A Study of Linke Turbidity Factor over Qena / Egypt

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Received December 4, 1995

ABSTRACT

Data on instantaneous atmospheric Linke turbidity factor $T_L(m)$ are reported for clear days at Qena / Egypt in the period from June 1992 to May 1993. $T_L(m)$ is determined using the values of irradiance of direct solar radiation (I), which are calculated from global (G) and diffuse (D) – solar radiation measurements. Monthly and seasonally variations of both diurnal and daily average values of $T_L(m)$ increases steadily in the direction of sunset in the months from June to December 1992 as well as Summer and Autumn seasons, while it falls generally in this direction for the months from January to March and Winter season. In April and May, $T_L(m)$ fluctuates obviously through the day hours. It is also shown that the average values of $T_L(m)$ are particularly large during Summer months compared to other months of the year. This behavior of $T_L(m)$ is discussed in view of the variations of some weather elements, which affect the content of water vapor and dust particle in the atmosphere of the study region. It seems *t* be of similar trends to that of other locations inside and outside Egypt. The virtual variation of $T_L(m)$ is eliminated by reducing its value to relative optical air mass $m = 2$, according to Kästen formula. The resulting $T_L(2)$ is more representative for the content of dust particles and water vapor in the atmosphere.

Key words: Turbidity, Linke turbidity factor, Seasonal and diurnal variations, Solar radiations components measurements, Aerosol

I. INTRODUCTION

The intensity of direct solar radiation reaching the earth's surfaces at clear days is attenuated by the various constituents of the atmosphere, namely gases, liquid and solid particles. Attenuation of solar radiation through a real atmosphere versus a clean dry atmosphere gives an indication to the atmospheric turbidity, which its study is important in meteorology, climatology and for monitoring of atmospheric pollution (Louchi et al., 1986). It is also important in biological studies as well as studies of biophysical of inadvertent modifications in the atmospheric composition (Abdelrahman et al., 1988).

There are different coefficients, which are used to measure the atmospheric turbidity, such as Linke turbidity factor, Angstrom and Shueppe turbidity coefficients. The suitable choice of one of these coefficients depends on the kind of application under the question. For characterizing the total extinction by the atmosphere (the main point of this work), Linke turbidity factor $T_L(m)$ is an appropriate measure (Kasten, 1980). It refers to the whole spectrum, that is overall spectrally integrated attenuation, which includes presence of water vapor and aerosols and is also considered as a first simple measure of its contents in the atmosphere.

The aim of this paper is to study the characteristics of atmospheric turbidity $T_L(m)$ measured perheliometrically at clear days in the period from June 1992 to May 1993 at Qena / Upper Egypt. This study includes:

(i)–Monthly and seasonally variations of both diurnal and daily averages of $T_L(m)$ in view of the corresponding changes of some weather elements, affecting the content of water vapor and dust particles in the atmosphere, such as temperature (T), relative humidity (RH), wind speed (WS) and vapor pressure (VP).

(ii)– Relation between $T_L(m)$ and diffuse fraction (D / G).

(iii)–A try to eliminate the virtual variations of $T_L(m)$ by reducing it to relative optical air mass $m=2$ using Kasten formula (Kasten, 1988), which derived for this purpose. The resulting reduced turbidity factor $T_L(2)$ may be more convenient than $T_L(m)$ in representing the contents of water vapor and dust particles in the atmosphere.

II. METHODOLOGY

1. *Theoretical Background and Computation Techniques*

The extinction of direct normal solar radiation measured over the whole solar spectrum is described by (Kasten, 1980):

$$I = I_0 e^{-[m(\bar{\delta}_R + \bar{\delta}_D + \bar{\delta}_W)]}, \tag{1}$$

where I is the irradiance of direct normal solar radiation measured by the perheliometer. It is given by:

$$I = \int_{\lambda=0}^{\infty} I_\lambda \cdot d\lambda \tag{2}$$

in which I_λ is its spectral irradiance ($\mathrm{wm^{-2}\mu m^{-1}}$) at earth's surfaces.

I_0 is the irradiation of the extraterrestrial solar radiation, it is given by:

$$I_0 = \int_{\lambda=0}^{\infty} I_{0\lambda} \cdot d\lambda, \tag{3}$$

in which $I_{0\lambda}$ is its spectral irradiance.

