NOTES AND CORRESPONDENCE


Linke and Unsworth–Monteith turbidity parameters in Athens

By H. D. KAMBEZIDIS*, D. H. FOUNDAL and N. S. PAPANIKOLAOU
Institute of Meteorology and Physics of the Atmospheric Environment, National Observatory of Athens, P.O. Box 20048, GR-118 10 Athens, Greece

(Received 23 July 1991; revised 21 August 1992)

SUMMARY

The Linke, \( T_L \), and Unsworth–Monteith, \( T_U \), turbidity parameters were chosen to be studied in Athens. They were derived from measurements of direct solar radiation and exhibit similar behaviour in the decade 1975–84. The increase in air pollution in Athens does not reveal itself in the trends of the turbidity coefficients during the decade. An interdependence between \( T_L \) and \( T_U \) was found and is expressed as a mathematical function. The seasonal, monthly and daily variation of the parameters are investigated.

1. INTRODUCTION

The solar radiation is attenuated by the earth’s atmosphere owing to absorption and scattering by molecules. This attenuation is not constant even from one hour to the next. The concentrations of oxygen (\( O_2 \)) and carbon dioxide (\( CO_2 \)) in the atmosphere remain more or less constant in time and space, but the total vertical amount of ozone (\( O_3 \)) varies with latitude and season. Average values of \( O_3 \) are found in the literature (Robinson 1966) or can be calculated (Van Heuklon 1979). [For Greece, Bais et al. (1988) have shown that total \( O_3 \) shows a slight increase during the period 1982–88 (0.02% per year). A \( \pm 10\% \) variation in the total vertical amount of \( O_3 \) gave a corresponding variation of \( \pm 2\% \) in the turbidity values.] The total vertical amount of water vapour (\( H_2O \)) is even more variable. A sensitivity analysis showed that a \( \pm 25\% \) variation in the amount of precipitable water gave a corresponding variation in the turbidity values in the range –5 to +5%.

The amount of aerosol in the atmosphere is represented by the turbidity parameters. The two studied in this work are the following.

1. The Linke turbidity factor, \( T_L \). This parameter was introduced by Linke (1922, 1929) and shows how many dry atmospheres would be necessary to produce the attenuation of the extraterrestrial radiation that is produced by the real atmosphere. Typically \( T_L \) varies from 1 to 10. An expression for \( T_L \) is given by:

\[
T_L = P(m) (\log I_{\text{at}} - \log I_d - \log S)
\]

where \( P(m) \) is defined in appendix A and its expression as a function of \( m \) is given in appendix B, Eq. (B.4).

2. The Unsworth–Monteith turbidity coefficient, \( T_U \). This parameter was introduced by Unsworth and Monteith (1972) and expresses the absorption of the solar rays by a dust-laden atmosphere relative to a dust-free one with a specified water vapour content. Typically \( T_U \) varies from 0 to 1. An expression for \( T_U \) is given by:

\[
T_U = -(\ln I_d + \ln S - \ln I_{d'})/m.
\]

The major physical difference between what is measured by \( T_L \) and \( T_U \) is that \( T_L \) is focused on the attenuation of the solar rays produced by a water-vapour-free atmosphere, while \( T_U \) is referred to an atmosphere with a specified water vapour content.

The values of \( T_L \) have been divided by ten throughout this work so that they fit on the same ordinate scale as \( T_U \). An explanation of all the symbols used in this work is given in appendix A.

2. CALCULATION OF TURBIDITY PARAMETERS

This section refers to the methods used in this study for the calculation of \( T_L \) and \( T_U \). All the calculations concern turbid atmospheres over Athens for 1974–85 which on certain occasions may contain air pollutants in addition to the natural atmospheric aerosols. This period was chosen for two reasons: (i) it is a complete decade, and (ii) for technical reasons afternoon pyrheliometric measurements ceased at the National Observatory of Athens (NOA) after 8 April 1985; they

* Corresponding author.
started again in October 1990. The NOA total-spectrum direct (0.3–3 μm) solar-radiation data bank was used. The NOA is situated on a hill of 107 m height above mean sea level (a.m.s.l.) near Athens city centre; and because of its location all turbidity-parameter values calculated at the NOA site are likely to be representative of the Athens mean turbidity.

The Linke and Unsworth–Monteith turbidity parameters were chosen to be studied together in this work for two reasons: (i) both $T_L$ and $T_U$ are calculated from measurements of total-spectrum direct solar radiation, and (ii) $T_U$ has never been investigated in Greece before. [Studies of the Ångström, Linke and Schüepp turbidity coefficients before 1975 have been carried out for Athens (Karalis 1976; Katsoulis 1977).]

