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A new method for the determination of atmospheric turbidity

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ABSTRACT

Atmospheric turbidity is usually measured using either a pyrheliometer fitted with a red RG630 filter or a Volz sun photometer, the turbidity coefficients so determined being designated as β and B, respectively. Both techniques are subject to error, the former in underestimating high turbidities and the latter in giving rise to errors at low turbidities. The present paper describes a new, simpler and less expensive method of evaluating β from measurements of direct and diffuse solar radiation, made as a routine at principal radiation stations.

Using a theoretical model for determining the attenuation of solar radiation due to absorption and scattering by water vapour and other gases, dust and aerosols in the atmosphere, an expression for the ratio of diffuse to direct solar radiation D/I_H is derived as a function of β. Then, from the hourly mean values of global and diffuse solar radiation routinely recorded at principal radiation stations, D/I_H is calculated. β can now be readily evaluated using a special nomogram based on the formula relating β to D/I_H. The values of β derived for Indian stations using the above technique show remarkable internal consistency and stability, proving its utility and reliability.

1. Introduction

Atmospheric turbidity is a measure of the total vertically integrated particulate load in the atmosphere, an important factor influencing the energetics of solar radiation in the earth's atmosphere. It is a useful index of atmospheric pollution, particularly in studies of long-term secular changes in the composition of the atmosphere and resultant global climatic changes. Despite its importance, there does not appear to be any universally-adopted definition of turbidity or any universally-accepted technique for its measurement.

Two parameters of turbidity are currently used. The Ångström turbidity coefficient β is expressed by

$$\tau_\lambda = \exp(-\beta/\lambda^\alpha), \tag{1}$$

where τ_λ is the transmission factor for solar radiation at a wavelength λ for dust scattering for a vertical path, and α is a constant characterized by the size distribution of dust particles, usually assumed to have a mean value of 1.3. The precise value of α prevailing under a given condition of the atmosphere can be determined from measurements of τ_{λ_1} and τ_{λ_2} at two convenient wavelengths λ_1 and λ_2 in the solar spectrum, free from the absorption effects of water vapour and ozone, using the relation

$$\alpha = \frac{\log \tau_{\lambda_1} - \log \tau_{\lambda_2}}{\log \lambda_2 - \log \lambda_1}. \tag{2}$$

However, it was shown by Mani et al. (1969) that very careful measurements alone would yield α and therefore β values of sufficient reliability, and that routine observations made at network stations may not be sufficiently accurate for the purpose.

The other turbidity parameter is B which is the decadic extinction coefficient for dust scattering at a wavelength of 0.5 μ. It is measured with a sun photometer (Volz, 1961) using the relation

$$I_\lambda = I_{0\lambda} S \cdot 10^{-(a_n + B)^m}, \tag{3}$$

where $I_{0\lambda}$ is the intensity of the extraterrestrial solar radiation at a wavelength λ that falls on the sensor of the photometer, S the conversion factor for the varying earth–sun distance, m the relative air mass and a_R the decadic extinction coefficient at λ for Rayleigh scattering. β and B are related to each other through the equation (Ångström, 1961)

$$B = 0.4343\beta \cdot 2^{\alpha}. \tag{4}$$

Measurements of both β and B have been made at a number of stations in India for many years (Mani et al., 1969). Using these values and Hoyt's (1978) model, attempts were made to compute direct and diffuse solar radiation during the course of a programme of computations of solar radiation for 145 stations in India (Mani and Rangarajan, 1982). It was found that the diffuse solar radiation values, calculated using the model, were much lower than those actually measured, the differences being large during the highly turbid pre-monsoon months March to June. This could arise from errors in measurement of D and β and the inadequacy of Hoyt's model for Indian conditions. The errors in measurement of D may arise from inadequate corrections for the area of the sky shaded by the ring. Under highly turbid conditions, the shading ring correction factor has to be augmented for the anisotropy in sky radiation. Errors in β can be caused by the influence of the sun's aureole which is not taken into account in the evaluation of β (Ångström and Rodhe, 1966). Measurements made to test Hoyt's model (Mani and Chacko, 1980) showed that the computed values of I and D agree within a few %, well within the accuracy possible for measurements with thermoelectric pyranometers. The low computed values of D could therefore result from the too low values of β used in the computation. Allowing for the inherent uncertainties in the measurement of β, which are much greater than those for D, an attempt was made to compute β from values of D and I, using Hoyt's model.

