THE ATTENUATION OF SOLAR RADIATION IN URBAN ATMOSPHERES

BY

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ABSTRACT

Unsworth and Monteith's (1972) aerosol attenuation coefficient τ_A was calculated with hourly cloudless data at four North American and four European stations for varying time periods. Monthly and seasonal turbidity trends were examined. Annual cycles were observed with summer maximums and winter minimums. The North American stations were less turbid and had more pronounced trends than the European stations. Both air mass origin and local weather affect the turbidity. Local sources of pollution have a significant effect on turbidity most notably in large urban centres.

ACKNOWLEDGEMENTS

I wish to thank my supervisor Dr. J.A. Davies for giving me the idea for this thesis and mostly for the continued guidance he offered throughout the year. I would also like to thank Susan Swan for her support and friendship.

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LIST OF SYMBOLS

ET	Equation of time (minutes)
$I_{(\lambda)}$	Spectral flux density of direct beam solar radiation
I_0	Cloudless sky direct beam solar radiation
$I_{0(\lambda)}$	Incoming flux at the top of the atmosphere by a plane normal to the sun
I_0^*	Extraterrestrial radiation
$I^*_{(\lambda)}$	Spectral flux density below an aerosol free atmosphere
I^0	Corrected solar constant
LAT	Local Apparent Time
LS	Longitude of station
LSM	Standard meridian for time zone
LST	Local Standard Time
RV	Radius vector (rv)
S	Solar constant = 1376 Wm^{-2}
S.E. $\bar{x}$	Standard error of the mean
T_{03}	Transmission due to ozone
T_r	Transmission due to Rayleigh scatter
UO3	Depth of ozone in the atmosphere
UH_2O	Depth of water vapour in the atmosphere
a_{H20}	Absorption due to water vapour
h	Solar hour angle (degrees)
m	Optical air mass
n	Number of observations

LIST OF SYMBOLS (continued)

z	Solar zenith angle
δ	Solar declination
σ	Standard deviation
$\tau_{a(\lambda)}$	Spectral optical depth for aerosol attenuation
τ_A	Aerosol attenuation (turbidity) coefficient
$\tau_{H2O(\lambda)}$	Spectral optical depth for water vapour absorption
$\tau_{O3(\lambda)}$	Spectral optical depth for ozone absorption
$\tau_{r(\lambda)}$	Spectral optical depth for Rayleigh scattering
$\varnothing$	Latitude of station

CHAPTER ONE

INTRODUCTION

Atmospheric aerosols reduce the amount of solar energy that reaches the surface due to absorption, which increases the radiative heating of the atmosphere, and scattering (Unsworth and Monteith, 1972). Aerosol attenuation can affect the climate at both the global and local scales. Davies and al. (1988) and Ball and Robinson (1982) indicated that the attenuation of global radiation due to aerosol is very significant and can approach the magnitude of attenuation by water vapour in atmospheres that are significantly affected by mankind's pollution.

Aerosols can be defined as the system of colloidal particles (solids and liquids, excluding water vapour) dispersed in the atmosphere. The terms turbidity coefficient and aerosol attenuation coefficient are interchangeable.

The concern surrounding changes in global climate has fostered interest in the radiative effects of aerosols (Unsworth and Monteith, 1972). Rasool and Schneider (1971) calculated that an increase in aerosol content will decrease the mean surface temperature on the earth.

This study examines the effect that aerosols have on solar radiation in urban atmospheres. In order to examine these effects a turbidity coefficient (τ_A) is used as a measure of the attenuation.

Values of turbidity coefficients calculated from radiation measurements at eight stations in North America and Europe will be examined on a monthly and seasonal basis. Six of the eight stations are located within or near (≤40km) a major urban centre.

CHAPTER TWO

BACKGROUND

2.1 Theoretical Background

The turbidity coefficient that will be used is Unsworth and Monteith´s (1972) aerosol attenuation coefficient as modified by Uboegbulam and Davies (1983). The turbidity coefficient τ_A is related to the spectral optical depth. The coefficient is derived from Beer´s Law which determines the spectral flux density of direct beam solar radiation, I_λ , after atmospheric attenuation (Uboegbulam and Davies, 1983). The spectral flux density is given as

$$I_{(\lambda)} = I_{0(\lambda)} \exp\{-(\tau_{03(\lambda)} + \tau_{r(\lambda)} + \tau_{H20(\lambda)} + \tau_{a(\lambda)})m\} \qquad (1)$$

From equation 1, $I_{0(\lambda)}$ is defined as the incoming flux density at the top of the atmosphere by a plane normal to the sun, $\tau_{03(\lambda)}$, $\tau_{r(\lambda)}$, $\tau_{H20(\lambda)}$, $\tau_{a(\lambda)}$, are spectral optical depths for ozone absorption, Rayleigh scattering, water vapour absorption, and aerosol attenuation respectively, and m is the optical air mass.

In order to separate $\tau_{a(\lambda)}$ from the other spectral components equation (1) can be rewritten as

$$I_{(\lambda)} = I_{0(\lambda)} \exp\{-(\tau_{03(\lambda)} + \tau_{r(\lambda)} + \tau_{H20(\lambda)})m \cdot \exp\{-\tau_{a(\lambda)}m\} \quad (2)$$

or,

$$I_{(\lambda)} = I'_{(\lambda)} \exp\{-\tau_{a(\lambda)}\, m\} \qquad (3)$$

where

$$I^{*}_{(\lambda)} = I_{0(\lambda)} \exp\{-(\tau_{O3(\lambda)} + \tau_{r(\lambda)} + \tau_{H2O(\lambda)})m\} \quad (4)$$

which is the spectral flux density below an aerosol free atmosphere. Unsworth and Monteith (1972) extended (3) to the whole solar spectrum,

$$I = I^{*} \quad \exp(-\tau_{A} m) \quad (5)$$

By rearranging equation (5), the aerosol attenuation coefficient can be calculated as

$$\tau_{A} = -\frac{1}{m} \ln \left(\frac{I}{I_{0}} \right) \quad (6)$$

where τ_A is the weighted mean aerosol optical depth defined as

$$\exp(-\tau_{A}\, m) = \frac{\int I^{*}_{(\lambda)} \exp[-\tau_{a(\lambda)} m] \; d\lambda}{\int I^{*}_{(\lambda)} \; d\lambda} \quad (7)$$

and I is the measured direct beam flux density for the entire spectrum and I_0 can be calculated from

$$I_{0} = S\ (rv^{2})\ [T_{O3(mUO3)} \quad T_{r(m)} \; - \; a_{H2O(mUH2O)}\ ] \quad (8)$$

where S is the solar constant, rv^2 is the radius vector, m is the optical air mass, T_{O3} is the transmission due to ozone, T_r is the transmission due to Rayleigh scatter, a_{H2O}

is the absorption due to water vapour and U_{O3} and U_{H2O} are the depths of ozone and water vapour respectively in the atmosphere.

2.2 Literature Review

The topic of atmospheric turbidity has only been dealt with briefly in the literature.

By measuring the attenuation of the direct beam solar radiation, an atmospheric turbidity coefficient can be inferred for a given site (Unsworth and Monteith,1972; Uboegbulam and Davies,1983; Darby and Hay,1984). There have also been studies on turbidity based on Linke´s turbidity factor (Flowers et al. (1969; Polavarpu, 1978). Both methods of determining atmospheric turbidity display a definite seasonal variation. Unsworth and Monteith (1972) recorded larger values in summer then in winter, in which there was no pronounced change. They attribute the higher atmospheric turbidity coefficient values to the greater frequency of continental air masses rather then to local sources of aerosols which would be produced from domestic or industrial sources.

Uboegbulam and Davies (1983) reported that atmospheric turbidity over Eastern Canada also showed a pronounced annual cycle with a maximum occurring during the summer. This coincides with turbid continental air mass flow from the south. Hay and Darby (1984) found that the typical monthly variation pattern, with summer maximums, occurred for Vancouver, Canada. This was shown to be related to the synoptic conditions and air mass characteristics.

Peterson, et al. (1981) found that there was an annual pattern of highest turbidity and daily variation happening in the summer and during the winter it was the opposite, with lowest turbidity and daily variation occurring. The largest turbidity occurred when air masses had a southern origin.

The importance of the urban environment's effect on the aerosol attenuation coefficient will also be examined in the research paper. Unsworth and Monteith's (1972) variations in turbidity are due to changes in air mass origins and to urban effects, including local industrial sources. Urban effects, most notably pollution, appear to influence turbidity. Uboegbulam and Davies (1983) found that τ_A for Montreal, Quebec and Woodbridge, Ontario are larger than those for Goose Bay, Newfoundland. This is more evident in summer than in winter. (Uboegbulam and Davies, 1983). Flowers, et al. (1969) found that there were increases in the turbidity levels for the larger heavily industrialized cities, most notably New York, Baltimore and Philadelphia.

From previous studies (Unsworth and Monteith, 1972, Uboegbulam and Davies, 1983,) it has been shown that eastern

Canada has a smaller aerosol loading in the atmosphere then Britain's.

Uboegbulam and Davies (1983) found that the long term variation of τ_A decreased during a ten year period (1968-78) in eastern Canada. This decrease may correspond to changes in the air quality and the 1971 Canadian Clean Air Act law. In the United States, Peterson, et al. (1981) found an 18% per decade (1965-1979) linear increase in turbidity over central North Carolina. On the other hand, Flowers et al. (1969) did not find any pronounced trend in atmospheric turbidity over the period of 1961-1966.

