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ABSTRACT

A large set of illuminance and irradiance data has been collected for four years at 15-minute intervals in San Francisco. This data set has been used to investigate the impact of atmospheric turbidity on daylight calculations. Existing predictive formulae for Linke turbidity, T_L, provide moderate agreement to measured values of T_L when using nominal design values for the Angström scattering coefficient, β, and precipitable water vapor, w. When average measured values for β and w are used, the agreement improves. We suggest the use of an illuminance turbidity, T_{il}, to calculate direct normal illuminance directly. We derive a simple approximate solution, $T_{il} = 1 + 21.6\ \beta$. T_{il} appears to be a better parameter to describe atmospheric conditions since, unlike T_L, it is insensitive to air mass and thus solar altitude or time of day. We present and compare plots of T_{il} and T_L vs. solar altitude, time of day, and month. Finally, we examine and compare several alternative pathways to derive direct normal illuminance from irradiance and luminous efficacy (dependent on β and w), or directly from β.

INTRODUCTION

Predicting the illuminance from direct sun and sky on clear days is essential to any study of daylighting strategies in buildings. During the past two decades, a number of researchers have developed predictive models that account for atmospheric turbidity. These models quantify the influences of atmospheric aerosols, gases, and water vapor on direct and diffuse radiation at the Earth's surface. The models enable researchers to estimate irradiance or illuminance on a given surface at a specific time, day, and location.

Although theoretically derived algorithms for illuminance data can be developed, atmospheric scattering and absorption processes are sufficiently complex and microclimate-dependent that a measured data base is essential. Researchers have reported results of availability studies in Australia, Japan, South Africa, and several European countries, but there is a paucity of data for the United States.

For most daylight calculations with clear skies, it is essential to know the mean clarity of the atmosphere, which is commonly expressed in terms of turbidity factor. The currently used parameters are Linke's turbidity factor (T_L) and the Angström coefficient (β). In this paper we compare these calculation methods with our measurements and introduce a new concept, which we call illuminance turbidity (T_{il}).

DATA COLLECTION

Data were collected in San Francisco, California, at 38° north latitude, 123° longitude within the Pacific time zone. The station is situated on a peninsula that separates the large San Francisco Bay from the Pacific Ocean. Sea fog and low cloudy skies in the early mornings are characteristic of this area. There is a great variation of local climates within the Bay Area [1]. The data were collected on top of the Pacific Gas and Electric (PG&E) building in the city's financial district at 140 m (450 ft) above sea level. The instrumentation at the station consists of seven illuminance sensors, of which two measure global and diffuse illuminance on a horizontal plane and four measure vertical illuminance at each cardinal orientation. The seventh is used as a luminance sensor for zenith luminance measurements. Two pyranometers measure global and diffuse irradiance on a horizontal surface. For detailed information about the station and the type of data collection system, see Refs. [2] and [3]. Additional short-term measurements made at this location are described later in this paper.

THEORETICAL BACKGROUND

The solar flux in the direct beam that reaches the Earth's surface as a function of wavelength, λ, is given by the Bouguer-Lambert law. Direct normal irradiance is calculated by the integral over the entire range of wavelengths (λ):

$$E_{esn} = \frac{1}{S} \int_0^{\infty} E_{eo}(\lambda) e^{-a(\lambda)m} \, d\lambda , \tag{1}$$

where E_{eo} (λ) is the solar constant at the mean sun-Earth distance as a function of λ, E_{esn} is the direct normal irradiance at the Earth's surface, m is the absolute air mass, S is a parameter that normalizes for the variations of E_{eo} within the sun-Earth distance as defined later, and a(λ) is the overall extinction coefficient per unit air mass as a function of λ. The absolute air mass is a product of a pressure correction factor and a geometric factor called the relative air mass, $m_r(\epsilon_s)$:

$$m = (p/1000)\, m_r(\epsilon_s), \tag{2}$$

where p is the pressure in mb and ϵ_s is the angular distance of the sun from the zenith. For $\epsilon_s < 80°$, m_r can be approximated by SEC(ϵ_s), where SEC(ϵ_s) is the slant height through a plane-parallel atmosphere [4]. For larger angles a more exact formula should be used [see Eq. (7)]. The normalization parameter, S, is given as the square of the ratio of the actual sun-Earth distance, R_a, to the mean sun-Earth distance, R_m [5,6]:

$$S = \frac{R_a^2}{R_m^2} \quad . \tag{3}$$

The overall extinction coefficient per unit air mass, a(λ), has a number of complex constituent terms. In a perfect (non-absorbing) Rayleigh atmosphere, scattering occurs from molecules of atmospheric gases that are much smaller than the wavelength of light; this produces a strong wavelength dependence ($1/\lambda^4$) in the scattering coefficient. However, it is well known that even in the clearest atmosphere, there is significant additional scattering and absorption due to atmospheric aerosols and other natural and man-made atmospheric components. Accordingly, several models have been developed to predict quantitatively the attenuation of direct beam radiation in a turbid atmosphere. We review these in the sections that follow, describe the results of measurements and calculations using each model, and comment on the relative merits of each.

MODELS OF TURBIDITY

Linke turbidity factor

The intensity of solar radiation reaching the Earth's surface through a turbid atmosphere depends on several scattering and absorption processes in which gases, water vapor, dust, and aerosols are the major variables in addition to solar altitude. Equation (1) can be rewritten with $\overline{a}_r$, the mean extinction coefficient integrated over all wavelengths, where $\overline{a}_r$ accounts for scattering and absorption processes. Linke proposed to express $\overline{a}_r$ in terms of the product of a Rayleigh scattering term for an ideal clear atmosphere, α_r, and the Linke turbidity factor, T_L, where T_L can be interpreted as the number of equivalent Rayleigh atmospheres required to produce the same extinction. Thus,

$$E_{esn} = E_{eo} e^{-\overline{a}_r m} \, , \tag{4.1}$$

and

$$E_{esn} = E_{eo} e^{-\alpha_r m \, T_L} \, . \tag{4.2}$$

where $T_L > 1$. T_L can then be calculated from Eq. (4.2):

$$T_L = (\ln E_{eo} - \ln E_{esn})(\alpha_r m)^{-1}, \tag{5}$$

where $E_{eo} = 1370\ \mathrm{W/m^2}$ and:

$$\alpha_r = (9.4 + 0.9m)^{-1} \qquad \textit{[from Kasten, Ref. 7]} \tag{6}$$

$$m = \frac{-\sin\gamma_s + [\sin^2\gamma_s - 1 + (1.001572)^2]^{1/2}}{0.001572} \qquad \textit{[from Mahotkin, Ref. 7]} \tag{7}$$

and $\gamma_s = 90 - \epsilon_s$.

The integrated Rayleigh scattering coefficient, α_r, depends primarily on m. The Linke turbidity factor, T_L, depends on precipitable water vapor and scattering aerosols as well as on m. Although T_L is useful for comparing atmospheric properties for a variety of clear sky conditions, it has one drawback: measurements suggest that, under constant clear conditions, T_L varies with m (see, for example, Ref. 5). This shows that the absorption and scattering processes based on aerosol and water vapor content are also wavelength-dependent. Thus T_L by itself is not a good indicator of the atmospheric aerosol and water vapor concentrations that influence daylight availability.

The dependance of T_L on m creates a problem when analyzing long-term solar irradiation data. In the analysis of our data, we used the criterion, recommended by the World Meteorological Organization (WMO), that a clear sky condition exists when the direct normal irradiance equals or exceeds 200 watts per square meter (W/m^2). This irradiance corresponds to the minimum intensity required to scorch the paper strip of a Campbell-Stokes sunshine recorder. The criterion limits the maximum value of T_L for clear conditions and makes it a function of solar altitude. Equations (5) and (6) can be solved for T_L with $E_{esn} \geqslant 200$ W/m^2 as a function of solar altitude (through m), resulting in Eq. (8):

$$T_L(max) = (ln\ E_{eo} - \ln 200)\ \frac{9.4 + 0.9\ m}{m} \tag{8}$$

These results are presented in Fig. 1. They suggest that a comparison across seasons of calculated values of T_L based on measured data may be misleading because the maximum value of T_L obtainable at low altitudes is in part an artifact of the clear sky radiation criterion. Changing the clear sky "definition" to $E_{esn} \geqslant 400$ W/m^2, so as to analyze only very clear days, shifts the maximum T_L line lower, as shown in Fig. 1, further constraining the "allowable" T_L at low solar altitudes. Furthermore, any threshold type definition of a "clear sky" will still allow some cirrus cloud contaminated data into the "clear" conditions. Direct visual observation or additional test criteria (e.g., diffuse/global ratio) may help in further analysis. This suggests that alternative clear sky definitions may be desirable.

