Mausam, (1985), **36**, 1, 51-56

551.51 : 551.551

A simple method for assessment of atmospheric turbidity

N. C. MAJUMDAR
Defence Institute of Physiology and Allied Sciences, Delhi
and
N. C. WALMIK & B. K. AGARWAL
M. M. H. College, Ghaziabad (U. P.)

(*Received 17 August 1982*)

सार — लिक (1942) के आविल गुणांक T की अवधारणा के आधार पर मजूमदार (1978) तथा उनके साथियों ने एक युक्तियुक्त आविल गुणांका T_r को परिभाषित किया । उसमें 'वायुराशि प्रत्यक्ष विचरण' को न्यूनतम किया गया । इन दोनों ही गुणांकों में वायुमण्डल में स्थित परिवर्तनशील जल-वाष्प भी शामिल थी, अतः इसे स्काट रेड फिल्टर, RG-2 की सहायता से दूर किया गया । अब आविलता का एक नया सूचकांक τ को प्रत्यक्ष और तीव्रता के फिल्टर सहित और उसके बिना प्राप्त दो मापों के आधार पर परिभाषित किया गया है ।

ABSTRACT. Based on the concept underlying Linke's (1942) turbidity factor, T, a rational turbidity factor, Tr, was defined by Majumdar *et al.* (1978), in which "Virtual variation with air mass" was minimised. Since both these factors include the effect of variable water vapour in the atmosphere, the same has been eliminated with the help of Schott red filter, RG-2. A new index of turbidity, τ, has now been defined on the basis of two measurements of direct solar intensity with and without filter.

1. Introduction

The intensity of direct solar radiation received on the surface of the earth depends on its depletion along its path through the atmosphere. The depletion is small in pure dry air, but increases with the amount of contamination, pollution, or turbidity associated with variable components such as water vapour, dust, smoke, haze or, in general, aerosols. It should, therefore, be possible to determine the turbidity of the atmosphere from measurements of direct solar radiation.

A number of difficulties arose in this connection, because solar radiation comprises a wide range of spectral wavelengths, with varying distribution. There would have been no difficulty with monochromatic radiation for which Beer's law holds good (Robinson 1966, p. 47) According to this law if a parallel monochromatic beam of intensity $I_{0\lambda}$ is incident upon absorbing medium then the intensity $I_{\lambda(m)}$ of the beam after passing through the medium is given by :

$$I_{\lambda(m)} = I_{0\lambda} \exp(-a_\lambda . m) \tag{1}$$

where 'm' is the absolute air mass in the path of the beam, and a_λ is the extinction coefficient (per unit air mass). From Eqn. (1) :

$$\ln I_{\lambda(m)} = \ln I_{0\lambda} - a_\lambda . m \tag{2}$$

It follows from Eqn. (2) that $\ln I_{\lambda(m)}$ plotted against 'm' should yield a straight line with a slope equal to $-a_\lambda$. It also follows that :

$$a_\lambda = m^{-1} \ln [I_{0\lambda} / I_{\lambda(m)}] \tag{3}$$

In analogy with Eqn. (3) a mean extinction coefficient 'a' was defined for solar radiation, so that :

$$a = m^{-1} \ln (I_0 / I) \tag{4}$$

where I_0 is the solar constant, and I is the measured value of the intensity of direct solar radiation at normal incidence, for air mass 'm'. However, an actual plot of $\ln I$ against 'm' even for a pure and dry atmosphere, yields a curve which is slightly concave upwards, indicating that 'a' decreases with increasing air mass. As a result, the mean extinction coefficient could not be taken as a reliable measure of atmospheric turbidity due to its so called "virtual variation with air mass".

2. Linke's turbidity factor (T)

The first rational approach to the problem was perhaps due to Linke (1922), who defined a turbidity factor (T), which indicates the number of standard atmospheres of pure and dry air which produce the same total depletion or give rise to the same intensity of direct solar radiation as in the given turbid atmosphere. This was based on the consideration that scattering and absorption by pure, dry air is the basic atmospheric effect.

