Illuminance turbidity parameters and atmospheric extinction in the visible spectrum

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SUMMARY

Expressions for the Linke, $T_L$, and Unsworth–Monteith, $T_U$, turbidity parameters in the visible region of the solar spectrum are derived for application in illumination studies. They are based on new determinations of the visible transmittances of various atmospheric constituents, obtained by spectrally integrating the atmospheric transmittances calculated with a spectral irradiance research code. The influence of pressure, of circumsolar radiation, and of different atmospheric constituents (including both stratospheric and tropospheric nitrogen dioxide) on the interpretation of turbidity is discussed. $T_U$ is shown to be virtually free of parasitic variations with air mass or other factors, and to approximate the aerosol optical thickness at 555 nm quite closely.

The experimental $T_U$ has been obtained from measurements of direct illuminance for two days at Albany, New York, and compared to the aerosol optical thickness measured with a multil wavelength radiometer. The observed fair agreement between $T_U$ and the experimental optical thickness at 555 nm suggests that the latter may be accurately approximated by the former, using equipment that can be easily calibrated in the laboratory—unlike sun photometers. The use of $T_U$ in subsequent measurement campaigns is recommended rather than that of $T_L$.

KEYWORDS: Aerosols Nitrogen dioxide Optical thickness Pollution Solar radiation

1. INTRODUCTION

Solar radiation in the visible spectrum (defined here as the interval 380–770 nm, corresponding to the limits of sensibility of the average human eye) is attenuated by the earth’s atmosphere, because of absorption and scattering by various constituents. This attenuation is not constant, even from one hour to the next, partly because of the sun’s apparent motion. Moreover, whereas the distribution of scattering molecules and absorbing mixed gases in the atmosphere remains roughly constant with time and location, the total vertical column of ozone ($O_3$) varies with location and time. The total vertical column of nitrogen dioxide ($NO_2$) varies considerably with time (due to photochemical reactions) and with atmospheric pollution. As with $O_3$, there can be natural stratospheric $NO_2$ and tropospheric $NO_2$ in polluted areas, but, unlike $O_3$, the tropospheric concentration can be some orders of magnitude larger than its stratospheric counterpart. Worldwide average values of the total vertical amount of ozone can be obtained from ground or satellite measurements, from tabulations in the literature (Robinson 1966; Keating et al. 1990), or can be estimated through a general formula (Van Heuklon 1979). The total vertical amount of water vapour ($H_2O$), conventionally expressed in terms of precipitable water, is also highly variable (Iqbal 1983). In contrast to the infrared part of the spectrum, water vapour absorbs only very weakly in the visible.

The last—and potentially dominant—source of extinction is atmospheric aerosols. These particulates deplete the incident solar beam, both by scattering (very effective in the forward direction) and by absorption. Strong aerosol scattering results in a high sky-radiance, considerably more than the background radiance, which is due to Rayleigh scattering only. This compensation effect between direct and scattered (diffuse) visible radiation is of particular importance in illumination studies (e.g. the internal lighting of buildings by day), pollution studies (e.g. visibility reduction), and in plant biology, as

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humans and other living organisms are strongly dependent on the quality and quantity of daylight.

Figure 1 shows the different atmospheric transmittances just reviewed in the visible spectrum, as predicted by a recent solar spectral irradiance code called SMARTS2 (Gueymard 1994a, 1995). This code calculates the downward flux of direct-beam and sky-diffuse irradiance, as well as the circumsolar radiance and irradiance, at a resolution of 1 nm from 280 to 1700 nm, and 5 nm up to 4000 nm. In a comparison with reference measured data and rigorous calculations, the detailed performance of the code was assessed as successful. In particular, Gueymard (1995) found that the code’s predictions of direct irradiance were in good agreement with those of MODTRAN2 (Berk et al. 1989), from which it borrows some algorithms and gaseous absorption data, and with those of BRITE, a Monte Carlo rigorous code (Bird and Hulstrom 1979). Circumsolar radiation calculations are based on the conventional single-scattering approximation, corrected for molecular multiple scattering and backscattering between a non-black ground and the atmosphere (Box and Deepak 1979, 1981). Predictions made using this method are in good agreement with other calculations performed by Tomasi et al. (1990) and with predictions from BRITE, for different atmospheric conditions and radiometer apertures.

The calculations shown in Fig. 1 have been made for atmospheric conditions corresponding to a typical mid-latitude cloudless summer day at 30° solar elevation (air mass 2): pressure 1000 mb, 0.3434 atm-cm O₃, 0.0003 atm-cm stratospheric NO₂ (no tropospheric NO₂ is considered here, although it may be present in large quantities (see subsection 4(d)), 2.4 cm precipitable water, and a rural aerosol whose optical characteristics reproduce those of a conventional model (Shettle and Fenn 1979). The Ångström turbidity coefficient β selected for this aerosol is 0.15, corresponding to optical thicknesses of 0.405 at 500 nm and 0.349 at 555 nm. The transmittances of the three prominent extinction processes (Rayleigh scattering, and aerosol scattering and absorption) are shown in Fig. 1(a), whereas Fig. 1(b) presents the transmittances for the absorption by O₃, NO₂, H₂O and uniformly mixed gases (primarily oxygen, carbon dioxide and methane). The latter effects appear to be of secondary importance compared with those shown in Fig. 1(a), but present very distinct spectral patterns because of selective absorption by gases.

The purpose of this paper is to account for the aerosol effects on visible radiation, and, more specifically, on illumination, using newly defined atmospheric parameters specific to the visible spectrum and including the contributions of O₃, NO₂, uniformly mixed gases and water vapour.