The differences which occur between the United States and Canadian studies is most likely a result of localized turbidity trends for each region. The different turbidity factors and coefficients used in the literature place some restrictions on how the data results can be interpreted.

CHAPTER THREE

METHODOLOGY

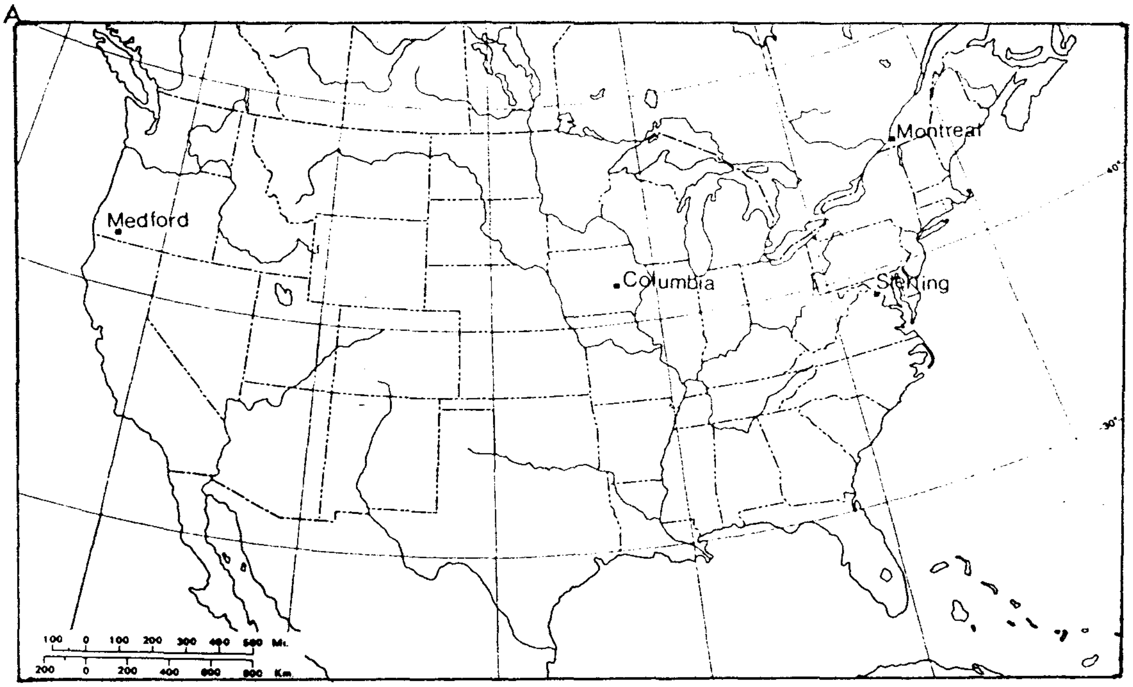
3.1 Data Source

The measured radiation data were extracted from the 1988 International Energy Agency (IEA) solar heating and cooling program. The data were originally used in a study to validate several models for estimating solar radiation on horizontal surfaces as part of the solar radiation and pyranometry studies.

The data used were only the cloudless hours from the IEA data set. Only cloudless data are used because clouds attenuate solar radiation significantly. The data included hourly values of global and diffuse solar radiation, visibility, optical air mass, and the zenith angle. The direct solar radiation was calculated as the difference between the global and diffuse radiation.

This study used data from eight mid-latitude stations (Fig. 1.) of which four were North American and four were European. The North American stations are Montreal, Quebec (45.50N, 73.62W), Columbia, Missouri (38.82N, 92.22W), Medford, Oregon (42.37N, 122.87W) and Sterling, Virginia (38.98N, 77.47W). The European stations were De Bilt, Netherlands (52.10N, 5.18E), Hamburg, West Germany (53.63N, 10.00E), Kew, United Kingdom (51.48N, 0.30W), and Zurich, Switzerland (47.48N, 8.53E).

A



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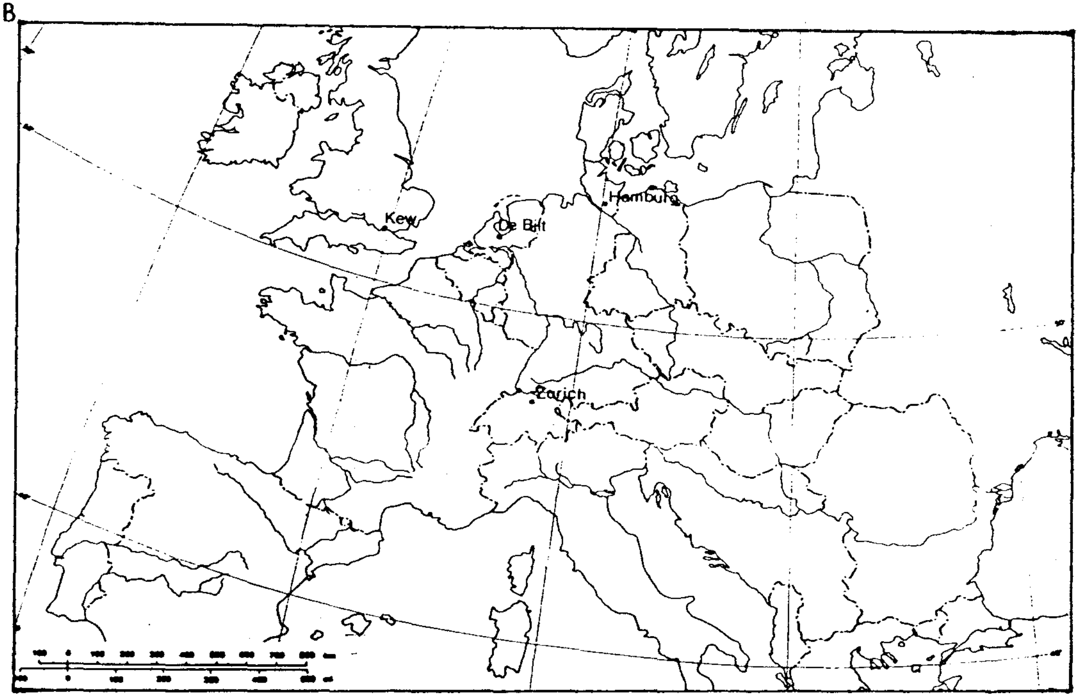

FIG. 1. Location of the eight stations, A. North America and B. Europe.

There were at least two years of data available for each station. The stations and the years from which the data were collected are shown in Table 1.

TABLE 1. Stations and data periods

Station	Data period
Montreal, Quebec	1972 - 1974
Columbia, Missouri	1979 - 1980
Medford, Oregon	1978 - 1980
Sterling, Virginia	1979 - 1980
De Bilt, Netherlands	1971,1976,1979
Hamburg, W.Germany	1976 - 1978
Kew, United Kingdom	1975 - 1977
Zurich, Switzerland	1964 - 1965

3.2 Procedure

A number of astronomical calculations were required before the turbidity coefficient could be calculated. The calculations were all done using a Fortran 77 program (Appendix 1). This program ran the extracted data set from the IEA study and calculated a turbidity coefficient for each hour.

The following summarizes the astronomical calculations and parameters used in the Fortran 77 program. The algorithm used to compute the astronomical parameters is from Michalsky (1988). For all calculations the solar constant (S) has a value of 1376 Wm^{-2}. The solar constant was multiplied by the radius vector (RV) squared in order to correct for the departure of the sun-earth distance from the mean value.

The extraterrestrial radiation can be calculated as

$$I_0' = I^0 \cos z \tag{9}$$

where I^0 is the corrected solar constant and cos z is the cosine of the solar zenith angle, z. This is calculated from

$$\cos z = \sin\phi \sin\delta + \cos\phi \cos\delta \cosh \tag{10}$$

where ϕ is the latitude of the station, δ is the solar declination and h is the solar hour angle (degrees). The value h is defined as

$$h = 15|12 - LAT| \tag{11}$$

where LAT is the local apparent time. The LAT is calculated from

$$LAT = LST + ET/60 + (LSM - LS)/15 \tag{12}$$

where LST is the local standard time, ET is the equation of time (minutes), LS is the longitude of the station and LSM is the standard meridian for the time zone.

3.3 Data Management

The data provided for this study were in a Fortran 77 results file. The initial step was to edit the data set for negative τ_A values and for obvious errors in the global or diffuse solar radiation data.

The next step involved editing out the days that had only single hours of cloudless data. This removed the possibility of cloud contaminated data. Due to the limited data set, it was decided that there would be no monthly minimum number of hours required. Only in the cases of extreme results were the data eliminated from the set.

The third step was to determine mean monthly and mean seasonal τ_A values for each station. Also the number of hours and standard deviations for each month and season were calculated.

3.4 Error Analysis

The standard deviation was calculated for each monthly and seasonal mean. Using the standard deviation, the standard error of the mean was calculated as follows,

$$S.E.\ \bar{x} = \frac{\sigma}{\sqrt{n}} \qquad (13)$$

where σ is the standard deviation of the sample and $\sqrt{n}$ is the square root of the number of observations. The standard error of the mean is represented as plus and minus error bars on the plotted figures.

CHAPTER FOUR

RESULTS AND DISCUSSION

4.1 Calculation of τ_A

The data sets available for each station were somewhat limited, the minimum number of consecutive cloudless hours in one day that were required for this analysis was set at two hours. The total number of cloudless hours was variable for each station and their periods (Table 2). The monthly and seasonal τ_A values may be unrepresentative for months with small data samples.