Let 'I_0' be the solar constant, 'I' the intensity of direct solar radiation at normal incidence, and 'm' the absolute air mass at the time and place of observation. If $a(m)$ be the mean extinction coefficient

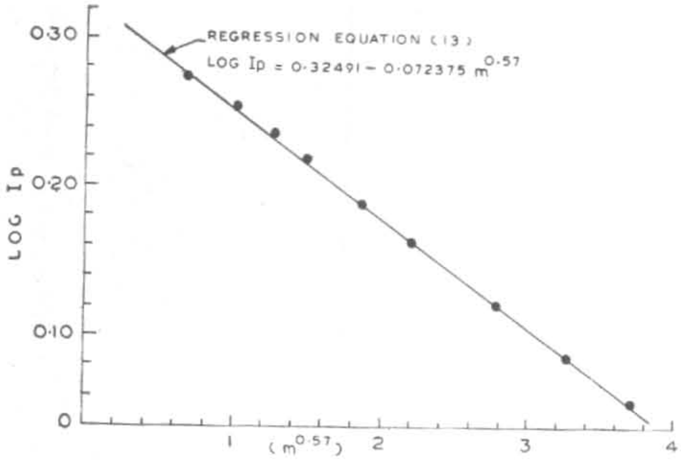

Fig. 1. Comparison of regression equation with available data (Majumdar *et al.* 1978 b)