Dogniaux [8] and Valko [9] have both suggested empirical fits of the Linke turbidity factor (T_L) as a function of aerosol and water vapor content and relative air mass. Dogniaux's formula [8], valid for $5° < \gamma_s < 65°$, is:

$$T_L = \left[\frac{(\gamma_s + 85)}{(39.5e^{-w} + 47.4)} + 0.1 \right] + (16 + 0.22w)\beta\ . \tag{9}$$

Valko's formula [9], valid for $5° < \gamma_s < 65°$, is:

$$T_L = (B + 0.54)[1.75\ log(w/m + 0.1) + 14.5] - 5.4. \tag{10}$$

The variables in both equations are:

T_L = Linke's turbidity factor,
m = air mass,
γ_s = solar altitude (deg.),

β = Angström turbidity coefficient (see below),
w = water vapor content in the atmosphere (cm), and
B = 1.07 β (by Schuepp) [9].

Dogniaux provides nominal annual average values of β and w that can be used to estimate T_L for sites where β and w have not been measured. San Francisco is an urban temperate zone, which gives (from Dogniaux) $\beta = 0.1$ and $4 > w$ (cm) > 2. We calculate T_L at 15-minute intervals from our measured data, using Eq. 5 and values of E_{esn} derived from measured global irradiance, measured diffuse irradiance, and shadow band corections based on Ref. 11. Figure 2 plots our monthly averaged measured turbidities against values calculated from Dogniaux's and Valko's formulae with $\beta = 0.1$ and $w = 2$. The formulae do not give constant values over the year because the distribution of clear sky solar altitudes changes over the year. As an annual summary of the adequacy of the formulae, we have calculated the average monthly ratios T_L (calculated)/T_L(measured) and standard deviation to be 1.19 $\pm$ 0.03 (Dogniaux) and 1.18 $\pm$ 0.04 (Valko). Figure 2 shows that single monthly ratios will differ by a larger factor.

If we repeat the calculation using $w = 4$ cm, the formulae give a less accurate answer in comparison to our measured data. The T_L(calculated)/T_L(measured) ratio for Dogniaux is 1.27 and for Valko 1.29.

To better determine the accuracy of the basic formulae, we need additional data on β and w. In the next sections, we discuss how these parameters can be estimated from measured data. We then again compare the formulae to our measurements.

Angström turbidity coefficient, β

The Angström turbidity coefficient, β, provides a means of estimating the scattering due to aerosols. The overall extinction coefficient, $a(\lambda)$, can be written as a weighted sum of three extinction coefficients:

$$a(\lambda) = \alpha_R(\lambda) + \frac{m_r}{m} \cdot \alpha_D(\lambda) + \frac{m_r}{m} \cdot \alpha_W(\lambda), \qquad (11)$$

where $\alpha_R(\lambda)$ is the Rayleigh scattering coefficient per unit m, absolute air mass; $\alpha_W(\lambda)$ is the selective absorption coefficient for gases within the atmosphere (O_2, O_3, H_2O, CO_2, etc.); and $\alpha_D(\lambda)$ is the aerosol extinction coefficient. The aerosol extinction coefficient is weighted by m_r/m because the dust concentration depends on the path length but not the air pressure, p. This ratio is also used to weight the absorption coefficient for gases because of the dominance of water vapor (which is independent of p) in this term.

According to Angström the aerosol extinction coefficient can be calculated by the following formula:

$$\alpha_D(\lambda) = \beta/(\lambda^{\alpha}), \qquad (12)$$

where β is the Angström turbidity coefficient, and the wavelength exponent α depends on the size of atmospheric particles. The most commonly used value is $\alpha = 1.3$, which was recommended by Angström. Given a value for α, it is necessary to determine α_D in order to find β.

For our daylighting applications, the effect of $\alpha_w(\lambda)$ on the total extinction coefficient can be largely eliminated by noting that $\alpha_w(\lambda)$ is dominated by absorption of radiation by water vapor, which occurs primarily in the far red and infrared. Thus the Rayleigh coefficient and aerosol extinction coefficient can be estimated from measurements of the solar irradiance in the visible region of the spectrum. To make these measurements, we used a Schott RG2 filter having a nominal cutoff at 630 nm. The filter was mounted on a normal-incidence pyrheliometer with a sun-tracking system.

Coulson provides a figure to determine β (for α = 1.3) as a function of measured irradiance (for $\lambda < 630$ nm) and air mass, m (Fig. 3.5 in Ref. [6]). To facilitate computer-assisted analysis of our measured data, we used Coulson's data to fit the aerosol coefficient, α_D, as a function of β and m:

$$\alpha_D = \frac{\beta \, \alpha_{R_c}}{(C_c + d_c m)}. \tag{13}$$

Using this equation for α_D, we can substitute Eq. (11) into Eq. (4.1) and derive the following expression for β:

$$\beta = \frac{[ln(E_c/E_{esn}) - \alpha_{R_c} m] \cdot [C_c + d_c m]}{\alpha_{R_c} m}, \tag{14}$$

where E_c is the extraterrestrial solar constant for $\lambda \leqslant 630$ nm, E_{esn} is the measured flux, α_{R_c} is the Rayleigh coefficient for $\lambda \leqslant 630$ nm (0.1865 from Ref. [6]), and C_c and d_c are the coefficients we derived to fit the air mass dependence of the mean aerosol extinction coefficient.

Because the actual cutoff wavelength, λ_c (λ at which T = 50%), will vary from the nominal value (λ = 630 nm for the RG2 filter), each of the wavelength-dependent parameters in Eq. (14) must be adjusted to reflect this $\Delta\lambda = \lambda_c$ - 630 nm. Using Coulson's tabular data for this $\Delta\lambda$ effect, we developed a series of equations to correct the nominal values in Eq. (14):

$$E_c = E_{630} + \Delta E \; \Delta\lambda = 499 \; + 1.6 \, \Delta \, \lambda \; (W/m^2) \tag{14.1}$$

$$\alpha_{R_c} = \alpha_{R_{630}} + \Delta \, \alpha_R \; \Delta \, \lambda = 0.1865 - 0.00062 \, \Delta \, \lambda \tag{14.2}$$

$$C_c = C_{630} + \Delta \, C \; \Delta \, \lambda = 0.0696 - 0.00008 \, \Delta \, \lambda \tag{14.3}$$

$$d_c = d_{630} + \Delta \, d \; \Delta \, \lambda = 0.00272 - 0.0000340 \, \Delta \, \lambda \tag{14.4}$$

We used a spectrophotometer to measure the spectral transmittance of our filter and determined that $\Delta \lambda$ = 0.5 nm. Thus our correction terms were very small.

The resultant values of β estimated by applying Eq. (14) to hourly data collected during a 3-month period (December 1982 - February 1983) show no obvious dependance on air mass. Figure 3 shows the frequency distribution of our results, which have a mean of 0.088, a standard deviation of the distribution, σ, of 0.03 and a standard deviation of the mean of 0.002 ($\sigma/\sqrt{n}$). This compares to the nominal value of 0.1 suggested by Dogniaux for urban temperate climates.

Atmospheric water vapor

Water vapor content, w, for this area was calculated from mean daily dew-point temperatures, t_d, taken from Ref. [1] and are based on measurements from 7 am to 4 pm. These values

$$\ln w(cm) = 0.981 + 0.0341\ t_d\ (°F). \tag{15}$$

Water vapor can also be calculated from atmospheric pressure and dewpoint using an ASHRAE algorithm [4, Chapter 5]. This approach differed by no more than 3% from the results in Eq. (15). Monthly average values of w from throughout the four-year measurement period are shown in Fig. 4. The distribution $\overline{w}$ = 1.58 cm through 2.2 cm has a somewhat lower average than the range recommended by Dogniaux for a typical temperature climate (2 to 4 cm).

