THE EFFECT OF EL CHICHÓN ON ATMOSPHERIC TURBIDITY AT WOODBRIDGE

by

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ABSTRACT

Monthly median and annual mean values of optical depth and the ratio of diffuse to direct solar radiation for 1981-1983 were calculated using integrated values of global and diffuse radiation and calculations of precipitable water, under cloudless conditions. Results indicate that El Chichón's volcanic dust cloud has affected turbidity over southern Ontario. This is reflected in an increase in optical depth and the ratio of diffuse to direct solar radiation.

ACKNOWLEDGEMENTS

I am deeply indebted to my supervisor, Dr. John Davies for giving me the idea for this thesis and for his continued guidance until its completion. My thanks are also extended to Dr. Magdy Abdelwahab for his generous advice and help with the computations. Finally, I thank God whose grace continually supports me.

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INTRODUCTION

The volcano El Chichón erupted in southeast Mexico ($17.33^{o}N$, $93.2^{o}W$) during late March, early April of 1982, ejecting $3.3 \times 10^{12}g$ of sulphur dioxide into the stratosphere (Barth et al, 1983). By early fall, the resultant volcanic dust cloud had moved north and may have affected turbidity over eastern Canada.

Volcanic aerosols cause an increase in atmospheric turbidity. Dyer (1974) found that volcanic eruptions cause short-term (1-2 years) increases in global turbidity. The increase can be calculated from the attenuation of direct beam solar radiation and the change in diffuse solar radiation.

The eruptions of Mt. Agung (1963) and Mt. St. Helens (1980) have caused detectable short-term increases in turbidity (Pollack et al, 1976, DeLuisi et al, 1983). The eruption of El Chichón is considered to be greater than that of Mt. Agung and Mt. St. Helens but it is not as great as that of Krakatoa (1885).

The purpose of this thesis is to determine the affect of El Chichón's dust cloud on atmospheric turbidity at Woodbridge.

THEORETICAL BACKGROUND

In this study, two indices of turbidity are used; Unsworth and Monteith's (1962) aerosol attenuation coefficient τ_A, which is related to optical depth and the ratio of diffuse to direct beam solar radiation, D/I. Optical depth is derived from Beer's Law which defines the extinction of the solar beam through the atmosphere as

$$\tau_\lambda = - m^{-1} \ln [I_\lambda \;/ I_\lambda(o)] \qquad -(1)$$

where τ is optical depth at wave length λ, m is the optical airmass, I_λ is the measured spectral direct beam radiation flux density at the ground and $I_{\lambda\,(o)}$ is its unattenuated value at the top of the atmosphere.

Letting κ and σ be spectral mass absorption and scattering coefficients, can be written as

$$\tau_\lambda = \int_0^{zt} (\kappa_\lambda + \sigma_\lambda)\, dz \qquad -(2)$$

where Zt is the height of the aerosol atmosphere, It is the sum of four main components:

$$\tau_\lambda = \left| \tau_{(\lambda)R} + \tau_{\lambda(o)} + \tau_{\lambda(w)} + \tau_{\lambda(a)} \right| \qquad -(3)$$

where R, w, o represent Rayleigh scatter, water vapour absorption, ozone absorption and aerosol attenuation respectively. Rearranging (1),

$$I_\lambda = I_{\lambda\,(o)} [\exp^{-(T_{R\lambda} + T_{o\lambda} + T_{w\lambda})} \exp^{-T_\lambda m}] \qquad -(4)$$

and

$$I_\lambda = I_\lambda^* \quad e^{-\tau_{a_\lambda} m} \qquad -(5)$$

where $$I_\lambda^* = I_\lambda \ (o) \exp \left[- (T_{R\lambda} + To_\lambda + Tw_\lambda \ ) \ m\right] \qquad -(6)$$

in this way, τ_{a_λ} is separated from the other components.

Therefore,

$$e^{-\tau a_\lambda m} = I_\lambda / I_\lambda^* \qquad -(7)$$

integrating over wavelength,

$$\exp^{-\tau_{Am}} = \frac{\int I_\lambda \, d_\lambda}{\int I_\lambda^* d_\lambda} = \frac{\int I_\lambda^* \quad e^{-\tau a_\lambda m} \, d_\lambda}{\int I_\lambda \quad d_\lambda}$$

This shows that τ_A is a weighted mean aerosol optical depth. Davies and Stewart (1984) showed that τ_A and $\tau_\lambda = 0.69$ m are well correlated and of very similar magnitude.

τ_A has been used more frequently than D/I to measure turbidity. Using τ_A, Davies and Uboegbulam (1983) studied turbidity over eastern Canada. Uboegbulam (1981) also used τ_A when studying the attenuation properties of aerosols.

Because τ_A is calculated after the effects of other attenuating factors have been removed, it is superior to other turbidity indices such as Kondratyer's atmospheric transparency (Kondratyev, 1969) and Linke's turbidity factor (Linke, 1942).

The D/I ratio has not been widely used. It was suggested by Deirmendjian (1980) as a very sensitive measure of turbidity since as turbidity increases, D increases, I decreases and therefore D/I varies. Wesely and Lipschutz (1976) used this index in a study of the effects of aerosols on solar radiation and found that it is a very sensitive indicator of turbidity.

DATA ACQUISITION AND ANALYSIS

Hourly integrated values of measured global and diffuse solar radiation along with hourly observations of drybulb and dewpoint temperature, atmospheric pressure and cloud amount were obtained from the Atmospheric Environment Service for Woodbridge, Ontario (43.30°N, 79.23°W).

Direct beam solar radiation was calculated as the difference between global and diffuse radiation. The accuracy of this determination is important since error in direct beam radiation will introduce an error in the calculated value of τ_A. To assess the accuracy of direct beam radiation, hourly values were compared with direct pyrheliometric measurements for Woodbridge for 1981 and 1982. The two sets of values were well correlated ($r = 0.94$). The mean bias error was 3.4%. For similar data, Uboegbulam and Davies (1983) found that the residual was very well correlated with the measured direct beam, with a standard deviation of 3%.

Probable absolute error in τ_A was determined using the method of propagation of errors by Bevington (1969) where:

$$\Delta\tau_A^{\,2} = \left(-\frac{1}{m}\ \frac{\Delta I}{I}\right)^2 + \left(-\frac{1}{m}\ \frac{\Delta Io}{Io}\right)^2 \qquad -(9)$$

Values of 0.034 and 0.0 (Davies and Hay, 1980) were used for $\frac{\Delta I}{I}$ and $\frac{\Delta Io}{I}$ respectively. The error in τ_A was 0.013. Error in direct beam radiation estimates contributed to most of the error τ_A. This amounted to 0.125 whereas for ΔIo it was 0.003. This value is well within the probable error of direct beam measurement.