Aerosols are conventionally assessed in terms of spectral or broadband turbidity parameters. The two selected for further development here belong to the broadband category: the Linke and Unsworth–Monteith parameters. The dependence on climatology of these parameters has been investigated, over the whole solar spectrum (e.g. Unsworth and McCartney 1973, Karalis 1976; Katsoulis 1977; Rawlins and Armstrong 1985; Kambezidis et al. 1993), in just the visible portion (e.g. McCartney 1977; McCartney and Unsworth 1978), and for illumination studies (e.g. Navvab et al. 1984; Molineaux et al. 1995). The definition of the appropriate spectrum needs to be clearly stated here. It is essential to distinguish the visible turbidity parameters, derived by McCartney (1977) and McCartney and Unsworth (1978) from the average value of the atmospheric optical thickness in the visible band, from the illuminance turbidity parameters introduced by Navvab et al. (1984), considering the same optical thickness as before, but convolved with the standard spectral luminous efficiency function for photopic vision (hereafter ‘photopic curve’; Illuminating Engineering Society of North America (IES), 1984). In other words, illuminance turbidity

* In its sense as a defined mathematical operation.
Figure 1. Spectral variation, under reference conditions, of the transmittance of all the atmospheric extinction processes considered in this study: (a) molecular and aerosol extinction; (b) ozone, nitrogen dioxide, uniformly mixed gases, and water-vapour absorption.
parameters refer to the detection of illuminance by either an unaided eye or an equivalent photometer. In practice, experimental data of beam illuminance at normal incidence, as obtained with a tracking illuminometer, are needed to evaluate these turbidity parameters.

The rest of this paper concentrates on the illuminance turbidity parameters. They will be derived and then the reports of some experimental tests will be presented. It will be shown that the effect of the convolution with the photopic curve, which has a sharp peak at 555 nm as shown in Fig. 1, is to transform the conventional broadband turbidity-parameters into illuminance parameters that appear quasi spectral (at 555 nm). This finding has very important implications (see section 5).

2. The Linke Turbidity Factor

This parameter was introduced by Linke (1922, 1929) and expresses how many idealized clean dry atmospheres (CDA) would be necessary to produce the same attenuation of the extraterrestrial radiation as the real atmosphere produces. Typically, the Linke turbidity factor, $T_L$, varies between 1 and 10, and is given by

$$T_L = (\ln E_0 - \ln E_b + \ln S)/(m_R \tau_c),$$

(1)

where $\tau_c$ is the broadband optical-thickness of the CDA, a function of the relevant air mass $m_R$. Appendix A provides a list of all symbols used here. An expression for $\tau_c$ is given in appendix B. The illuminance counterpart of $T_L$ is denoted as $T_{li}$ (i denotes illuminance), and, by analogy with Eq. (1), it is given by

$$T_{li} = (\ln E_{0i} - \ln E_{bi} + \ln S)/(m_R \tau_{ci}),$$

(2)

where $\tau_{ci}$ is the illuminance optical-thickness of the CDA, and its expression is given in appendix B. It is remarkable that $\tau_{ci}$ (in contrast to $\tau_c$) is only weakly dependent on $m_R$, as noticed by Navvab et al. 1984. However, $\tau_{ci}$ is significantly dependent on pressure (or site altitude), a result that the derivation of Navvab et al. (1984) did not show. The present determination of $\tau_{ci}$ is compared to that of Navvab et al. in Fig. 2, for two surface pressures: the standard value, $p_0 = 1013.25$ mb (corresponding to sea level), and 810.6 mb (or 0.8 $p_0$, corresponding to a site at about 1840 m). The present determination of $\tau_{ci}$ at sea level is approximately 20% larger than that of Navvab et al. This difference translates into a variation of opposite sign in the values of $T_{li}$ obtained experimentally with the two methods. When the pressure in the present model is decreased by 20%, a reduction of about 20% is observed in $\tau_{ci}$. This is because $\tau_{ci}$ is predominantly a function of Rayleigh scattering, for which the effective optical thickness of the atmosphere above a given level is proportional to the pressure of the air at that level.

3. The Unsworth–Monteith Turbidity Coefficient

The parameter $T_U$ was originally introduced (Unsworth and Monteith 1972) to express the absorption of solar rays by an aerosol-laden atmosphere relative to an aerosol-free one (or ‘clean moist atmosphere’, CMA) with a specified water vapour content. $T_U$ is simply a broadband aerosol optical depth having a value which varies typically from 0 to 1. Its expression is

$$T_U = -(\ln E_b - \ln E_{bw})/m_R.$$
Figure 2. Variation of $\tau_{\lambda}$ with air mass and site pressure, according to derivations from this work and that of Navvab et al. (1984).

For illumination studies, $T_U$ above is modified to obtain $T_{U\lambda}$. This derivation, never proposed before to the authors’ knowledge, is detailed in appendix C. From the respective definitions of $T_{L\lambda}$ and $T_{U\lambda}$, it is easily found that

$$T_{U\lambda} = \tau_{cl}(m_R/m_w)(T_{L\lambda} - 1) - \tau_{wi}.$$  \hspace{1cm} (3)

Equation (3) shows that a simple linear relationship, with a variable slope resulting from the dependence of $\tau_{cl}(m_R/m_w)$ on the air mass, and a marginally variable intercept (through $\tau_{wi}$), exists between $T_{L\lambda}$ and $T_{U\lambda}$ so that conversions of climatologically average values between coefficients can be performed. It appears that $T_{U\lambda}$ is almost proportional to $(T_{L\lambda} - 1)$ since $\tau_{wi}$ is relatively small. The linear relationship will be verified experimentally in section 5.