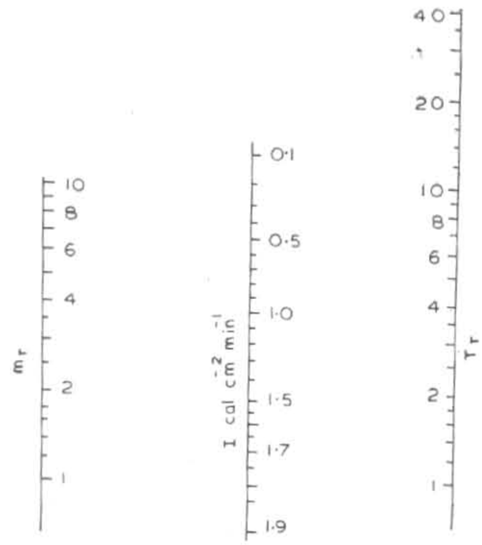

Fig. 2. Nomogram for evaluation of T_r

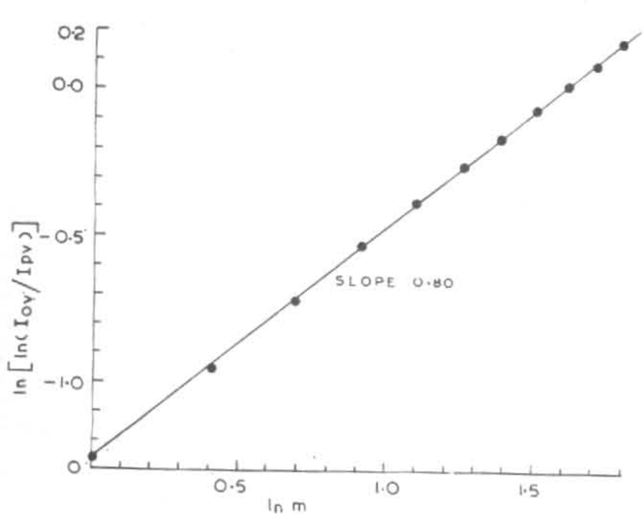

Fig. 3. Linear relation between ln [ln (I_{ov}/I_{pv})] and ln m

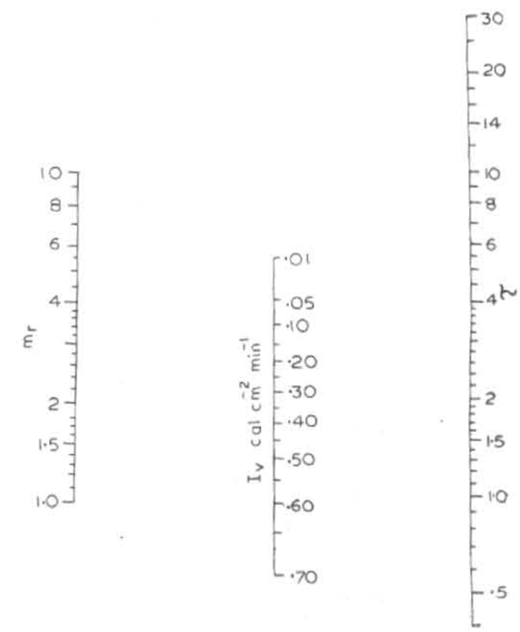

Fig. 4. Nomogram for estimation of the new index of turbidity (τ) from relative air mass (m_r) and observed radiation intensity (I_v) below 0.63 μm reduced to mean sun-earth distance

for the given turbid atmosphere, and $a_p(m)$, that of pure, dry atmosphere for the same air mass 'm', we should have according to Linke :

$$I = I_0 \exp[-a(m).\, m] = I_0 \exp[-a_p(m).\, m\, T] \quad (5)$$

since an air mass 'm' of turbid air of turbidity factor 'T' is, by definition, equivalent to air mass 'mT' of pure, dry air.

From Eqn. (5), we obtain :

$$T = P(m) \log(I_0 / I) \quad (6)$$

where $P(m)$ is a factor which is inversely proportional to the extinction produced by an absolute air mass m and hence is a function of air mass as well.

Further, $$P(m) = 2.303/m.\, a_p(m) \quad (7)$$

With $T=1$ (pure, dry atmosphere), Eqn. (6) leads to

$$P(m) = 1/\log(I_0 / I_p) \quad (8)$$

Computed values of $P(m)$ with $I_0 = 1.980$ cal cm^{-2} min^{-1} were tabulated by Robinson (1966, p. 99) for air mass varying from 0.5 to 10. Knowing $P(m)$ and observed I, Linke's turbidity factor T, can be estimated from Eqn. (6). It may be noted, however, that the observed I must be reduced to the mean sun-earth distance by multiplying it by the factor $(R/R_0)^2$, where R is the actual sun-earth distance on the day of observation, and R_0, the annual mean. Computational work is avoided with the help of an alignment chart devised for the purpose.

Linke's turbidity factor, T, was found to be a distinct improvement on earlier measures of turbidity. It was, however, found to suffer from one great disadvantage. Even under constant turbidity conditions 'T' was found to exhibit a diurnal variation, or the so called virtual variation with air mass. In other words, 'T' is not strictly independent of air mass. This was ascribed to the dependence of the extinction coefficient on the wavelength of radiation (Robinson 1966, p. 100). It may also be noted that 'T' is a measure of total turbidity, including the contribution by water vapour, since it is based on insolation measurements without filter.

3. Angstrom's turbidity coefficient (β)

In order to eliminate the effect of water vapour from solar radiation measurements, Angstrom introduced the Schott red filter, RG-2, having transmission from 0.63 to 2.8 μ m. Two measurements are made, one without filter (I_t), and one with the filter (I_R). The latter is multiplied by the reduction factor, DR2, of the particular filter supplied by the manufacturer in order to correct for transmission losses within the filter. The difference (I_t — DR2 $\times$ I_R), therefore, gives the radiation intensity over all wavelengths < 0.63 μm, thereby eliminating the effect of variable water vapour which is almost entirely confined to the infrared region of the solar spectrum. We are thus left with two basic effects, *viz.,* molecular or Rayleigh scattering and scattering due to aerosols (dust, smoke, haze, etc).

The extinction coefficient '$a\lambda_D$', due to aerosols is a complex function of the particle number and size distribution. However, Angstrom (1929, 1930) approached the problem in an empirical way, and derived an expression for $a\lambda_D$ on an experimental basis. It may be put in the form :

$$a\lambda_D = \beta\lambda^{-\alpha} \quad (9)$$

where β is a constant characteristic of the aerosol content, and α is a number between 0 and 4, characteristic of the particle size distribution. The constant β is known as Angstrom turbidity coefficient.