$\bar{\delta}_R, \bar{\delta}_D, \bar{\delta}_W$ are the vertical optically thicknesses of the atmosphere, with respect to Rayleigh scattering, dust extinction and water vapor absorption, weighted by the extraterrestrial solar spectrum as follow:

$$e^{-m[\bar{\delta}_R + \bar{\delta}_D + \bar{\delta}_W]} = \frac{\int_{\lambda=0}^{\infty} I_{0\lambda} e^{\{-m[\delta_R(\lambda) + \delta_D(\lambda) + \delta_W(\lambda)]\}} \cdot d\lambda}{\int_{\lambda=0}^{\infty} I_{0\lambda} \cdot d\lambda}, \tag{4}$$

in which $\delta_R(\lambda), \delta_D(\lambda)$ and $\delta_W(\lambda)$ are the spectral vertical thickness of the atmosphere, which are given by the integrals of the corresponding extinction coefficient (in $\mathrm{m^{-1}}$) $\sigma_R, \sigma_D, \sigma_W$ at vertical height z above the earth's surface as:

$$\delta_{r,D,W}(\lambda) = \int_{z=0}^{\infty} \sigma_{R,D,W}(\lambda, z) dz. \tag{5}$$

$T_L(m)$ is defined as:

$$T_L(m) = \frac{\bar{\delta}}{\bar{\delta}_R}, \tag{6}$$

where $\bar{\delta}$ is the total vertical optical thickness of the atmosphere. It is given by:

$$\bar{\delta} = \bar{\delta}_R + \bar{\delta}_D + \bar{\delta}_W. \quad (7)$$

From Eqs. 6 and 7

$$T_L(m) = 1 + \frac{\bar{\delta}_D + \bar{\delta}_W}{\bar{\delta}_R} \quad (8)$$

substituting by Eq. 8 in Eq. 1 we have

$$I = I_0 e^{-[m \cdot \bar{\delta} \cdot T_L]}, \quad (9)$$

from which

$$T_L(m) = (1 / m\bar{\delta}_R) * \ln(I_0 / I). \quad (10)$$

It is important here to say that the effect of ozone in the above calculation is left out and compensated by including an average ozone absorption optical thickness in the Rayleigh scattering optical thickness $\bar{\delta}_R(\lambda)$.

Equation 10 is used to determine $T_L(m)$ with the aid of experimental values of I. The extraterrestrial solar irradiance I_0 is calculated by:

$$I_0 = I_{sc}[1 + 0.033\cos(360n / 365)], \quad (11)$$

in which n is the day number of the year and I_{sc} is the solar constant (1376 wm^{-2}). In this paper, the relative optical air mass m is taken as $m = 1 / \sin h$, which closely holds for solar elevation angle (h) of value $h > 10°$(Kasten, 1988). The used values of $\bar{\delta}_R$ were obtained by Kasten formula (Kasten, 1980):

$$\bar{\delta}_R = 1 / (9.4 + 0.9\ m). \quad (12)$$

2. *Measurements*

The direct solar radiation I, which is used in Eq. 10 to calculate $T_L(m)$, is computed from global (G)–and diffuse (d)–solar radiation measurements according to the equation:

$$G = I\ \sin\ h + D. \quad (13)$$

The measurements were carried out using two Kipp and Zonen precision pyranometers (Model GM 6B), which complies with the specification for "first class" pyranometers (WMO, 1983). One of them was used to measure G and the second was shaded by a shadow band (60 mm width, 610 mm radius), constructed by the author, to measure the diffuse component by eliminating the direct beam. The centering of the sun image on the band was checked daily ensuring adequate declination and azimuth tracking by the band. A two–channel solar integrator (Kipp and Zonen Model cc 12) was connected to the pyranometers to give the solar irradiance (G and D) in wm^{-2}. Because the shadow band intercepts not only the direct solar beam but also a small part of the diffuse sky radiation, the measured values of D were multiplied by a correction factor given by (Latimer and Mac Dowall, 1971):

$$1 / (1 - F / D), \quad (14)$$

where $F / D = (2\omega / n\gamma)\cos^3\delta(\sin\varphi\sin\delta * H + \cos\delta\sin H)$,

φ is the latitude of the station.

H is the hour angle of the sun at sunrise or sunset, given by

$$H = \cos^{-1}(-\tan\varphi\tan\delta), \tag{15}$$

δ is the declination angle of the sun.

γ, ω are the radius and the width of the shadow band, respectively.

The calculated correction factor is found to be varying from 1 to 1.14, which is in a very good agreement with its values determined for other diffusographs (Latimer and Mac Dowall, 1971).

Weather elements, which used in this paper, were measured simultaneously with G and D. The vapor pressure VP in hPa and relative humidity in % were determined from the recorded values of air temperature and wet bulb temperature using tables prepared by the meteorological Authority of A.R. Egypt, 1984 following the WMO rules. Wind speed was measured using air meter (Griffen & Geogre Model 1749, England).

III. RESULTS AND DISCUSSION

As we mentioned above, the value $T_L(m)$ denotes to large extent to the content of dust particles and water vapor in the atmosphere. Consequently its variation with time and place is attributed to changes occurring in the content of these atmospheric components and parameters affecting it, such as some of weather elements and local prevailing synoptic situations.