TABLE 2. Number of cloudless hours for stations, a.North America and b.Europe.

a.

Year	Montreal	Medford	Columbia	Sterling
1972	113	-	-	-
1973	72	-	-	-
1974	69	-	-	-
1978	-	864	-	-
1979	-	1104	481	621
1980	-	1225	354	744
Total	254	3193	835	1365

b.

Year	De Bilt	Hamburg	Kew	Zurich
1964	-	-	-	188
1965	-	-	-	124
1971	163	-	-	-
1975	-	-	87	-
1976	292	205	119	-
1977	-	70	23	-
1978	-	67	-	-
1979	47	-	-	-
Total	502	342	229	312

In general, the North American stations tend to have a larger number of observations then the European stations with the exception of Montreal. The station with the greatest sample size is Medford, Oregon and the station with the smallest sample size is Kew, United Kingdom. The variability in sample size may be due to instrumentation or the location of the station within the city.

4.2 Variability in Turbidity

Unsworth and Monteith (1972) indicate there are two important factors which can affect the turbidity of an atmosphere. The first one is the local weather, this can determine the aerosol input from the both domestic and industrial sources. The second factor is the synoptic history of the prevailing air mass which can determine the input of aerosol from the more distant sources. Air masses also affect the distribution of aerosols through the troposphere. Of the two factors Unsworth and Monteith (1972) consider the latter of the two the dominating factor and that the former is less important at most sites.

4.3 Daily variation

The preliminary analysis of the data investigated the diurnal trend in turbidity for a number of days at two stations (Montreal, Quebec and Hamburg, West Germany). Each station displayed the characteristic diurnal trend noted by Unsworth and Monteith (1972) and Peterson et al. (1980). The turbidity trend had morning and late afternoon minima and daily maxima around solar noon. Due to the inconsistent data periods available for each station direct comparisons (same days) between stations were not possible.

4.4 Monthly variation

The monthly variation of τ_A was calculated from the averages of the daily hourly τ_A values (minimum 2 hours) for each month. A major problem faced in the monthly analysis was the lack of data for the winter months. Most of the stations show a pronounced annual cycle of April to September maximums and October to March minimums. The North American stations tend to have lower monthly mean turbidity values then the European stations.

Montreal, Quebec shows an annual cycle for both 1972 and 1973 (Fig. 2). The data for 1974 was limited and does not display the maximum and minimum pattern observed in 1972 and 1973. Montreal is considered a large urban/industrial station by Canadian standards and has a population of over a million people. The monthly variation of Medford, Oregon shows very consistent results (Fig. 3). The mean turbidity value is around 0.08 for each month for the three years. This station could be used as a background value for an area

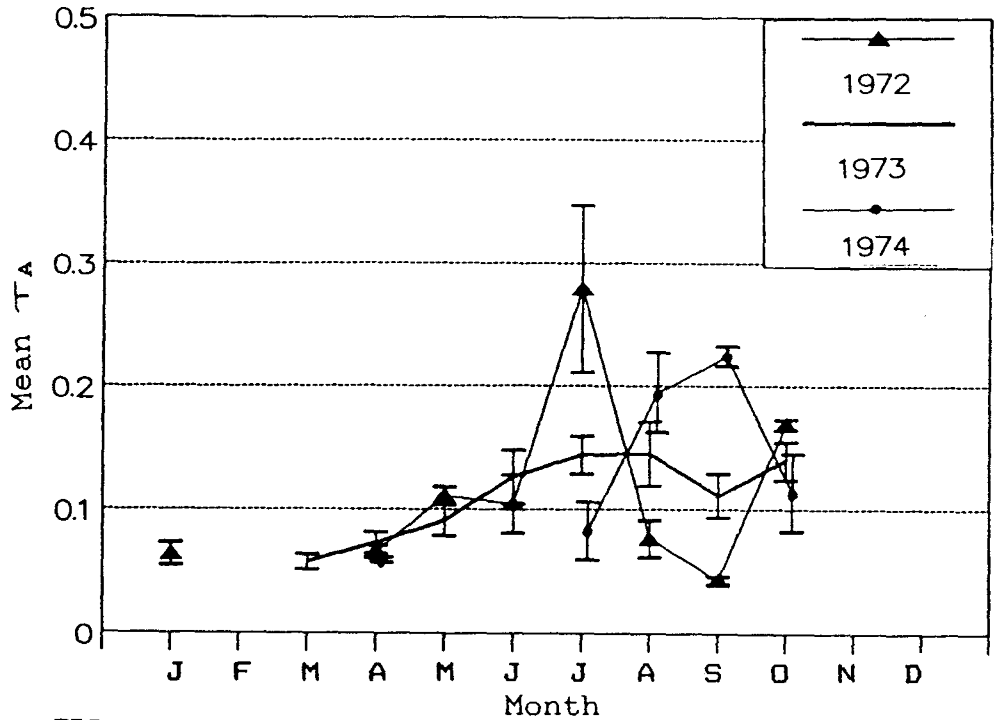

FIG. 2. Monthly variation of mean τ_A for Montreal, Quebec. The bars indicate plus or minus one standard error.

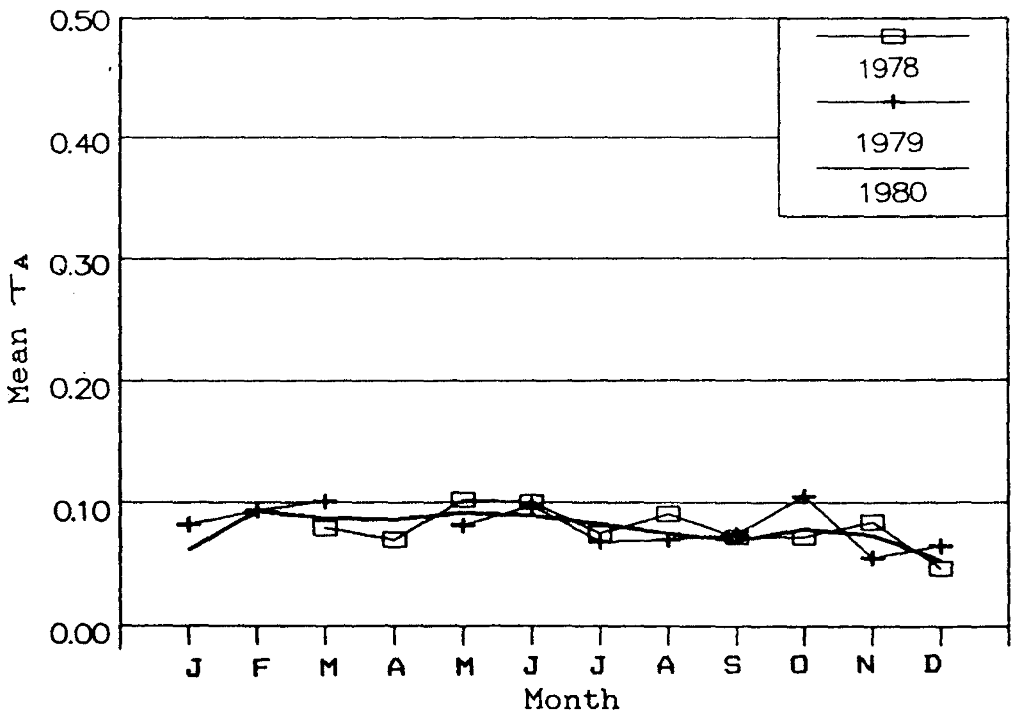

FIG. 3. Monthly variation of mean τ_A for Medford, Oregon. The bars indicate plus or minus one standard error.

with limited pollution and population (50,000).

In Columbia, Missouri the monthly variation of atmospheric turbidity also displays the annual cycle characteristics for both periods (Fig. 4). This station is located approximately sixty kilometres west of St.Louis and has a population of less than 100,000. The graph of Sterling, Virginia (Fig. 5) shows a very distinct annual cycle of summer maximums and winter minimums. Sterling has the highest turbidity coefficient (τ_A =0.338) value of the four sites in North America. It is located approximately forty kilometres north west of Washington, D.C., a large urban centre.