On the basis of spectral investigations, a reasonable, average value of the exponent 'α' in Eqn. (9) was found to be 1.3, so that the equation could be simplified to the form :

$$a\lambda_D = \beta.\, \lambda^{-1.3} \quad (10)$$

With $\alpha = 1.3$, elaborates tables have been prepared from which β can be readily estimated from pyrheliometric observations.

'β' has been in wide use by meteorological organisations all over the world as a rough guide to the measure of dust, smoke and haze of the local atmosphere.

The principal drawback of 'β' is the assumption of a constant value for 'α', which may widely differ from 1.3, thereby giving rise to the so-called virtual variation of β with air mass even under steady turbidity conditions of the atmosphere. This is particularly true in tropical countries like India, where particle size distribution has wide seasonal variations. In northern India, larger particles predominate during summer months, so that $\alpha \rightarrow 0$, and occasionally becomes negative (Ramanathan and Karandikar 1949).

4. Schuepp's generalized turbidity coefficient (B)

Schuepp (1949) has defined a turbidity coefficient 'B' referred to the base 10 of ordinary logarithms, and by replacing the term $\lambda^{-1.3}$, by $(2\lambda)^{-\alpha}$ so that Schuepp's generalised turbidity coefficient, B, is referred to the wavelength 0.5 μm in the central part of the visible spectrum. It will be seen that Angstrom's turbidity coefficient 'β' is referred to as wavelength of 1 μ m, well outside the visible spectrum, and its actual mangitude is, therefore, only of academic interest.

Schuepp's generalised turbidity coefficient, 'B' together with the wavelength exponent, α, and the precipitable water, W, gives the most valuable information because the size spectrum of the aerosols can be inferred from 'α'. Graphical procedures were devised by Schuepp to simplify computational work. The method, though ingeneous in the academic sense, is of little practical use (Robinson 1966, p. 107). Firstly, it needs measurements with different filters, and is too cumbersome in its practical applications. Secondly, it is subject to large errors caused by a long series of radiation measurements. Thirdly, the formula is based on the ratio of small differences between several measured values so that the total error is multiplied manifold.

5. A simplified approach

In view of the foregoing, we considered it worthwhile to explore the possibilities of defining an index of total atmospheric turbidity based on measurements without filters, by overcoming the defect of Linke's turbidity factor T, so as to minimise virtual variation with air mass (Majumdar *et al.* 1978 b).

A critical study of Linke's turbidity factor T, leads to the conclusion that there is nothing materially wrong with the basic definition of T, according to which 'T' is the number of standard atmospheres of pure, dry air (Rayleigh atmosphere) which produce the same total depletion of direct solar radiation, as the given turbid atmosphere. We have traced the real error to the formulation of the quantitative expression for T given Eqn. (5), in which $a(m)$ is the mean extinction coefficient of the given turbid atmosphere of air mass 'm', while $a_p(m)$ is that of pure, dry air of the same air mass 'm'. Since $a_p(m)$ is a function of 'm', while $m_r T$ is the equivalent air mass of pure, dry air, it follows that $a_p(m)$ in Eqn. (5) should be replaced by $a_p(m_r T)$ where $a_p(m_r T)$ is the extinction coefficient of an equivalent air mass of pure, dry air. The modified equation we have proposed is as follows :

$$I = I_0 \exp\left[-a(m).\, m\right]$$

$$= I_0 \exp\left[-a_p(m_r T_r).\, m_r T_r\right] \qquad (11)$$

where I_0 is the solar constant, and I is the measured value of the intensity of direct solar radiation at normal incidence after it has passed through an equivalent air mass, $m_r T_r$, of pure, dry air with the mean extinction coefficient $a_p(m_r T_r)$. We have termed the turbidity factor, T_r, defined by Eqn. (11) as the "Rational turbidity factor", in order to distinguish it from Linke's turbidity factor, T. It follows, therefore, that

$$T_r = \ln(I_0/I)/m_r \cdot a_p(m_r T_r)$$

$$= 2.303 \log(I_0/I)/m_r \cdot a_p(m_r T_r) \qquad (12)$$

The main difficulty in applying Eqn. (12) for practical computation of T_r, arises from the fact that data on direct solar radiation in a pure and dry atmosphere are not available beyond an optical air mass of 10, while the equivalent pure air mass, $m_r T_r$ of a given turbid atmosphere may be many times the above value. No simple formula is known which would enable extrapolation much beyond $m_r = 10$.

