. β_g values on the other hand are quite low, 0.055 in October and register a sharp increase of 27 per cent to 0.077 in November. The time lag in building up of the aerosols in the atmosphere explains the sharp increase in βg.

A careful study for individual years 1987 and 1988 alongwith rainfall distribution gives support to the possibility of having β_{rr}/β_g more than 1.7 and in every case the role

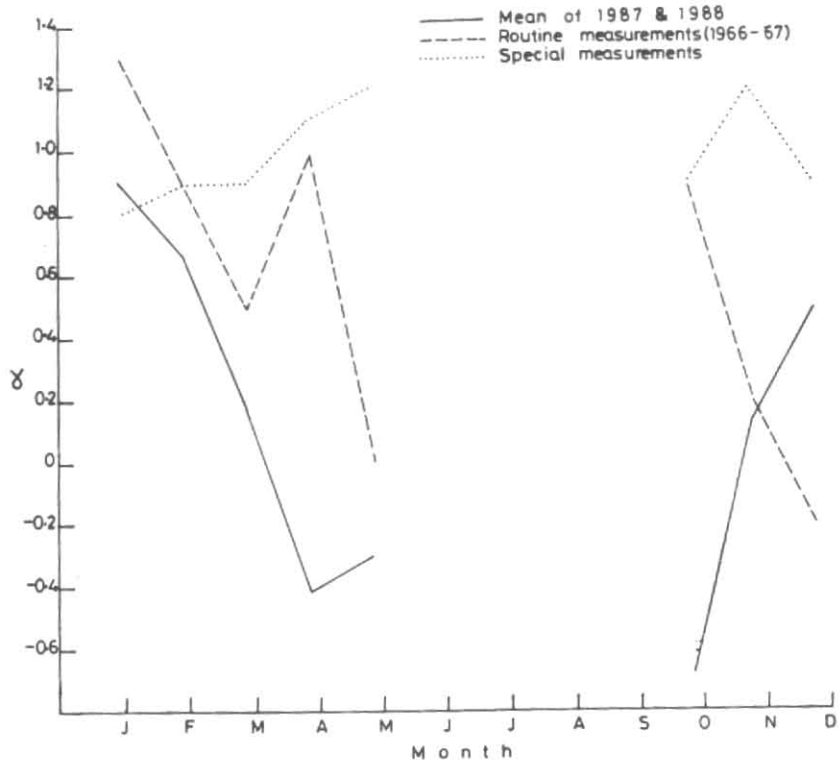


Fig.8. Comparative study of α obtained at different times

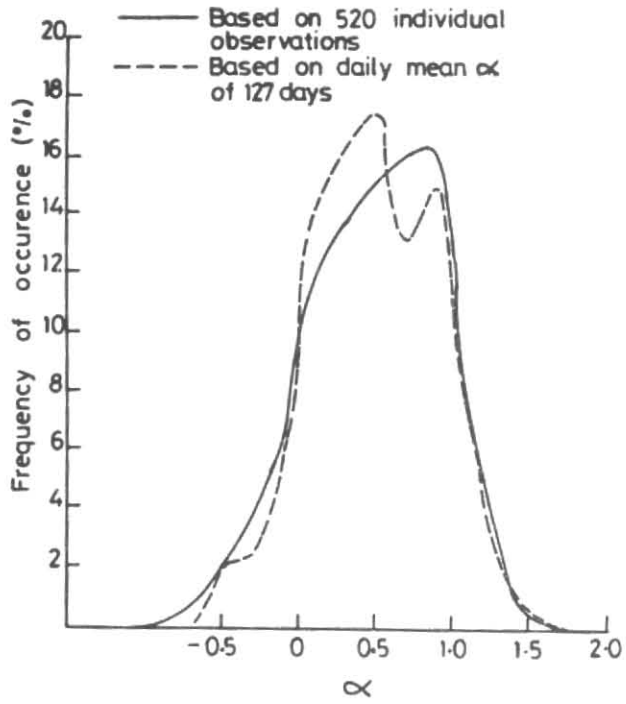


Fig.9. Frequency distribution curve for α during one year period (Rangarajan 1972)

of higher incidence of moisture could be inferred. Fig.5 shows the values of β_g for different air masses during 1987 and 1988. It is seen that β_g values are consistently low for 1987 when compared with 1988 values. The only exception is February when 1988 values are lower. A study of vapour pressure values indicates that February 1988 was more humid in the forenoons than the afternoons. 1987 on the other hand had a more uniform water vapour content in February (Table 2). It is also seen that β_g has somewhat closer values for different airmasses during 1987 as compared to the values in 1988. 1987 had rains nearly every month as compared to 1988 (Table 1). Also the values of β_g for 1987 are consistently higher than for 1988. Besides Fig.5 reveals that β_g is lowest for $m = 3.0$, more probably due to the removal of smaller hygroscopic particles which grow larger in the presence of available moisture. β_g values are generally higher in the forenoons than in the afternoons, both in 1987 and 1988. The difference in 1988 is more marked because of the lack of enough rainfall and lower prevailing water vapour in the atmosphere. A reference to Krishnamurthy (1991) indicates that the irradiances in the range 300-525 nm during the same period varied in many cases in the opposite sense to those values in the infrared range.