Calculation of $T_L$. This is calculated through Eq. (1). The only parameter required in the equation is the direct total solar irradiance which is measured at NOA. Appendix B gives a list of all intermediate formulae used to derive $T_L$.

Calculation of $T_U$. This is of similar nature to $T_L$. Appendix C gives all the intermediate equations used to derive $T_U$.

The direct (0.3–3 μm) and spectral (0.525–2.8 μm, 0.630–2.8 μm and 0.710–2.7 μm) solar irradiance measurements at NOA under a cloudless sky are taken manually by observers with a Kipp and Zonen CM5 actinometer at 0820, 1120, 1420 and 1720 LST (LST = GMT + 2 hours). Since it is up to the observer’s naked eye to judge whether the sky is cloudless in order to perform the measurements, sky conditions with sub-visual cirrus are easily misjudged as cloudless ones. Correction of the direct solar irradiance measurements due to the actinometer body temperature is taken into account, and compatibility of the results to the World Radiometric Reference is also assured, i.e. a solar constant equal to 1367 W m$^{-2}$ is used.

3. RESULTS AND DISCUSSION

The variation of the average annual values of $T_L$ and $T_U$ for the period of study is shown in Fig. 1. Two peaks are detected, one in 1979 and another in 1983. The second peak is in agreement with the higher $T_U$ values recorded in the United Kingdom during that year, as reported by Rawlins and Armstrong (1985), while the first peak may be caused by the La Soufrière and Sierra Negra eruptions. Aerosol in the atmosphere comes from many sources, for example emission of pollutants,
and wind blown dust. However, at stratospheric heights, a major perturbation in the aerosol optical depth occurred after the El Chichón eruptions in Mexico during April 1982 (Pollack et al. 1983) causing an increase in turbidity levels detectable world-wide. Large amounts of volcanic debris entered the stratosphere and gradually formed a veil of dust over the northern hemisphere, mostly during the subsequent months. Dutton and DeLuisi (1983) found from aircraft measurements that a stratospheric aerosol optical depth exceeding 0.1 at 0.5 μm at 50°N during December 1982, compared with the background value of about 0.005 at 0.55 μm (Toon and Pollack 1976).

In Fig. 2 the seasonal variations of $T_L$ and $T_U$ are shown. The straight lines represent regression analysis fits with corresponding equations:

$$T_L = 3.6 \times 10^{-4} NS + 0.444$$

$$T_U = -6.1 \times 10^{-4} NS + 0.218$$

where $NS = 1, \ldots, 39$, 1 being spring 1975. [The positive and negative slopes are probably not significant in the above equations since they are so close to zero.] Minimum turbidity levels occur in winter and maximum ones in summer. This is so because the high synoptic winds during the wintertime play a depollution role in the Athens basin (Kambezidis et al. 1986), while in the summertime longer periods with light winds and local sea-breeze circulations favour higher levels of pollution aerosols (Lalas et al. 1983; Kambezidis and Papanikolaou 1988). Generally throughout the year Athens $T_U$ levels are higher than those reported by Rawlins and Armstrong (1985) for Lerwick, Eskdalemuir and Bracknell in the United Kingdom.

Figure 3 shows the variation in mean monthly values for the period under study. Three curves are shown for each turbidity parameter: one for 1975, a second for 1984 and a third for the whole decade; these were chosen so that the turbidity levels at the beginning, the end and for the whole period can be compared. It is noteworthy that the values for the decade lie almost exactly between those for the first and last year. Both $T_L$ and $T_U$ show maximum values in the summer and minimum ones in the winter, as already shown in Fig. 2. This is not surprising for the reasons given above; also the subsidence inversion is stronger in the summer and tends to trap the pollution in the boundary layer, leading to the worst problems in this season. Similar conclusions for $T_L$ are reported by Katsoulis (1977) for earlier periods in Athens, and Rawlins and Armstrong (1985) for the three above mentioned locations in the United Kingdom.

![Figure 2](image)

Figure 2. Average seasonal values of $T_L$ and $T_U$ at NOA for 1975–84. $T_L$ values have been divided by 10.
Figure 3. Averages of monthly average values of $T_L$ and $T_U$ at NOAA for 1975–84. The values in the years 1975 and 1984 are shown separately. $T_L$ values have been divided by 10.