The error in D/I was calculated in a similar way:

$$\left(\Delta D/I\right)^2 = \left(\frac{\Delta D}{D}\right)^2 + \left(\frac{\Delta I}{I}\right)^2 \qquad -(10)$$

There is no accepted standard measurement of D. Error in D is likely to be greater than error in I. In this study, $\Delta D/D$ is assumed to be 0.05. The errors in the measurement of I and D therefore, cause a probable absolute error of 0.084 in D/I.

Following Paltridge and Platt (1976), Io values were calculated from:

$$Io = [\, S(Tr_{(m)}\ To_{(m,u_o)}) - Aw\ (m,u_w)] \qquad -(11)$$

where S is the solar constant corrected for departure of the sun-earth distance from the mean value; To and Tr are transmittances of ozone absorption and Rayleigh scatter, A_w is water vapour absorption, m is optical airmass and u_o and u_w are the atmospheric path length of ozone and water respectively. Davies and Hay (1980) show that use of this equation gives results that are within 1% of those obtained from Braslau and Dave's (1973) detailed radiative transfer calculations for an aerosol free atmosphere.

Following Thekawkara and Drummond (1971) 1353 Wm^{-2} was used as the solar constant when calculating Io. Io was also adjusted to represent the extraterrestrial irradiance on a horizontal surface by:

$$Io = Io^{*} \cos\theta \qquad -(12)$$

where $\cos\theta$ is the zenith angle and is calculated from

$$\cos\theta = \sin\phi \sin\delta + \cos\phi \cos\delta \cos H \qquad -(13)$$

in which ϕ is the station latitude, δ is solar declination and H is the hour angle (degrees) given by

$$H = 15 \left| 12 - LAT \right| \qquad -(14)$$

LAT, local apparent time is determined by

$$LAT = LST + ET/60 + (LSM - LS)/15 \qquad -(15)$$

where LST is local standard time, ET is the equation of time (in minutes) and LS and LSM are longitudes of the station and the standard meridian appropriate to the time zone respectively.

Values of δ and ET were calculated from

$$\delta = 0.006918 - 0.399912 \cos\theta_o + 0.070257 \sin\theta_o - 0.006759 \cos\theta + 0.000907 \sin\theta_o - 0.002697 \cos\theta_o 3 + 0.001480 \sin\theta_o 3 \qquad -(16)$$

and $ET = 0.000075 + 0.001868 \cos\theta_o - 0.032077 \sin\theta_o - 0.14615 \cos\theta_o - 0.040840 \sin\theta_o 2 \qquad -(17)$

Atmospheric water pathlength required to obtain water vapour absorptivity was calculated using Won's (1977) empirical formula:

$$U'w = \exp(2.2572 + 0.05454\ Td) \qquad -(18)$$

where Td is the dewpoint temperature in Celsius and U'w values are in millimetres. Values of U'w are corrected for dependence on temperature and pressure after Paltridge and Platt (1976) by

$$U'w - U'w\ (P/Po)^{\frac{3}{4}}\ (To/T)^{\frac{1}{5}} \qquad -(19)$$

where T is surface temperature in ^{o}K, To is standard sea level temperature, $273^{o}K$, P is station pressure (kPa) and Po is standard sea level pressure 101.3 kPa. Hourly dewpoint and drybulb temperatures and station pressure are used to estimate U'w.

Atwater and Ball (1976) showed that the precipitable water estimated from a dewpoint based formula introduced little error in estimates of global irradiance. Uboegbulam and Davies showed that error in τ_A due to error in estimating precipitable water using Won's (1977) formula ranged between 0.069 and 0.037 for dewpoint temperatures greater than or equal to $-20^{o}C$, for Woodbridge. The absorptivity of water vapour is calculated from Lacis and Hansen's (1974) formula

$$Aw = 0.29X_2/[(1 + 14.15X_2)^{0.635} + 0.5925X_2] \qquad -(20)$$

where $X_2 = M_{uw}$

The relative optical airmass (m) is calculated from Roger's (1967) formula

$$m = 35/(122\ \mathrm{Cos}\,\theta^{\ 2} + 1)^{\frac{1}{2}}$$

This formula allows for refraction effects at large zenith angles. Multiplying by P/Po corrects m for atmospheric pressure. In caclulating
Tc

τ_A, airmasses greater than 3 were neglected to eliminate large zenith angles. For large zenith angles, small errors in measurement time and variations in the vertical distribution of aerosol can introduce significant errors to airmass calculations (Thomason et al, 1982).

The transmissivity of ozone is also calculated from a formula given by Lacis and Hansen (1974)

$$T_o = 1 - a_o \tag{22}$$

$$a_o = \frac{0.1082X_1}{(1 + 13.86X_1)^{0.805}} + \frac{0.00658X_1}{1 + (10.36X_1)^3} + \frac{0.002118X_1}{1 + 0.0042X_1 + 0.00000323X_1^{\,2}}$$

where X_1 = mu.

A constant value of 3.5 mm was assumed in all the calculations. The error introduced by using a constant value for ozone is negligible because ozone transmittance is insensitive to large variations in ozone amounts (Uboegbulam and Davies, 1983).

Transmissivity after Rayleigh scattering was based on calculation given in Davies and Hay (1980). Cloudless sky data was used in all instances.

RESULTS AND DISCUSSION

τ_A

Median values of τ_A were calculated for 1981-1983 since distributions of τ_A were skewed. These were plotted in Figure 1. In the pre-El Chichón period, maximum values occur in summer and minimum values in winter. This variation is similar to that found by Uboegbulam and Davies (1983) for Montreal and Woodbridge and by Hay (1983) for Vancouver. It is typical of southern Canada and is related to the dominant airmass type (Uboegbulam and Davies, 1983).

The largest value of τ_A for the period under study was recorded in July 1982, four months after the eruption. However, this is not due to El Chichón although Hay (1983) stated that the effects of the volcanic aerosol may have been felt as early as one month after the eruption and at least by July 1982. Other evidence suggests that the body of the dust cloud did not reach southern Canada until about November 1982 (Pollack et al, 1983).

It is also unlikely that the "mystery cloud" that appeared shortly before El Chichón's eruption (DeLuisi et al, 1983) due to an earlier unrelated volcanic explosion could have caused this anomalously large July value because this cloud was smaller than the El Chichón cloud. The July 1982 median value was calculated from only eleven hours of data; six of these were greater than 0.268. They represent extreme tropospheric aerosol conditions due to existing weather conditions.

