THE EFFECT OF ATMOSPHERIC AEROSOL ON THE NET SOLAR RADIATION BALANCE OF THE SURFACE-LOWER ATMOSPHERE SYSTEM

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by

HOWARD BARKER

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TITLE: The Effect of Atmospheric Aerosol on the Net Solar Radiation Balance of the Surface-Lower Atmospheric System

AUTHOR: Howard Barker

SUPERVISOR: Dr. J.A. Davies

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ABSTRACT

A simple method for computing the effect of atmospheric aerosol on the net solar radiation balance of the surface-lower atmosphere system is presented. It was found that in clear sky conditions at Goose Bay, Toronto and Winnipeg, for the period 1977 - 1982, the presence of aerosol made the systems 10 - 20% more efficient at absorbing radiation than if the aerosol was absent. Furthermore, surface albedo is shown to be the most important parameter governing the effect of aerosol on the net solar radiation balance in an aerosol system, while the effect of volcanic aerosol produced by El Chichon had a minor influence on the net solar radiation balance.

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a	aerosol absorption coefficient (multiple reflections)
a_T	total aerosol absorption coefficient
a_w	water vapour transmission coefficient
b	aerosol backscatter coefficient (Robinson's method)
f	aerosol forward scatter coefficient (multiple reflections)
t	aerosol transmission coefficient (multiple reflections)
$^a/b$	absorption to backscatter ratio
$(^a/b)crit$	$^a/b$ critical value
A	aerosol absorption coefficient (Robinson's method)
B	aerosol backscatter coefficient (Robinson's method)
C	forward-to-back scatter ratio
D_m	measured diffuse beam radiation
D_o	theoretical diffuse beam radiation
F	excess diffuse beam radiation (normalized to G_o)
G_m	measured global beam radiation
G_o	theoretical global beam radiation
I_o	measured direct beam radiation
R	aerosol recovery rate (multiple reflections)
R_R	aerosol recovery rate (Robinson's method)
S'	adjusted solar constant
T_o	ozone transmission coefficient
T_R	Rayleigh transmission coefficient
Z	solar zenith angle
α_E	effective albedo

α_s	surface albedo
γ_A	total absorption by aerosol system
ρ_A	ratio between radiation gained and lost by aerosol system
ρ_c	ratio between radiation gained and lost by clean system
ρ_R	aerosol sensitivity parameter (Robinson's method)
ρ	aerosol sensitivity parameter (multiple reflections)

PART 1

- INTRODUCTION -

Atmospheric aerosol*, whether of anthropogenic or non-anthropogenic origin, has the potential to change both local and global climate. However, the direction and magnitude of such change is uncertain. Hence, there is a need to accurately and efficiently determine the radiation attenuation properties of aerosol for an understanding of how aerosol affects climatic change.

Until the early 1960's, it was thought that aerosols only scattered radiation, thus cooling the Earth-Atmosphere system. By the early 1970's, it was believed that the ratio of absorbed to backscattered radiation was approximately unity (Robinson, 1962; Charlson and Pilat, 1969; Schneider, 1971). Currently it is realized that the absorption to backscatter ratio of aerosols, for spectrally integrated solar radiation, is highly variable both spatially and temporarily, but in general, exceeds unity (Ensor et al., 1971; Joseph and Wolfson, 1975; Davies and McArthur, 1980).

Early determinations of aerosol properties from surface measurements of solar radiation neglected multiple scattering in the aerosol layer (eg., Robinson, 1962). However, the effects of multiple reflections between the surface and the aerosol are easily incorporated. (c.f. Schneider, 1971; Schneider and Dickenson, 1976; Davies and Hay, 1980).

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*solids and liquids suspended in the atmosphere excluding water vapour

Since multiple reflections increase the path length of radiation through the aerosol, one might expect difference between results obtained by the traditional method and the multiple reflection method, particularly with respect to the absorption and radiative heating of aerosols.

By including multiple reflections, a simple procedure for estimating the efficiency of aerosols in perturbing the solar energy balance of the surface-lower atmosphere system is developed.

This study uses measured solar radiation data from Winnipeg, Toronto and Goose Bay for the period 1977 - 1983 with model values (MAC model) to determine aerosol perturbations to the solar radiation balance. Effects of volcanic aerosol from the Mexican volcano, El Chichon, are also shown.