Comparison of calculated and measured T_L

The calculated monthly average water vapor content of the atmosphere, w, and the mean annual average β value derived from our measured data and the γ_s for which $E_{esn} > 200$ W/m^2, were used in Dogniaux's and Valko's equations to calculate the monthly variation in turbidities. Figure 5 shows the ratio of the calculated values to our measured values. Some of the error we found when using design values for w and β (see Fig. 2) is eliminated when using more appropriate values for w and β. A summary measure of the degree of the fit is again expressed by the average ratio and standard deviation of T_L(calculated)/T_L (measured) = 1.12 ± 0.03 (Dogniaux) and 1.13 ± 0.03 (Valko).

Both formulae provide a good rough estimate of measured T_L. We need additional data and analysis to determine if Eqs. 9 and 10 will accurately predict T_L for the San Francisco climate. There are several potential experimental and analytical sources which could reduce accuracy and account for the discrepancies between measured and calculated values. These include: 1) errors in our estimates of w and β; 2) the use of average γ_s rather than average air mass, m; 3) the shadow band correction techniques [11]; and 4) error introduced by use of the 200 W/m^2 criterion for a clear day. Changing this criterion might add more low-altitude points, which would lower the calculated mean turbidities. The effect on measured values is less clear, so it is not certain that it would improve the formulae.

Illuminance turbidity

The concept of turbidity factor has been developed for irradiance measurements. In the daylighting field, Linke's turbidity factor (T_L) is used to calculate irradiances that are then converted into illuminances, using the luminous efficacy of radiation, K. Both T_L and K depend on m, β, and w, which increases the opportunity for error and the complexity of the calculation. The calculation of an illuminance turbidity, T_{il}, provides a way around these problems. The concept is analogous to that of irradiance turbidity. Direct normal illuminances would be calculated by Eq. (4.1) with the irradiance parameters E_{esn}, E_{eo}, α_r, and T_L replaced by their illuminance counterparts, E_{lsn}, E_{lo}, α_{il}, and T_{il}.

Because water vapor absorption occurs predominately in the infrared, T_{il} should be insensitive to w. In addition, because illuminance strongly weights a narrow band of wavelengths, the extinction coefficient should be relatively insensitive to m. We now derive simple approximations for the illuminance-weighted Rayleigh coefficient, α_{il}, and the illuminance turbidity, T_{il}.

Because water vapor absorption is not important for illuminances, we approximate the extinction coefficient, $\overline{a}_{il}$, in terms of $\overline{\alpha}_{il}$ and an aerosol term:

$$\overline{a}_{il} = \alpha_{il} + \beta/(\overline{\lambda})^{\alpha}, \tag{16}$$

where $\overline{\lambda}$ = 0.5527 μm is the mean wavelength weighted by the V(λ) curve [6]. Solving Eqs. (4.1) and (4.2) for T_{il} gives:

$$T_{il} = \overline{a}_{il} / \alpha_{il}. \tag{17}$$

Substitution of Eq. (17) into Eq. (16) gives

$$T_{il} = 1 + \beta / \alpha_{il} \, (\overline{\lambda})^{\alpha}. \tag{18}$$

The recommended value of the exponent α is 1.3. An expression for α_{il} can be derived from Eqs. (1) and (4):

$$\alpha_{il} = \frac{1}{m} \ln \left[\frac{E_{lsn}}{\int_o^{\infty} E_{lo}(\lambda) e^{-\alpha(\lambda) m} \, d\lambda} \right], \tag{19}$$

where $E_{lo}(\lambda) = E_{eo}(\lambda)$ V(λ), and V (λ) is the CIE photoptic sensitivity curve. The integral in Eq. (19) can be estimated from the best fitting Gaussian (the method of steepest descents). This gives a reciprocal expression for α_{il} [see also Eqs. (6) and (13)]:

$$\alpha_{il} = 0.1/(1 + 0.0045 \, m). \tag{20}$$

Equation (20) suggests that α_{il} can be assumed to be 0.1 to about 5% accuracy for $\gamma_s > 5°$, (m < 10). Substituting the values of $\overline{\lambda}$, α, and α_{il} into Eq. (18) yields the following simple expression:

$$T_{il} = 1 + 21.6\beta. \tag{21}$$

Now let us reconsider the problem of computing the direct normal illuminance, E_{lsn}. Dogniaux has listed design values of β and w from which design turbidities and efficacies are computed as a function of air mass. Using the illuminance turbidity concept, we can replace this procedure with three design illuminance extinction coefficients: $\overline{a}_{il}$ = 0.21, 0.32, and 0.53 for rural, urban, and industrial areas, respectively. (Coulson quotes an uncertainty on α of $\pm$ 0.2, which translates into about a 10% uncertainty on these design values). Calculation of E_{lsn} using design values for $\overline{a}_{il}$ follows directly from the illuminance analogue of Eq. (4).

The use of annual average design values of $\overline{a}_{il}$ (or any related parameter such as β) may provide adequate average estimates for illuminance but will normally not be a good indicator of differences due to atmospheric effects on a shorter time scale (e.g., hourly, daily, monthly). For many daylighting calculations, values for irradiance (E_{esn}) may be available so that illuminance can be calculated if K, the luminous efficacy, is known. K, however, is a complex function of solar altitude, γ_s, water vapor content, w, and atmospheric scattering effects, β. In Tables 1 and 2 below we compare hourly/monthly values of beam luminous efficacy obtained in two different ways.

In Table 1, we use our measured annual average value for β (0.088) and monthly average daytime values of water vapor content to calculate T_L based on Eq. (9). T_L is then used in Eq. (4.2) to calculate E_{esn}. For the illuminance, we use β to calculate T_{il} [Eq. (21)], and then T_{il} is used to calculate E_{lsn} using the illuminance analog of Eq. (4.2). The values in Table 1 show K = E_{lsn} / E_{esn} on an average hourly basis for each month. In Table 2, we present values of K obtained by dividing measured beam illuminance by measured beam irradiance. We note in

Table 1 that the use of a single annual value for β (and monthly w) results in values for K which show a strong dependance on solar altitude but otherwise little month to month variation. The rapidly decreasing beam efficacy at low solar altitudes is expected and results from the wavelength dependent scattering properties of the atmosphere.

Comparing the measured data in Table 2 to those of Table 1 illustrates the differences to be expected when comparing average measured hourly/monthly data to annual results. May and October show lower and higher values respectively than we predicted for Table 1. We are trying to determine the source of the atmospheric effects that caused those differences.

We conclude that the traditional approach of using design values of β and w to calculate E_{lsn} from E_{esn} using a luminous efficacy works well as long as design values are accurately known. However, if β can be measured or accurately estimated, then Eq. (21) can be used directly to calculate E_{lsn}, bypassing the requirement to convert irradiance to illuminance.

The fact that T_{il} derived from measured illuminance is essentially independent of m and thus γ_s is an advantage in evaluating atmospheric conditions. As mentioned earlier, a plot of T_L against time of day or solar altitude is difficult to interpret because T_L is dependent on γ_s and atmospheric conditions. A plot of T_{il} against time of day or solar altitude gives more information about the variation in atmospheric conditions.

Figure 6 plots T_{il} derived from measured illuminance as a function of solar altitude. The reader should remember that the values at low altitudes are biased due to the 200 W/m^2 criterion (see Fig. 1).

T_{il} should be closely related to particulate concentrations over the course of a day, whereas T_L will be more dependent on changing airmass. Figure 7 plots T_{il} and T_L as a function of time of day. The results indicate that there are substantial variations in T_L as a function of time of day due in part to its dependence on air mass, which varies with the month of the year. T_{il} is relatively stable during the central portion of the day but we also noticed what may be a local peak in T_{il} caused by afternoon rush-hour traffic or the fog that often covers the Bay in the early evening. This suggests that illuminance turbidity may indeed be more sensitive to atmospheric conditions than is irradiance turbidity.

The hourly average data in Fig. 7 do not show monthly climatic trends. So in Fig. 8 we plot illuminance and irradiance turbidities as a function of time of year. The average value over the year and standard deviation for illuminance turbidity is 2.6 $\pm$ 0.4, and for irradiance turbidity is 3.4 $\pm$ 0.5. The atmospheric conditions that result in a maximum average monthly value for T_L in May cause an even sharper peak in T_{il} for May.