We have overcome this difficulty by defining an effective air mass, m_e, such that $\log I_p$ plotted against m_e should yield a straight line, thereby enabling extrapolation to much larger values of 'm' with reasonable accuracy.

Our analysis of available data (Majumdar *et al.* 1978 b) has led to a reasonably reliable extrapolation formula, namely,

$$\log I_p = 0.32491 - 0.072375\, m^{0.57} \qquad (13)$$

so that the effective air mass is given by

$$m_e = m^{0.57} \qquad (14)$$

Here I_p is the intensity of direct solar radiation at normal incidence in cal cm^{-2} min^{-1} for a pure and dry atmosphere. The equation has been compared against available data in Fig. 1. The difference does not exceed $\pm 0.5\%$ over the entire range. If I be the observed intensity (reduced to mean sun-earth distance) in a turbid atmosphere (T_r) of relative air mass, m_r, then it follows from the definition of T_r, and Eqn. (13), that :

$$\log I = 0.32491 - 0.072375\, (m_r T_r)^{0.57},$$

whence we obtain

$$T_r = \frac{1}{m_r}\left[\frac{0.32491 - \log I}{0.072375}\right]^{\frac{1}{0.57}} \qquad (15)$$

In order to avoid tedious and time consuming computational work, we have devised a nomogram (Fig. 2) for quick evaluation of T_r.

The proposed measure of turbidity, T_r, is made up of three components, namely, (*i*) pure and dry air (basic effect), (*ii*) precipitable water, W and (*iii*) aerosols (dust, smoke, haze, etc). In order to assess the relative contribution of aerosols to the total turbidity, it is essential to have a reliable estimate of precipitable water. We have evolved a new method for estimating 'W' from surface humidity (Majumdar *et al.* 1977) by introducing the concept of a "Lapse parameter", characteristic of the vertical moisture profile. We have also proposed an improved method for computation of 'W' from upper air data (Majumdar *et al.* 1978 a).

However our subsequent studies (Majumdar *et al.* 1979 a, 1979 b) have revealed that the effect of W cannot be isolated from that of aerosols. This may be explained as due to strong interaction between scattering by aerosols and absorption by water vapour in the near-infrared region of the solar spectrum. This effect is likely to be more pronounced with larger particle size.

The chief merits of the "Rationalised turbidity factor" T_r are :

(*i*) Measurement with filter is not required,

(*ii*) It does not assume any fixed particle size distribution,

(*iii*) It can be easily and quickly estimated with the help of the nomogram in Fig. 2,

(*iv*) Virtual variation with air mass is likely to be negligible and

(*v*) It is applicable to all altitudes.

6. Development of a new index of turbidity

Since, however, the contributions of aerosols towards the total turbidity cannot be reliably estimated in the presence of water vapour, we have defined a new index of turbidity based on two pyrheliometric measurements, one without filter (I_t) and the other with Schott red filter RG-2, (I_R), as required for the estimation of Angstrom's turbidity coefficient, β. If DR2 be the reduction factor of the filter then the difference (I_t — DR2 $\times I_R$) gives the radiation intensity I_v, over all wavelengths below 0.63 μm.