Monthly variation of mean turbidity for the European stations tends to be much more variable. This may be explained by the smaller sample sizes available to work with. The monthly variation for Zurich, Switzerland is uniform and does not display any pronounced cycles considering the amount of available data (Fig. 6). Zurich is located in north Switzerland with a approximately 400,000 people. The graph of De Bilt, Netherlands shows a weak annual cycle with mean turbidity maximums usually occurring earlier in the year (Fig. 7). De Bilt is located in west central Netherlands near Utrect which has a population over 500,000. This station has the highest turbidity value (τ_A =0.564) of all the stations with average values equal to

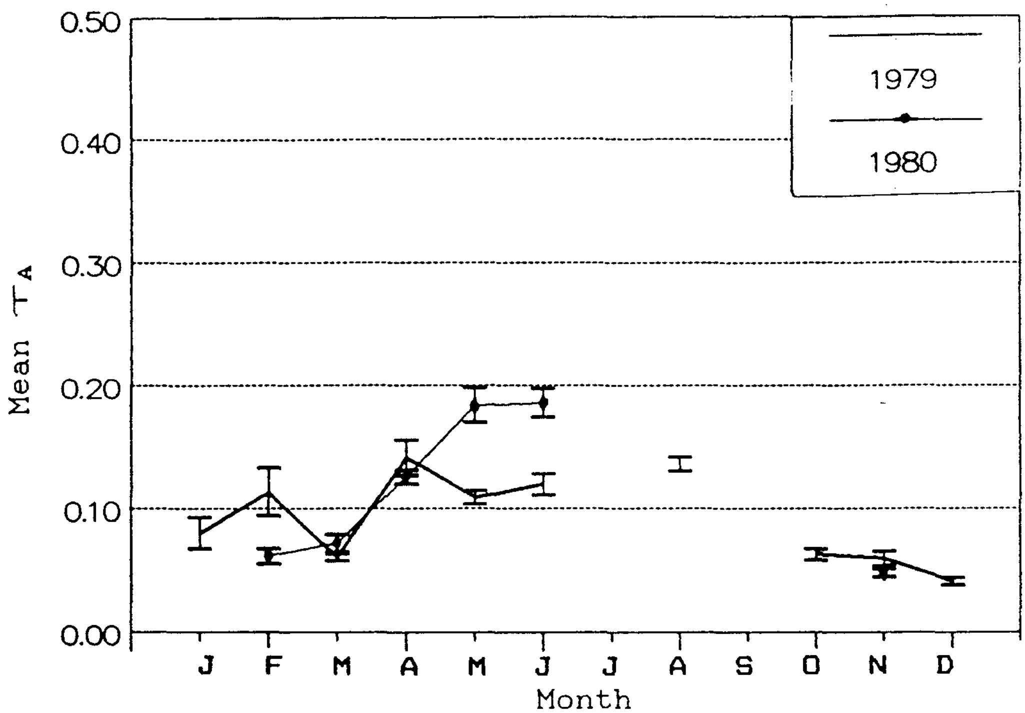

FIG. 4. Monthly variation of mean τ_A for Columbia, Missouri. The bars indicate plus or minus one standard error.

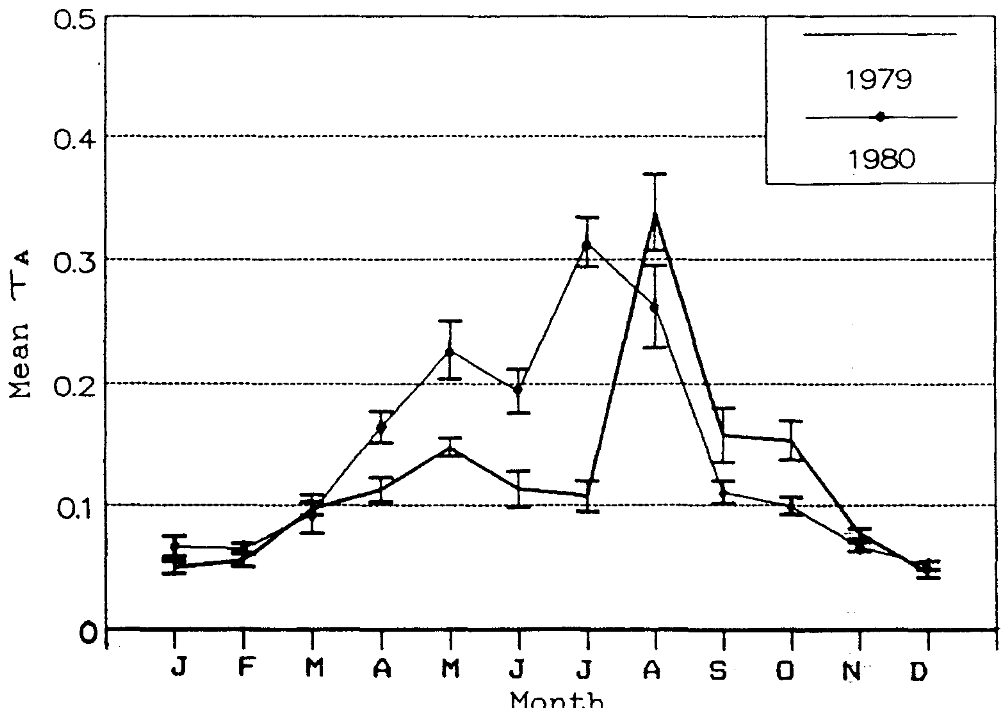

FIG. 5. Monthly variation of mean τ_A for Sterling, Virginia. The bars indicate plus or minus one standard error.

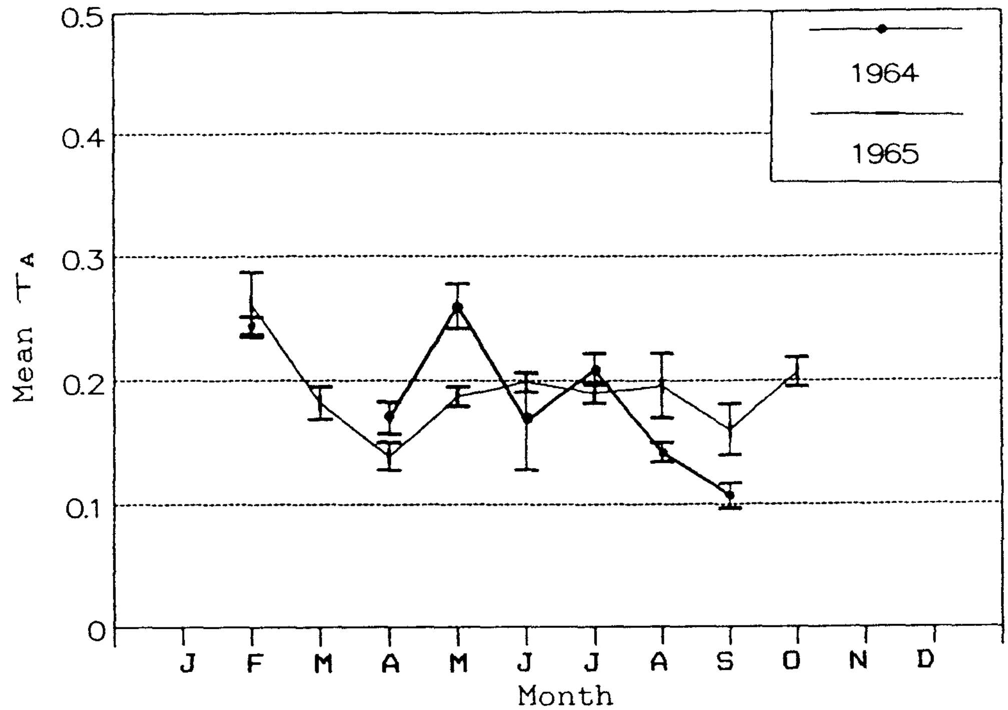

FIG. 6. Monthly variation of mean τ_A for Zurich, Switzerland. The bars indicate plus or minus one standard error.

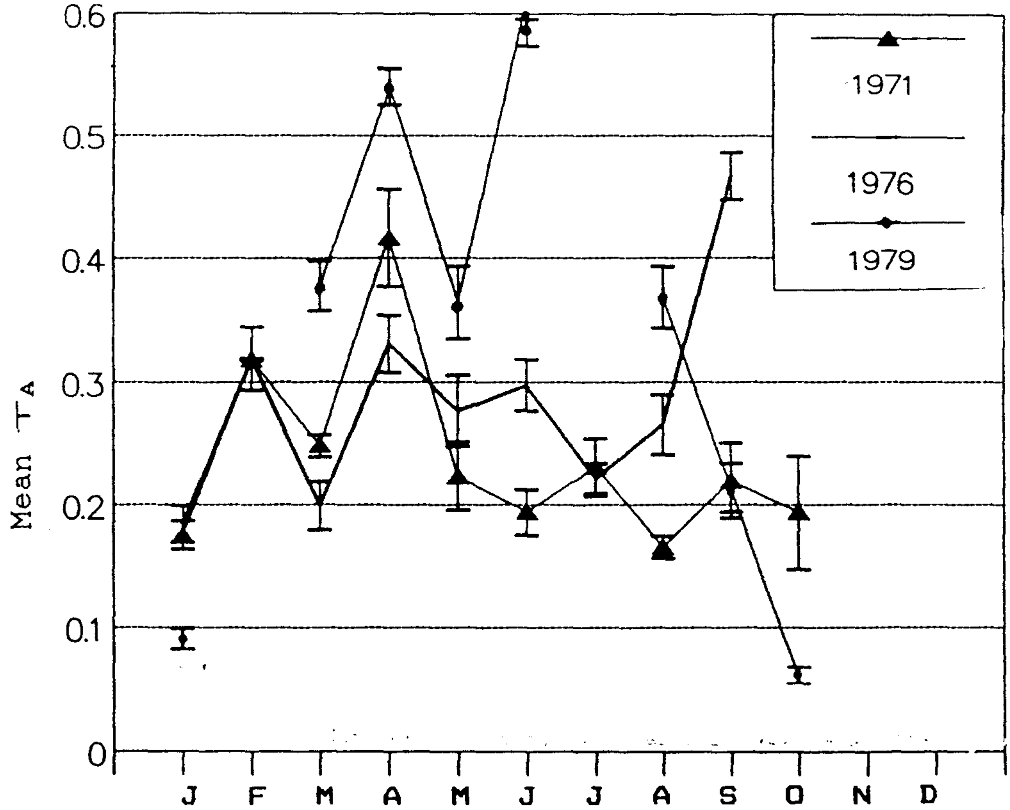

FIG. 7. Monthly variation of mean τ_A for De Bilt, Netherlands. The bars indicate plus or minus one standard error.

0.25.

