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ON THE EVALUATION OF ATMOSPHERIC TURBIDITY PARAMETERS FROM ACTINOMETRIC DATA

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RESUMEN

Bajo condiciones promedio de turbidez se obtuvieron grandes variaciones en los parámetros de turbidez atmosférica (exponente de longitud de onda α y coeficiente de turbidez β) adoptando la técnica usual de Ångström, empleando la representación gráfica de valores de intensidad en bandas espectrales amplias. Con la finalidad de mejorar la exactitud en la determinación del parámetro de turbidez α se llevó a cabo la integración numérica de los datos actinométricos del espectro solar y la transmisibilidad de los filtros clásicos de Schott. Este trabajo demuestra que las amplitudes del parámetro de turbidez α así derivados, son reales y dan valores más exactos de turbidez en la atmósfera.

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ABSTRACT

Under average turbidity conditions, large variations in atmospheric turbidity parameters, namely the wavelength exponent α and turbidity coefficient β, were obtained when the usual Ångström's technique employing the graphical representation of intensity values in broad spectral bands was adopted. With a view to improving the accuracy in the determination of turbidity parameter α, numerical integration of the actinometric data was performed using most recent values of the solar spectrum and the transmissivity of the classical Schott filters. This paper shows that the amplitudes of turbidity parameter α thus derived are real and yield more accurate values of turbidity in the atmosphere.

INTRODUCTION

The turbidity of the atmosphere is defined as the reduced transparency of the atmosphere caused by the scattering of incoming solar radiation by solid or liquid particles, other than clouds, held in suspension. This scattering causes a certain degree of extinction of incoming radiation which, according to Ångström (1961), is characterised by an extinction function:

$$f(\lambda) = \beta \, . \, \lambda^{-\alpha} \qquad (1)$$

where λ denotes the wavelength and α and β are the two turbidity parameters which represent the characteristics of scattering and absorbing aerosols.

Ångström (1964), for a simple way to determine the turbidity coefficient β, assumed the average value of the exponent α to be constant, of the order of 1.3. It was established later that this assumption was only approximately true and that the large variations of the character of dust in tropics, particularly in India, result in highly variable α -values. Useful information, under these conditions, can only be gained through a determination of β as well as α.

Determination of atmospheric turbidity parameters were made by the author (1971 a, b) at twelve pyrheliometric stations in India. The irradiations measured without filter and those measured behind the Schott Glass filters OG_1, RG_2, and RG_8 were utilised to calculate

the irradiations within the wavelength intervals $300 < \lambda < 525$ nm and $630 < \lambda < 700$ nm. In general, the irradiations defined by the integrals $\int_{0.3}^{0.525} F(\lambda)\, d\lambda$ and $\int_{0.630}^{0.700} F(\lambda)\, d\lambda$ are regarded as monochromatic groups of radiation and β values derived from them are denoted through the indices g and rr viz., β_g and β_{rr} respectively. For our purpose, β_g and β_{rr} values were derived against different airmass values to observed values of irradiation from the two charts prepared with the help of above two energy integrals and then the expressions

$$\alpha_0 = 1{\cdot}3 - 6 \cdot 02 \log (\beta_{rr}/\beta_g) \tag{2}$$

and,

$$\log \frac{\beta_0}{\beta_{rr}} = (\alpha_0 - 1 \cdot 3) \,.\, 0 \cdot 177 \tag{3}$$

were used to obtain appropriate values of turbidity parameters. This study revealed that the wavelength exponent α undergoes seasonal varation approaching to zero or even negative during summer months. Nevertheless the high frequency of negative values of α were difficult to understand. It appeared probable that the derived amplitude of α were somewhat larger than the real ones. In fact, since the mean values of the two wavelength intervals considered for the tabulation of β_g and β_{rr} are relatively close together, i.e., $\lambda_1 = 454$ nm and $\lambda_2 = 669$ nm, small errors in β_g and β_{rr} values could result in large variations of α. With a view to improve the accuracy in determination of turbidity parameter α, a series of experiments have been made using the Ångström Pyrheliometer (model 169 type) which was previously calibrated using a standard Ångström Compensation Pyrheliometer in Central Radiation Laboratory, Poona. The numerical integration of the basic equation has been performed by the author utilizing the actinometric data together with most recent values of the solar spectrum and transmissivity of filters and the results are presented in this paper.