Figure 4 shows the mean variation of $T_L$ and $T_U$ throughout the day for the summers of 1975–84. Neither parameter shows any important diurnal variation, as would be expected on days with an almost static synoptic situation (Unsworth and Monteith 1972) typical of Greek summers and autumns. A relative maximum is detected around noon. In contrast, the variability in weather systems during winters and springs seems to average out the $T_L$ and $T_U$ daily variations. In these cases a relative minimum occurs at 1420 LST.

Figure 5 shows the variation of $T_U$ with $T_L$ for all optical air masses encountered and for the 120 months of this study. The straight line is a regression line with equation:

$$T_U = 0.07 T_L - 0.11$$

or

$$T_L = 14.37 T_U + 1.62.$$  \(5a\) \(5b\)

The above equation has an $R^2$ equal to 0.882. Unsworth and Monteith (1972) gave a diagram (Fig. 1 in their paper) with $T_L$ against $T_U$ for $m = 1, 3, 5$, from which the following equations can be derived:

$$T_L = 11.50 T_U + 2.85 \quad \text{for } m = 1$$

$$T_L = 15.00 T_U + 2.00 \quad \text{for } m = 3$$

$$T_L = 12.00 T_U + 2.00 \quad \text{for } m = 5.$$  \(5c\) \(5d\) \(5e\)

Comparing Eq. (5b) with Eqs. (5c)–(5e) it is seen that their coefficients and constants are comparable for $m = 1, 3, 5$.

Another interesting feature is the frequency distribution of mean monthly values of $T_L$ and $T_U$ for 1975–1984. Table 1 shows that $T_L$ has an almost constant distribution in the range 2.6–6.5 and $T_U$ has the highest frequency in the range 0.08–0.32.

From an air-mass classification analysis it was found that $T_L$ and $T_U$ exhibit a diurnal cycle with maximum values around noon in all cases except for a continental tropical air mass in the cold period of the year (November to March) where a minimum occurs at 1420 LST. $T_L$ and $T_U$ values are in all cases higher in calm weather.
Figure 4. Average summer daily values of $T_L$ and $T_U$ at NOA for 1975–84. $T_L$ values have been divided by 10.

| TABLE 1. FREQUENCY DISTRIBUTION OF TURBIDITY PARAMETERS $T_L$ AND $T_U$ |
|---------------------------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| Frequency of occurrence (%)    | 1.0–1.3         | 1.3–2.6         | 2.6–3.9         | 3.9–5.2         | 5.2–6.5         |
| $T_L$ range                     | 0.8             | 0.8             | 30.0            | 37.6            | 30.8            |
| $T_U$ range                     | 0–0.08          | 0.08–0.16       | 0.16–0.24       | 0.24–0.32       | 0.32–0.40       |
| Frequency of occurrence (%)    | 2.5             | 25.8            | 35.8            | 30.8            | 5.1             |

4. CONCLUSIONS

There is a growing worldwide interest in atmospheric turbidity as this is related to air pollution studies in cities such as Athens and because of its wider significance in tropospheric chemistry and climate studies. In this work, motivated by the above considerations, the Linke and Unsworth–Monteith turbidity parameters were studied.

On investigating the average annual variation of the above parameters for 1975–84 two peaks were found; one in 1979, probably caused by the La Soufriere and Sierra Negra eruptions, and another in 1983, possibly as a result of the El Chichon eruptions in 1982 (Pollack et al. 1983). It was also shown that minimum turbidity levels occur in the winter and maximum ones in the summer as found by other workers locally (e.g. Karalis (1976)) and internationally (e.g. Rawlins and Armstrong (1985)). This is attributed to the high synoptic winds and more frequent rains during wintertime, which have a cleansing effect on the atmosphere over Athens, and the light winds and longer dry periods with higher concentrations of dust-airborne particles during summertime. The mean summer and autumn daily variation of the above turbidity parameters exhibited a smooth course probably due to the invariability in weather systems (Unsworth and Monteith 1972) that characterize these seasons in Greece. Nevertheless, the mean daily variation of the parameters in the rest of the year was found to be quite similar to these of summer and autumn; this implies that variable weather can average out turbidity-parameter values, leading to little mean daily variation. The investigation of the relation between the turbidity parameters and the air pollution problems in Athens was left as a future work; $T_L$ and $T_U$ must be recalculated using solar radiation measurements in the wavelength bands mentioned in section 2.
Figure 5. Variation of monthly mean values of $T_U$ with $T_L$. $T_L$ values have been divided by 10.