The variations in the values of β_{rr} should be more because it involves the selective absorption by ozone and by water vapour in the wings of the spectrum, 630 to 700 nm. The scattering effect will be of a lower order as compared to the wavelengths less than 525 nm. β_{rr} steadily increases from January (0.110) to May, April making a quantum jump over March, due to moisture incursions. Fig.6 shows the cleansing effect by the monsoon activity. The October values of β_{rr} drops by 28 per cent from its May value. β_g drops by 19 per cent only in the corresponding period.

β_{rr} has large variations for $m = 1.5$ and shows a steep drop of 40 per cent after the monsoon. The peak value of 0.162 of April is more than 55 per cent over the January value. The variations are least (10 per cent) for $m = 2.0$ and they are 26 per cent for $m = 3.0$.

The values from January to May are uniformly low during 1987 (excepting February) as compared to 1988 values. 1988 had an extended drier spell during this period, thus releasing more large sized particles from the soil. The lower β_{rr} values of 1987 must be due to the good rainfall distribution, thereby keeping down the dust load. Similarly the absence of rain during the post monsoon period of 1988 has allowed the continued presence of large sized particles especially during October leading to higher β_{rr} values.

In the mean, the ratio β_{rr}/β_g is always greater than unity, indicating thereby that $\alpha = 1.3$ is not obtainable at Pune. Except for the period February to May, β_{rr}/β_g is found to increase with optical path length, the value being highly consistent for $m = 3.0$. In the case of February to May, the mean values of the ratio are higher for $m = 1.5$ than that for $m = 2.0$ (Fig.7). The ratios in 1987 are invariably lower than those in 1988, directly attributable to equitable rainfall distribution during 1987. April 1987, however, has higher ratio than April 1988 as the latter had more rains during the month. The postmonsoon values, October to January, also indicate the effects of rain, 1987 having lower ratio values when compared with the values for the corresponding period in 1988.

β_{rr}/β_g is found to be always higher in the afternoons than in the forenoons. This is perhaps due to the loosening of the soil by irradiation and subsequent lifting up of these dust particles by convection processes and wind, as reasoned by Rangarajan (1970). A careful study of individual values show that there have been occasions when $\beta_{rr} \approx \beta_g$. And on several other occasions where there had been rains, the ratios were even less than one, indicating that almost all large particles had been washed out. For instance, β_{rr}/β_g became 0.600 on 28th May 1989 in the forenoon after heavy rains the previous day evening. However, this was only for a brief period; C_i and haze covered the skies in the afternoon. When observations were available throughout the day, the ratios were seen to be increasing as the day progressed. Mani *et al.* (1969), however, concluded that there are considerable observational errors that led to larger values of β_{rr}/β_g. Their study, however, were restricted to the measurements at a fixed time, 1130 h IST.

Routine measurements carried out at different times are given in Fig.8. The effect of atmospheric pollution is self-evident. The particle sizes are becoming larger due to the industrial activities in and around Pune whereas α values are taking a rather sharp dip. Rangarajan (1972) reports that α values at Pune are largely much less than 1.3, mostly around 0.5-0.6. Taking ten-day averages, he found α to be less than 0.5 during the first half of the year and to be of the order of 0.75 during October to January. Fig.9 reproduces the diagram presented by Rangarajan (1972) in which the distribution of α shows negative skewness, the mode being greater than the median. The frequency of occurrence of daily mean value of α for the 127 days are derived from the individual values recorded during each day. The median value is about 0.5 and the mode is about 0.8 as can be seen from the continuous line obtained from the individual 527 observations. However in appreciable number of cases, Rangarajan found α to be very near zero. While two maxima, one at 0.5 and the other at 0.9 Rangarajan argues that a value of 0.5 is more representative as this coincides with the median. Rangarajan concludes his work suggesting the use of $\alpha = 0.6$ instead of 1.3. This may give a closer correspondence with the turbidity factor B.

Tomasi *et al.* (1983) computed α for different particle sizes and for various values of refractive index. They found that the "non-absorbing" Aitken particles give a value between 3.7 and 4.1. The value of α for small "absorbing" particles was found to be between 0.4 and 1.3. For continental and urban particles, they found α to be very close to 0.5. They obtained zero and even negative values for large particles, producing almost neutral extinction. Urban particle size distribution, they state, are frequently characterised by both small and giant particles which give rise to rather low values of α. "Usually these particles are mainly composed of water soluble and carbonaceous substances so that they absorb considerable visible and near infrared radiation".

4. Conclusion

The study of atmospheric aerosols and its role in the modification of solar irradiation field are very complex. It is, however, possible to determine the attenuation of the solar irradiance in different spectral regions and indirectly infer the role which the various constituents play in transmitting the solar irradiance through the atmosphere. The higher incidence of water vapour with hygroscopic aerosol nuclei present in the atmosphere is conducive for growth of the particles and can cause higher degree of attenuation of solar irradiation. Because of this feasibility as evidenced by

large values of β_{rr} whenever the water vapour content is higher, α values tend to be much lower than the classical value, 1.3. An average value of the order of 0.4-0.6 would be more appropriate for α over Pune. The feasibility of negative values of α is very high given the large size hygroscopic particles injected into the atmosphere by industrial activities. The attenuation may be caused by selective absorption in the red wavelength, besides by scattering. Further studies and determination of α are being made taking into consideration the temperature and water vapour content at the time of each measurement.

The spectral radiometric studies enable one to infer the likely changes that take place in the atmosphere. While it is more apparent that moisture controls the size of the particles that are hygroscopic in nature, radiometric studies alone may not be able to provide any definite inference on the changes that take place in the size distribution of the non-hygroscopic particles. Further collaborative studies on the size distribution of aerosols are called for.

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