COMPUTATIONAL TECHNIQUE

In the following a simplified procedure which has been adopted by the author for processing of pyrheliometric data for a reliable determination of turbidity parameters employing most recent values of solar spectrum and absorption coefficients for ozone, water vapour and carbon-di-oxide is outlined:

(i) The primary input data consist of the following information: mV-output obtained for total radiation and that behind OG1, RG2 and RG8 filters respectively, instrument temperature, total ozone content of the atmosphere, atmospheric pressure, sun-earth distance, standard meridian time, equation of time and a code for the state of sky.

(ii) The integration of an equation similar to

$$J_\lambda = \frac{1}{S}\int_0^\infty J_{o\lambda} \cdot \exp\left(-m_R \cdot \frac{p}{p_o}\, a_R \cdot \lambda^{-4.09} - m_\Omega \cdot a_{\Omega,\lambda} - m_D \cdot a_D^{(\lambda)}\right) \cdot A_\lambda(H_2O, CO_2)\, d\lambda \qquad (4)$$

was performed numerically in 50 steps (20 nm from 300 to 1000 nm and 200 nm from 1000 to 4000 nm). Values of extra-terrestrial spectrum were adapted to those of Labs and Neckel (1968) which are given in absolute scale and have to be reduced by 1.8% to accord with International Pyrheliometric Scale of 1956. Ozone absorption coefficients are those of Elsasser et al (1960). The transmission of the three Schott filters were also used in the same 50 intervals as regards their absorption and reflection.

(iii) Since for the evaluation of filter measurements the spectral extinction by dust and water vapour is needed, the basic equation was solved by successive approximation. In our program $a_{D,\lambda}$ (dust

extinction) and $a_{\Omega,\lambda}$ (Ozone extinction) were used as constants ($a_{D,\lambda}$: previously determined graphically) and also the two limits of approximation ϵ_1 and ϵ_2. For the case a_D (300-525 nm) $< a_D$ (630-700 nm), the assumption $a_{D,\lambda} = \beta.\lambda^{-\alpha}$ seems not justified hence a different function $a_{D,\lambda} = \beta.\ P(\lambda)$ was used. a_D, $P(\lambda)$ and A_λ (H_2O) are the preset values to be approximated in a later step.

(iv) There follows a number of preparatory computations: True solar time, zenith distance, airmass (for Rayleigh, ozone and dust). The calibration factor of the instrument is adapted to the instrument temperature. The Rayleigh scattering coefficient is adapted to the measured atmospheric pressure and Rayleigh and ozone scattering coefficients are then stored for the remainder of the program.

(v) Absorption by water vapour and carbon dioxide must be known before the intensities of radiation, measured behind filters, are substracted from total radiation measured without a filter, as they change the small intensity between 2.8 and 4.0 microns. For this purpose the analysis by Elsasser et.al (1960) were used by the author.

(vi) The filter factor F(T) was computed taking temperature into account according to the equation:

$$\int_A^B J_\lambda \, d\lambda = F(T) \left[\int_0^\infty J_\lambda \cdot \tau_\lambda (O°C)\, d\lambda + \bar{\tau}_\lambda \cdot C \cdot T \left(\frac{dJ_\lambda}{d_\lambda} \right)_A \right] \qquad (5)$$

where, τ_λ is the filter transmission at 0°C

$\bar{\tau}_\lambda$ is the mean transmission in the Continnum

C is the temperature coefficient of the filters and $(dJ_\lambda/d\lambda)_A$ refers to the wavelength of the lower cutoff.

(vii) Broad band intensities were computed by

$$J\ (300.525\ nm) = J\ (total) - J\ (OG\ 1) - J\ (2.8\text{-}4.0\ \mu) \qquad (6)$$

where J (2.8-4.0 μ) is a function of water vapour and carbon dioxide. Broad band extinction coefficients a_D were then obtained by comparing

measured intensities to those computed for all extinction processes except turbidity. This led to the first comparison of computed a_D and preset a_D^*.

(viii) For different wavelength bands, the effective wavelength was evaluated by the following expressions:

$$\exp.(-m_D \cdot \beta \cdot p(\lambda_{eff.})) \int_A^B J_\lambda \, d\lambda = \int_A^B J_\lambda \exp.(-m_D \cdot \beta \lambda^{-\alpha}) \, d\lambda \qquad (7)$$

This led to the comparison of computed α and preset α^*.

The above procedure gives us three α-values for the individual spectral bands which are averaged for the entire spectrum and then the average α is used for computing the Ångström's turbidity coefficient β which, together with the turbidity parameter α, provides meaningful informations as regards the transparency of the atmosphere. The pyrheliometric measurements for the hours 0830, 1000, 1130, 1300, 1430, 1600, and 1730 Indian standard time were made on each day of clear sky from February to April, 1972 at the Central Agri-Met. Observatory, Poona (18°32'N, 73°51'E, RL. 555 m) by the author. These data have been used for the analysis, the results of which are presented here.