PART 2

THEORETICAL BACKGROUND

The introduction of an aerosol into an otherwise clean atmosphere* perturbs the solar energy balance of the surface-lower atmosphere and alters the quality and quantity of radiation incident at the surface.

Therefore, by comparing measured global and diffuse radiation to theoretical values for an aerosol-free atmosphere, as determined by the MAC model, the optical properties of aerosol are obtained. Once this is done, the effect of aerosol on the radiation balance can be obtained.

2.1 The MAC Model (Davies and Hay, 1980)

Solar radiation incident at the top of the atmosphere is attenuated in its passage through the atmosphere by absorption and scattering. In a cloudless atmosphere, the primary attenuaters are ozone, water vapour and dry air molecules. In this study, the MAC Model is used to estimate direct, I_o, and diffuse, D_o, radiation incident at the surface under a clean, cloudless atmosphere using,

$$I_o = S' \text{Cos } z \ (T_o T_R - a_w) \tag{1}$$

$$D_o = (1 - T_R)/2 \tag{2}$$

$$G_o = I_o + D_o \tag{3}$$

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*an atmosphere devoid of aerosol

where G_o is global radiation, S' is the solar constant corrected for Earth-Sun distance, z is the solar zenith angle, T_o, T_R and a_w are respectively the ozone, Rayleigh and water vapour transmission coefficients which are determined from observed meteorological conditions.

2.2 Robinson's Method (1962)

To estimate aerosol effects on the radiation balance, aerosol optical properties must first be determined. The method used is based on Robinson's procedure.

The difference between measured diffuse, D_m, and the expected amount obtained by the MAC model is attributed to forward scattering by aerosol. The "excess" of diffuse radiation, normalize by G_o, is expressed as,

$$F = \frac{D_m}{G_o} - \frac{D_o}{G_o} \qquad (4)$$

Using experimentally-integrated values of measured angular distribution of scattered radiation, Robinson tabulated the forward to backscatter ratio of an aerosol as a function of solar zenith angle (Fig. I). Hence, the backscatter coefficient of an aerosol is defined as,

$$B = F/C(z) \qquad (5)$$

where B + F is the total amount scattered by aerosol.

The difference between the normalized amounts of global radiation expected at the surface in an aerosol-free atmosphere and the measured amount, G_m, defines the total attenuation of solar radiation