2. The new method for computing β

Hoyt's (1978, 1979) model enables direct (I_H), diffuse (D) and global (G) solar radiation to be computed using the following equations valid for clear sky noon conditions:

$$I_H = I_0\left\{1 - \sum_{i=1}^{5} \alpha_i\right\}(1 - S_a)(1 - S_d)\cos z, \tag{5}$$

$$D = I_0\left\{1 - \sum_{i=1}^{5} \alpha_i\right\}(0.5S_a + 0.75S_d)\cos z, \tag{6}$$

$$G\downarrow = R(I_H + D)\left\{1 - \sum_{i=1}^{5} \alpha_i'\right\}(0.5S_a' + 0.25S_d'), \tag{7}$$

$$G = I_H + D + G, \tag{8}$$

$$S_a = 1 - f(m^*)^{m^*}, \tag{9}$$

$$S_d = 1 - g(\beta)^m, \tag{10}$$

where I_0 is the solar constant, $\sum_{i=1}^{5} \alpha_i$ represents absorption effects due to water vapour, carbon dioxide, ozone, oxygen and dust, $f(m^*)$ the transmission function for pure air due to Rayleigh scattering for an absolute air mass m^*, and $g(\beta)$ the transmission function due to dust scattering. α_i' represents the absorption evaluated for air mass value of $(m^* + 1.66P/P_0)$, where P is the station barometric pressure and $P_0 = 1013.2$ mb.

Dividing eq. (6) by eq. (5), the absorption effects of dust and other atmospheric constituents are eliminated and the ratio D/I_H is given in terms of the scattering functions of pure air and dust, S_a and S_d,

$$D/I_H = \frac{0.5S_a + 0.75S_d}{(1 - S_a)(1 - S_d)}. \tag{11}$$

S_d can be obtained if D/I_H and S_a are known; S_a is precisely calculable through its dependence on Rayleigh scattering. From eq. (10), $g(\beta)$ can be obtained from S_d. Using Hoyt's (1978) table relating β and $g(\beta)$, a nomogram was prepared (Fig. 1) for direct evaluation of β from D/I_H.

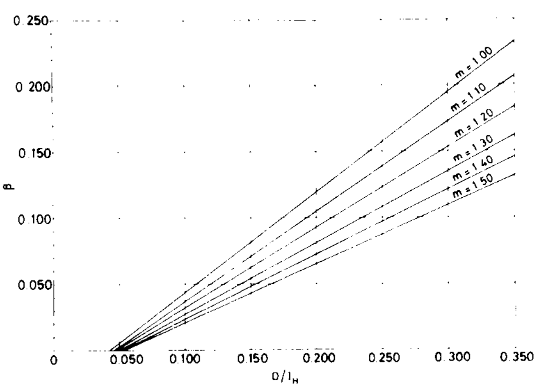

Fig. 1. Nomogram for deriving β from the ratio D/I_H.

Choudhury (1981) has recently derived the following relationships,

$$S_a = \frac{0.606m^*}{643 + m^*}, \tag{12}$$

$$g(\beta) = 1 - 1.375\beta \quad \text{for } \beta < 0.08,$$
$$g(\beta) = 0.9843 - 1.1429\beta \quad \text{for } \beta \geqslant 0.08. \tag{13}$$

By coupling eqs. (12) and (13) with eq. (11), we have a simple alternate way of constructing the nomogram of Fig. 1. Comparison of S_a values obtained through eq. (12) with those of Hoyt (1979) showed very close agreement to within 1%, while a similar comparison of $g(\beta)$ values using eq. (13) also showed agreement to within 1%.

It should be pointed out that D in eq. (11) is the diffuse radiation arising from scattering during the course of the first pass of solar radiation in the atmosphere, whereas the diffuse radiation actually measured by a pyranometer with a shading ring or disc is $D + G\downarrow$. Since eq. (11) is valid only for D, it