4. Sensitivity Analysis

The major physical difference between what is measured by $T_L$ or $T_{L\lambda}$ and $T_U$ or $T_{U\lambda}$ is the water-vapour dependence for $T_L$ and $T_{L\lambda}$, while $T_U$ and $T_{U\lambda}$ are closer to the true concept of turbidity as they depend almost entirely on aerosol loading, with only a slight ‘parasitic’ dependence on air mass and other factors, as will be discussed below.

(a) Air mass

The quasi-spectral nature of $T_{L\lambda}$ and $T_{U\lambda}$ is such that their dependence on $m_R$ is far more limited than that of $T_L$ and $T_U$. This is illustrated in Fig. 3, which shows the results of calculations of spectral transmittance and irradiance carried out with SMARTS2 (Gueymard 1995) for the same fixed atmospheric conditions as in Fig. 1. The rural-aerosol model is used with constant optical characteristics (in particular, $\beta = 0.15$) during this
whole (fictitious) day. It can be seen that $T_{LI}$ and, even more so, $T_{UI}$ have negligible variations despite the large variation in solar elevation, whereas $T_L$ and $T_U$, calculated using the method of Kambezidis et al. (1993), display a non-negligible parasitic variation. It is recommended that $T_{LI}$ and $T_{UI}$ be reported using a reference solar elevation of 30° (air mass 2), as has been generally agreed to circumvent the problem caused by the parasitic diurnal variations of $T_L$ (Katsoulis and Tselepidaki 1986; Kasten 1988; Grenier et al. 1994). The corresponding values thus calculated for Fig. 3 are $T_{LI} = 3.89$ and $T_{UI} = 0.346$. The latter value corresponds roughly to $\tau_{555}$, the aerosol spectral optical-thickness at the peak of the photopic curve, 555 nm. (More detailed calculations show that the effective wavelength would be precisely 558 nm in this case.) This means that the aerosol optical-thickness can be estimated at other wavelengths if its spectral variation is known, or can be assumed using Ångström’s equation ($\tau_{\lambda} = \beta \lambda^{-\alpha}$). For instance, an average-wavelength exponent $\alpha = 1.3$ is often used to characterize rural aerosols. This value is close to the 500–2000 nm average of 1.432 calculated for Shettle and Fenn’s rural-aerosol model at 45% relative humidity, or 1.335 calculated for the Standard Radiation Atmosphere (hereafter SRA) continental aerosol (IAMAP 1986); see Gueymard (1995).

Ångström’s $\beta$ can then be easily extrapolated from $T_{UI}$ and $\alpha$. Assuming that the above-mentioned value of $T_{UI}$ (0.346) corresponds to 555 nm, that it is deduced from perfect illuminance measurements, and that $\alpha = 1.3$, a rough estimate of $\beta$ in the case of Fig. 3 would be $0.364 \times 0.555^{-1.3} = 0.161$, only 7% above the exact value of 0.15.

This ‘quasi-spectral’ behaviour of $T_{UI}$ (and in a slightly looser way in the case of $T_{LI}$) can be attributed to the smoothness and relatively limited spectral variations of the main extinction processes in the visible (Rayleigh scattering, and aerosol scattering and absorption, see Fig. 1(a)) which compensate for the relatively large band of the illumination sensors’ filter. (Its ideal full width at half maximum should be about 100 nm according to the photopic curve.) If the reference values of $T_{LI}$ and $T_{UI}$ at $\gamma = 30^\circ$ cannot be experimentally

Figure 3. Variation of turbidity coefficients with solar elevation during a typical mid-latitude summer’s day.
obtained, it is still possible to avoid any significant parasitic diurnal variation with $\gamma$ if the latter is limited to values above $10^\circ$. It is also possible to calculate the reduced $T_{Li}$ at air mass 2, denoted by $T_{Li}(2)$. Different methods have been proposed to reduce $T_L$ at air mass 2 (Katsoulis and Tselepidaki 1986; Kasten 1988; Grenier et al. 1994). By analogy with the method proposed by Kasten (1988), chosen here for its simplicity,

$$T_{Li}(2) = T_L(m_R)\tau_{cl}(m_R)/\tau_{cl}(2).$$

Similarly, the following is proposed to reduce $T_{Uli}$ at air mass 2:

$$T_{Uli}(2) = T_{Uli}(m_R) + [\tau_{cl}(m_R) + \tau_{wi}(m_w)] - [\tau_{cl}(2) + \tau_{wi}(2)].$$

This reduction at air mass 2 prevents the parasitic variations of turbidity caused by time of day, season, or latitude and thus allows easier comparisons of the aerosol content of the atmosphere from turbidity data obtained for different periods or locations.

(b) Pressure

As was shown in section 2 and Fig. 2, $\tau_{cl}$ is significantly affected by variations in pressure. Its effect on $T_{Uli}$ and $T_{Li}$ thus needs to be assessed. When simulating the atmosphere with either a moderate or heavy continental aerosol ($\beta = 0.091$ and 0.319 respectively), and 'reference' conditions ($m_R = 2$, $u_3 = 0.3434$ atm-cm, $u_n = 0.0003$ atm-cm, and $w = 1.419$ cm), it is found that a decrease of pressure from sea level (1013.25 mb) to 810.6 mb (a 20% decrease) translates into an equivalent relative change of opposite sign in $T_{Li}$ (16 to 23%), but affects $T_{Uli}$ only marginally (0.5 to 1%). The latter result is understandable because $T_{Uli}$ is mostly affected by the total aerosol loading, which is not a function of pressure. The large sensitivity of $T_{Li}$ to changes in pressure is generally not a problem because the local pressure is normally available with good accuracy and remains stable on a daily basis, at least under cloudless conditions.