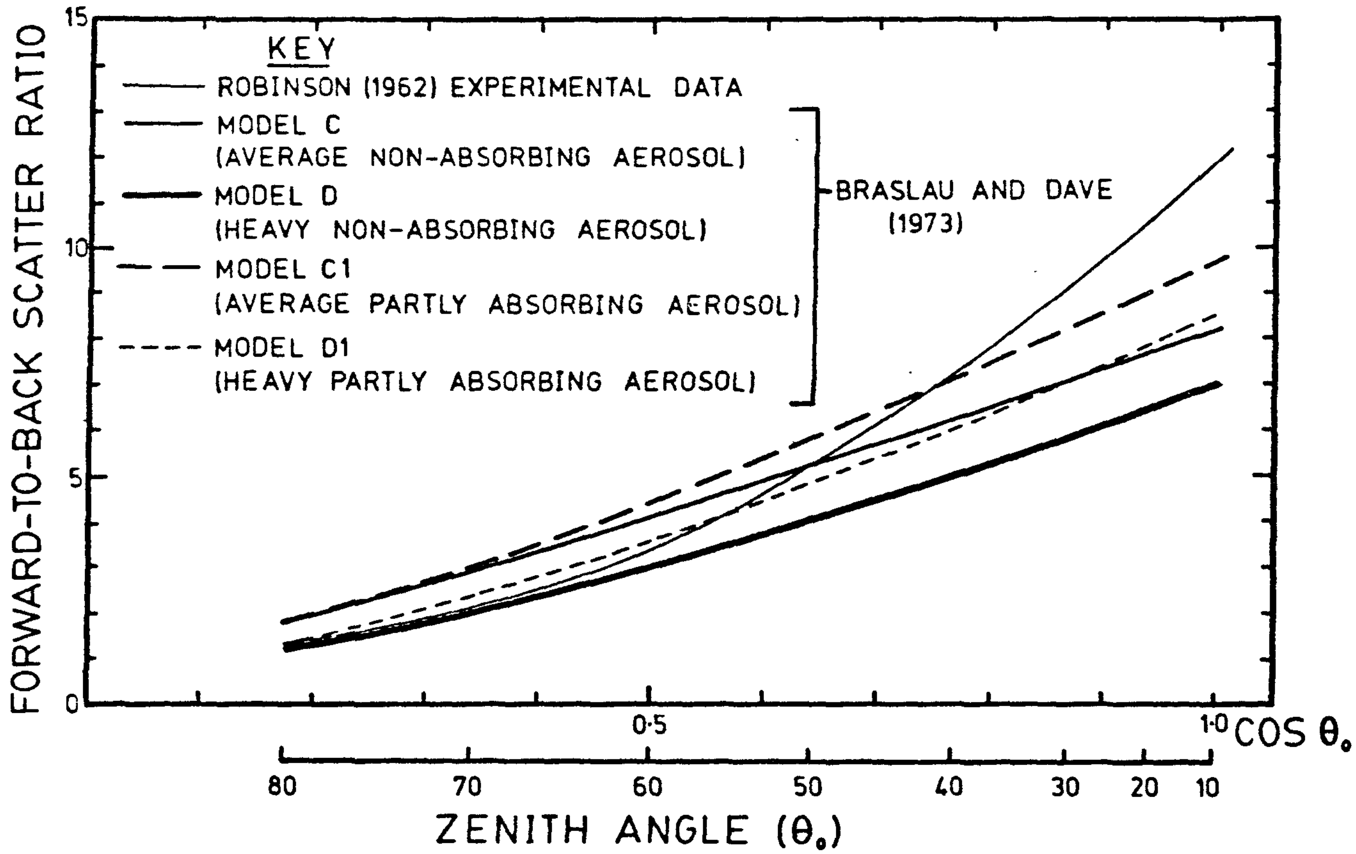

Fig. 1 Aerosol forward-to-back scatter ratio.

by aerosol. Upon subtracting from this difference the corresponding value of B, the remaining quantity is an estimate of aerosol absorption, A,

$$A = 1 - G_m/G_o - F/C(z) \tag{6}$$

2.3 Multiple Reflection Method

Robinson's method assumes that radiation passes only once through the aerosol layer. The multiple reflection method assumes an infinite number of passages. The two scenarios are shown in Fig. 2.

Fig. 2 shows that Robinson equates F with the forward scatter from the primary ray, f. However, if all internally-reflected rays are diffuse (Davies and Hay, 1980), F includes infinitely many rays. Therefore, Robinson's method attributes too much of the excess diffuse beam to aerosol forward scattering. Furthermore, by overestimating f, aerosol backscatter is also overestimated and absorption is underestimated. Hence, multiple reflections should be included to determine aerosol properties, (c.f. Leighton, 1979).

To derive expressions for absorption and backscatter coefficients using the multiple reflection method, the following assumptions are made:

Referring to Fig. 2(a)

$$\left.\begin{array}{l} a_i = a \\ b_i = b \end{array}\right. \quad \forall\, i \in [1,\infty) \tag{7}$$

Equation (7) states that, for all passages of radiation through an aerosol layer, the absorption and backscatter coefficients are equal to those associated with the primary passage. It is also assumed that all

Fig. 2

reflected rays have a zenith angle equal to that of the primary beam. Thus, the value of C(z) is constant for each passage of radiation through the aerosol. The latter assumption is supported by the following; (i) for many surfaces, the primary reflected beam is most concentrated near the solar zenith angle (Eaton and Dirnhirn, 1979), and (ii) for most surfaces, over 98% of the total radiation absorbed by aerosol occurs before the second surface reflection. Hence, the angle at which subsequent rays are reflected is of relatively minor importance to total absorption by aerosol.

Consider first the total absorption coefficient of an aerosol. It is expressed as the sum of the absorption terms in Fig. 2(a),

$$a_T = a + a\alpha_s t + a\alpha_s^2 bt + a\alpha_s^3 b^2 t + \ldots. \quad (8)$$

$$= a + at \sum_{i=1}^{\infty} \alpha_s^i b^{i-1}$$

where $t = 1-a-b$ is the aerosol transmission coefficient and α_s is the surface albedo.