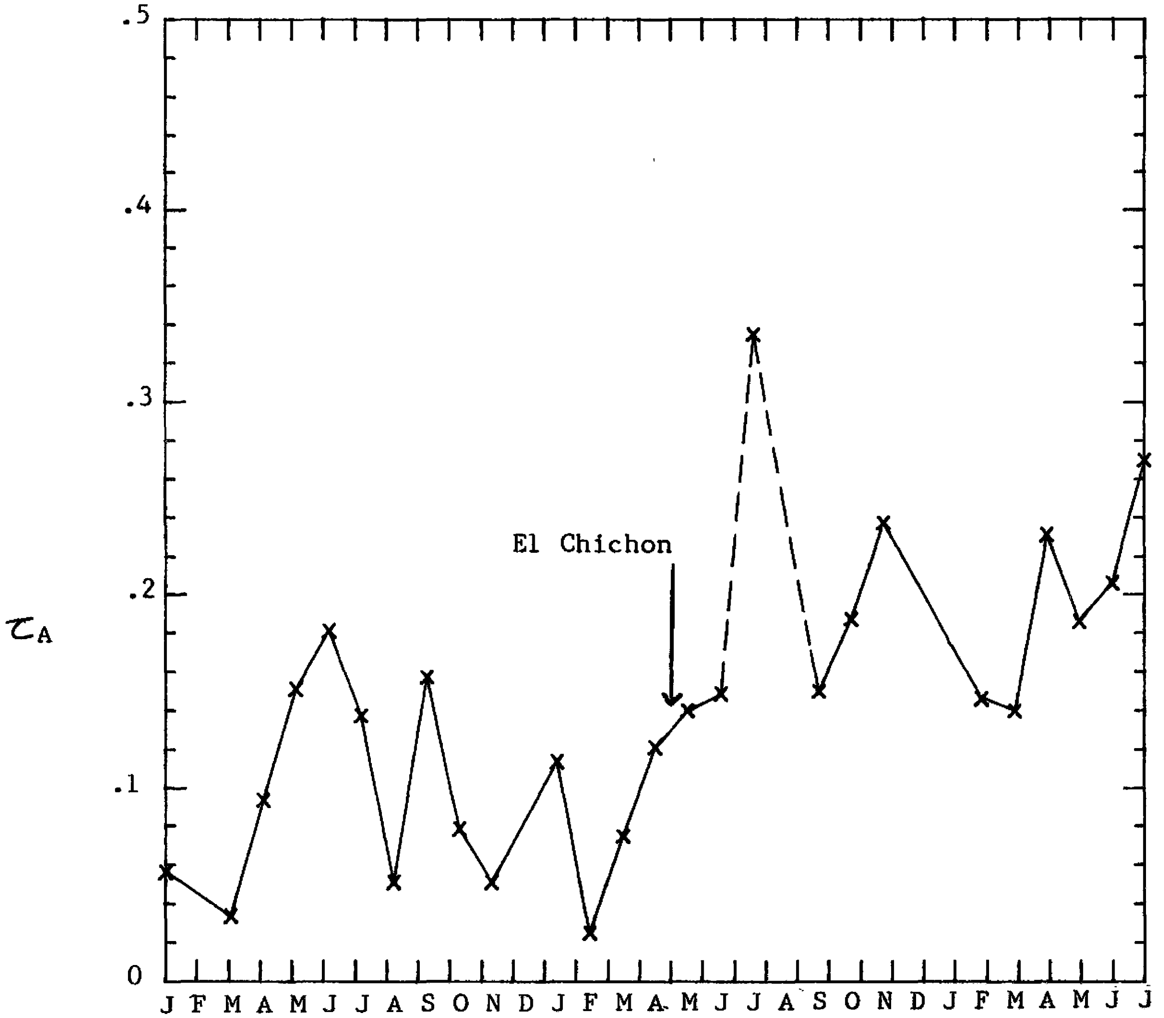

Figure 1: Monthly Variation of τ_A

Table 1

MONTHLY MEDIAN VALUES OF τ_A AND D/I

	1981				1982				1983			
Month	#Obs.	τ_A	#Obs.	D/I	#Obs	τ_A	#Obs.	D/I	#Obs.	τ_A	#Obs.	D/I
January	22	.045	25	.184	5	.091	5	.255				
February	-	-	7	.113	12	.020	14	.179	15	.117	15	.418
March	5	.027	9	.115	25	.060	25	.179	44	.112	44	.283
April	23	.075	29	.153	51	.097	53	.168	13	.185	13	.304
May	42	.121	43	.235	15	.112	15	.198	18	.149	18	.270
June	14	.145	14	.309	15	.119	8	.300	47	.165	47	.415
July	46	.110	51	.210	11	.268	11	.590	43	.216	43	.427
August	16	.041	17	.221	-	-	-	-				
September	9	.126	9	.210	9	.120	9	.245				
October	14	.063	15	.160	19	.150	20	.251				
November	5	.041	6	.186	9	.190	9	.367				
December	-	-	-	-	-	-	-	-				