The next station examined was Hamburg, West Germany, a large industrial city with a population of approximately 1.8 million people. Hamburg was also very variable (Fig. 8), fluctuating throughout the year, with maximums occurring in June and October. Kew, United Kingdom had the smallest sample size, but the monthly variation of mean turbidity does display the annual cycle (Fig. 9). For all three years the maximums occur in July and August. Kew is a London suburb, located within a heavily populated and industrialized city.

Mean monthly turbidity values were calculated for each year and station from the monthly values. The results in Table 3. are a summary of each station. When the data periods are averaged they show a more noticeable trend.

TABLE 3. Average monthly turbidity means

Month	Montreal	Medford	Columbia	Sterling	DeBilt	Hamburg	Kew	Zurich
Jan	0.065	0.072	0.080	0.055	0.142	0.157	-	-
Feb	-	0.094	0.088	0.057	0.305	0.097	0.310	0.253
Mar	0.058	0.089	0.066	0.090	0.259	0.218	0.238	0.182
Apr	0.065	0.078	0.133	0.130	0.404	0.183	-	0.155
May	0.101	0.092	0.146	0.176	0.270	0.124	0.243	0.223
Jun	0.115	0.096	0.153	0.143	0.343	0.285	0.238	0.183
Jul	0.167	0.075	0.406	0.195	0.217	0.183	0.301	0.200
Aug	0.134	0.079	0.136	0.287	0.249	0.085	0.253	0.169
Sep	0.121	0.072	-	0.128	0.281	0.165	-	0.134
Oct	0.139	0.085	0.063	0.121	0.123	0.221	0.150	0.207
Nov	-	0.071	0.054	0.068	-	0.185	-	-
Dec	-	0.055	0.041	0.046	-	0.047	0.150	-

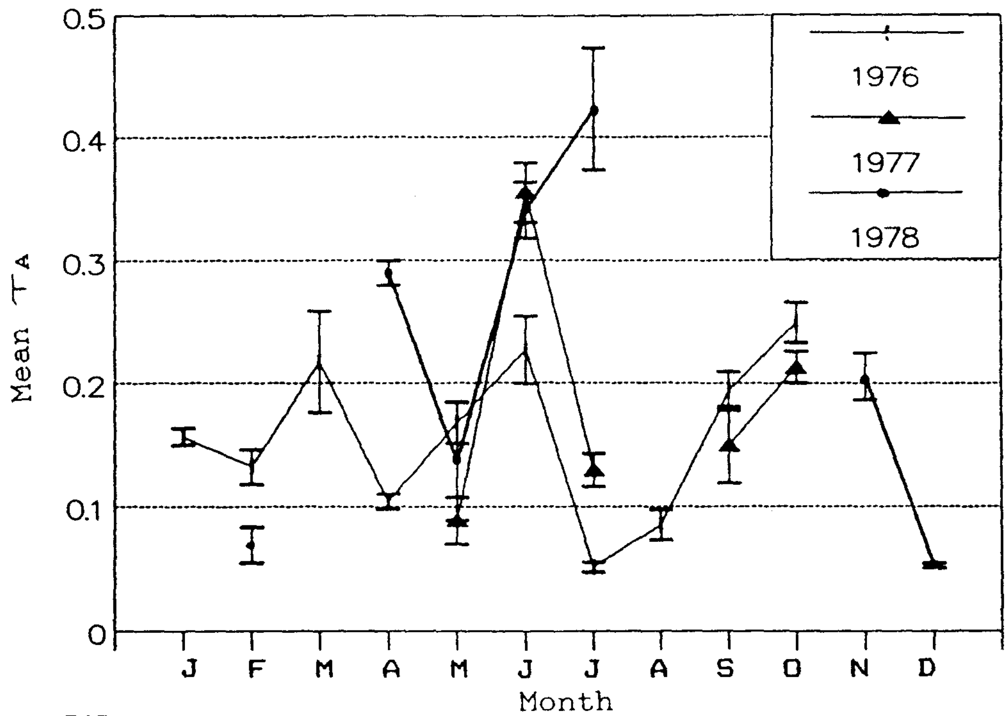

FIG. 8. Monthly variation of mean τ_A for Hamburg, West Germany. The bars indicate plus or minus one standard error.

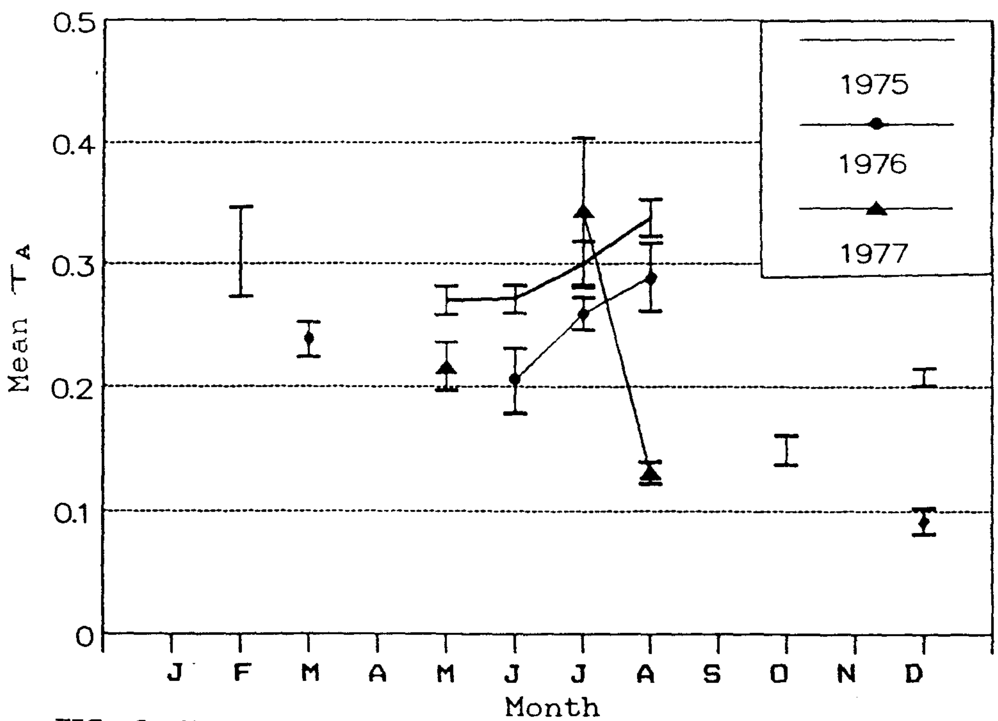

FIG. 9. Monthly variation of mean τ_A for Kew, United Kingdom. The bars indicate plus or minus one standard error.

4.5 Seasonal variation

In order to eliminate the problem of missing data for different months, seasonal turbidity means were calculated. From the monthly mean turbidity values seasonal variations were examined. The data were divided into four seasons as follows spring (March, April, May), summer (June, July, August), fall (September, October, November), and winter (December, January, February).The expected trend from the seasonal data is similar to the monthly data.

Figure 10. shows the characteristic seasonal variation from 1972 to 1974 of Montreal, Quebec. This station displays a summer maximum in all three years and a winter minimum for 1972 (no data for 1973-74). The seasonal variation of Medford, Oregon (Fig.11) is very consistent over the three years with spring maximums and winter minimums. Columbia, Missouri (Fig.12) and Sterling, Virginia (Fig.13) show similar seasonal trends. Both stations have the characteristic trend with summer maximums and winter minimums and Sterling on average has a higher turbidity then Columbia.

The European stations show similar seasonal variations. The seasonal variations for De Bilt, Netherlands (Fig.14) and Zurich, Switzerland (Fig.15) do not display pronounced trends. De Bilt, Netherlands shows spring maximums for the three years but, it should be noted that

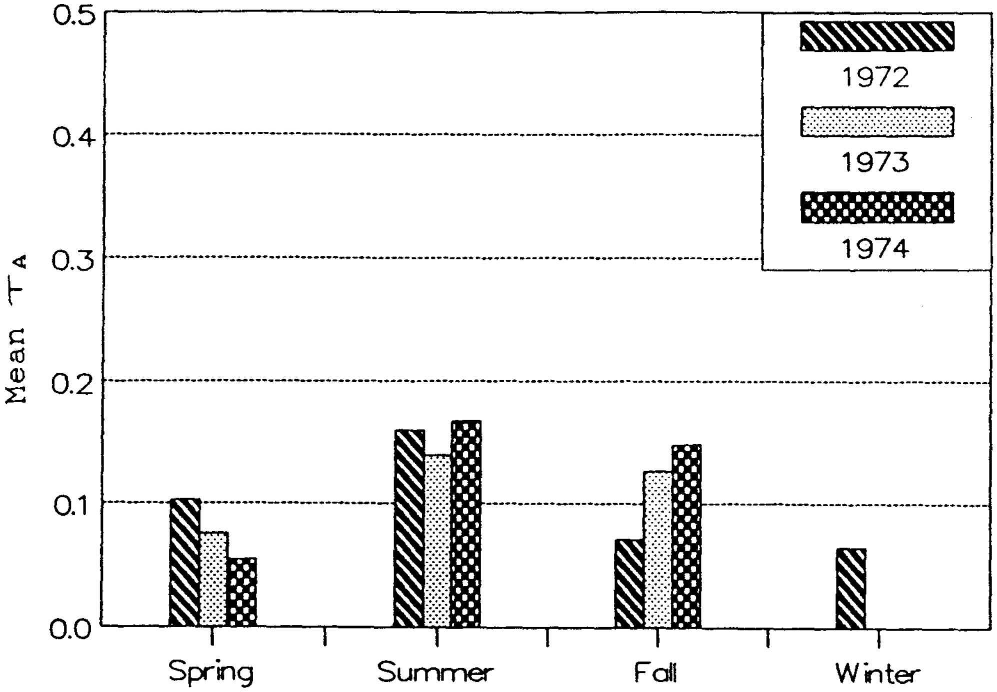

FIG. 10. Seasonal variation of mean τ_A for Montreal, Quebec.