Multiplying eq (8) by $\alpha_s b$ gives,

$$\alpha_s b a_T = \alpha_s ba + at \sum_{i=1}^{\infty} \alpha_s^{i+1} b^i \quad (9)$$

Upon subtracting eq (9) from eq (8), and solving for a_T, the generalized form of a_T is obtained,

$$a_T = \frac{a(1 + \alpha_s(t - b))}{1 - \alpha_s b} \quad (10)$$

Using the same summing procedure, t is defined as,

$$t = \frac{G_m}{G_o}(1 - \alpha_s b) \quad (11)$$

while a is defined as,

$$a = 1\text{-}t\text{-}b \qquad (12)$$

The remaining coefficient to be determined is b. To do this, f must first be defined. Assuming that all rays received at the surface are diffuse save for t, the excess measured diffuse (eq (4)) is defined as,

$$\begin{aligned} F &= f + \alpha_s bt + \alpha_s^2 b^2 t + \ldots. \\ &= f + t \sum_{i=1}^{\infty} (\alpha_s b)^i \\ &= \frac{f - \alpha_s bt(1-f)}{1 - \alpha_s b} \end{aligned} \qquad (13)$$

Modifying eq (5) such that,

$$f = bC(z) \qquad (14)$$

then substituting eq (11) and eq (14) into eq (13) gives a quadratic equation in b,

$$b^2\{-\alpha_s(C(z) + \frac{\alpha_s G_m}{G_o})\} + b\{C(z) + \alpha_s(\frac{G_m}{G_o} + F)\} - F = 0 \qquad (15)$$

Neglecting second order terms,

$$b = \frac{F}{C(z) + \alpha_s(\frac{G_m}{G_o} + F)} \qquad (16)$$

2.4 Thermal Effects of Aerosol on the Surface-Lower Atmosphere System

2.4.1 Absorption to Backscatter Ratio (Schneider, 1971)

Traditionally, the absorption to backscatter ratio has been used to estimate the local thermal effect of aerosol on the surface-lower atmosphere system. By comparing the surface albedo to the effective albedo of the system, α_E, it may be estimated whether an aerosol is warming or cooling the system.

The effective albedo is defined as the albedo of the surface-aerosol system after an infinite number of internal reflections between the surface and the aerosol layer have occurred. It is derived in the same manner as eq (10) and is expressed as,

$$\alpha_E = \frac{b + \alpha_s(t^2 - b^2)}{1 - \alpha_s b} \tag{17}$$

The critical boundary between warming and cooling of the system occurs when aerosol neither increases nor decreases the systems reflectivity, that is, when $\alpha_E = \alpha_s$.

By setting eq (17) equal to α_s and solving for $^a/b$, the critical value is obtained,

$$(\tfrac{a}{b})_{crit} = \frac{2a\alpha_s + (1-\alpha_s)^2}{\alpha_s(2-a)} \tag{18}$$

where if a = 0

$$(\tfrac{a}{b})_{crit} = \frac{(1-\alpha_s)^2}{2\alpha_s} \tag{19}$$

and if a = 1

$$(\tfrac{a}{b})_{crit} = \frac{1 + \alpha_s^2}{\alpha_s} \tag{20}$$

Therefore, the approximate region of $^a/b$ for which aerosol has no net heating or cooling effect on the system is described as,

$$\frac{(1 - \alpha_s)^2}{2\alpha_s} \leq (\tfrac{a}{b})_{crit} \leq \frac{1 + \alpha_s^2}{\alpha_s} \tag{21}$$

Although the logic behind this procedure is clear, the difficulty in visualizing the critical condition where $\alpha_E = \alpha_S$, particularly after infinitely many internal reflections, prompted the development of the following procedure.

2.4.2. Aerosol Sensitivity Parameter

The aerosol sensitivity parameter defines how efficiently the surface-aerosol atmosphere system absorbs solar radiation relative to how efficiently the system would absorb solar radiation in the absence of aerosol.