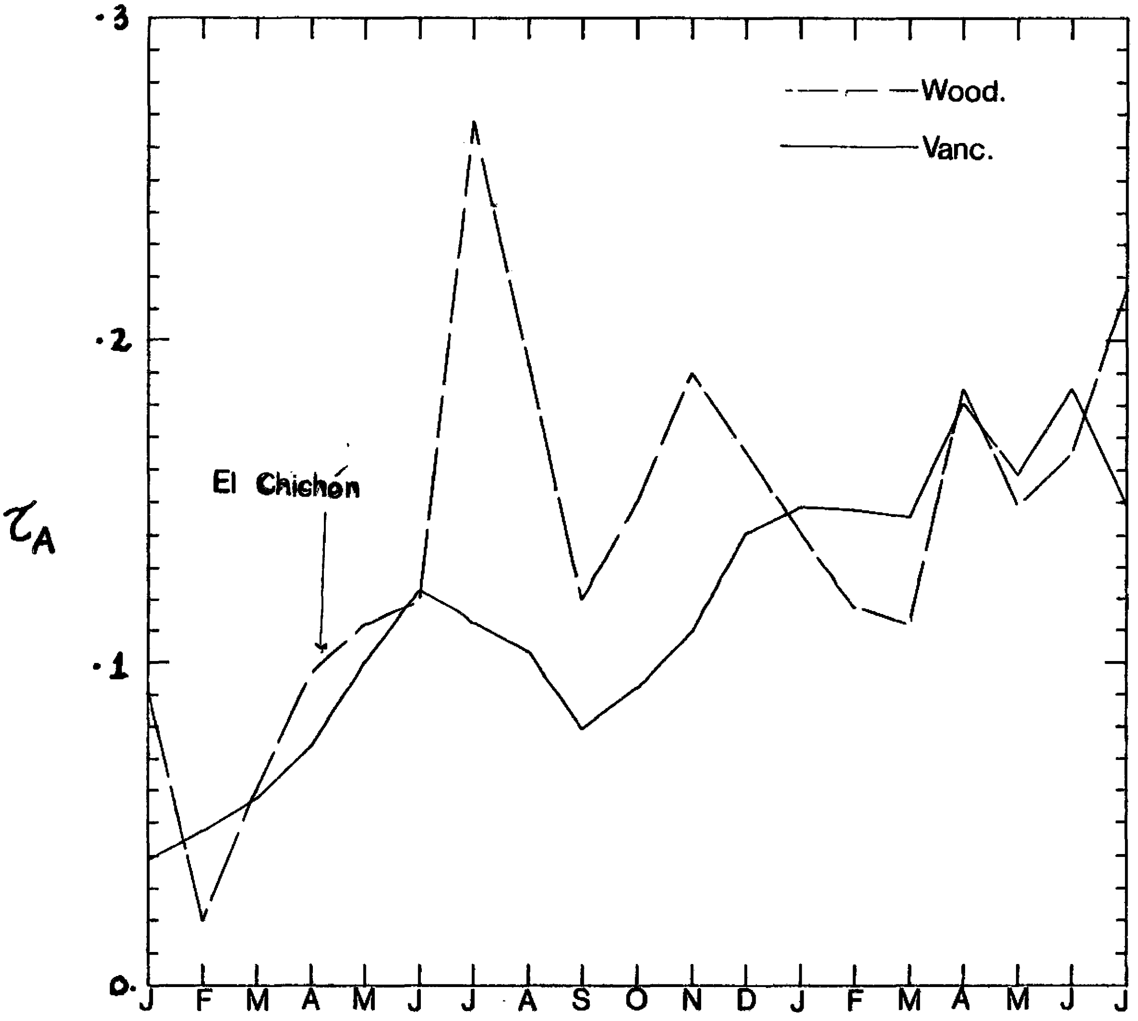

Figure 2 : Monthly variation of τ_A for Woodbridge and Vancouver

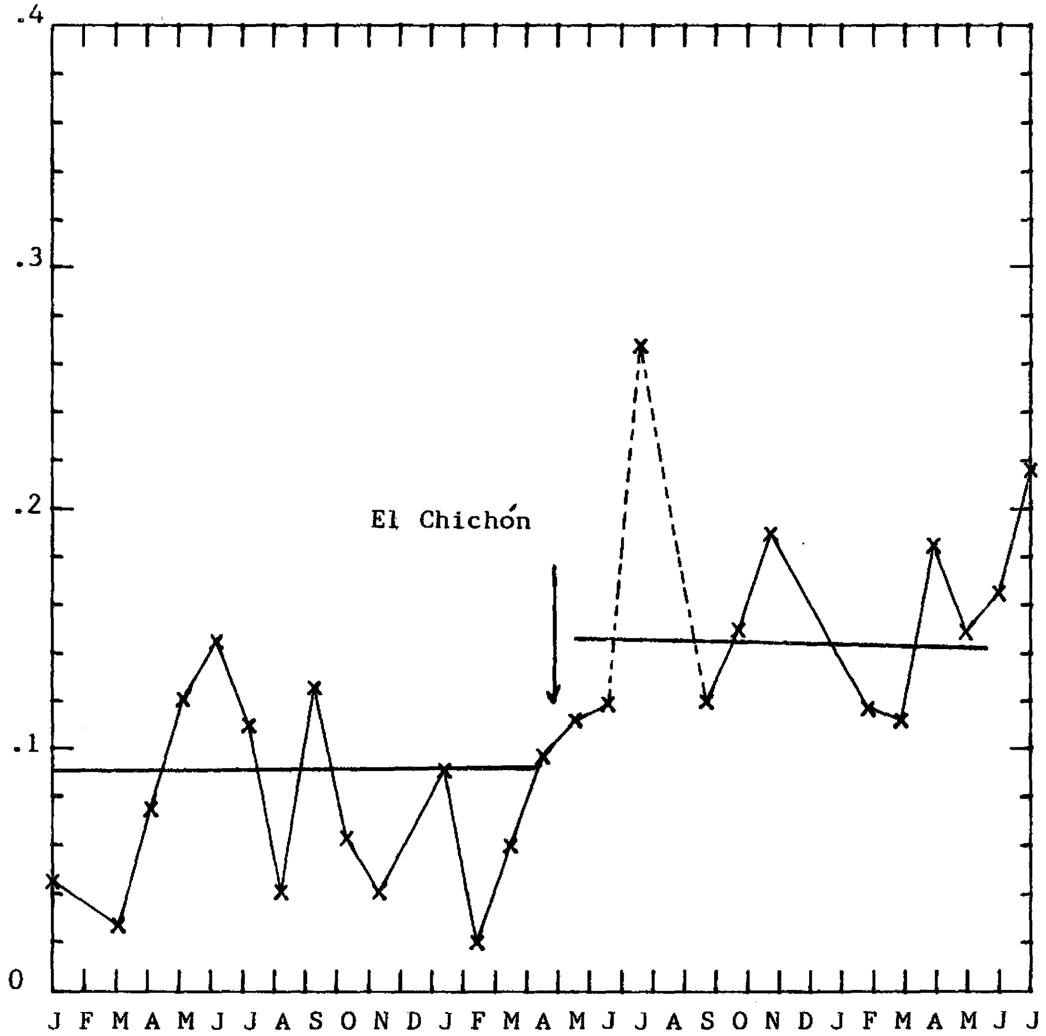

Figure 3: Trend of τ_A

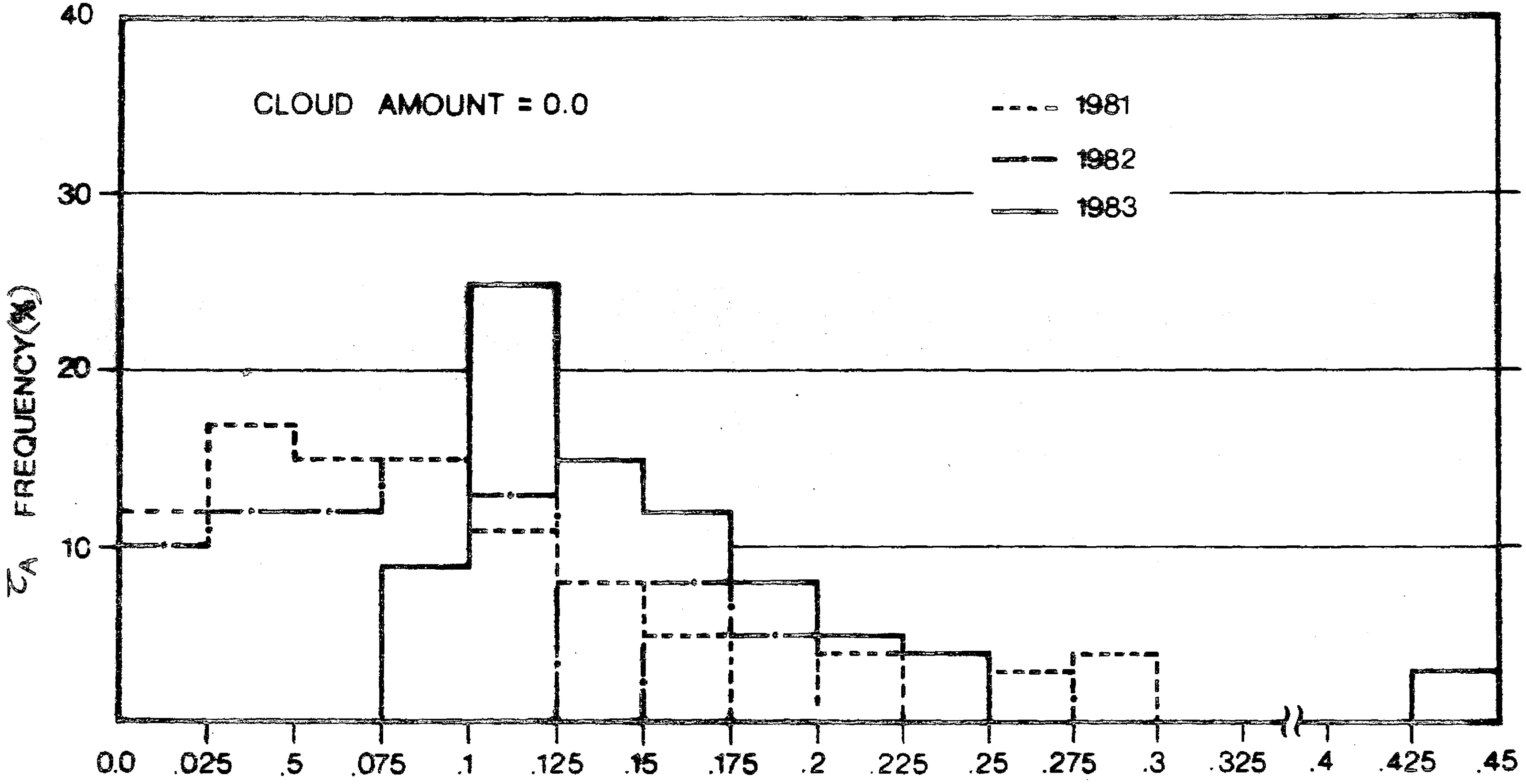

Figure 4. Frequency of occurrence of τ_A values for 1981 – 1983.

τ_A values increase from September to a maximum in November 1982 which is opposite to the normal seasonal cycle (Figure 1). This suggests influence from the El Chichón cloud which reached these latitudes at this time (Pollack et al, 1983). This 'arrival time' is one month later than the October peak estimated by Hay (1983).

τ_A values continued to increase towards a summer maximum. They are higher than the corresponding values in 1981 and 1982. However, several very turbid days in May and June due to large amounts of tropospheric aerosol enhanced the τ_A values (Davies et al, 1984).

In general, results in Figure 1 can be interpreted either as a linear increase in τ_A or as a step increase; the step occurring shortly after the volcanic eruption (Figure 3). An increase in τ_A would be expected after the eruption but the nature of the increase could be either gradual or stepwise. Although there is a difference in the estimated peaks, variation in τ_A for Vancouver is similar to that for Woodbridge (see Figure 2). Annual average τ_A values for 1981, 1982 and 1983 are 0.147, 0.142 and 0.213 respectively. These suggest an increase of about 0.066 in τ_A due to El Chichón. This value is in good agreement with the value of 0.06 obtained for Hamilton by Davies et al, 1984, and with aircraft surveys by DeLuisi (1983). Figure 4 also supports the idea of a stepwise increase. There is an increase in the frequency of occurrence of large τ_A values.

D/I

Figure 5 shows the monthly variation of median values for D/I. Variations in D/I and τ_A are similar (Figure 6). However, fluctuations in D/I are larger, which indicates that D/I is a more sensitive measure of turbidity than τ_A.

Although the curve suggests that the main impact of the El Chichón dust cloud occurred in February 1983, this may not have been the case because there were no data for the two preceding months. The high values in May and June may also be related to the high tropospheric turbidity noticed by Davies et al (1984) in May and June, 1983.

The annual average D/I values were 0.279, 0.305, and 0.477 for 1981, 1982, and 1983 respectively. Figure 7 indicates that this may be a combination of a linear and a step increase. The trend before El Chichón's eruption was relatively constant. However, shortly after the eruption, the values of D/I experienced an abrupt increase which then continued linearly.

Figure 8 shows an increase in the frequency of occurrence of large D/I values. It is reminiscent of τ_A (Figure 3).

Since Figures 1 and 4 show similarities between τ_A and D/I, the relationship between the two turbidity indices was investigated. A scatter plot of τ_A vs D/I (Figure 9) suggests a positive relationship as expected, but this relationship is not very strong and not linear. The linear correlation between the indices was only 0.13.