As a check on the illuminance turbidities, we can compare the value of β derived from our illuminance measurements and from illuminance turbidity, Eq. (21), to the value of β we calculate from our irradiance turbidities, which are based on measured irradiance, monthly average w, and either Dogniaux's or Valko's fits [Eqs. (9) and (10)]. The irradiance-derived βs consistently underpredict the illuminance βs with the average annual ratios, r = 0.87 $\pm$ 0.03 (Dogniaux) and r = 0.82 $\pm$ 0.04 (Valko). Figure 9 shows the month by month values and Figure 10 shows monthly ratios that indicate the magnitude of the monthly differences. Further measurement and analysis are required to determine how much of the discrepancy is due to the use of monthly average w, to inaccuracies in Valko's and Dogniaux's formulae, or to measurement error.

CONCLUSIONS

A comparison of calculated T_L (using expressions derived by Dogniaux and Valko) and measured T_L shows good agreement (approximately 12% difference) when the correct local values of β and w are used in the formulae. The use of nominal design values for β and w gives correspondingly rougher agreement (18-28% difference). The Campbell-Stokes clear sky criterion of 200 W/m^2 affects the maximum calculated turbidity at very low solar altitudes. The results at these altitudes are therefore not comparable to results at higher altitudes. Additional criteria to unambiguously define clear sky conditions would be desirable.

We present a simple equation for an illuminance turbidity T_{il} (analogous to T_L), and suggest that the use of T_{il} may simplify some daylighting calculations. This illuminance turbidity also has the advantage that it better indicates atmospheric conditions than does Linke turbidity. The illuminance turbidity has an heuristic advantage in that an approximate analytical expression for it can be derived directly from the Angström formula.

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LUMINOUS EFFICACY OF DIRECT BEAM

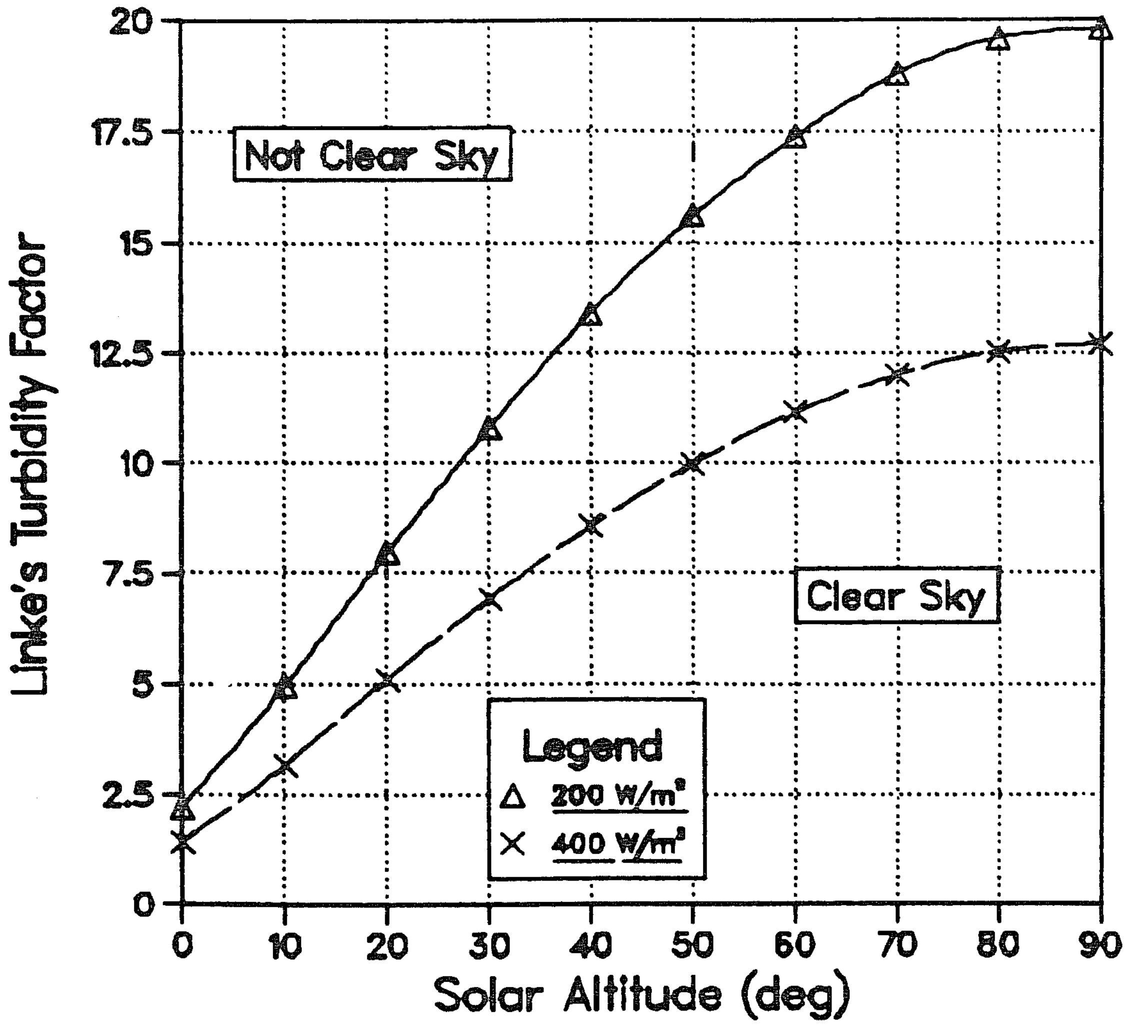
ALT.	JAN	FEB.	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
5	[illegible]	25	25	26	26	26	27	27	27	27	26	25
10	[illegible]	67	68	68	69	69	70	70	70	70	69	68
15	[illegible]	86	86	86	87	88	89	89	89	88	87	86
20	[illegible]	95	95	95	96	96	97	97	97	97	96	95
25	[illegible]	99	99	99	100	101	101	102	102	101	100	99
30	101	102	102	102	103	103	104	104	104	104	103	102
35	101	103	104	104	104	105	105	106	106	105	104	104
40		105	105	105	105	106	107	107	107	106	106	
45		106	106	106	106	107	107	107	108	107		
50			106	107	107	107	108	108	108	108		
55			107	107	108	108	109	109	109			
60			108	108	108	109	109	109	110			
65				108	109	109	110	110	110			
70				109	110	110	111	111				
75				110	110	111	111	111				

1 - Luminous efficacy of direct normal radiation as a function of solar altitude using ...age $\beta = 0.088$.