In an aerosol-free system, the ratio between the amount of solar radiation gained by the system (excluding ozone and water vapour) and the amount lost by the system is defined as,

$$\rho_C = \frac{(1 - \alpha_S)}{\alpha_S} \qquad (22)$$

In an aerosol system, the amount of radiation gained by the system, γ_A, is the sum of surface absorption and total aerosol absorption,

$$\gamma_A = \frac{t(1 - \alpha_S) + a(1 + \alpha_S(t - b))}{1 - \alpha_S b} \qquad (23)$$

The total amount of radiation lost from the system is α_E. Therefore, dividing eq (23) by eq (17) gives the ratio between radiation gained and lost by the aerosol system,

$$\rho_A = \frac{t(1 - \alpha_S) + a\ (1 + \alpha_S(t - b)}{b + \alpha_S(t^2 - b^2)} \qquad (24)$$

The quotient obtained by dividing eq (24) by eq (22) defines the aerosol sensitivity parameter,

$$\rho = \frac{\alpha_s}{1 - \alpha_s} \frac{t(1 - \alpha_s) + a\ (1 + \alpha_s(t - b))}{b + \alpha_s(t^2 - b^2)} \tag{25}$$

If $\rho = 1$, this is equivalent to $^a/b = (^a/b)_{crit}$, where the aerosol has no effect on the net solar radiation balance of the system. If $\rho > 1$, this implies that the presence of aerosol has caused the system to be more efficient at gaining radiation. Conversely, if $\rho < 1$, the aerosol has made the system less efficient at gaining energy.

PART 3

DATA REQUIREMENTS

All measured radiation and meteorological data were obtained from the Atmospheric Environment Service.

Measured global and diffuse radiation were obtained in the form of integrated hourly totals centred on the half hour in local apparent time.

The meteorological data needed to determine the atmospheric transmission coefficients used in the MAC model are; dry bulb temperature, wet bulb temperature, air pressure, precipitable water vapour, total cloud amount and ozone amount.

Precipitable water vapour for Winnipeg was estimated from the dew point temperature, while for Toronto, values obtained from radiosonde ascents at Buffalo International Airport were used. Goose Bay has its own radiosonde.

Ozone amounts were measured at Toronto and Goose Bay, while for Winnipeg, values obtained at Bismarck, North Dakota were used.

Surface albedo was assigned on a daily basis as a function of dry bulb temperature and location. Undoubtedly, this method of albedo determination introduces errors on the daily scale, but on the annual scale, the errors appear to be minimal.

Clouds present problems when attempting to isolate the scattering and reflective properties of aerosol. This is because clouds enhance diffuse beam radiation through scattering and reflecting as well. Because total cloud amount is estimated once per hour and radiation values are integrated hourly totals, the occurrence of unreported cloud within the hour contaminates the radiation value. To avoid this problem, only days with five or more hours of continuous cloudless conditions were used. As such, the data are by no means randomly selected.

PART 4

RESULTS AND DISCUSSION

The requirement for five consecutive hours of uninterrupted sunshine reduced the data set drastically. Thus, only annual trends could be represented. This is important when considering the effects of El Chichon.

Based on the findings of Hay and Darby (1984), it was late in 1982 before the stratospheric aerosol produced by El Chichon noticeably attenuated solar radiation at Vancouver. As well, since maximum attenuation occurred in the winter and spring of 1983, this conveniently isolates 1983 as a test year to be compared to the period 1977 - 1982.

In order to reduce the noise caused by the eruption of Mt. St. Helens in 1980 (although Hay and Darby showed it to be quite minor), seasonal variations and the slight effect of El Chichon in 1982, the years 1977 - 1982 were grouped together. It is therefore assumed that the median values of a, b, $^a/b$ and ρ for 1977 - 1982 (Table 1) represents the characteristics of aerosols which typically occur during periods in which protracted stratospheric aerosol anomalies are absent, while the means in Table 2 show the influence of tropospheric noise.

4.1 1977 - 1982

Table 1 shows a remarkable consistency in the median and both quartiles of a, b, $^a/b$ and ρ for all stations. This may be due to the data selection criterion. In general, five hours of continuous sunshine is associated with high pressure systems which are characterized by clear skies and relatively clean tropospheres due to the influx of northern air.

Hence, the assumption that what is being seen in Fig. 4 - 7 is largely the background stratospheric aerosol.