APPENDIX A

List of symbols

$e_M$ = partial water vapour pressure (cm Hg)
$e_s$ = saturated water vapour pressure (cm Hg)
$H$ = height a.m.s.l. (m)
$I_g$ = normal incidence total solar radiation at the earth’s surface (W m$^{-2}$)
$I_d$ = normal incidence total solar radiation at the bottom of a dust-free atmosphere (W m$^{-2}$)
$I_{sx}$ = total solar radiation constant = 1367 W m$^{-2}$
$k_R$ = mean optical thickness due to Rayleigh scattering and gas absorption
$m$ = optical air mass
$m'$ = relative optical air mass
$NS$ = increasing number of season in the decade 1975–84
$P$ = atmospheric pressure (mbar)
$P_0$ = reference atmospheric pressure = 1000 mbar
$P(m)$ = a function of optical air mass
$R_2$ = square of correlation coefficient
$R_a$ = actual sun–earth distance (km)
$R_m$ = mean sun–earth distance (km)
$RH$ = relative humidity (%) = $100 \frac{e_M}{e_s}$
$S$ = correction factor for the sun–earth distance
$T_L$ = Linke turbidity factor
$T_U$ = Unsworth–Monteith turbidity coefficient
$U_O$ = total $O_3$ in a vertical column from surface (cm at s.t.p.)
$U_W$ = total water vapour content in the atmosphere (cm)
$XN$ = Julian day
$z$ = apparent solar zenith angle (deg)
$\tau_O$ = absorption transmittance of O$_3$
$\tau_R$ = absorption transmittance of Rayleigh scattering
$\tau_{UM}$ = absorption transmittance of CO$_2$ and O$_2$
$\tau_W$ = absorption transmittance of water vapour
NOTES AND CORRESPONDENCE 373

APPENDIX B

Calculation of $T_L$

The relative optical air mass, $m'$, is given by Kasten (1966):

$$m' = \{\cos z + 0.15 (93.885 - z)^{-1.23}\}^{-1}$$  \(\text{(B.1)}\)

and the (absolute) optical air mass, $m$, by:

$$m = m' (P/P_0).$$  \(\text{(B.2)}\)

The acceptable range for $m$ in our calculations was $m \leq 6$.

The correction factor, $S$, for the sun–earth distance is given by Duffie and Beckman (1980):

$$S = R_e^2/R_m^2 = 1 + 0.033 \cos(2\pi XN/365).$$  \(\text{(B.3)}\)

$P(m)$ referred to in Eq. (1) is defined as follows:

$$P(m) = (\bar{k}_R m \log e)^{-1}.$$  \(\text{(B.4)}\)

According to Louche et al. (1986):

$$\bar{k}_R = (6.5567 + 1.7513 m - 0.1202 m^2 + 0.0065 m^3 - 0.00003 m^4)^{-1}.$$  \(\text{(B.5)}\)

From our analysis no $T_L$ values greater than 10 were found.

APPENDIX C

Calculation of $T_U$

The Unsworth–Monteith turbidity coefficient is calculated via Eq. (2), in which $I_d$, $S$ and $m$ are estimated as in appendix B. $I_d$ is the normal incidence irradiance at the earth’s surface through a dust-free atmosphere with a specified water vapour content. According to Bird and Hulstrom (1981) the direct solar beam reaching the surface of the earth and passing through such an atmosphere can be estimated through the relationship:

$$I_d = 0.9751 I_{ex} \tau_R \tau_O \tau_{UM} \tau_W$$  \(\text{(C.1)}\)

where

$$\tau_R = \exp\{-(0.0003 m^{0.84}) (1 + m - m^{1.01})\}$$  \(\text{(C.2)}\)

$$\tau_O = 1 - (0.1611 x_o (1 + 139.48 x_o)^{-0.3035} - 0.002715 x_o + 0.0003 x_o^2)^{-2}$$  \(\text{(C.3)}\)

where

$$x_o = U_O m$$  \(\text{(C.4)}\)

and

$$U_O = 327.0 + 22.4 \sin(0.9865 XN - 29.595)$$  \(\text{(C.5)}\)

according to Van Heuklon’s (1979) Eq. (4) with application to Athens,

$$\tau_{UM} = \exp(-0.0127 m^{0.26})$$  \(\text{(C.6)}\)

$$\tau_W = 1 - 2.4959 U_W m \{(1 + 79.034 U_W m)^{0.8828} + 6.385 U_W m\}^{-1}$$  \(\text{(C.7)}\)

$$U_W = 2.3 e_M 10^{-n/22000} \quad \text{(according to Gates 1962)}$$  \(\text{(C.8)}\)

$$e_M = e_S \text{ RH}/100.$$  \(\text{(C.9)}\)

In the analysis of these data no $T_U$ values greater than 1 were found.

REFERENCES