LUMINOUS EFFICACY OF DIRECT BEAM

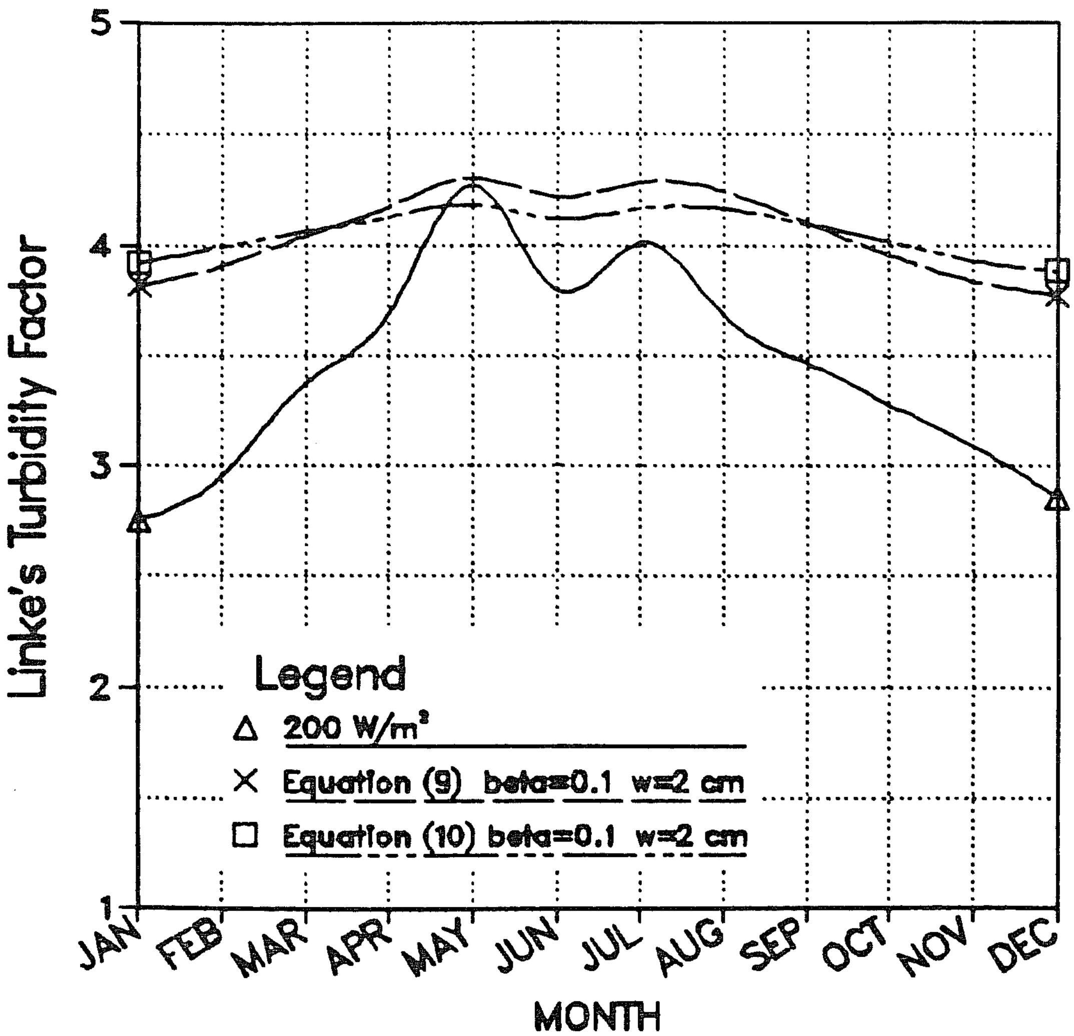
Alt.	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
5	37	30	36	28	28	40	21	34	39	37	34	32
10	65	59	52	53	57	61	58	70	67	65	64	60
15	87	80	78	80	82	86	80	90	89	89	87	88
20	99	94	90	90	91	97	93	105	100	103	100	101
25	104	102	95	98	95	104	97	109	106	109	107	108
30	107	105	101	102	99	110	107	108	110	114	112	111
35	110	106	103	103	99	110	108	112	111	116	112	114
40		105	102	102	97	110	106	110	109	114	112	
45		101	102	102	98	106	105	109	108	112		
50			102	102	98	108	105	108	109	113		
55			103	102	97	106	105	110	108			
60			106	104	98	106	106	109	104			
65				101	98	106	106	111	106			
70				102	98	106	107	110				
75				0	102	106	105	104				

Table 2 - Luminous efficacy of direct normal radiation as a function of solar altitude based on measured E_{esn} and E_{lsn}.



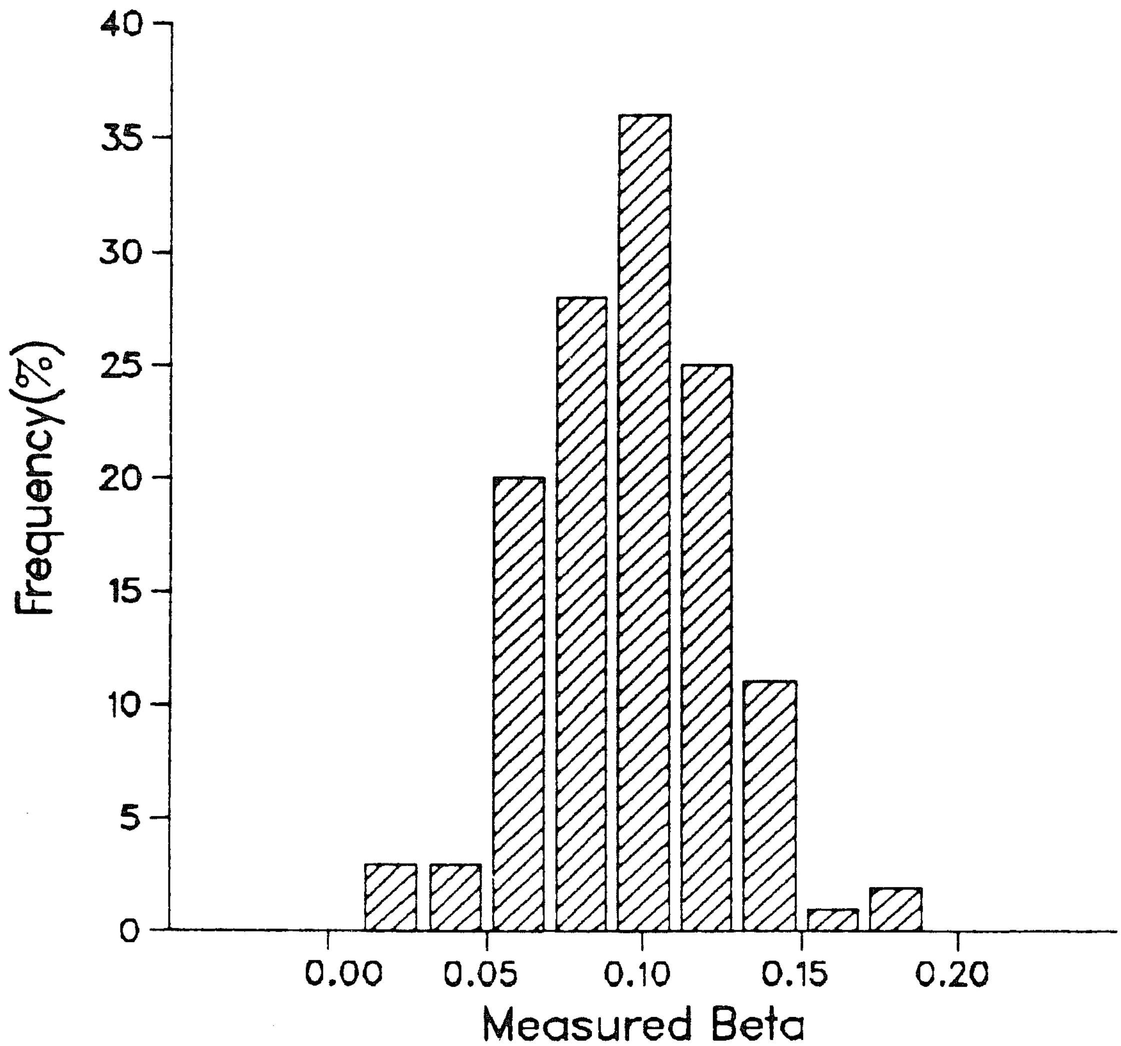
XBL 846-2330

Figure 1 - Maximum turbidity as a function of solar altitude.