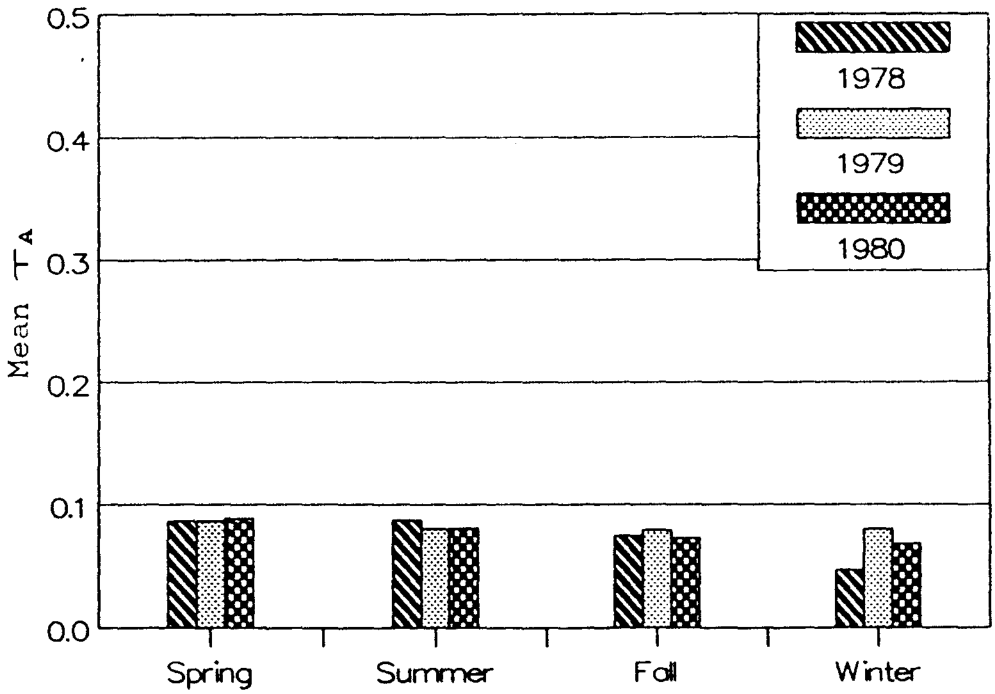

FIG. 11. Seasonal variation of mean τ_A for Medford, Oregon.

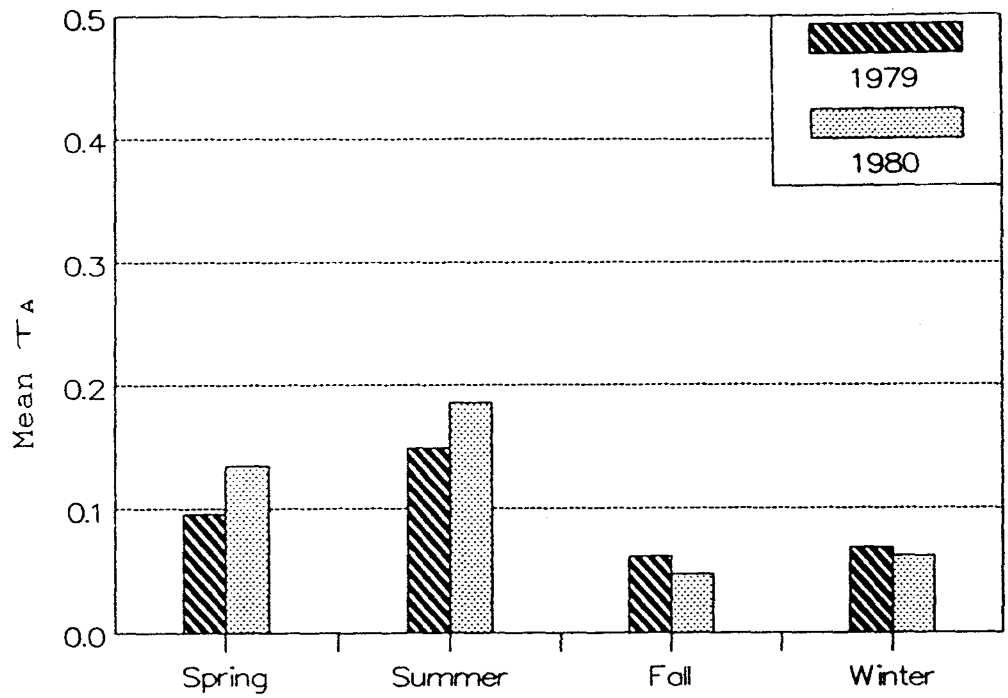

FIG. 12. Seasonal variation of mean τ_A for Columbia, Missouri.

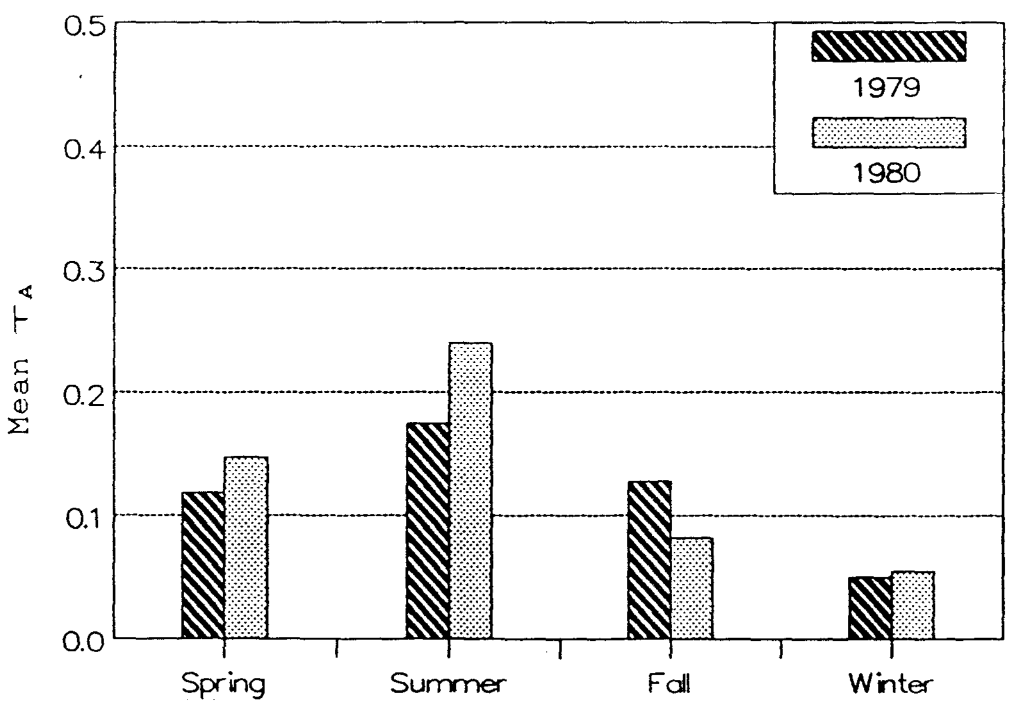

FIG. 13. Seasonal variation of mean τ_A for Sterling, Virginia.

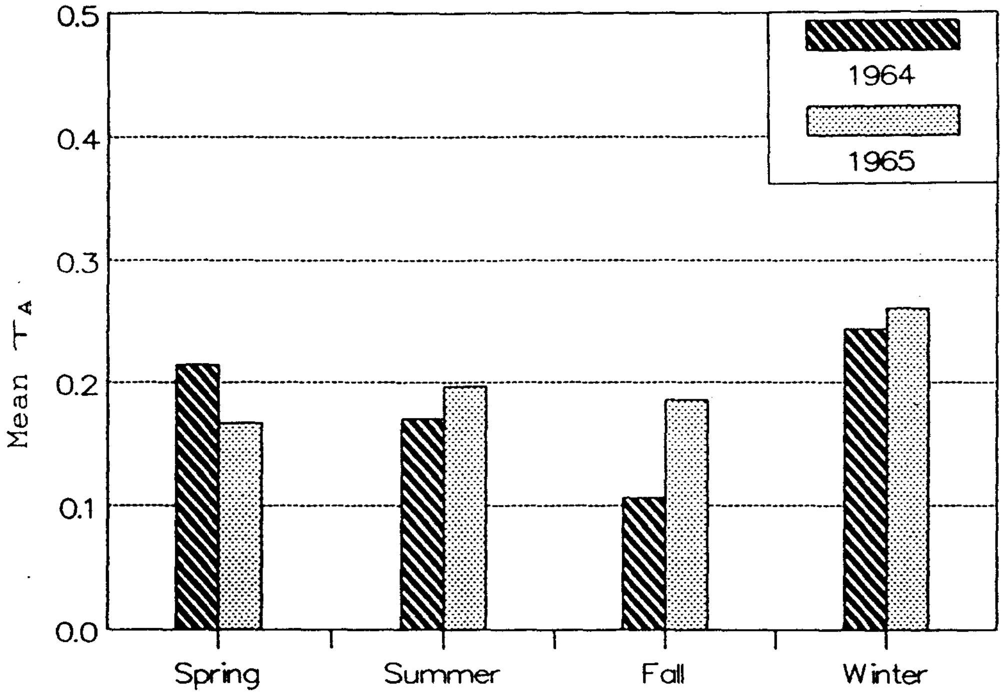

FIG. 14. Seasonal variation of mean τ_A for De Bilt, Netherlands.