The most striking aspect of this set of results is that the aerosol system most efficient at "trapping" radiation is Goose Bay, where $\rho = 1.198$. This means that the surface-lower atmosphere system at Goose Bay is approximately 20% more efficient at gaining radiation when aerosol is present than if the aerosol was absent. As well, at the 95% confidence level ($\alpha = 0.05$), the system at Goose Bay is insignificantly more efficient at gaining radiation than Winnipeg, but significantly more efficient than Toronto.

Tables 1 and 2 show an apparent paradox between a, b and ρ. One might intuitively expect that an aerosol with a low absorption coefficient, such as Goose Bay, should be relatively inefficient at gaining solar radiation. This, however, is not the case and the explanation is simple.

The numerator of ρ consists of the surface and total aerosol absorption terms. If the surface co-albedo, $1 - \alpha_s$, is large ($\sim$ 0.7 to 0.9), then it generally exceeds its aerosol counterpart, a, whose largest value in this study was 0.22, and whose mean was about 0.06. When α_s is small, only a small portion of the primary global beam is reflected while most of it is absorbed. Therefore, regardless of how large a is, the surface absorption term in ρ is the dominant term. Consequently, as α_s increases, so does the importance of aerosol absorption while the importance of surface absorption decreases.

To illustrate how important large values of α_s are at enhancing a_T, Fig. 3 shows $^a/a_T$ plotted against α_s. Going from a surface albedo of 0.2