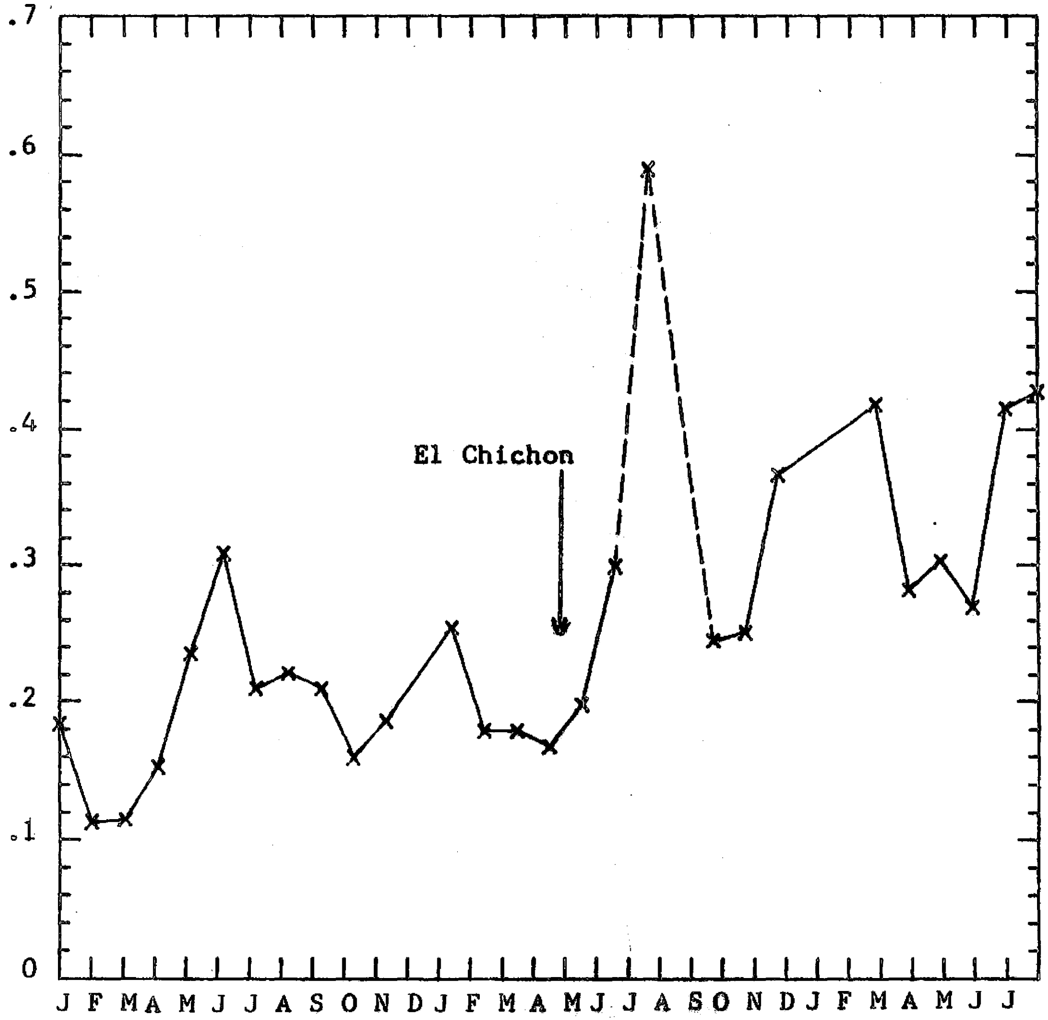

Figure 5: Monthly Variation of D/I

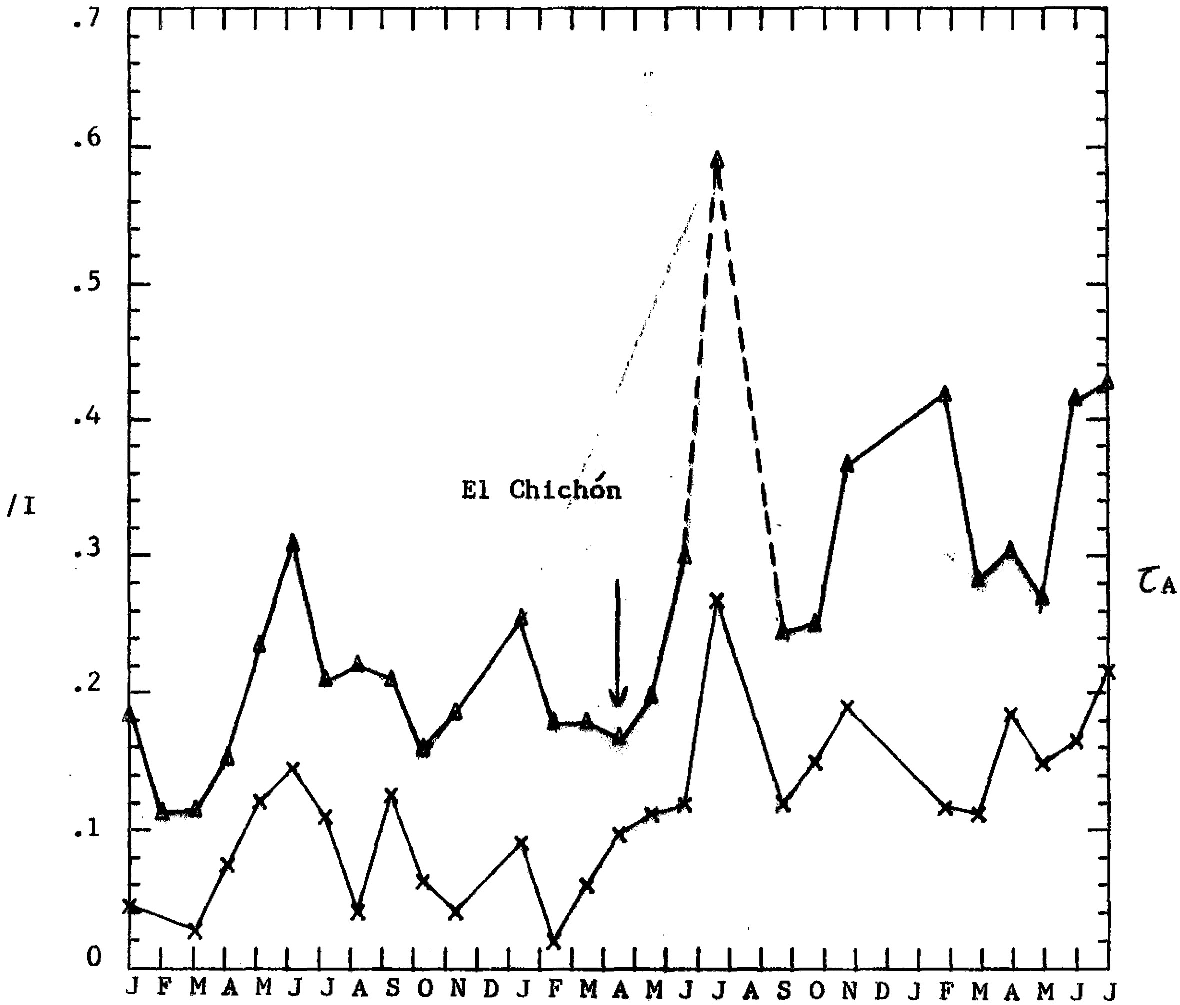

Figure 6: Comparison of the variations and magnitude of D/I and τ_A.