1. *Diurnal Variation of* $T_L(m)$

Fig. 1 represents the diurnal variation of $T_L(m)$ for clear days at Qena / Egypt averaged monthly, seasonally and over the whole period from June 1992 to May 1993. Also important feature of the diurnal variation of $T_L(m)$ for forenoon, noon and afternoon averages through this period is shown in Fig. 2. From both figures, one can deduce the following:

i) For the months from June to December 1992 as well as in Summer and Autumn seasons, the value of $T_L(m)$ increases steadily from sunrise to sunset. In these months and seasons, the avarage values of $T_L(m)$ in the afternoon time are larger than those in the forenoon one (see Fig. 2). This increase differs from month to month. The diurnal patterns of $T_L(m)$ in this period of the year is due to the high dust content in the lower atmospheric layers arising from the well developed vertical mixing of dust particles owing to the high temperature especially in the afternoon hours. Very good positive correlations, which are found between diurnal values of $T_L(m)$ and temperautre in Summer (0.98) and Autumn (0.93) support this conclusion.

ii) For the months from January to March and Winter season, $T_L(m)$ value falls generally in the direction of sunset, with high average values in the forenoon time compared to the afternoon time. This behavior may be explained as a result of the evaporation of water droplets in the atmosphere at noon time (El-Hussainy, 1989).

iii) In April and May, the value of $T_L(m)$ fluctuates obviously through the day, the matter which appears in its high values of standard deviation found in the two months, especially in the afternoon time. Average daily values of standard deviation are $\pm$ 1.5 in April and $\pm$ 1.2 in May, while its average values in the afternoon time are $\pm$ 1.98 and $\pm$ 1.54 in April and May, respectively. This reflects the instability of the atmosphere in the study region with respect to aerosol particles and water vapor in these two months especially in April due to the dusty khamassin winds blowing in it. In Spring the diurnal variation of $T_L(m)$ is not strong along the most hours of the day (9–15 LAT), in which somewhat closed values of $T_L(m)$ were

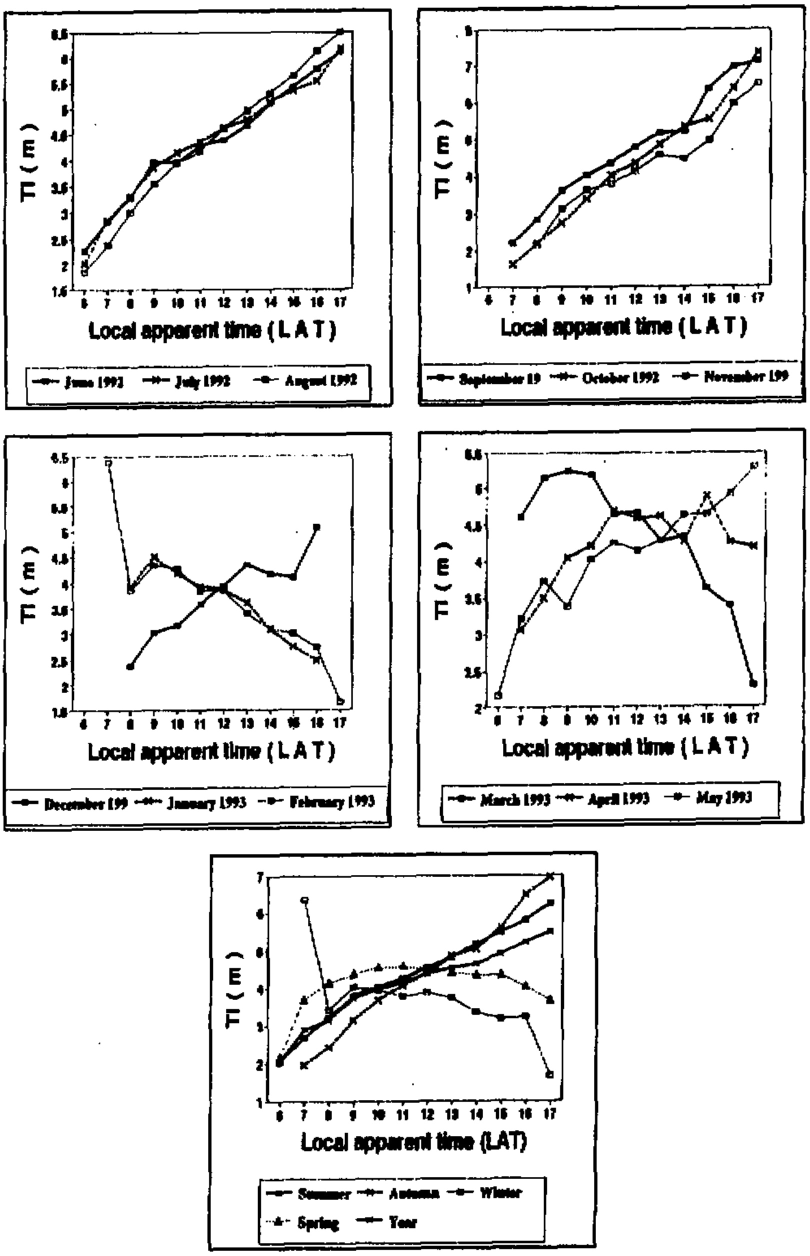

Fig.1. Diurnal variation of $T_L\,(m)$ at clear days in the period from June 92 to May 93.