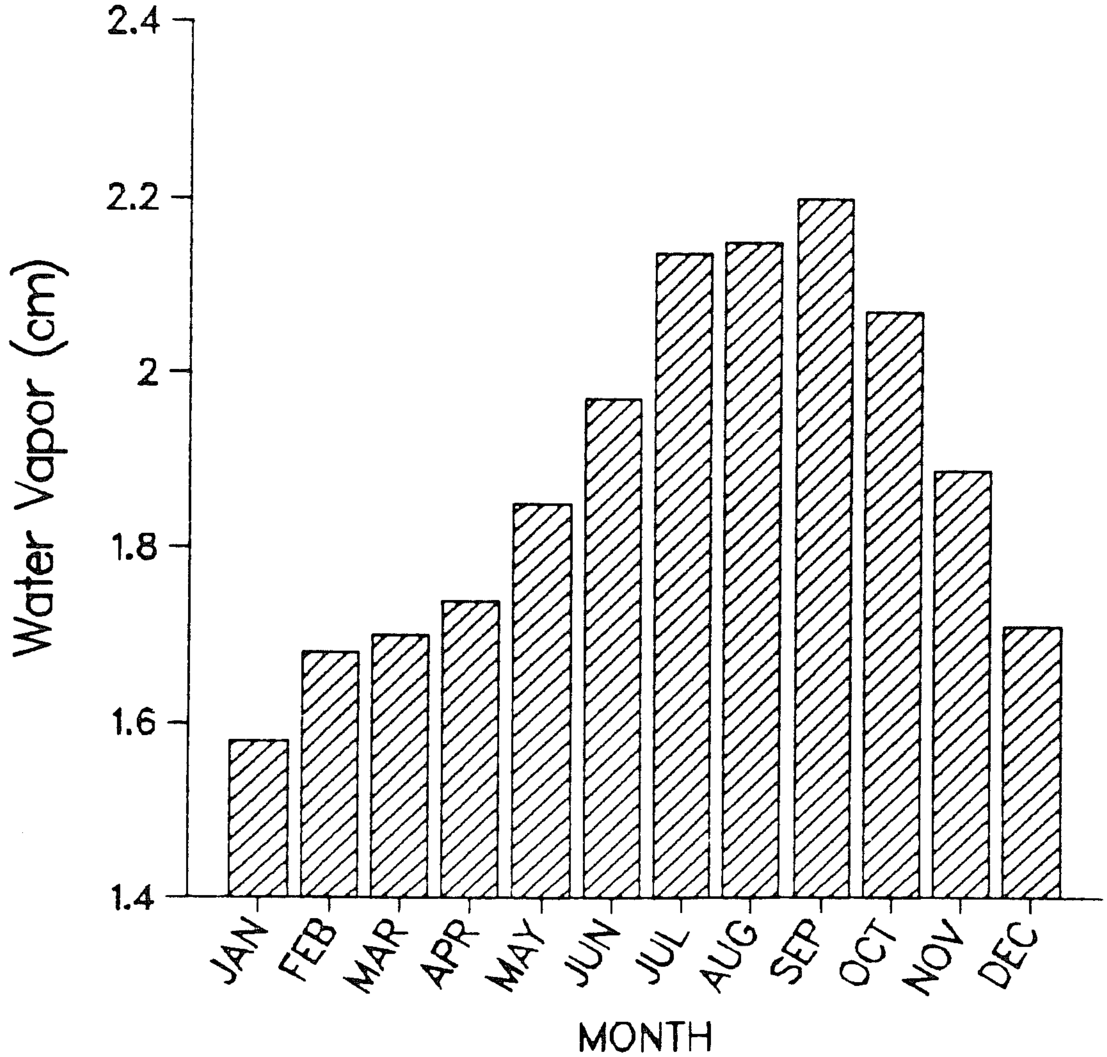
XBL 846-2331

Figure 2 - Monthly average values of Linke turbidity as measured vs. predictions from Dogniaux's [7] and Valko's [8] equations, using $\beta = 0.1$, $w = 2.0$.



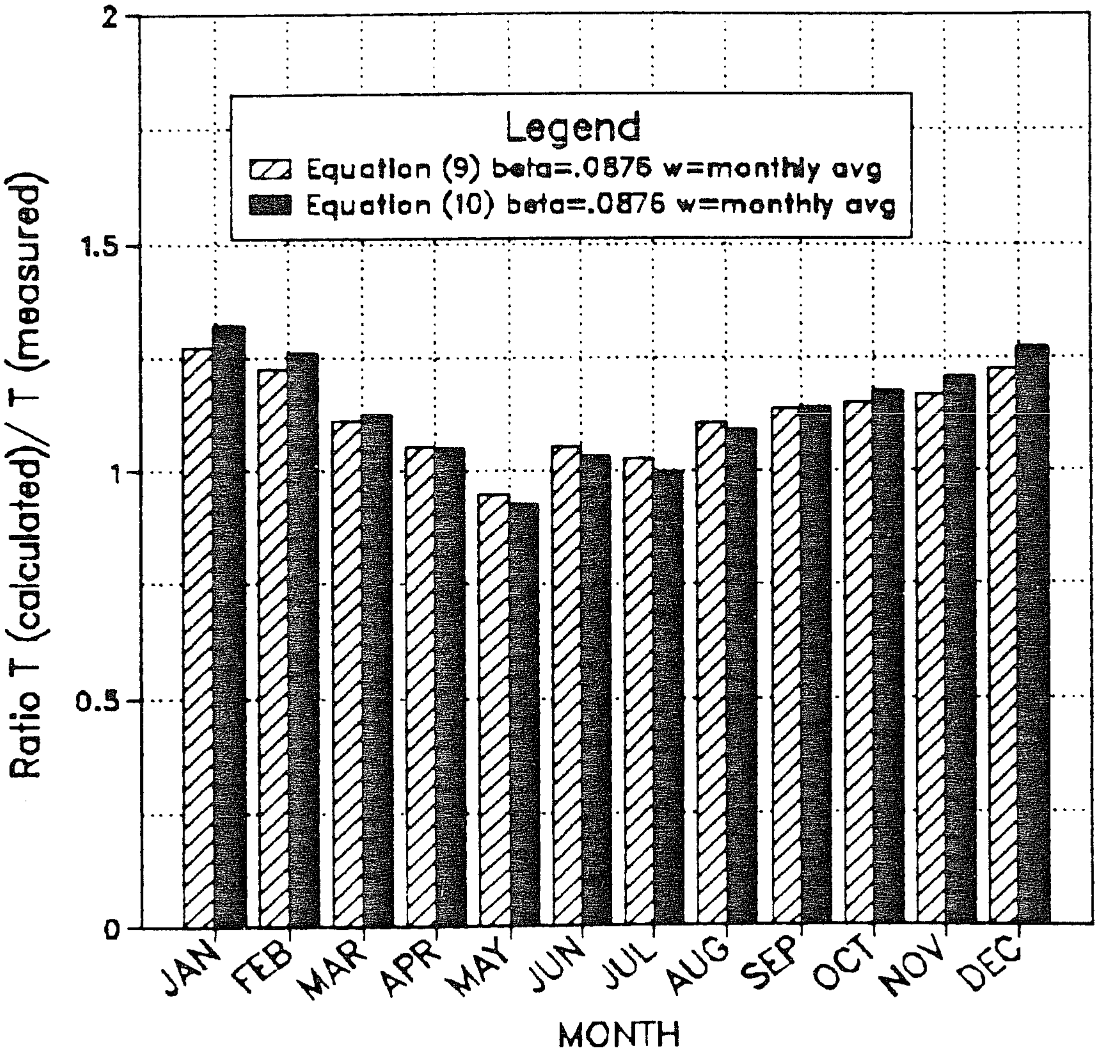
XBL 845-2050

Figure 3 - Frequency distribution of Angström turbidity coefficient (β).



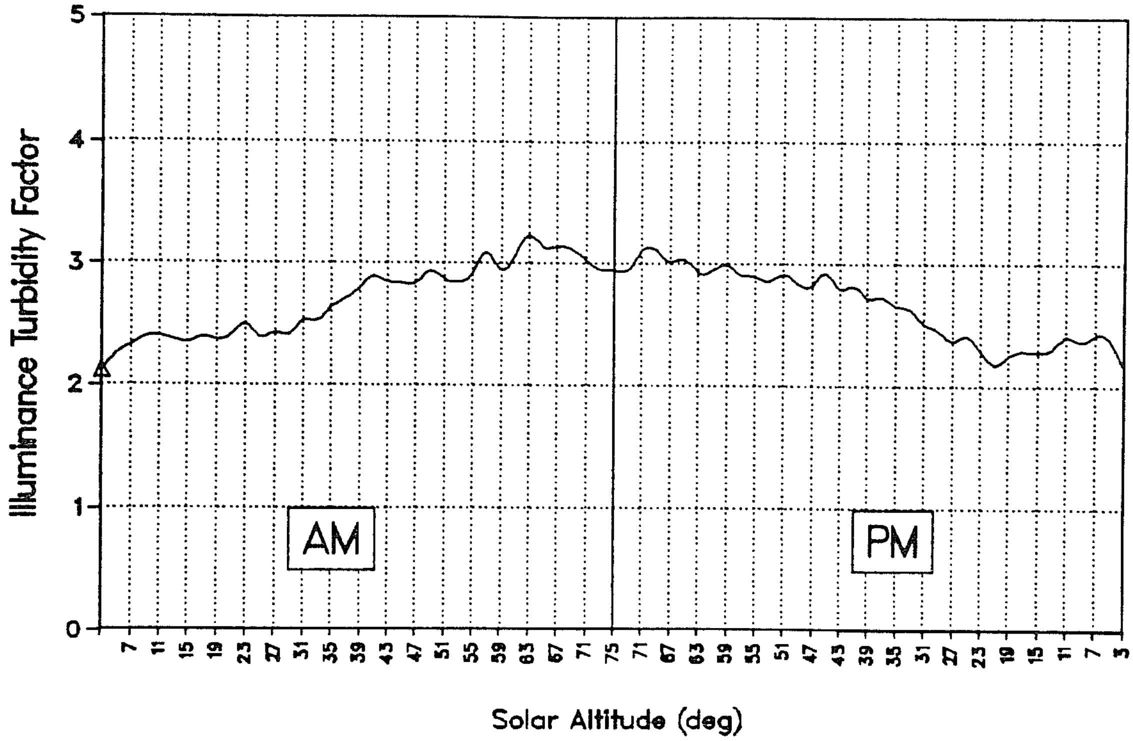
XBL 845-2051

Figure 4 - Monthly average values of water vapor content of the atmosphere for San Francisco.