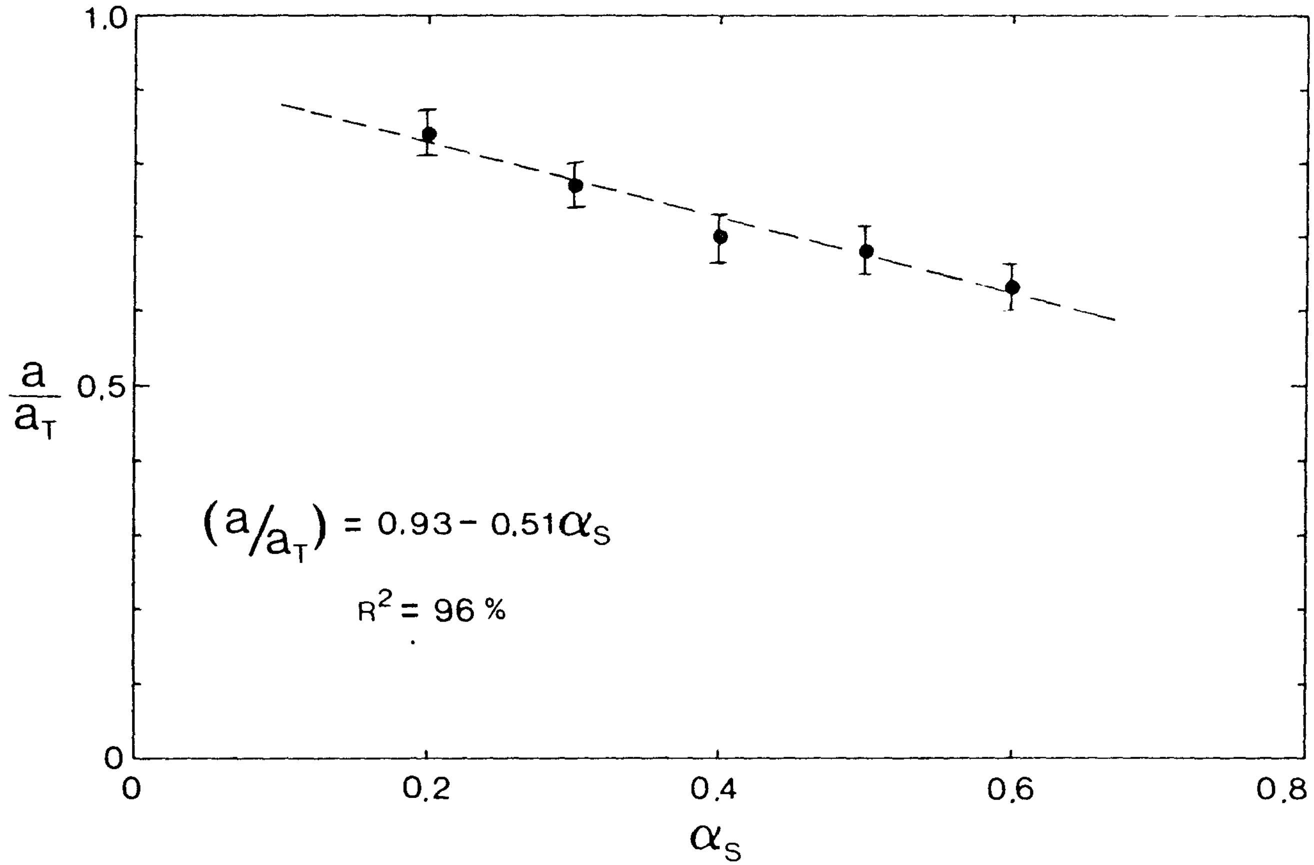

Fig. 3 Relationship between the efficiency of aerosol total absorption and surface albedo

to 0.6, the fraction of a_T made up of a, decreases from about 0.84 to 0.63. Furthermore, each point in Fig. 3 consists of 10 randomly drawn samples from each station. Thus, the small standard deviation bars imply that a_T has a relatively weak dependence on a and b.

The backscatterance of aerosol plays two parts in determining ρ. First, a quantity of radiation is reflected back to space, thus cooling the system. This becomes important when b is large, a is large, and α_s is small, for when total attenuation, a + b, is large, significantly less radiation is incident at the surface. This decreases the effectiveness of the important surface absorption component. Secondly, b becomes increasingly important at trapping radiation as α_s increases, for both b and α_s govern the efficiency of multiple reflections which greatly influences ρ through increasing both a_T and G_m (see Appendix for the effectiveness of multiple reflections on absorption by aerosol and ρ).

Therefore, in the case of Goose Bay, Tables 1 and 2 show that b is virtually constant for all stations and has little effect on ρ. However, α_s at Goose Bay is approximately 50% larger than at Winnipeg and Toronto. Therefore, at Goose Bay, not only does surface absorption carry relatively little importance, but multiple reflections, and thus a_T, are enhanced. Hence, it is the larger surface albedo which makes the surface-lower atmosphere system at Goose Bay more efficient at gaining radiation than Toronto and Winnipeg. In fact, if at Goose Bay α_s was to change from 0.44 to 0.30, and a and b were to remain constant, ρ would decrease to only 1.122 which is necessarily less than both Toronto's and Winnipeg's ρ value, for Goose Bay has the least absorbing aerosol (Fig. 4).

Before discussing the $^a/b$ ratio, it should be pointed out that two different methods of finding the mean produced enormously different answers. The first method divides hourly occurring values of a by b, whereas the second method divides the mean annual value of a by the mean annual value of b, thereby giving an annual mean for $^a/b$ independent of hourly quotients.

When the mean is taken in the first case, it is very much affected by a few large values of $^a/b$ (primed figures in Fig. 8). In the second method, the occurrence of large a and small b values were far less frequent and less significant than the occurrence of large relative differences in corresponding hourly values of a and b. Since the median of the hourly quotients of $^a/b$ (Fig. 6, Table 1) agree with the means obtained from the second method, the second method is assumed to be a more accurate representation of typical aerosols and is therefore used in this study.

The ambiguity in interpreting $^a/b$ is shown in Tables 1 and 2, where Winnipeg has the largest $^a/b$ and Goose Bay the smallest. One may therefore be tempted to conclude that Goose Bay's aerosol produces the smallest warming effect on the system which would directly contradict what was implied by the aerosol sensitivity parameter. However, it is not the absolute magnitude of $^a/b$ which is important, but rather the relative positioning of $^a/b$ with respect to the line representing weak extinction in Fig. 8. This now complies with the results of ρ, since $^a/b$ for Goose Bay is 11 times that of the corresponding weak extinction limit, compared to Winnipeg where $^a/b$ is only 7.5 times that of the corresponding weak extinction limit.

Hence, this section has shown that the solar radiation balance of the surface-lower atmosphere system at Goose Bay is most affected by the presence of aerosol. This is primarily because of the high mean surface albedo. On the other hand, the system at Toronto is least affected due solely to a relatively low surface albedo.