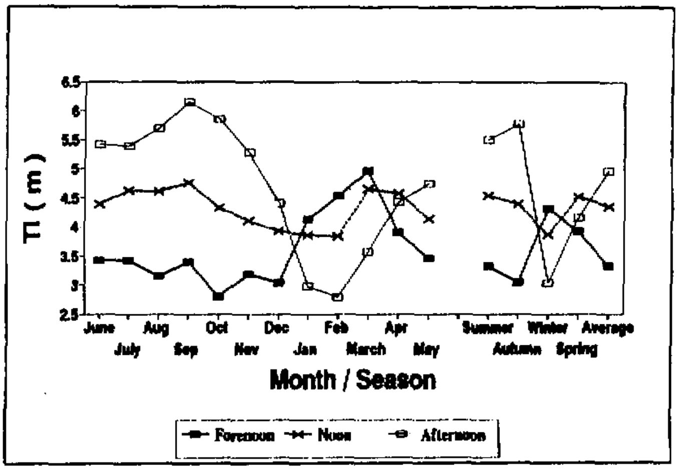

Fig. 2. Average $T_L(m)$ at different day times from June 92 to May 93.

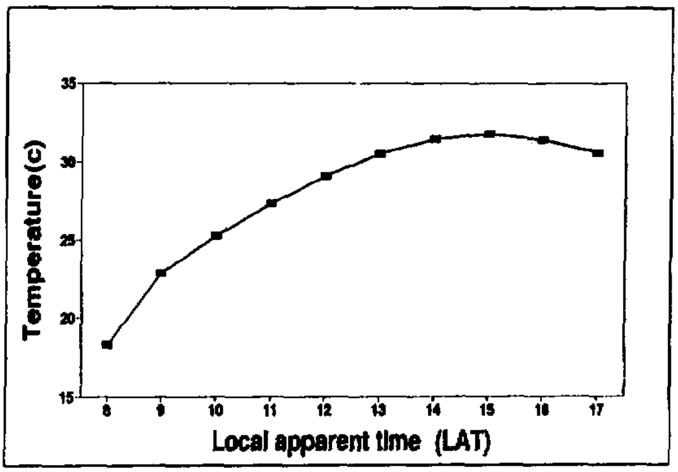

Fig. 3. Diurnal variation of temperature at Qena / Egypt (June 92–May 93).

observed. In this period the average value of $T_L(m)$ in the afternoon time is higher than in the forenoon time.

iv) On the whole measurement period, the diurnal variation of $T_L(m)$ in the study region shows steadily increase in the direction of the sunset with higher values in the afternoon hours than in the forenoon ones. This behavior is logic in view of the general climatic feature of the study region, which is characterized with higher value of temperature in the afternoon time accompanied with higher values of $T_L(m)$ in this time (see Fig. 3).

The above diurnal pattern of $T_L(m)$ is generally in a good agreement with that observed in many other studies (El–Hussainy, 1989).

2. *Daily Variation*

The daily values of $T_L(m)$ averaged over each month and season in the measurement period is represented in Fig. 4. The figure indicates high average values of $T_L(m)$ in summer months (June–August) (4.35 ± 0.52) in comparison with winter months (December – February) (3.64 ± 0.32). The increase of $T_L(m)$ from winter to summer is generally associated with an increase in temperature, vapor pressure, wind speed and decrease in relative humidity. This change in these weather elements leads to an increase in the content of dust and water vapor in the atmosphere and consequently large values of $T_L(m)$. A general feature of the variation of these weather elements in the study region is represented graphically in Fig. 5. A correlation study between each of them and $T_L(m)$ shows very good correlations between $T_L(m)$ and temperature (0.89), vapor pressure (0.78) and moderate correlation between it and wind speed (0.55). However a weak correlation was obtained between $T_L(m)$ and relative humidity (−0.35). The reverse weak influence of RH on $T_L(m)$ is due to its low values and variance in the study region in the most months of the year (<0.30%). It varies from 21% in June to 52% in December and January. The low values of RH limit to large extent its expected effect on increasing $T_L(m)$ through its role in developing the percentage of water droplets condensing on the aerosol particles in the atmosphere (Hanel, 1971; Takada et al., 1986). Accordingly, the effect of RH on $T_L(m)$ in Qena region follows the effect of temperature in a reverse manner. An exception from that is the maximum value of $T_L(m)$ in September which is attributed to the anomalous relative high value of RH in this month (Av. 59%, Max. 81%) in addition to the effect of temperature (Av. 32.5, Max. 38.7). Also the somewhat high values of wind speed recorded to September (Av. 1.54 ms^{-1}& Max. 4.5 ms^{-1}), support this maximum value of $T_L(m)$.