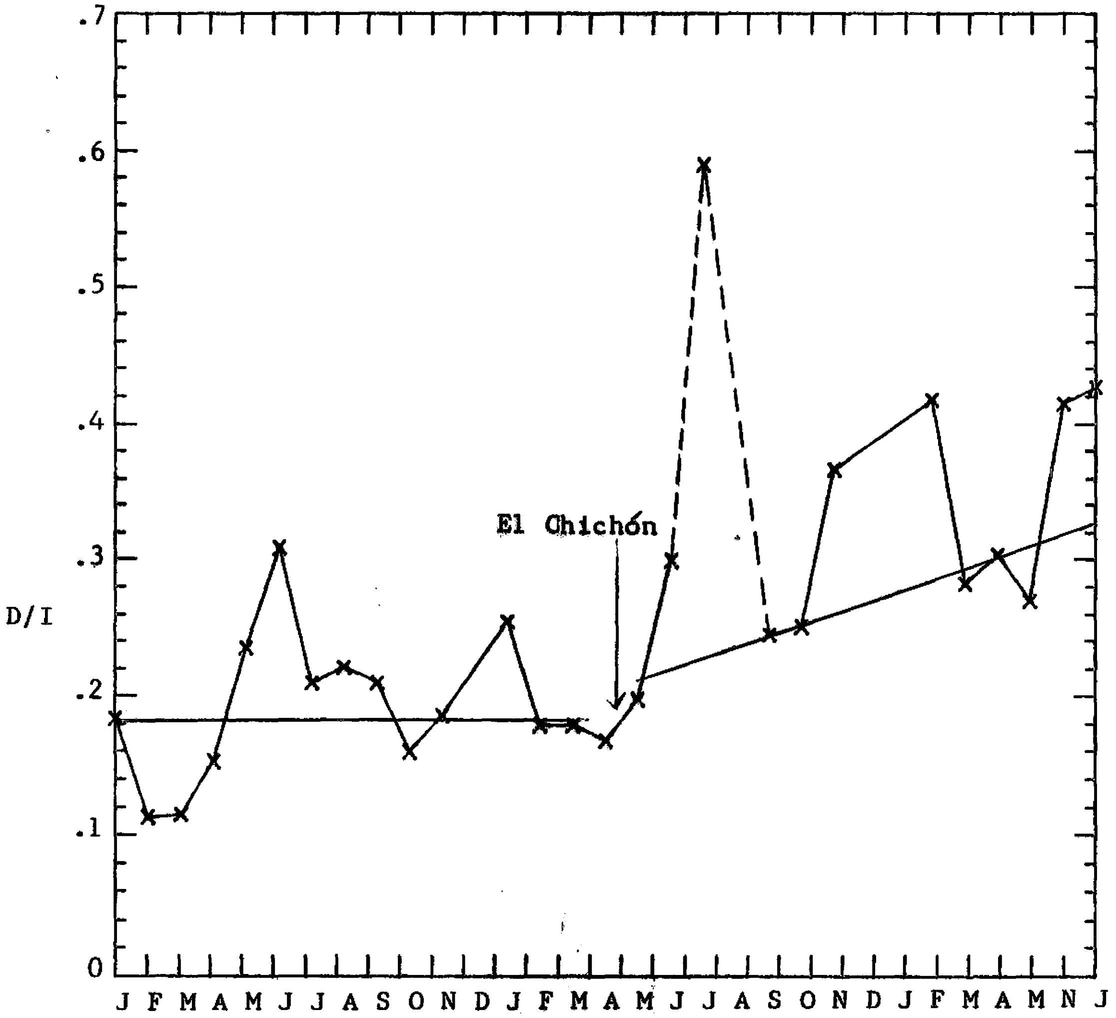

Figure 7: Monthly Trend of D/I

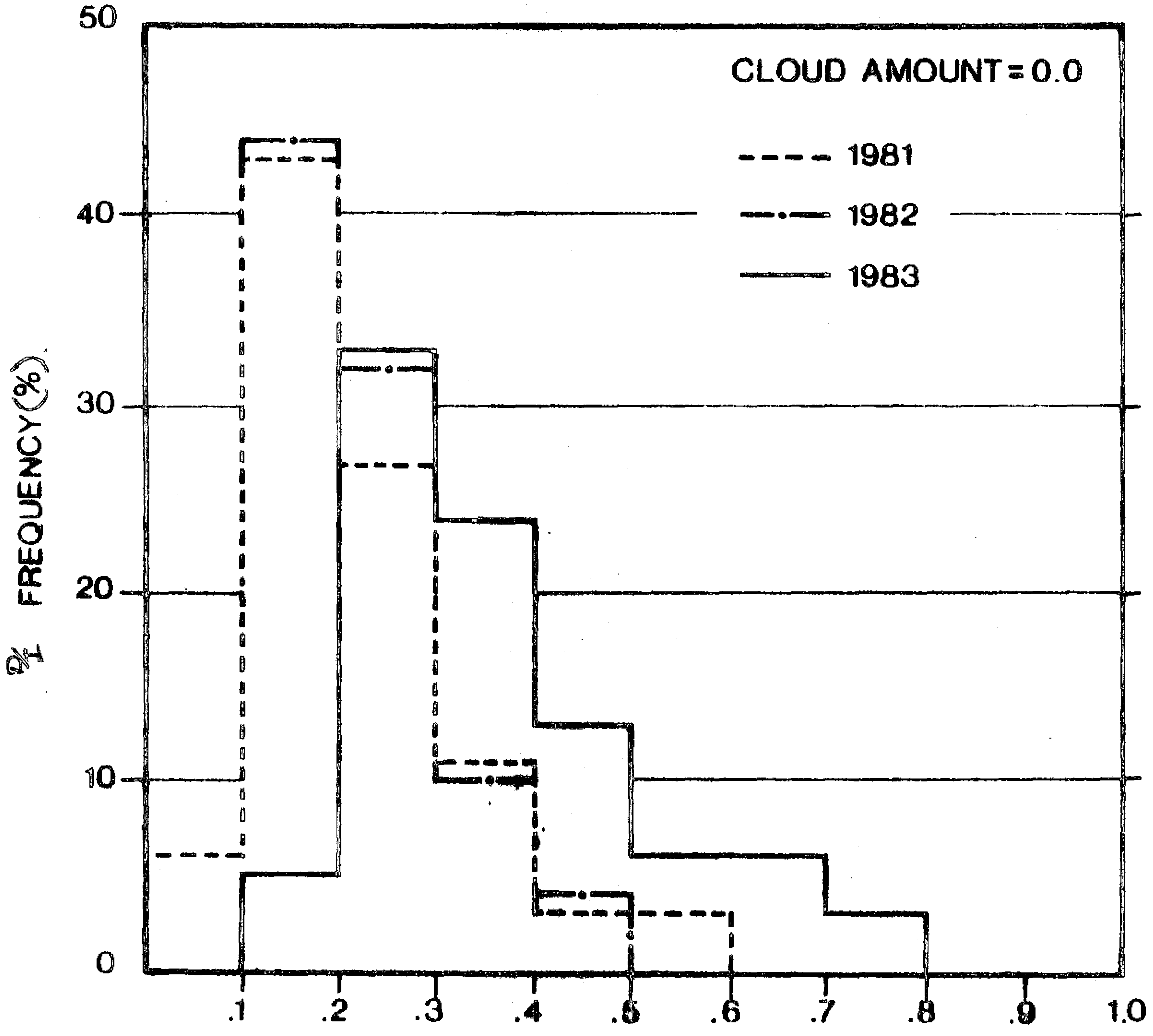

Figure 8: Frequency of occurrence of D/I values for 1981-1983

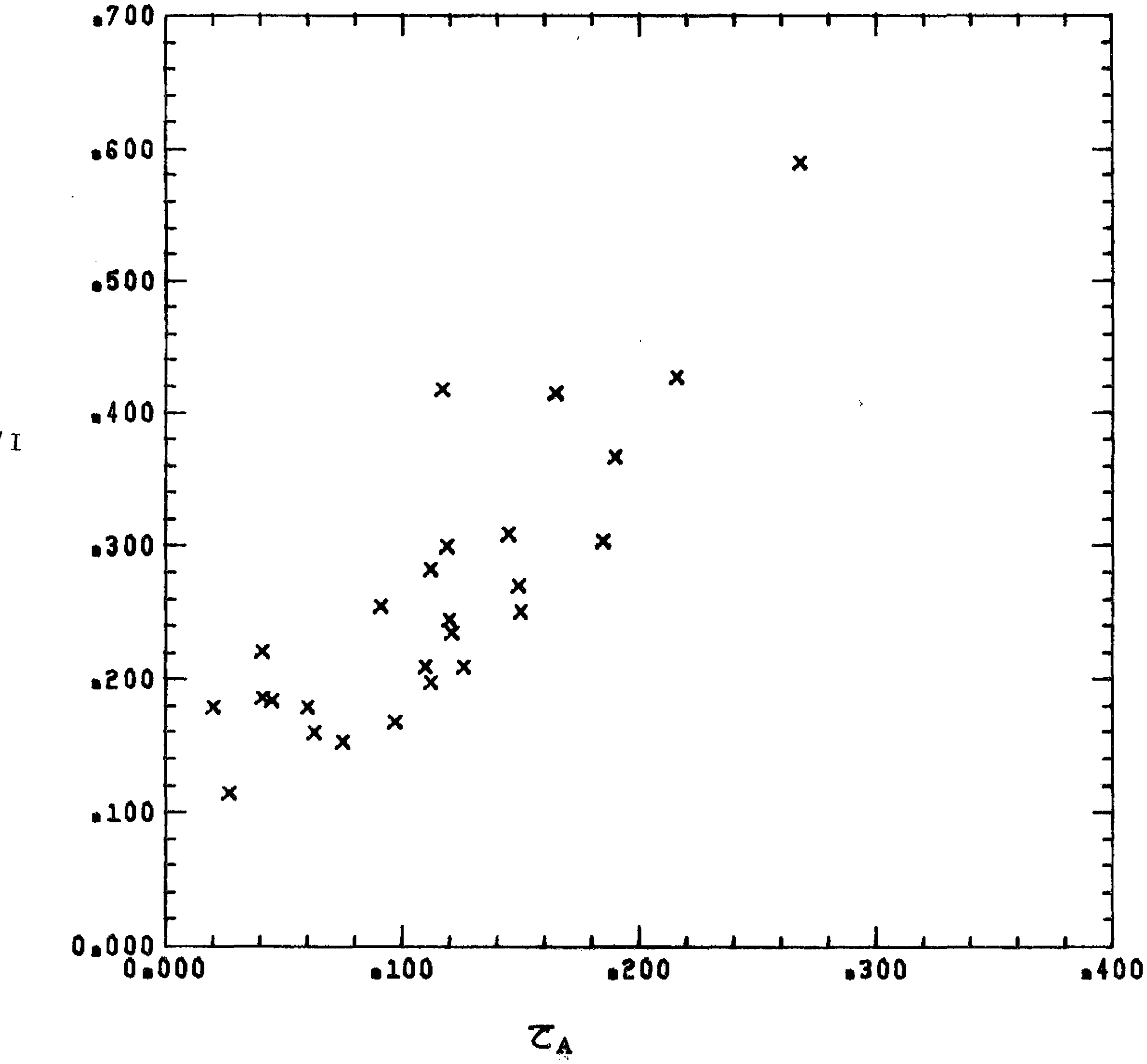

Figure 9: Correlation of τ_A and D/I

CONCLUSIONS

The eruption of El Chichon with its resultant volcanic dust cloud has affected turbidity over southern Canada. Its main impact was felt in November 1982 and is reflected in an increase in annual and monthly mean and median values of τ_A and D/I.

τ_A increased by 0.06 which is in good agreement with values obtained for Hamilton by Davies et al (1984) and from aircraft surveys by Ellsworth and DeLuisi (1983) and Sprinhirne (1983). Monthly variation in τ_A is similar to that observed by Hay (1983) at Vancouver. However, the main impact of El Chichon on turbidity at Vancouver occurred in October 1982, while for Woodbridge it occurred a month later.

The average annual value of D/I increased by 0.2 which is three times the calculated increase in τ_A. Thus for this study, D/I is the more sensitive index of turbidity. It is therefore recommended that D/I should be used more widely. It is simply calculated, more sensitive than τ_A and independent of model calculations.

Appendix 1

LIST OF SYMBOLS

Upper Case Roman

Symbol	Definition	Unit
D	Diffuse solar radiation	W/m^2
ET	Equation of time	minutes
H	Hour angle	degrees
I	Direct beam solar radiation	W/m^2
I_λ	Measured spectral direct beam solar radiation at the surface	W/m^2
$I_{o\lambda}$	Spectral direct beam solar radiation at the top of the atmosphere	W/m^2
I_o	Cloudless sky direct beam solar radiation	W/m^2