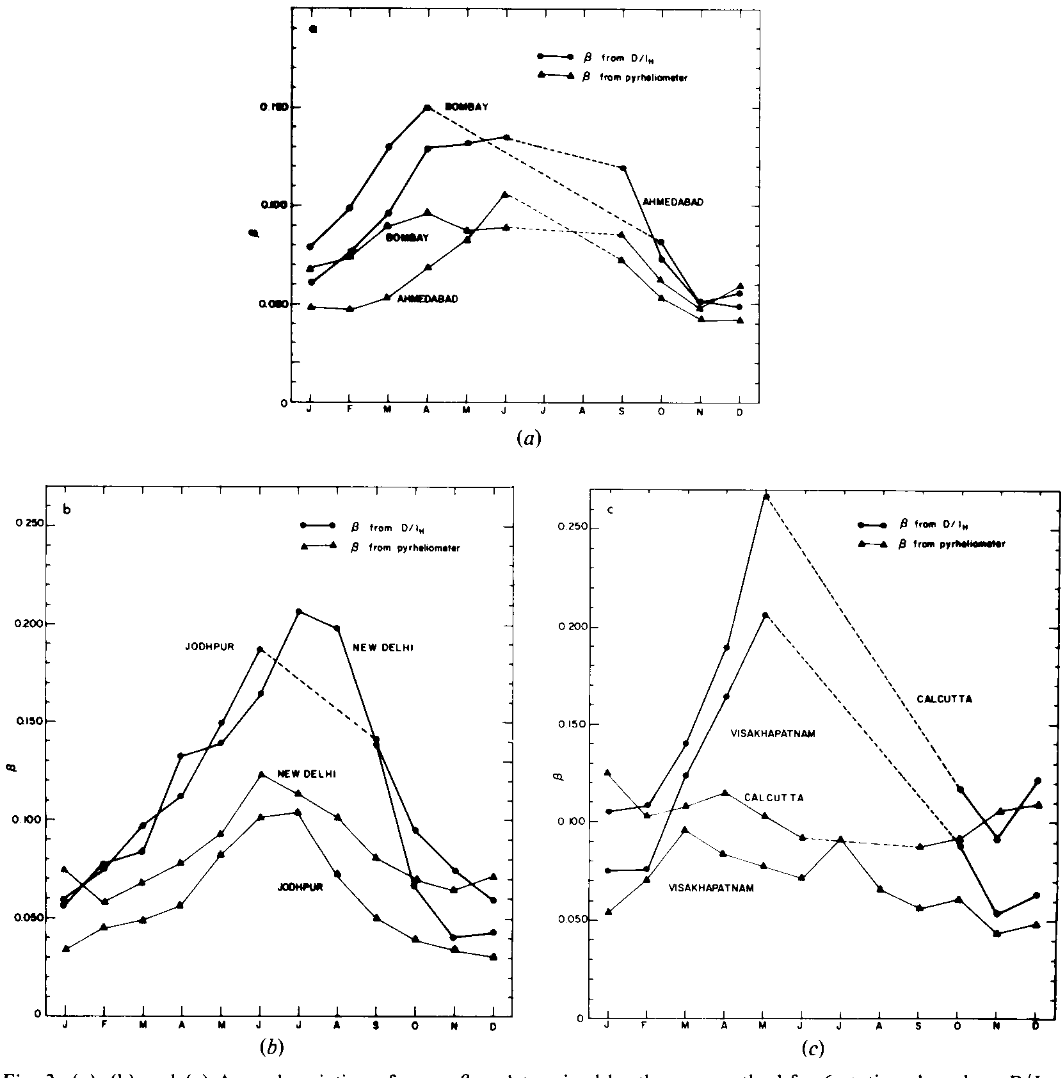

Fig. 2. (a), (b) and (c) Annual variation of mean β as determined by the new method for 6 stations based on D/I_H (1958–1975) with corresponding β measured by pyrheliometer.

is necessary to subtract $G\downarrow$ from the observed values of diffuse radiation before using the nomogram, or calculating $g(\beta)$ and β from eq. (11). Fortunately this is not difficult, because in almost all cases for which computations of D, I_H, $G\downarrow$ and G were made for different inputs of water vapour and turbidity, $G\downarrow$ was found to be numerically equal to a nearly constant fraction of 9 to 11% of $(D + G\downarrow)$. Hence the first step towards computation of β using the nomogram is to reduce the observed diffuse radiation through multiplication by a constant factor 0.9 to obtain true D. Since both diffuse and global radiation include the component $G\downarrow$, the difference between global and diffuse radiation will be equal to I_H in eq. (11).

3. Presentation of results

From the observed mean values of D and I_H for clear sky noon, climatological mean values of β were calculated for each month using the nomogram for 6 stations. These are presented in Fig. 2a, b and c, with the corresponding β values determined by the phyrheliometer. It will be seen that the β values determined from the D/I_H ratios show a marked increase during the pre-monsoon summer months at all stations, particularly at New Delhi, Jodhpur and Calcutta.

Measurements of diffuse solar radiation at principal radiation stations in the India network are made with carefully calibrated pyranometers provided with shading rings. Appropriate corrections are made (Drummond, 1956) for the small fraction of the diffuse radiation obstructed by the shading ring under the general assumption of isotropy of the sky radiation. Under conditions of clear sky, a further correction to account for the anisotropy of clear sky diffuse radiation will have to be made, the magnitude of this being dependent on the prevailing atmospheric turbidity. This additional correction is not, however, made at the network stations. Experiments made at Pretoria and elsewhere (Drummond, 1956) indicate that for average conditions, the additional correction is of the order of 6–8% and is independent of solar altitude. The experimentally-determined corrections are 3–4% for low turbidity, 6–8% for moderate turbidity and 10–12% for high turbidity. Similar experiments carried out at the Central Radiation Laboratory, Poona (results unpublished) have indicated corrections of the same order for Indian stations.