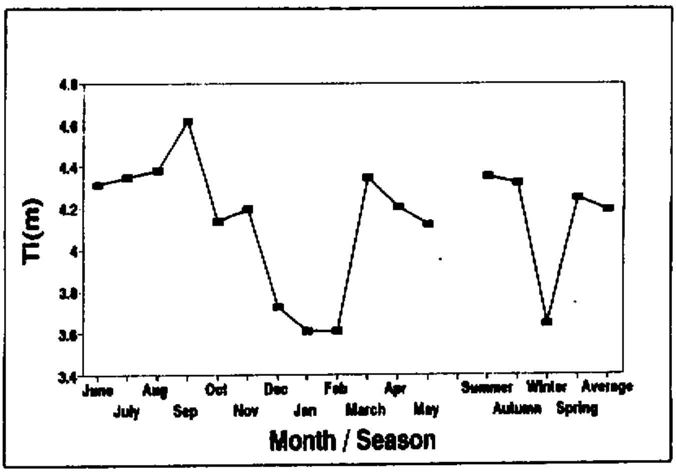

Fig. 4. Daily averages of $T_L(m)$ at clear days from June 92 to May 93.

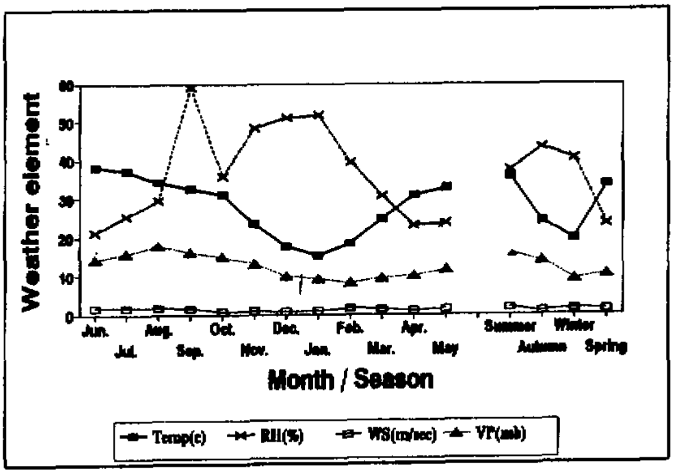

Fig. 5. Variation of some weather elements from June 92 to May 93.

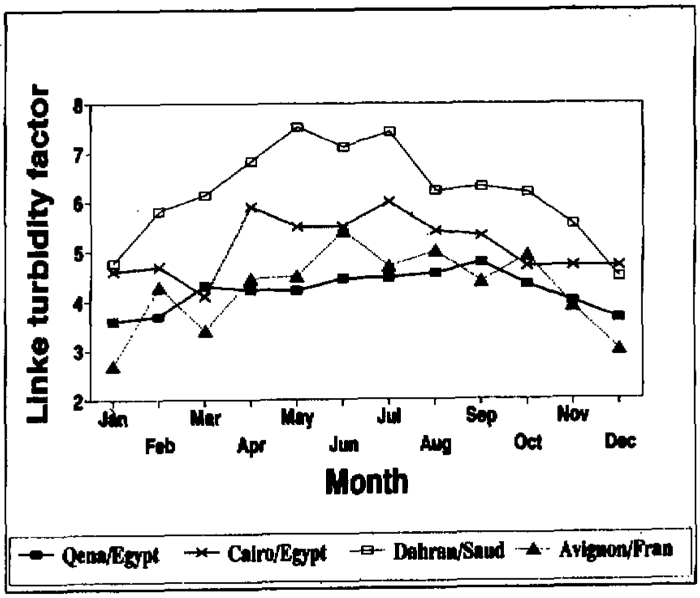

Fig. 6. Monthly averages of Linke turbidity factor in different locations.