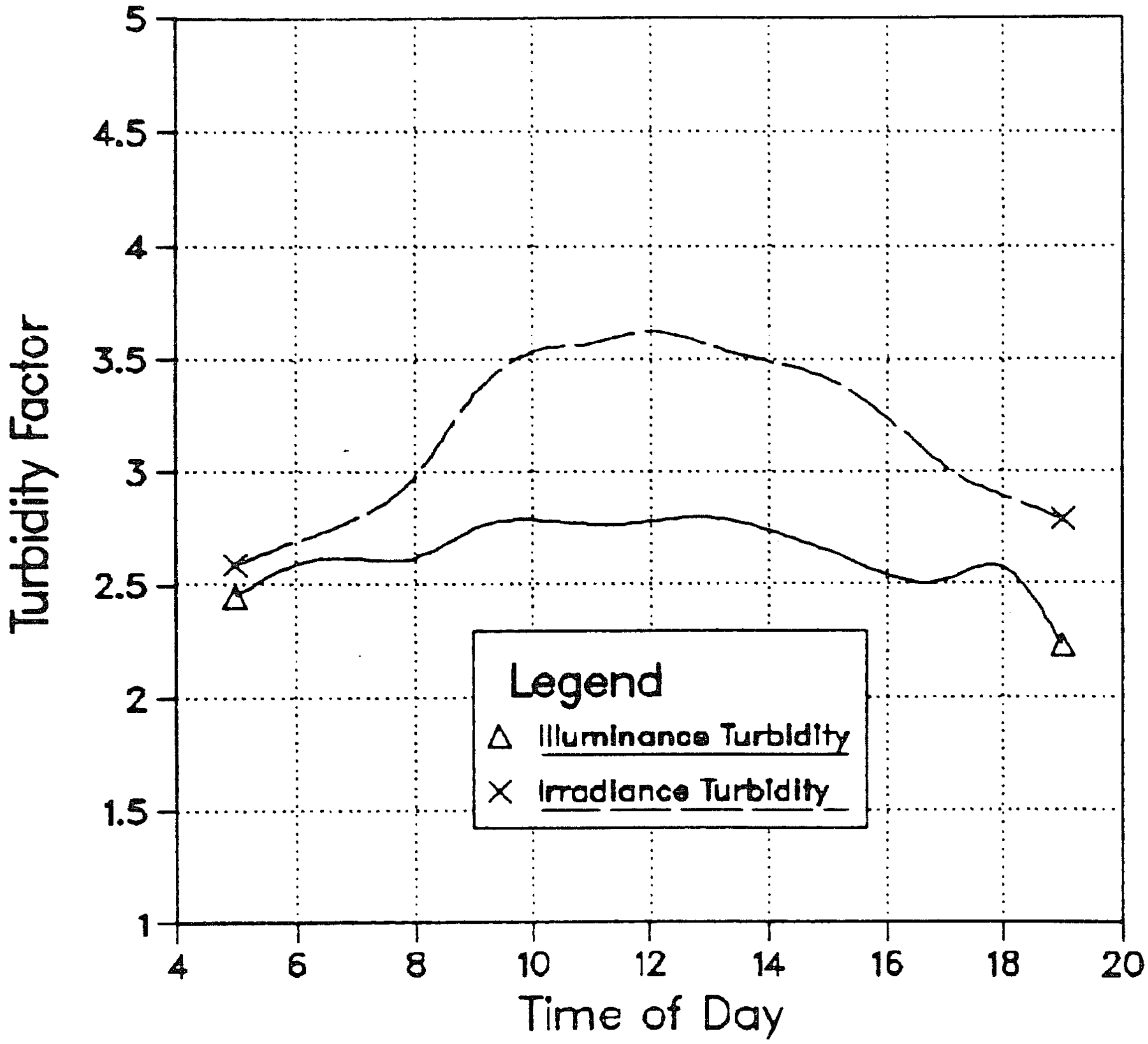
XBL 845-2052

Figure 5 - Monthly ratio of calculated Linke's turbidity using Dogniaux's and Valko's equations to measured turbidity; $\beta = 0.09$, w from Fig. 4.



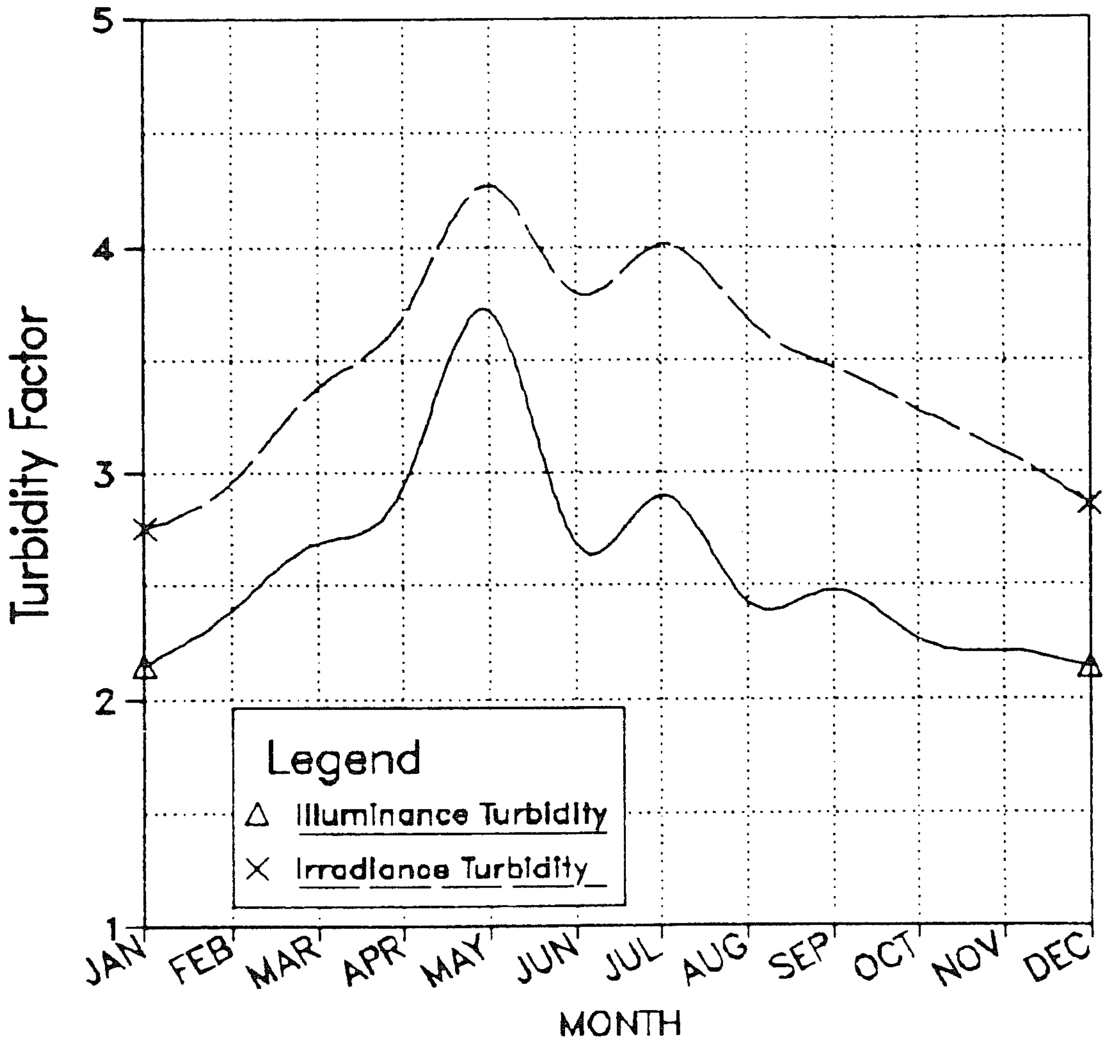
XBL 845-2053

Figure 6 - Illuminance turbidity, T_{il}, as function of solar altitude for hourly data for all data meeting the "clear day" criterion.