I_*	Extraterrestrial solar radiation on a horixontal surface corrected for the sun-earth distance	W/m^2
LAT	Local apparent time	hour
LS	Longitude of station	degree
LSM	Longitude of standard meridan	degree
LST	Local standard time	hour
P	Station pressure	kPa
P_o	Standard sea level pressure	kPa
S	Solar constant corrected for departure of the sun-earth distance from the mean	W/m^2

Appendix 1 (continued)

LIST OF SYMBOLS

Upper Case Roman

Symbol	Definition	Unit
T	Surface temperature	^oK
T_a	Transmittance after extinction	dimensionless
T_d	Dewpoint temperature	^oC
To	Standard sea level temperature	^oK
Toz	Transmittance after absorption by ozone	dimensionless
T_R	Transmittance after scattering by dry air molecules	dimensionless
U_w	Pressure and temperature corrected water vapour path length	mm
U'w	Uncorrected water path length	mm
	Lower Case Roman	
a	Aerosol absorption	W/m^2
a_w	Water vapour absorptivity	dimensionless
m	optical airmass	dimensionless
u_o	Atmospheric path length of ozone	dimensionless

Appendix 1 (continued)

LIST OF SYMBOLS

Lower Case Greek

Symbol	Definition	Unit
δ	Solar declination	degree
κ	Spectral mass absorption coefficient	m^2/Kg
λ	Wavelength	m
θ	Zenith angle	degree
ϕ	Station latitude	degree
σ	Spectral mass scattering coefficient	m^w/Kg
τ	Optical depth	dimensionless
τ_A	Aerosol attenuation coefficient	dimensionless
τ_λ	Spectral optical depth	dimensionless
τ_{a_λ}	Spectral optical depth of aerosol	dimensionless
τ_{O_λ}	Spectral optical depth of ozone	dimensionless
$\tau_{R\lambda}$	Spectral optical depth of Rayleigh scatter	dimensionless
τ_{w_λ}	Spectral optical depth of water vapour	dimensionless

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