According to the above discussion, one see that the seasonal variation of $T_L(m)$ at clear days is mainly dependent on the variation of ambient temperature. This is generally in a good agreement with many other studies, which show approximately the same seasonal pattern as at Qena taking in consideration the different climates and locations. Fig. 6 represents a comparison between the variation of monthly averages of Linke turbidity factor in Cairo / Egypt (El-Hussainy, 1989), Daharan / Saudi Arabia (Abdelrahman, 1988), Avignon / France

(Katz et al., 1982) and in the study region. It is clear that the turbidity level in Qena is smaller than Daharan (Av. ≈ 6.28) and Cairo (Av. 5), while it is higher than Avignon (Av. ≈ 4.22). Also the variation of turbidity factor in Qena is not strong in the most months of the year in comparison with the other cities, reflecting the relative stability of the atmosphere in this region with respect to aerosol and water vapor content. The maximum values of $T_L(m)$ along the year months found in Daharan are due to its short distance from the Gulf, leading to prevailing high level of humidity and consequently water droplets in the atmosphere and also due to the winds blowing in winter and summer from desert area and are highly loaded with dust particles (El-Hussainy, 1989). However in Cairo, the high values of $T_L(m)$ are due to the very high pollution level resulting from industrial traffic and high man activities. Avignon is semi rural site falling in a temperate climate zone and dominated by vegetation and lack of dust storms (Katz et al., 1982), so the turbidity level in this city is smaller than that in the other cities.

3. *Frequency Distribution of $T_L(m)$*

A more clear picture of atmospheric turbidity is obtained by investigating its frequency distribution in the study region. Fig. 7 presents such a distribution for the data measured through the measurement period. It is seen that the frequency distribution over the whole period is asymmetric with a peak frequency of about 16.2% representing the class interval (3.5–4) of turbidity factor. A similar study of frequency distribution of $T_L(m)$ at Cairo (El-Hussainy, 1989) shows a peak frequency of about 20% representing the class interval (4.5–5), which when compared to that at Qena reflects the developed urban effect at Cairo.

4. *The Variation of Turbidity Factor $T_L(m)$ with Diffuse Fraction (D / G)*

Fig. 8 represents the relation between the daily average values of $T_L(m)$ and (D / G)

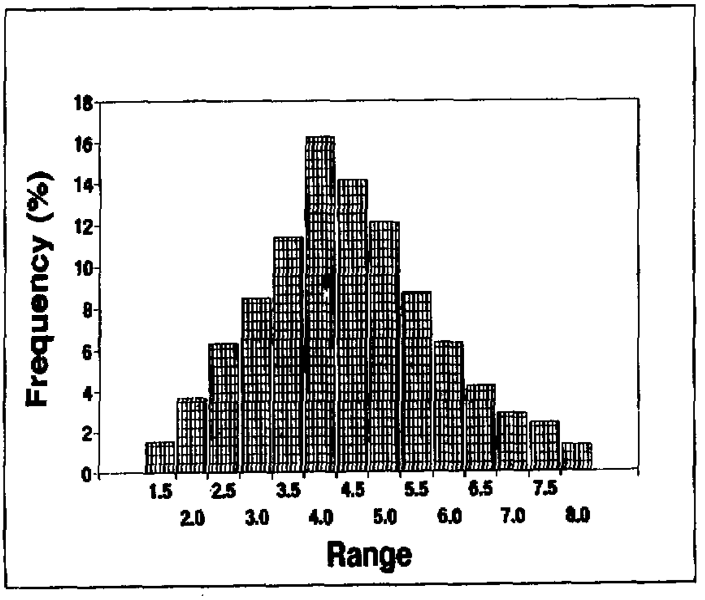

Fig. 7. Percentage frequency distribution of $T_L(m)$ at clear days (June 92–May 93).

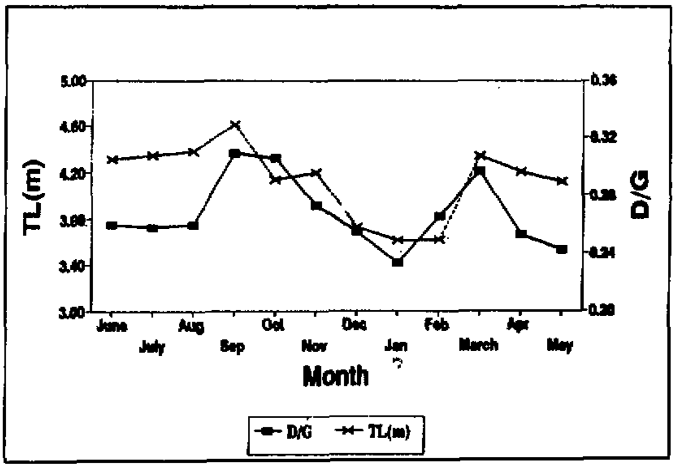

Fig.8. Variation of $T_L(m)$ and D / G at clear days from June 92 to May 93.