(c) Ozone

Unlike pressure, the $O_3$ content of the total vertical column is not always known with appropriate accuracy, and has significant daily variations. A 20% inaccuracy in $u_3$ has no effect on $T_{Uli}$ and a small but non-negligible effect (2.6 to 3.6% depending on $\beta$, or about one sixth of the $O_3$ variation, with opposite sign) on $T_{Li}$. The methodology and reference conditions used in subsection 4(b) are also used here.

(d) Nitrogen dioxide

The brownish cloud of air pollution, which sits over many industrialized cities during pollution episodes, is clearly seen by the naked eye and therefore affects the illuminance turbidity factors more than the corresponding broadband $T_L$ and $T_{Uli}$ which extend far beyond the visible. Incidentally, this brownish pollution veil signals both a high level of NO$_2$ pollutant and of man-made aerosols (Husar and White 1976). To the authors' knowledge, this is the first methodology incorporating an explicit dependence of turbidity on NO$_2$ absorption. This dependence is essential in the present case, as both NO$_2$ and illumination applications are likely to be important in large cities.

It seems appropriate here to explain how densities of NO$_2$ are related to turbidity. $T_L$ and $T_{Uli}$ are defined with reference to a 'clean' atmosphere (CDA and CMA respectively), but a standard precise definition of this concept is still lacking. At the time of its inception (Linke 1922, 1929), a CDA was essentially a Rayleigh atmosphere with a crude correction for ozone absorption. Absorption by mixed gases was added to the definition of the CDA
Figure 4. Overestimation of turbidity that results if the absorption by tropospheric NO$_2$ in a polluted environment is neglected. Reference conditions as in Fig. 3 (see section 4 of text), except that NO$_2$ here amounts to 0.01 atm-cm.

only recently (Louche et al. 1986; Grenier et al. 1994). Now that spectral and broadband NO$_2$ extinction calculations are possible, it becomes necessary to know if this gas should be considered as a constituent of the CDA/CMA, or if its resulting extinction should be considered as ‘turbidity’ even though it is neither an aerosol nor a scatterer. It is proposed here that the NO$_2$ effects be split between the stratospheric and the tropospheric layers. The former corresponds to natural and relatively minute amounts of NO$_2$, which can be estimated from climatological normals, and for this reason should be considered as part of the CDA/CMA. A vertical column of 0.0003 atm-cm, corresponding to $8.1 \times 10^{15}$ molecules cm$^{-2}$, is used here because it appears to be representative of the average maximum at middle and high latitudes (N oxidon 1979; Solomon and Garcia 1983; Coffey 1988). A 20% variation of the stratospheric NO$_2$ loading around this reference value of 0.0003 atm-cm has a totally negligible incidence (< 0.1%) on both $T_L$ and $T_U$.

Tropospheric NO$_2$ is essentially a sign of pollution, closely paralleling an increase of man-made aerosols, and results in an alteration of visibility, as do tropospheric aerosols. Moreover, as the amount of tropospheric NO$_2$ is highly variable and generally unknown, it is proposed here that tropospheric NO$_2$ should be assimilated into the aerosol turbidity and excluded from the CDA/CMA definition, unlike its stratospheric counterpart. In a large and polluted city, this also means that the calculated values of $T_L$ or $T_U$, and even more specifically of $T_{LI}$ or $T_{UI}$, will tend to overestimate the real aerosol burden slightly, particularly during severe pollution episodes. In the case of $T_{UI}$, the magnitude of this overestimation is about equal to the value of $\tau_m$ for the actual total column of NO$_2$ aloft. As shown in Fig. 4, drawn for the same general conditions as Fig. 3, the overestimation of $T_{UI}$ reaches 0.027 at air mass 2, or about 8% of $T_{UI}$ (2), as defined in Eq. (4), for a NO$_2$ amount of 0.01 atm-cm, representative of severe, near-peak, conditions in a Canadian industrial city (Schroeder and Davies 1987). Pure aerosol turbidity values cannot be derived unless the concomitant total amount of NO$_2$ is measured independently.
Figure 5. Example of circumsolar contribution to direct illuminance calculated for different air masses, different turbidities (in term of Ångström’s $\beta$), and continental (dashed curves) or maritime (continuous curve) reference aerosol models in the Standard Radiation Atmosphere. Only one maritime-aerosol curve is shown for clarity. Bold turbidity-numbers correspond to the same aerosol optical thickness value at 555 nm (0.22).

(e) Water vapour

A variation of 20% of precipitable water around its reference value of 1.419 cm has no effect on $T_{r\lambda}$ and a negligible effect on $T_{i\lambda}$ (less than 0.2%). This could be expected because absorption by water vapour is only marginal at visible wavelengths.

(f) Circumsolar radiation

Beam illuminance is normally measured with a sensor collimated in such a way that it admits radiation within an aperture of about 3° (half angle), and thus contains some parasitic circumsolar radiation from the solar aureole. Under certain atmospheric conditions, this aureole may become very bright and add significantly to the true-beam radiation. The spectral radiance originating from the circumsolar region decreases with wavelength and is thus relatively important in the visible. It is essentially determined by the air mass and the aerosol’s optical thickness and phase function. The latter is itself characteristic of the aerosol origin. Maritime aerosols, such as those modelled in the SRA for instance, are made in great part of oceanic particulates and have a steep phase-function reaching large values in the circumsolar region, thus generating significantly larger circumsolar radiance than aerosols of rural or continental origin (Fig. 5).