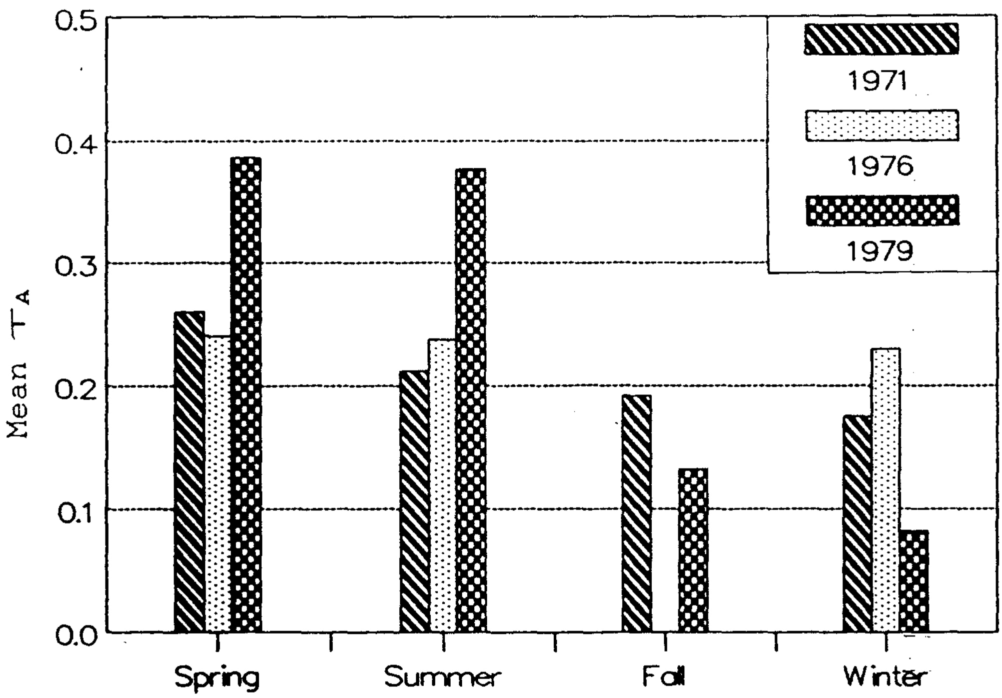

FIG. 15. Seasonal variation of mean τ_A for Zurich, Switzerland.

there were fewer hours available then the summer. The winter shows the characteristic minimum values. The seasonal variation for Zurich, Switzerland shows summer maximums. The winter data was not available for 1964 and the 1965 winter was calculated from a small sample size. The spring 19640 value also had a small sample.

Figure 16 shows the seasonal variation for Hamburg, West Germany. The results indicate summer maximums for 1977-1978 and winter minimums for 1976-1977 (no data for 1978). Turbidity for fall 1976 is highest for that year, but there were only five hours available. The seasonal variation for Kew, United Kingdom (Fig.17) shows summer maximums and winter minimums (no data 1977). The fall season for 1975 had a small sample size which may affect the turbidity calculation.

The averages of the seasonal turbidity means is summarized in Table 4 for each station. Seasonal trends are clearer for the North American stations.

TABLE 4. Average seasonal turbidity means.

Station	Spring	Summer	Fall	Winter
Montreal	0.078	0.156	0.116	0.065
Medford	0.087	0.083	0.076	0.065
Columbia	0.116	0.168	0.055	0.065
Sterling	0.133	0.207	0.105	0.053
De Bilt	0.295	0.262	0.163	0.162
Hamburg	0.144	0.241	0.189	0.095
Kew	0.242	0.285	0.149	0.244
Zurich	0.191	0.183	0.146	0.253

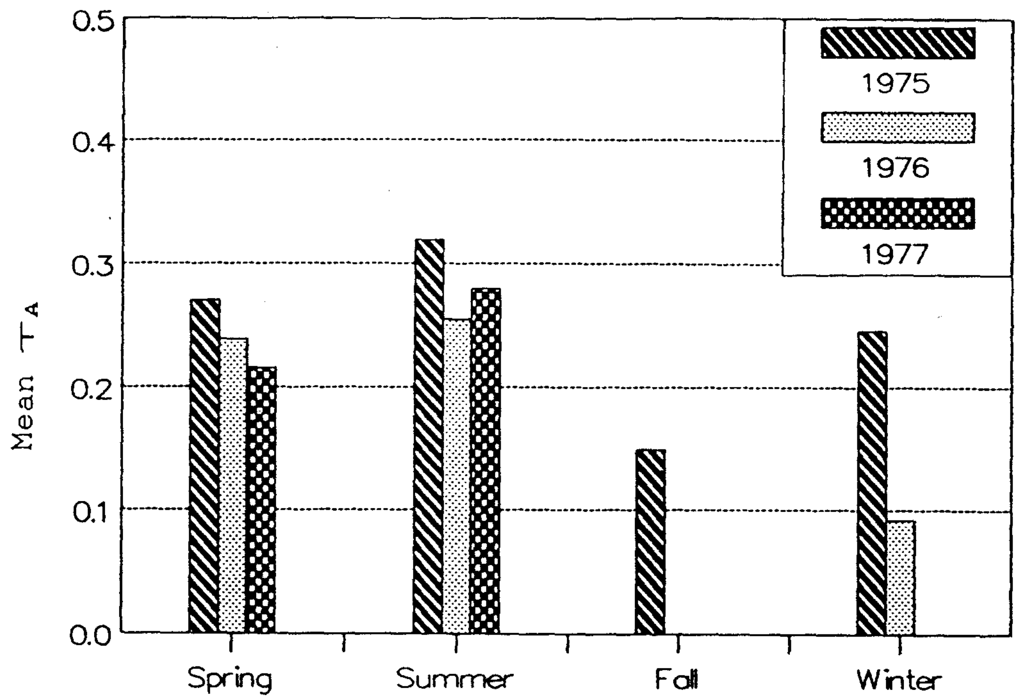

FIG. 16. Seasonal variation of mean τ_A for Hamburg, West Germany.

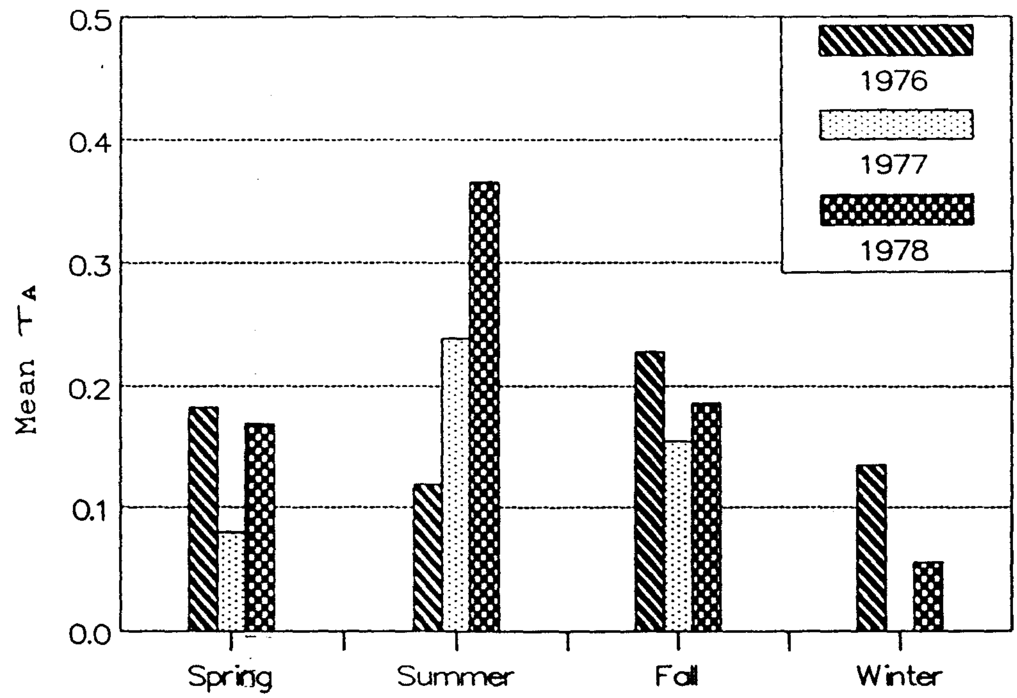

FIG. 17. Seasonal variation of mean τ_A for Kew, United Kingdom.

4.6 Discussion of results

The monthly and seasonal variations for both the North American and European stations compared well with results in the literature. The characteristic summer maximums and winter minimums were evident for most stations. Possible explanations for these trends can be based on Unsworth and Monteith (1972), as mention in 4.2 above.

Local sources of aerosol are important to the high turbidity levels in the summer months. There appears to be a greater amount of pollution from the major urban centres in this study. In the winter months, the mean turbidity values for the urban centres are lower then summer. A comparison of a small and large urban centre such as Medford, Oregon (Fig.11) and Montreal, Quebec (Fig.10) indicate average seasonal τ_A values to be the same in the winter. However in the summer Montreal´s turbidity is almost double that of Medford. This result was also found with comparisons of Medford to Sterling, Virginia and Columbia, Missouri. Turbidity can be heavily influenced by urban effects.