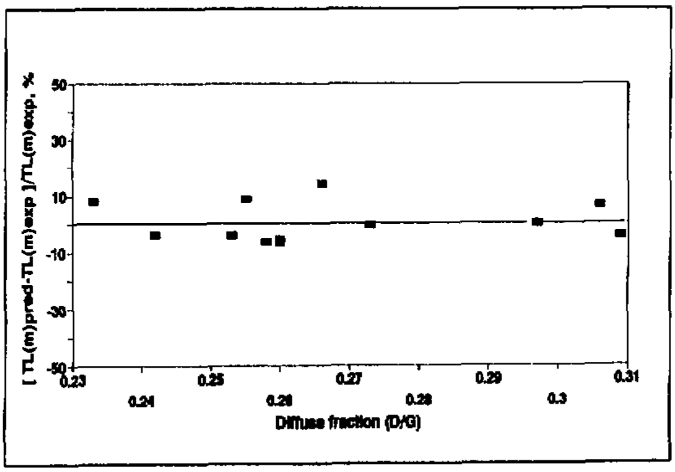

Fig. 9. Relative deviation of $T_L(m)$ pred from $T_L(m)$ exp vs diffuse fraction (D / G).

at different months. It is clear from this figure that the increase of (D / G) is generally accompanied by an increase in $T_L(m)$. An empirical equation between $T_L(m)$ and (D / G) is deduced over Qena. It is found to be in the form:

$$T_L(m) = 2.29 + 6.87(D / G) \tag{16}$$

with correlation coefficient = 0.52.

Fig. 9 represents the relative deviations of the daily averages of $T_L(m)$, predicted using Eq. (16), from the measured one $T_L(m)$ at different values of (D / G). Relative deviations in the range from −7.5% to 14%, which are shown in this figure, indicate the possibility of using (D / G) values to determine $T_L(m)$ according to Eq. (16) with acceptable accuracy.

5. *Elimination of the Virtual Variation of $T_L(m)$*

The value of $T_L(m)$ calculated according to equation 10 shows a reasonably dependence on m owing to the quite different dependence of $\bar{\delta}(\bar{\delta}_R, \bar{\delta}_D, \bar{\delta}_W)$ on the spectral distribution of the incident radiation, which varies with m (Kasten, 1988). This behavior is called the virtual variation of atmospheric turbidity. However the dependence of $T_L(m)$ on m is smaller than of $\bar{\delta}$ owing to the partial compensation of the wave length dependence of the numerator and denominator in Eq. (6). An elimination of the dependence of $T_L(m)$ on m is possible according to Katsoulis and Tselepidaki, 1986 and WHO 1981 by reducing the value of $T_L(m)$ measured perheliometrically to relative optical air mass $m=2$, at which it is independent of m (Kasten, 1988). In their calculations, they reduced $I(T_L(m), m)$ in equation 9 to the reduction optical air mass $m=2$ as:

$$I(T_L(m),2) = I_0[e^{(-T_L(m)\cdot m\cdot \bar{\delta}_R(m))}]^{2/m}, \tag{17}$$

and then introduced the reduced turbidity factor $T_L(m)$ by the:

$$I(T_L(m),2) = I_0 e^{(-T_L(2)\cdot 2\cdot \bar{\delta}(2))}. \tag{18}$$

Through some what laborious calculating procedure, the equating of Eqs. (17) and (18) boils down to

$$T_L(m)\cdot \bar{\delta}_R(m) = T_L(2)\cdot \bar{\delta}_R(2). \tag{19}$$

According to this equation, Kasten has derived formula for calculating the wanted reduced Linke turbidity factor $T_L(2)$ with the aid of the proposed function for $\bar{\delta}$ (Eq. 12) as follows:

$$T_L(2) / T_L(m) = 11.2 / (0.9m + 9.4) \tag{20}$$

that when substituted in Eq. 9, using the approximation $m = 1 / \sin h$, one obtains:

$$I = I_0 e^{(-T_L(2)/11.2\sin h)} \tag{21}$$

from which

$$T_L(2) = 11.2\sin h \ \ln (I_0 / I). \tag{22}$$

Table 1. Number of Observations of Global and Diffuse Solar Radiation Components at Clear Days from June 92 to May 93

Month	No. of observations
June 1992	208
July	280
August	254
September	201
October	210
November	149
December	87
January 1993	118
February	132
March	114
April	117
May	72
Total	1942

Table 2.Diurnal Variation of $T_L(2)$, $T_L(m)$, and m Averaged over the Whole Measurement Period

Time	7	8	9	10	11	12	13	14	15	16	17
$T_L(2)$	2.72	3.05	3.82	4.19	4.45	4.66	4.86	4.91	4.97	5.04	5.26
$T_L(m)$	2.91	3.15	3.72	3.98	4.16	4.37	4.55	4.65	4.90	5.22	5.50
m	2.55	1.73	1.39	1.24	1.20	1.25	1.44	1.94	3.07	3.33	1.84
$[T_L(2)-T_L(m)]/T_L(m)$ %	-6.31	-2.98	2.75	5.53	6.86	6.75	6.85	5.69	1.38	-3.51	-4.44