If beam-illuminance measurements are used to estimate the aerosol optical-thickness at 555 nm as suggested here, the experimenter faces the problem that a circumsolar correction may be needed, but depends on the variable under scrutiny, viz. the aerosol optical-thickness. The following sensitivity analysis has been performed with the optical circumsolar model in SMARTS2 simulating an instrumental half-aperture angle of 3°. The above fixed reference-conditions were used with three variables: SRA’s continental or maritime aerosol, Ångström’s $\beta$ varying from 0 to 0.3, and air mass varying from 1 to 10. The influence of the latter variable appeared negligible on both $T_{r\lambda}$ and $T_{i\lambda}$. The influence of
the aerosol optical-thickness (or Ångström’s $\beta$) is negligible on $T_{UI}$ and marginal on $T_{LI}$. This means that the most important factor to consider is the origin of the aerosol. For a continental aerosol, $T_{LIX}$, the experimental value of $T_{UI}$ should be multiplied by 1.031 to obtain the ‘true’ $T_{UI}$. For a maritime aerosol, this correction factor increases to 1.101, indicating a noticeable influence which can be explained by the very steep phase-function of this aerosol. The correction factor for the experimental value of $T_{LIX}$ would be such that

$$T_{LI} = T_{LIX}(1 + 0.0248k - 0.05634k \exp(-0.8713T_{LIX})),\,$$

where $k$ is equal to 1 for continental aerosols and to 3.26 for maritime aerosols. The difficulty with this specific correction is to obtain information on the origin of the aerosol or to measure its detailed optical characteristics, all of which are unknown or inaccessible in general. Because, at most inland sites, maritime aerosols are less frequent than rural or continental ones, it is suggested that turbidity measurements be corrected using the factors which have just been given for continental aerosols. (These calculations could be repeated for other aerosol models, and different correction factors would probably be obtained. A more detailed treatment is beyond the scope of this paper for the reason just mentioned—the origin of the aerosol is usually unknown.)

$(g)$ Climatological applications

The intrinsic daily stability of $T_{UI}$, even under changing atmospheric conditions, has been demonstrated in Fig. 3 and the previous sections. Also, using the previously defined reference-conditions and values of $\tau_{a555}$ from 0 to 0.7, the following simple relationship is obtained:

$$T_{UI} = 0.9903\tau_{a555}.$$

This near equality of $T_{UI}$ and $\tau_{a555}$ implies that important information on the aerosol optical-thickness at the peak of the photopic curve can be retrieved simply from experimental measurement of beam illuminance. Only correction for the circumsolar contribution is necessary for an accurate estimation of $\tau_{a555}$; easy monitoring of its variation over time is therefore possible, and can be useful in various applications dealing with aerosol climatology. For example, if simultaneous observations of horizontal visibility or meteorological visual range are available, correlations of aerosol concentration along horizontal and slant paths can be obtained.

All these findings are a direct consequence of $T_{UI}$'s quasi-spectral nature and gives it a decisive edge over $T_{LX}$. Therefore, the adoption of $T_{UI}$ as the standard illuminance turbidity parameter is recommended.

5. Experimental determination of turbidity parameters

This section presents the method used in this study for the illuminance measurements and the subsequent calculation of $T_{LI}$ and $T_{UI}$. As implied by Eqs. (2) and (4), $T_{LX}$ and $T_{UI}$ can both be obtained indirectly from measurements of direct illuminance. These parameters can provide the necessary information on aerosol loading without the need for more involved spectral measurements. They are particularly useful in many locations because of the lack of spectral data, and because of the renewed interest in daylighting applications, among other reasons. (If $T_{LX}$ and $T_{UI}$ are known for a given site or can be extrapolated from a nearby site, illuminances, a rarely measured quantity, can be predicted.) In this work, the calculations concern the atmosphere over Albany, New York (42° 42' N, 73° 50' W, 79 m a.m.s.l.) during two cloudless days in 1992. The method can, of course, be applied worldwide and for long periods of time, as will be shown in a future paper.
The experimental data used here were obtained from an illumination station set up on the roof of the Atmospheric Sciences Research Center building, as described by Dumortier et al. (1994). This station is equipped to measure direct-beam illuminance (with a Licor 210 SZ sensor) and atmospheric optical-thickness at 415, 500, 610, 665, and 862 nm (with an automated Multi-Filter Rotating Shadow-band Radiometer (MFRSR), described by Harrison et al. (1994)). The daylight station was set up, and the data quality-controlled, following CIE recommendations (Kendrick 1994). The MFRSR spectral data were analysed according to the objective algorithms of Harrison and Michalsky (1994). Only hourly data of illuminance and optical thickness were used.

The MFRSR spectral data can be used to test whether the aerosol optical-thickness follows the ideal Ångström power-law. It apparently does so for the two test days, as exemplified in Fig. 6. However, the wavelength exponent $\alpha$ was very low (about 0.41) on day 1 (13 April 1992), compared to a more typical value of about 1.09 on day 2 (4 June 1992). The first day’s low value of $\alpha$ may be the result of an incursion of maritime aerosols and/or a strong influence of stratospheric aerosols, which are known to be characterized by very low—and sometimes negative—values of $\alpha$ (Guermard 1994b). The latter explanation is highly probable because the stratospheric cloud of aerosol from Mount Pinatubo was particularly dense over the United States in early 1992 (see, for example, Michalsky et al. 1994).