The effect of pollution may have more influence on the European stations for both summer and winter. Most of western Europe is uniformly polluted and turbidity is much higher in summer and winter in comparison to North America.

Air mass origin may be more important to monthly and seasonal turbidity trends. There are three air masses which could affect the eight stations, arctic, polar, and continental, depending on the season. Air mass origin is not important to Medford, Oregon because it would not be affected by air masses that had long trajectories over the continent. The other North American stations are affected by polar and continental air masses from the south in summer and by polar and arctic air masses from the north in the winter. The air masses from the south tend to be more turbid then those from the north. The European stations are influenced by polar and tropical air masses from the west during summer periods and from the north west in the winter months. This would account for the variability in the annual cycles.

Seasonal variations of air masses directly influence the seasonal cycles of turbidity (Uboegbulam and Davies, 1983; Unsworth and Monteith, 1972).

The major obstacle in the analysis was the limited data available for each station. A more stringent control on the data in terms of minimum number of cloudless hours required would have reduced the sample sizes even further. For some months the turbidity values may be inaccurate due to the small sample size and therefore they are not representative of the month. For most stations the standard error of the mean calculated for monthly means tended to increase with τ_A values.

CHAPTER FIVE

CONCLUSION

Turbidity coefficients were calculated for eight North American and European stations. Using hourly values, monthly and seasonal variations were analyzed. European stations produced higher turbidity values then the North American stations. Annual cycles of turbidity displayed summer maximums and winter minimums.

Aerosols have a great effect on solar radiation in urban atmospheres. The effect is most noticeable for the large urban centres. Aerosols affected the turbidity values for most stations through local and distant sources of pollution. Local sources of pollution are very important during the summer for North American stations and throughout the year for European stations. The air mass origin is also an important effect on turbidity.

The attenuation of solar radiation due to aerosols is very significant at both the local and global scales. The most significant effect of increased turbidity is the reduction of solar radiation at the surface leading to a decrease in mean surface temperature on the earth.

Further study is needed to gain a better understanding of the direct effects of aerosols on the attenuation of solar radiation in urban atmospheres.

APPENDIX ONE

FORTRAN 77 Program used to calculate the turbidity values for each station. Written by Dr. J.A. Davies, the Astro subroutine is from Michalsky (1988).

```
c  Turbidity analysis
     IEA cloudless sky radiation data

     REAL lat, leap, jd, daymon (12)
     CHARACTER*20 infil, outfil, stanam
     D    A    T    A    d    a    y    m    o    n
/31.,28.,31.,30.,31.,30.,2*31.,30.,31.,30.,31./
     DATA solcon,oz /1367.0, 3.5/
     pi=2.0*acos(0.0)
      conv=pi/180.

c  Get names of input and output files
     write(*,*)' Enter input filename-------> '
     read*, infil
     write(*,*)' Enter output filename------> '
     read*, outfil
     OPEN (1,file = infil)
     OPEN (2,file = outfil)

c  Read data
     READ (1,*) stanam, lat, iyr
     WRITE (2,'(1x,a20, f10.2)') stanam, lat
     slat = lat*conv

     DO 1000 k =1,1000

     R              E              A              D
(1,'(4x,2i2,2x,f5.2,3f4.0,f4.1,f5.2,1x,3f4.1)',end=2000)mon,
     &    iday,st,g,d,s,dbt,stp,dpt,sh,vis

      IF (s .le. 0.0 .or. sh .lt. 1.0) GOTO 1000
     IF (mod(iyr,4) .eq. 0) daymon(2) = 29.

c  Compute astronomical parameters in subroutine ASTRO for 0
LAT
c
+-----------------------------------------------------------+
c    |Uses the algorithm given by J. Michalsky (1988) in
Solar Energy, V40,   |
c   |227-234.
         |
c   |    day    :  day of the month
         |
c   |    rat    :  ratio of actual to mean sun-earth  distance
```

```
        ¦
c   ¦   dec    :  solar  declination
        ¦
c   ¦   et     :  equation of time.
        ¦
c
+-----------------------------------------------------------+
     hour = 0.

c Get julian date
      sum=0.0
      DO 1 i=1,mon-1
    1     sum=sum+daymon(i)
      day=sum+iday
     delta=iyr-1949.
     leap=aint(delta/4.)
     jd=32916.5+delta*365.+leap+day+hour/24.
     time=jd-51545.0

     CALL ASTRO (time,rat,et,dec)

     rsq = (1./rat)**2

c     Solar  zenith  angle(deg),  optical   air  mass   and
extraterrestrial
c  solar radiation.
     CALL ZEN (dec,slat,conv,st,cz,z,am,stp,amc)

c  Precipiable water.
     CALL PRECIP (dbt,dpt,stp,pw)
     IF (pw. lt. 0.) GOTO 1000

c  Transmission functions.
     CALL TRANS (amc,pw,oz,aw,to,tr)

c  Turbidity
     CALL TAU (solcon,rsq,cz,aw,to,tr,s,am,taua)

c  Print results
     WRITE
(2,'(1x,i4,2i3,6f10.4)')iyr,mon,iday,rat,dec/conv,
     & 4*(et/conv),st,am,z
      WRITE (2,'(1x,5f10.4)')g,d,s,vis,taua

1000 CONTINUE

2000 STOP
     END
```

```
c***********************************************************
      SUBROUTINE ASTRO (time,rat,et,dec)
      IMPLICIT REAL (a-z)
       pi=2.0*acos(0.0)
      twopi = 2.*pi
       conv=pi/180.

c ecliptic coordinates
      mnlong=280.46+.9856474*time
      mnlong=mod(mnlong,360.)
      IF (mnlong .lt. 0.) mnlong=mnlong+360.

c mean anomaly in radians between 0 and 2*pi
      mnanom=357.528+.9856003*time
      mnanom=mod(mnanom,360.)
      IF (mnanom .lt. 0.) mnanom=mnanom+360.
      mnanom=mnanom*conv

c ecliptic longitude and obliquity of ecliptic in radians
      eclong=mnlong+1.915*sin(mnanom)+0.02*sin(2.*mnanom)
      eclong=mod(eclong,360.)
      IF (eclong .lt. 0.) eclong=eclong+360.
      oblqec=23.439-4.0e-7*time
      eclong=eclong*conv
      oblqec=oblqec*conv

c right ascension,declination,equation of time and the ratio
of actual
c to mean sun-earth distance
      num=cos(oblqec)*sin(eclong)
      den=cos(eclong)
      ra=atan(num/den)
      rat=1.00014-0.01671*cos(mnanom)-0.00014*cos(2.0*mnanom)

      IF (den .lt. 0.) THEN
           ra=ra+pi
      ELSE IF (num .lt. 0.) THEN
           ra=ra+twopi
      ENDIF
      et=mnlong*conv-ra
      dec=asin(sin(oblqec)*sin(eclong))

      RETURN

      END
c***********************************************************
      SUBROUTINE ZEN (dec,slat,conv,st,cz,z,am,stp,amc)
       aa=sin(dec)*sin(slat)
       bb=cos(dec)*cos(slat)
       ha=abs(12.-st)*15.*conv
```

```
      cz=aa+bb*cos(ha)
     IF (cz .gt. 1.0) cz=1.0
      z=acos(cz)/conv
      am=35./(sqrt(1224.*cz*cz+1.))
      IF (am.gt.30.0) am=30.0
      IF (stp.lt.0.0) stp=101.3
      amc=am*stp/101.3
      IF (amc.gt.30.0) amc=30.0

      RETURN
      END
c************************************************************
     SUBROUTINE PRECIP(dbt,dpt,stp,pw)
c
+----------------------------------------------------------+
c     |Precipitable  water formula contains  numerical values
determined |
c     |by  pooling  data  from  four  stations  (Davies  and
Abdel-Wahab, 1983|
c   |, p.14.)
        |
c
+----------------------------------------------------------+
      IF (dbt.lt.-99.) THEN
        ta=273.0
      ELSE
        ta=dbt+273.0
      ENDIF
      IF (dpt.gt.-9.0) THEN
        pw=exp(2.2572 + 0.05454*dpt)
        pw=pw*(stp/101.3)**.75*(273./ta)**0.5
      ELSE
        pw=-99999.0
      ENDIF

     RETURN
     END
c************************************************************
     SUBROUTINE TRANS (amc,pw,oz,aw,to,tr)

c  Ozone transmission
      x=oz*amc
      aa=2.118e-3*x/(1.+4.2e-3*x+3.23e-6*x*x)
      bb=0.1082*x/(1.+13.86*x)**0.805
      c=0.00658*x/(1.+(10.36*x)**3)
      ao3=aa+bb+c

      to=1.-ao3
```

```
c  Absorption by water vapour
      x=pw*amc
      aw=0.29*x/((1.+14.15*x)**0.635+0.5925*x)

c   Transmittance due  to Rayleigh  scattering is  estimated
from a
c  second-order logarithmic regression with 0.2% max error:
c
      x=8.688237*amc**(-0.806955 + 0.0279286*alog(amc))
      tr=x / (1.0+x)

      RETURN
      END
c************************************************************
      SUBROUTINE TAU (solcon,rsq,cz,aw,to,tr,s,am,taua)

      REAL i,istar

      i = s/(3.6*cz)
      istar = solcon*rsq*(to*tr-aw)
      taua = alog(istar/i)/am

      RETURN
      END
c************************************************************
```

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