Using the β values obtained by the new method, computations of I_H, D and G for clear sky noon conditions were repeated and the resulting values of D agreed with the observed values within 5% or better in most cases. The calculated G was understandably found to be not as sensitive to β as in the case of D.

A careful analysis was carried out with the hourly values of D and G measured at New Delhi for two randomly chosen years, 1960 and 1971 and using the new method, monthly mean values of β were obtained. These (Fig. 3) indicate not only their internal consistency but also a secular increase of β over a period of 11 years. The increase could be due to progressive industrialization and urbanization of the city and the consequent increase in the concentration of particulate matter in the atmosphere. Statistical tests for the significance of the secular trend based on the individual values that make up the mean β for each month confirmed the genuineness of the trend. The pyranometers used for the routine observations of D and G at New Delhi are compared annually with the national standards at Poona and as such the possibility of instrumental errors can be ruled out.

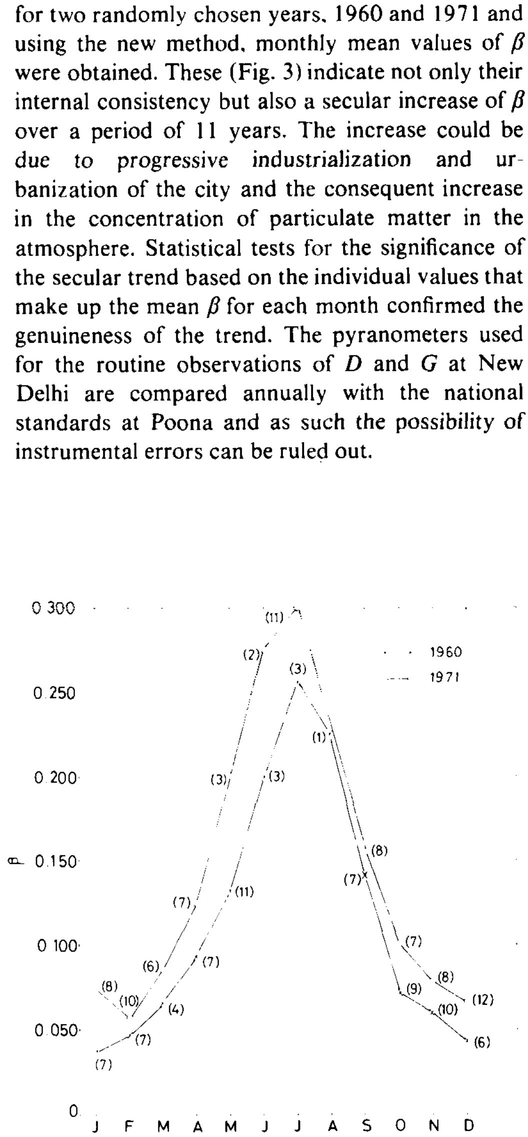

Fig. 3. Monthly mean β at New Delhi during the years 1960 and 1971 (numbers in brackets represent the number of observations used for averaging).

In eq. (5), it is assumed that for dust scattering, 0.75 of the total scattered radiation is propagated in the forward direction. This ratio could change with aerosol size distribution and there is no uniqueness about the value of the fraction. Calculations using an arbitrarily higher ratio 0.85 showed that the resulting β values were only about 10% less than those calculated with 0.75. Hence a uniform value of 0.75 can be assumed for the forward scatter ratio. Further, the assumption of a constant value of α in the preparation of the $g(\beta)$ tables does not lead to errors in the numerical computations related to dust scattering, because any value of β corresponding to a given prevailing α could be transformed to a new value corresponding to the value of 1.0 for α so as to give the same transmission function for dust scattering.

4. Concluding remarks

The usefulness of the technique outlined above lies in the fact that turbidity on clear days can readily be obtained on a day-to-day basis from readings of pyranometers that continuously record G and D at network stations. There is no need for additional equipment or observations for the measurement of β at radiation stations. Since solar radiation is integrated at hourly intervals, values determined from these will represent the mean for the hour, eliminating the transient and spurious variations that might be otherwise measured. This technique can also be used to monitor secular changes in atmospheric turbidity at locations where long-period records of global and diffuse solar radiation are available.

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