Figure 7 shows the experimental relationship between $T_{Li}$ and $T_{Ui}$ for the same two days. A continental aerosol has been assumed and the circumsolar correction calculated as described in subsection 4(f). A least-squares fit of these data points gives

$$T_{Ui} = 0.1124T_{Li} - 0.0968. \tag{5}$$

Equation (5) has a high coefficient of correlation (0.997) and shows a strong correspondence to Eq. (4), which, for the average atmospheric conditions during the two days of the experiment, would become

$$T_{Ui} = 0.1196T_{Li} - 0.1223. \tag{6}$$

Finally, Fig. 8 shows hourly values of $\tau_{555}$ as predicted from $T_{Ui}$ by the method discussed above and as obtained experimentally by wavelength interpolation of the MFRSR data. (A logarithmic interpolation was used, assuming an ideal Ångström power-law limited to the interval 500–610 nm.) The predicted values tend to follow the apparent slight diurnal variations of turbidity correctly. However, the very good agreement between predicted and experimental early-morning data has a tendency to deteriorate somewhat as the day progresses. Besides the possible shortcomings of the method used to predict $\tau_{555}$, the overall uncertainty of the measurements has to be taken into account. It results from a combination of (i) the calibration, linearity, and temperature dependence errors of the illuminance sensor; (ii) the calibration, linearity, and temperature dependence errors of two channels (500 and 610 nm) of the MFRSR; (iii) the uncertainty associated with the data-reduction technique; (iv) the uncertainty of the wavelength interpolation used to obtain the aerosol optical-thickness at 555 nm from observations at 500 and 610 nm; and

* The experimental aerosol optical-thickness at 500 and 610 nm were obtained by subtracting the calculated contributions from Rayleigh scattering and $O_3$ and NO$_2$ absorption from the observed total atmospheric optical-thickness. Total column abundances of 0.3 atm-cm for $O_3$ and 0.0005 atm-cm for NO$_2$ were assumed, but no convolution of their transmittance with the (unknown) radiometer’s filter-function was attempted. No absorption from water vapour or mixed gases was considered at these wavelengths. A better agreement between the predicted and ‘observed’ values of $\tau_{555}$ would have been obtained with lower abundances of ozone and, secondarily, nitrogen dioxide. This possibility is plausible because the measurement period was still strongly affected by the stratospheric perturbations that followed Mount Pinatubo’s volcanic explosion.
Figure 6. Spectral variation of the aerosol optical-thickness during two cloudless days at Albany, New York, USA. Straight lines are power-law fits of the experimental data, modelled on Ångström’s equation. (a) Day 1 (13 April 1992); (b) day 2 (4 June 1992).

(v) the uncertainty of the circumsolar correction that was applied to the predicted values. Despite these uncertainties in individual measurements, the daily averages of the predicted and observed $\tau_{455}$ are in fair agreement: 0.280 vs 0.261 for day 1, and 0.320 vs 0.311 for day 2.

6. Conclusions

There is a growing worldwide interest in atmospheric turbidity since it is related to air quality and because of its wider significance in tropospheric chemistry, climate and biological and illumination studies. This work was motivated by the last consideration, and the Linke and Unsworth–Monteith turbidity parameters were used in an in-depth analysis,
including a revision of the 'clean dry atmosphere' concept upon which the Linke coefficient is based. The derivation of the illuminance Unsworth–Monteith turbidity parameter has been proposed. It is shown that it has decisive advantages over its Linke counterpart, namely its negligible daily fluctuations and its quasi-spectral estimate of the aerosol optical thickness at 555 nm. The use of $T_{UI}$ is therefore recommended for illuminance and aerosol-climatology studies.

Because of the way broadband turbidity-coefficients have been defined, a part of what is attributed to aerosol extinction (or 'true' turbidity) is in fact attributable to extinction by other constituents, such as NO$_2$. This particularly affects the calculation of illuminance turbidity because NO$_2$ absorbs only in the visible part of the solar spectrum, and because its absorption is incorrectly interpreted as aerosol extinction. Because tropospheric NO$_2$ is directly related to pollution which is found in large cities, the method presented here may have a slight bias towards an overestimation of the visible aerosol-extinction. Removal of this bias would require simultaneous measurements of the vertical amount of NO$_2$.

Another experimental bias arises from the circumsolar parasitic contribution. A constant factor has been proposed, to correct the apparent values of turbidity. It is non-negligible but, for $T_{UI}$, solely dependent on the origin of the aerosol.

The fair agreement observed between $T_{UI}$ and the experimental optical-thickness at 555 nm suggests that the latter may be accurately approximated by the former. This makes the measurement of $T_{UI}$ useful in monitoring aerosol effects in the visible spectrum, for example in visibility-degradation studies. A major benefit of the proposed technique is that it uses only standard equipment (a tracking collimated illuminance-sensor) which can be easily calibrated with a standard lamp in the laboratory, rather than sun photometers which need frequent and costly absolute recalibrations at remote sites of exceptional atmospheric clarity and stability.
ACKNOWLEDGEMENTS

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APPENDIX A

List of acronyms

CDA clean dry atmosphere
CIE Commission Internationale de l’Eclairage (International Commission on Illumination)
CMA clean moist atmosphere (or aerosol-free atmosphere)
MFRSR Multi-Filter Rotating Shadow-band Radiometer
SRA Standard Radiation Atmosphere

List of Symbols

$q_{jk}$ coefficient of Eq. (B.5)
$E_b$ measured direct (normal incidence) irradiance at the earth’s surface (W m$^{-2}$)
$E_{bw}$ predicted direct (normal incidence) irradiance at the bottom of a CMA (W m$^{-2}$)
$E_{bn}$ measured direct (normal incidence) illuminance at the earth’s surface (klux)
$E_o$ extraterrestrial irradiance at the mean sun-earth distance, or ‘solar constant’, 1367 W m$^{-2}$
$E_{Oi}$ extraterrestrial illuminance at the mean sun-earth distance, 133.25 klux
$m_j$ optical mass for constituent $j$
$p$ atmospheric pressure (mb)
$p_0$ reference atmospheric pressure, 1013.25 mb
$S$ correction factor for the actual sun-earth distance
$T_L/T_D$ broadband/illuminance Linke turbidity-factor
$T_U/T_D$ broadband/illuminance Unsworth-Monteth turbidity-coefficient
\( \frac{T_{Li}}{T_{Ua}} \) experimental values of \( T_{Li} \) and \( T_{Ua} \), before circumsolar correction