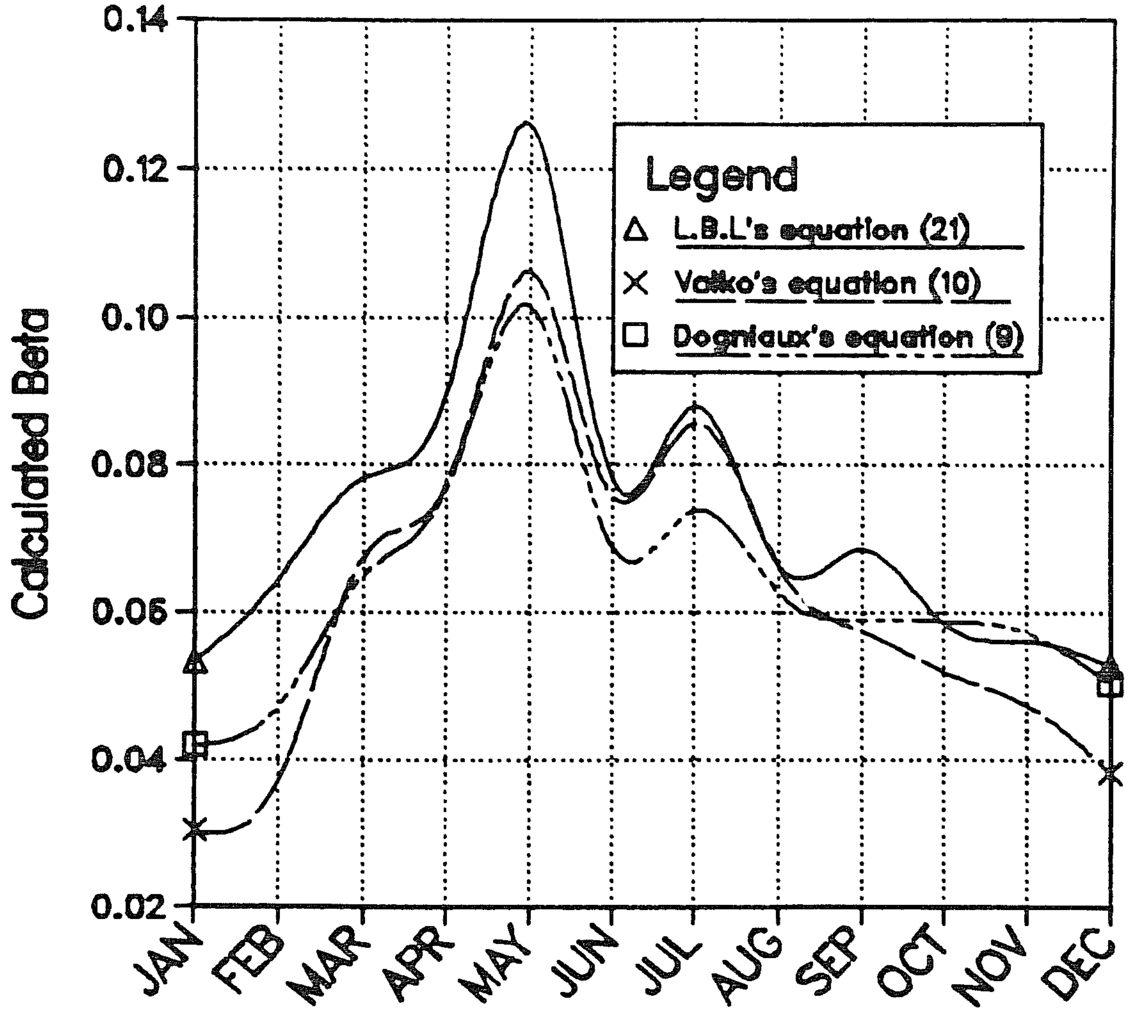
XBL 845-2054

Figure 7 - Average calculated illuminance and irradiance turbidity as a function of time of day, for hourly data for all data meeting the "clear day" criterion.



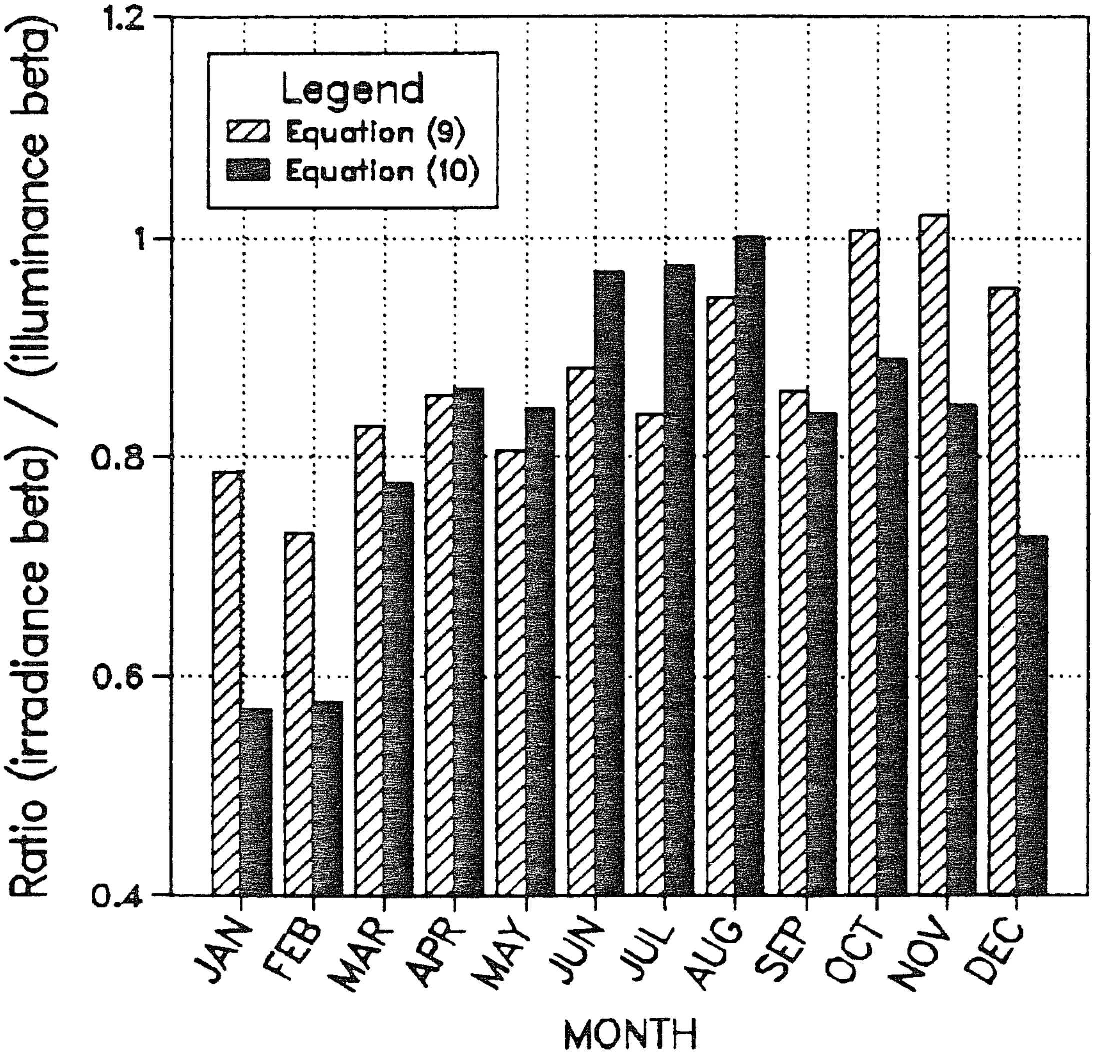
XBL 845-2055

Figure 8 - Monthly average value of illuminance T_{il} and irradiance T_L turbidities.



XBL 846-2332

Figure 9 - Monthly average of the irradiance-derived β using Eqs. (21), (10), and (9).



XBL 845-2056

Figure 10 - Ratio of irradiance-derived β to illuminance-derived β using Eqs. (21), (10), and (9).