DISCUSSION OF THE RESULTS

It is evident from the fig. 1 that the individual computations for α following the graphical interpolations for β_g and β_{rr} lack accuracy as the large variations in α-values derived from this method are rather unlikely to occur in the atmosphere. However, the nature of curve for numerically derived α shows variation of much less amplitudes which prevail in the case of real atmosphere. During dry summer season, the average monthly α-values obtained graphically and numerically range between 0.5 to 0.0 and 0.3 to 0.1 respectively when the

tropical continental airmass lies over Poona. Inspite of derived amplitudes of α being small, negative values of exponent α are obtained on some occasions of very clear sky. Rangarajan (1971), who claims that the sunphotometer measurements are selfconsistent, obtained negative values of the exponent α but expressed his doubts as to such cases were unlikely to occur in real atmosphere. It may be mentioned here that such cases depend largely on the microphysical and optical characteristics of the atmospheric haze. The negative values of the exponent α, which are mostly associated with various optical phenomena such as Bishop's rings, coloured clouds and milky canopy, signify that the aerosol extinction increases rather than decreases with increasing wavelength. This type of scattering behavior must be due to an aerosol size distribution different from that expressed by Junge's power law. The anomalous haze extinction has also been observed elsewhere by Quenzel (1970) in clean oceanic air during meteor expedition and Porch et al (1971). Porch and his colleagues analysed the observations of the extinction of star light by the atmosphere for nearly forty years and showed that astronomically determined measurements of aerosol extinction are consistent with pyrheliometric measurements. It is well known that, throughout the summer season, there persists a dense layer of haze in the atmosphere over north and central India. Further, the observations of the past twenty or more years suggest (1963) the presence of extremely tenuous cirrus layer at or just below tropopause in tropics which can only be observed visually with slant illumination. The negative α-values are attributed to the rather size particulates in the atmosphere and also to the presence of thin cirrus clouds.

As regards the turbidity coefficient β, it is seen from fig. 2 that the numerically derived β values range between the limits 0.1 and 0.3 whereas those obtained graphically lie within the interval $0.07 < \beta < 0.44$ for the period considered for this study. On most of the occasions, low values of the exponent α are found to be associated with high turbidity. A good proof for the real nature of diurnal turbidity variations observed by the author is their phase shift with

respect to noon; morning and afternoon β –values differ when compared at the same solar elevation.

CONCLUSIONS

The main results of this paper may be summarized as follows:

(1) The numerically derived exponent α show real variations which are likely to occur in a tropical atmosphere over Poona.

(2) Even on some occasions of clear weather, the negative values of the wavelength exponent α are obtained, which are attributed to rather larger size aerosols and tenuous cirrus clouds.

(3) A phase shift in diurnal turbidity variation is observed with respect to noon.

On the basis of these results, it may be concluded that the better use of pyrheliometric data with recent values of solar spectrum and transmission function of classical Schott filters and evaluating the data by numerical integration, would provide better estimates of atmospheric turbidity parameters in tropics.