\( u_n \) stratospheric nitrogen-dioxide amount (atm-cm)

\( u_3 \) total ozone amount (atm-cm)

\( \gamma_l \) spectral luminous efficacy (photopic curve)

\( w \) precipitable water in the atmosphere (cm)

\( \alpha \) wavelength exponent for Ångström’s turbidity-coefficient

\( \beta \) Ångström’s turbidity-coefficient

\( \gamma \) apparent solar altitude (°)

\( \lambda \) wavelength (nm or µm)

\( \tau_{555} \) optical thickness of aerosols at 555 nm

\( \tau_i/\tau_{ai} \) broadband/illuminance optical-thickness of aerosols

\( \tau_c/\tau_{ci} \) broadband/illuminance optical-thickness of a CDA

\( \tau_R/\tau_{ri} \) broadband/illuminance optical-thickness of mixed gases

\( \tau_a/\tau_{ai} \) broadband/illuminance optical-thickness of NO\(_2\)

\( \tau_O/\tau_{oi} \) broadband/illuminance optical-thickness of O\(_3\)

\( \tau_R/\tau_{wi} \) broadband/illuminance optical-thickness resulting from Rayleigh scattering

\( \tau_w/\tau_{wi} \) broadband/illuminance optical-thickness of water vapour.

The subscripts a, g, n, 3, R and w in any symbol refer to aerosols, mixed gases, NO\(_2\), O\(_3\), Rayleigh and water vapour atmosphere, respectively. The subscript i refers to the word ‘illuminance’; it modifies all the \( E \) symbols from irradiance (W m\(^{-2}\)) to illuminance (klux). The subscript \( \lambda \) refers to a spectral value.

**APPENDIX B**

*Calculation of \( T_{Li} \)*

For the broadband case, \( T_{Li} \) is classically defined by

\[
E_b/(SE_0) = \exp(-m_R \tau_c T_L), \tag{B.1}
\]

where the observed broadband atmospheric transmittance may alternatively be expressed as a function of broadband (instead of spectral) optical thicknesses by means of a pseudo-Bouguer’s law

\[
E_b/(SE_0) = \exp\{-m_R \tau_R + m_g \tau_g + m_3 \tau_3 + m_n \tau_n + m_w \tau_w + m_s \tau_s\}. \tag{B.2}
\]

The limiting case of the ‘clean dry’ atmosphere (CDA) corresponds to \( \tau_w = \tau_s = 0 \) and \( T_L = 1 \). Therefore, when combining Eqs. (B.1) and (B.2), the optical thickness of the ideal CDA is readily obtained as

\[
\tau_c = \tau_R + \tau_g + (\tau_3 + \tau_n) m_3/m_R,
\]

giving the following expression for \( T_L \)

\[
T_L = 1 + (m_w/m_R)(\tau_n + \tau_w)/\tau_c. \tag{B.3}
\]

In Eq. (B.3) it is assumed that \( m_g = m_R, m_n = m_w \) and \( m_n = m_3 \) (Gueymard 1995). Reduced paths of stratospheric NO\(_2\) up to about 0.0005 atm-cm with typical values of 0.0001 atm-cm in winter and 0.0002 atm-cm in summer can be assumed. Equation (B.1) or Eq. (1) may be rewritten for the illuminance case as

\[
T_{Li} = (\ln E_{bi} - \ln E_{ai} + \ln S)/(m_R \tau_{ci}), \tag{B.4}
\]

or, alternatively, from (B.3),

\[
T_{Li} = 1 + (m_w/m_R)(\tau_{ai} + \tau_{wi})/\tau_{ci}.
\]
TABLE B.1. Coefficients $a_j$ for the optical masses for use in Eq. (B.5).

<table>
<thead>
<tr>
<th>Extinction process</th>
<th>$a_{j1}$</th>
<th>$a_{j2}$</th>
<th>$a_{j3}$</th>
<th>$a_{j4}$</th>
<th>$m_j$ at $\gamma = 0^\circ$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rayleigh</td>
<td>4.5665 x 10^{-1}</td>
<td>0.07</td>
<td>96.4836</td>
<td>-1.6970</td>
<td>38.136</td>
</tr>
<tr>
<td>Ozone</td>
<td>2.6845 x 10^{2}</td>
<td>0.50</td>
<td>115.4200</td>
<td>-3.2922</td>
<td>16.601</td>
</tr>
<tr>
<td>Nitrogen dioxide</td>
<td>6.0230 x 10^{2}</td>
<td>0.50</td>
<td>117.9600</td>
<td>-3.4536</td>
<td>17.331</td>
</tr>
<tr>
<td>Mixed gases</td>
<td>4.5665 x 10^{-1}</td>
<td>0.07</td>
<td>96.4836</td>
<td>-1.6970</td>
<td>38.136</td>
</tr>
<tr>
<td>Water vapour</td>
<td>3.1141 x 10^{-2}</td>
<td>0.10</td>
<td>92.4710</td>
<td>-1.3814</td>
<td>71.443</td>
</tr>
<tr>
<td>Aerosols</td>
<td>3.1141 x 10^{-2}</td>
<td>0.10</td>
<td>92.4710</td>
<td>-1.3814</td>
<td>71.443</td>
</tr>
</tbody>
</table>