TABLE 1

Radiation data in relation to air mass for a pure and dry atmosphere

(I_{ov}=0.789 cal cm^{-2} min^{-1})

Air mass (m)	I_{pv} (cal cm^{-2} min^{-1})	ln m	ln [ln (I_{ov}/I_{pv})]	$m^{0.8}$	ln I_{pv}	ln I_{pv} est	Error
(1)	(2)	(3)	(4)	(5)	(6)	(7)	(8)
1.0	0.594	0.000	—1.259	1.000	—0.521	—0.519	+0.002
1.5	0.536	0.405	—0.949	1.383	—0.624	—0.627	—0.003
2.0	0.484	0.693	—0.715	1.741	—0.726	—0.729	—0.003
2.5	0.438	0.916	—0.529	2.081	—0.826	—0.826	0.000
3.0	0.397	1.099	—0.375	2.408	—0.924	—0.919	+0.005
3.5	0.364	1.253	—0.256	2.724	—1.011	—1.009	+0.002
4.0	0.336	1.386	—0.158	3.031	—1.091	—1.096	—0.005
4.5	0.307	1.504	—0.058	3.331	—1.181	—1.181	0.000
5.0	0.282	1.609	+0.029	3.624	—1.266	—1.264	+0.002
5.5	0.261	1.705	+0.101	3.911	—1.343	—1.346	—0.003
6.0	0.240	1.792	+0.174	4.193	—1.427	—1.426	+0.001

Water vapour and haze (dust) are the main sources of turbidity. It is possible to separate them out due to their widely different contributions in the visible and infrared parts of the spectrum. The difference (I_t—DR2 $\times$ I_R) is approximately the sum of the visible and ultraviolet portions of the solar radiation (Robinson 1966, p. 101). So with the removal of infrared radiation the effect of water vapour is eliminated. The new index of turbidity (τ) is then defined as the number of standard atmospheres of pure, dry air which produce the same depletion of direct solar radiation below 0.63 μm, as the given turbid atmosphere.

6.1. *Effective air mass* m_e ($\lambda < 0.63\ \mu$ m)

Our next task is to find the relationship between effective air mass 'm_e' and the actual air mass 'm' for pure dry air, so that log I_{pv} plotted against 'm_e' should yield a straight line of the form :

$$\ln I_{pv} = \ln I_{ov} - a.\ m_e \tag{16}$$

where I_{ov} is the solar constant or solar radiation intensity over all wavelengths below 0.63 μ m, and I_{pv} is the solar radiation intensity for the said wavelengths after their passage through pure and dry air. m_e is the effective air mass and a is the mean extinction coefficient due to aerosols (dust, haze etc). Assuming m_e=m^q where q is a positive fraction less than 1, Eqn. (16) becomes :

$$\ln (I_{ov}/I_{pv}) = a.m^q$$

whence,

$$\ln \left[\ln (I_{ov}/I_{pv}) \right] = \ln a + q.\ \ln m \tag{17}$$

Estimation of q has been done with the help of the data presented in the first two columns of Table 1. These values were computed by Angstrom's method, and have been extracted from India Meteorological Department, I. S. Circular No. 47a, dated 25-5-1958, Instruments Section, Meteorological Office, Pune, India. The second column of Table 1 gives the values of radiation intensity I_{pv} in cal cm^{-2} min^{-1} for pure and dry air, for $\lambda < 0.63\ \mu$ m (International Pyrheliometric scale, 1956, I_0 = 1.980 cal cm^{-2} min^{-1}, Solar spectrum according to Nicolet). We have taken I_{ov} = 0.789 cal cm^{-2} min^{-1} the value computed by Linke (1942).

The derived values in the third and fourth columns of Table 1 have been plotted in Fig. 3, which reveals an almost perfect linear relationship. The slope of the line does not differ significantly from 0.8. Thus the effective air mass may be given by:

$$m_e = m^{0.8} \tag{18}$$

6.2. *Correlation of radiation data with effective air mass for a pure and dry atmosphere*

In view of Eqn. (18), we can now write :

$$\ln I_{pv} = \ln I_{ov} - a.m^{0.8} \tag{19}$$

We no longer assume I_{ov} = 0.789 cal cm^{-2} min^{-1} as computed by Linke. The best value of ln I_{ov} and 'a' have been determined by the method of least squares, using the data in columns (5) and (6) of Table 1.