Table 3.Monthly Variations $T_L(2)$,$T_L(m)$ and m in the Period from June 1992 to May 1993

Month	$T_L(2)$	$T_L(m)$	m	$[T_L(2)-T_L(m)/T_L(m)$ %
June	4.46	4.31	1.861	3.467
July	4.48	4.35	1.930	3.039
August	4.56	4.38	2.434	4.036
September	4.78	4.62	1.957	3.540
October	4.33	4.14	2.610	4.778
November	4.00	4.20	3.430	−4.593
December	3.64	3.73	2.417	−2.366
January	3.59	3.61	2.223	−0.701
February	3.70	3.61	2.932	2.434
March	4.32	4.34	2.229	−0.532
April	4.23	4.21	1.760	0.485
May	4.22	4.12	2.127	2.467

Using Eq. (22), values of $T_L(2)$ were calculated at clear days from June 1992 to May 1993 in the study region. These values refer only to the effect of dust particles and water vapor on the depletion of the incoming solar radiation. Generally it show similar behavior to that of $T_L(m)$. Tables 2 and 3 represent the relation between $T_L(2)$ and $T_L(m)$ for computed 1942 values, averaged dirunally and monthly over the whole measurement period, respectively. It is seen that the average values of $T_L(2)$ are very closed to that of $T_L(m)$. The percentages of its relative deviations from $T_L(m)$ lie in the ranges 6.86% for diurnally variations and 4.6% for monthly ones. This indicates the weak effect of the virtual variation on the turbidity factor in Qena owing to the low deviations of m from considered value $m=2$ in this city, at which, the average value of m is 2.33 ± 0.46. Using linear regression technique, the following models have been found for Qena data of $T_L(2)$ and $T_L(m)$:

i) For diurnal variations

$$T_L(2)=1.001T_L(m)+0.070 \tag{23}$$

with correlation coefficient = 0.966.

ii) For monthly variations

$$T_L(2)=1.140T_L(m)-0.521 \tag{24}$$

with correlation coefficient = 0.956.

The predicted values of $T_L(2)$ using the above two Eqs. (23) and (24) are not

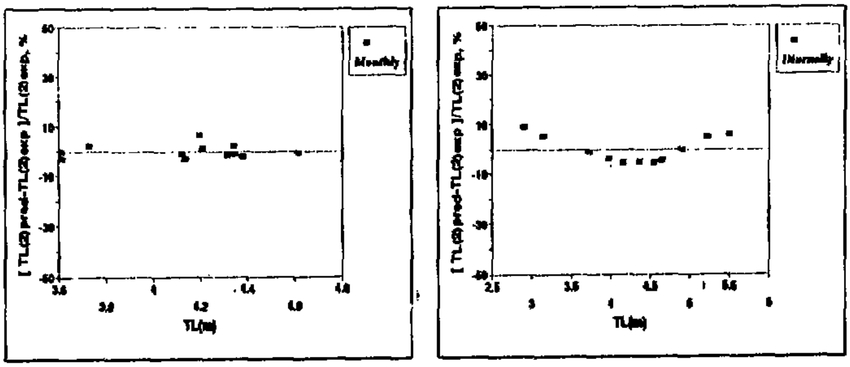

Fig. 10. Relative deviations of $T_L(2)$ pred from $T_L(2)$ Exp vs $T_L(2)$ Exp.

significantly different from that obtained experimentally by Eq. (22). Its relative deviations from $T_L(2)$ lie within the ranges (−4.94%−9.6%) and (2.97%−6.75%) for diurnally and monthly averages data, respectively (see Fig. 10).

IV. CONCLUSION

The effect of atmospheric aerosol particles and water vapor on the depletion of incident solar radiation in Qena / Egypt has been investigated, based on Linke turbidity factor data for clear days in the period from June 92 to May 93. The study leads to the following conclusions:

1. The diurnal variations of $T_L(m)$ are characterized with steadily increase in the direction of sunset in the period from June to December 92, while it shows general fall in this direction in the period from January to March 93. In April and May, the $T_L(m)$ values show an obvious fluctuation through the day hours from sunrise to sunset.

2. The values of $T_L(m)$ in the summer months are higher than its values in winter ones, which agrees with many other studies, at different locations.

3. The results of $T_L(m)$ is found to be highly dependent on the variation of the ambient temperature and vapor pressure and to some extend to the variation of wind speed. Effect of relative humidity seems to be weak and reversible, which is inconsistent with its expended rules.

4. The turbidity level in Qena is generally high. However it is smaller than in Cairo / Egypt and Daharan Saudi Arabia.

5. The partial dependence of $T_L(m)$ on the relative optical air mass m is eliminated by reducing its values to $m=2$ using kasten formula, which was introduced for this purpose. The resulting values of $T_L(m)$ are more representative for the content of water vapor and dust particles in the atmosphere.

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