For the calculation of the average optical thicknesses, version 2.7 of SMARTS2, a public-domain cloudless spectral solar-radiation code (Gueymard 1994a, 1995), was used. The spectral integration of the transmittance of each extinction process in the range 380–770 nm was performed by weighting them with the spectrum that results from extinction processes which have operated along the path of the solar beam earlier. For O$_3$ and stratospheric NO$_2$, the extraterrestrial spectrum is used for the weighting function because the bulk of all the other extinction processes occur below them. For the Rayleigh and mixed-gases extinctions, the weighting function should not be the extraterrestrial spectrum as has been considered conventionally (e.g. Navvab et al. 1984; Louche et al. 1986), but a spectrum altered by the preliminary absorption caused by O$_3$ and NO$_2$. In equation form

$$ T_{RI} = \left[ \int_{380}^{770} T_{RA} T_{3A} T_{nA} V_\lambda E_{0A}\ d\lambda \right] \left[ \int_{380}^{770} T_{3A} T_{nA} V_\lambda E_{0A}\ d\lambda \right]^{-1}, $$

where $T_{RI}$ is the average visible Rayleigh transmittance, $T_{RA}$ its spectral counterpart, $T_{3A}$ and $T_{nA}$ are the O$_3$ and NO$_2$ spectral transmittances, $V_\lambda$ is the spectral luminous efficiency (IES 1984), and $E_{0A}$ is the extraterrestrial spectrum, from tabulations at 1 nm intervals in the visible (Gueymard 1995). For the numerical integration needed to evaluate Eq. (B.4), a trapezoidal rule as well as typical values of $\mu_3$ (0.3434 cm, as obtained from the U.S. Standard Atmosphere) and $\mu_n$ ($\mu_3 = 3 \times 10^{-4}$ cm; see section 4) were used. The illuminance Rayleigh and mixed-gas transmittances are only marginally affected by variations around these values of $\mu_3$ and $\mu_n$. The illuminance Rayleigh optical-thickness is then obtained as $\tau_{RI} = -\ln T_{RI}/m_R$. The process is repeated for different air-masses and pressures so that a fit of $\tau_{RI}$ as a function of $m_R$ and $p$ can be obtained (see below). The same method is used to derive the illuminance mixed-gas optical thickness. The method of numerical integration described above is called a ‘multilayer spectrum weighting’ averaging process.

The general expression for the various air masses encountered is given by Gueymard (1995)

$$ m_j = \{\sin \gamma + a_{j1}(90 - \gamma)^{a_{j2}}(a_{j3} + \gamma - 90)^{a_{j4}}\}^{-1} \quad \text{(B.5)} $$

where the values of the coefficients are given in Table B.1 for each extinction process $j$.

Parameter $E_{RI}$ in Eq. (B.4) represents the measured direct-beam illuminance at normal incidence. The actual sun–earth distance $S$ and the solar declination, needed to obtain the sun’s position, can both be obtained precisely from ephemerides such as the Nautical Almanac, or may be approximated from simple algorithms (e.g. Iqbal 1983; Michalsky 1988).

The following functions were derived by application of the method described above, and are valid for $0 \leq \gamma \leq 90^\circ$ or $1 \leq m_R \leq 38$. 
• Rayleigh component

\[ \tau_{Ri} = 0.0966p/p_0 - (4.606 \times 10^{-5} + 2.669 \times 10^{-4} p/p_0) m_R \]

• Ozone component

\[ \tau_{3i} = 0.0768u_3 \]

• Uniformly-mixed-gas component

\[ \tau_{gi} = 2.569 \times 10^{-4} (p/p_0)^{0.3} / (1 + 0.32245m_R) \]

• Nitrogen dioxide component

\[ \tau_{ni} = 2.48u_n \]

where the stratospheric NO\textsubscript{2} amount, \( u_n \), is estimated as discussed earlier in this appendix.

**APPENDIX C**

**Calculation of \( T_{ui} \)**

The Unsworth–Monteith parameter is a broadband aerosol optical-thickness. This means that \( T_{ui} \) is the same as \( \tau_s \) in Eq. (B.2). By analogy with Eq. (B.2), the illuminance counterpart of \( T_{ui} \) may thus be obtained from

\[ E_{bi}/(SE_{bi}) = \exp\{- (m_R \tau_{Ri} + m_g \tau_{gi} + m_3 \tau_{3i} + m_n \tau_{ni} + m_w \tau_{wi} + m_s T_{ui})\} \]

Assuming \( m_s = m_w, m_g = m_R, \) and \( m_n = m_3 \) as in appendix B, the above equation can be rewritten

\[ T_{ui} = - (\ln E_{bi} - \ln E_{oi} - \ln S + m_R \tau_{ci} + m_w \tau_{wi})/m_w. \]  \hspace{1cm} (C.1)

A combination of this equation with Eq. (B.4) yields Eq. (3) after elimination of the term \( (\ln E_{bi} - \ln E_{oi} - \ln S) \) and simple manipulations. The water-vapour transmittance is obtained with the same ‘adjusted spectrum weighting’ averaging process as in appendix B. The following fit is finally obtained as a function of precipitable water

\[ \tau_{wi} = (1.506 \times 10^{-3} + 6.956 \times 10^{-4} p/p_0) w^{0.97} / (1 + 0.065m_w), \]

where \( w \) can be obtained from radiosonde data, or can be estimated from surface temperature and humidity using empirical models (e.g. Iqbal 1983; Gueymard 1994b).

The optical thickness \( \tau_{ci} \) and the optical masses \( m_R \) and \( m_w \) are obtained from appendix B.

**REFERENCES**