The values obtained are :

$$\ln I_{ov} = -0.2343 \text{ and } a = 0.2842,$$

so that I_{ov} = 0.791 cal cm^{-2} min^{-1}, which is practically the same as that of Linke.

Eqn. (19) now takes the form :

$$\ln I_{pv} = -0.2343 - 0.2842\ m^{0.8} \tag{20}$$

The observed values of ln I_{pv} have been compared with the values estimated from Eqn. (20), in columns (6) and (7) of Table 1. The maximum error is $\pm$0.005 as will be evident from column (8) of Table 1. This corresponds to a maximum error of $\pm$0.5% in the intensity values.

6.3. *The new index of turbidity* (τ)

As mentioned earlier, the new index of turbidity (τ) has been defined as the number of standard atmospheres of pure, dry air which produce the same depletion of direct solar radition below $0.63\ \mu$ m, as the given turbid atmosphere.

It follows from Eqn. (20) that :

$$\ln I_v = -0.2343 - 0.2842\ (m_r\tau)^{0.8}$$

whence,

$$I_v = 0.791 \exp\left[-0.2842\ (m_r\tau)^{0.8}\right] \qquad (21)$$

so that

$$\tau = \frac{4.819}{m_r}(\ln 0.791/I_v)^{1.25} \qquad (22)$$

where m_r is the relative air mass and I_v is the measured radiation intensity below $0.63\ \mu$ m in cal cm^{-2} min^{-1}, reduced to mean sun-earth distance.

In order to avoid tedious and time consuming computational work, a nomogram shown in Fig. 4 was devised for quick estimation of τ.

7. Discussion

It will be appreciated that the concept of optical air mass has proved to be very useful in eliminating or minimising virtual variation of the mean extinction coefficient '*a*' in Eqn. (19) which has now a constant value of 0.2842 independent of air mass. The proposed index of turbidity should, therefore, overcome the defects of earlier measures of turbidity.

Another advantage of the proposed index is that unlike Ångström's turbidity coefficient 'β', τ does not require the assumption of a fixed particle size distribution.

Acknowledgement

The authors are thankful to Gp. Capt. K. C. Sinha, Director, DIPAS, for his keen interest in the study, and permission to publish this paper.

References

Angstrom, A., 1929, On the atmospheric transmission of sun radiation and on dust in the air, *Geografiska Annaler*, **11**, 156.

Angstrom, A., 1930, On the atmospheric transmission of sun radiation, *Geografiska Annaler*, **12**, 130.

Linke, F., 1942, in *Handbuch der Geophysik*, Born trager, Berlin, **8**, 239-332.

Majumdar, N. C., Garg, O. P. and Agarwal, B. K., 1977, Water vapour content of the atmosphere in relation to surface humidity, *Def. Sci. J.*, **27**, 197-202.

Majumdar, N.C., Garg, O. P. and Agarwal, B. K., 1978 (a), On computation of precipitable water from upper air data, *Def. Sci. J.*, **28**, 7-12.

Majumdar, N.C., Garg, O. P. and Agarwal, B. K., 1978 (b), A fresh approach to the study of atmospheric turbidity, *Def. Sci. J.*, **28**, 167-174.

Majumdar, N.C., Garg, O. P. and Agarwal, B. K., 1979 (a), Further study on rational turbidity factor at unit air mass, *Def. Sci. J.*, **29**, 35-38.

Majumdar, N.C., Garg, O. P. and Agarwal, B. K., 1979 (b), Rational turbidity factor in relation to air mass, *Def. Sci. J.*, **29**, 85-90.

Ramanathan, K. R. and Karandikar, R. V., 1949, Effect of dust and haze on measurements of atmospheric ozone made with Dobson's spectrometers, *Quart. J. Roy. met. Soc.*, **75**, 257-267.

Robinson, N. (Ed.), 1966, *Solar Radiation*, Elsevier Publishing Company, London, pp. 47, 99-101, 107.

Schuepp, W., 1949, Die Bestimmung der komonenten der atmospherischen Trubung aus Aktinometer Messungen, Arch. Meteorol. Geophy. Bioklimatol., Ser. B, **1**: 257.