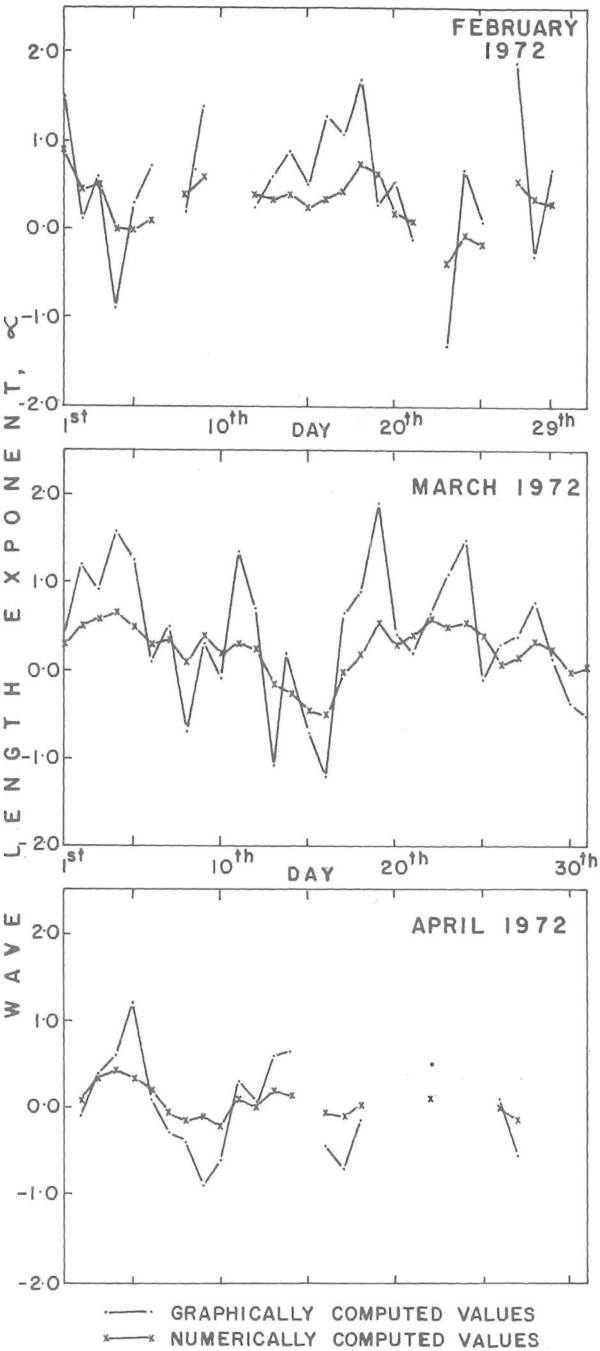

Fig. 1: Variation of wavelength exponent α at Poona; Feb.-April, 1972.

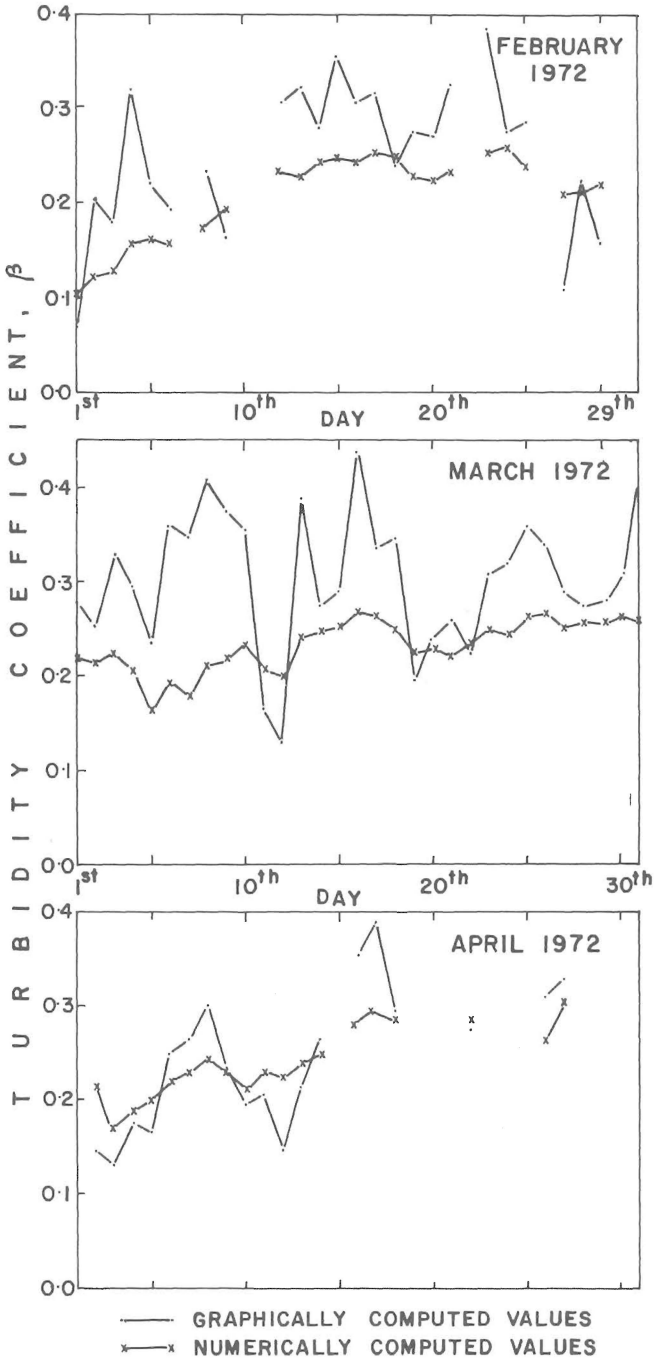

Fig. 2: Variation of Turbidity coefficient β at Poona; Feb.-April, 1972.

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