4.2 1983: El Chichon Effects

Table 2 shows that the inter-quartile range of a, b, $^{a}/b$ and ρ for 1983 are similar in size to those of 1977 - 1982. It is therefore assumed that Fig. 4 - 7 largely represent stratospheric aerosols, and as in Part 4.1, Table 2 reflects the effects of tropospheric anomalies.

Discussion of El Chichon's effects will begin by defining and examining the aerosol recovery rate, R. The recovery rate may be assessed in two ways; (i) as the fraction of the direct beam radiation depleted by aerosol which reappears as diffuse beam radiation, and (ii) as the following,

$$R = \frac{f}{f + a + b} \qquad (26)$$

The recovery rate is used to determine how efficiently an aerosol scatters radiation in the forward direction.

For Vancouver, Hay and Darby (1984) determined that R increased approximately 35% from 1977 - 1982 to 1983. This implies that the El Chichon aerosol substantially increased the relative importance of forward scattered radiation. In this study, all three locations showed the same trend (Table 2), suggesting that the effect of El Chichon is evident at all three.

Winnipeg's results are somewhat ambiguous. The data set of 1983 contains only 87 hourly measurements compared to 276 for 1980. However, the period between February - April was well represented by 50 of the 87 measurements. In this period, Hay and Darby recorded maximum attenuation in Vancouver at the same latitude. Results for Winnipeg show that absorption decreased while, elsewhere, absorption increased. However, changes in the recovery rate suggested an El Chichon effect.

A possible explanation for this is that the radiation backscattered by the El Chichon aerosol has reduced the effect of a significant absorbing aerosol(s) below the El Chichon aerosol (Pollack et al., 1976). This lower level aerosol(s) may have been responsible for the high value of a before 1983. It is therefore difficult to conclude what effect El Chichon had on the radiation balance at Winnipeg.

Of the two aerosol properties, a and b, which govern the radiation balance of the surface-lower atmosphere system, b is the most affected by El Chichon aerosol (Fig. 4, 5). Therefore, depending on α_s, the increase in b will either cool the system if α_s is small, or warm the system if α_s is large.

At Goose Bay, not only has b increased drastically, but both a and α_s have increased as well. All of these changes will affect ρ. First of all, the sum a + b is approximately 1.75 times the value of 1977 - 1982. Thus, less radiation reaches the surface. The large surface albedo, however, reduces the relative importance of the surface in determining ρ. Hence, not only is the total aerosol absorption increased because of a large surface albedo, but it is further accentuated by large values of a and b. In fact, when 1983 values of a and b, and the 1977 - 1982 value of α_s are substituted into ρ, the efficiency of the system at gaining radiation is only 2% greater than the actual 1977 - 1982 value. Therefore, at Goose Bay, the increase of about 13% in the aerosol sensitivity parameter in 1983 over 1977 - 1982 is mainly due to the increase in α_s, not by the increase in a and b.

If $^a/b$ is considered independent of the surface, the substantial decrease observed in 1983 at Goose Bay implies that El Chichon has a cooling effect on the system. Even when the surface is considered (by plotting $^a/b$ in Fig. 8), $^a/b$ is relatively closer to the weak extinction line than it was before 1983. Yet still, of the three stations, Goose Bay is relatively farthest from the weak extinction limit. This implies that the relationship between $^a/b$ and ρ is non-linear.

In the case of Toronto, b was substantially increased but only by half the amount observed at Goose Bay (Fig. 5). This may be attributed to the, more or less, monotonic increase in stratospheric aerosol concentration from 35^oN to 55^oN in December, 1982 (Dutton and DeLuisi, 1983). The increase in a was only slight. The surface albedo increased but is still within the range where multiple reflections are relatively inefficient (Fig. 3). Since ρ at Toronto remained virtually unchanged, the increase in total attenuation, a + b, which tends to cool the system, is cancelled by the slight increase in α_s, and the efficiency of multiple reflections which tends to warm the system. This conclusion is complemented by the change in $^a/b$ from 1977 - 1982 to 1983. In Fig. 8, both values are of approximately equal relative distance from the weak extinction limit.

The 1983 data for Winnipeg is here examined to gain more understanding of how ρ is affected by a, b and α_s. Changes are not attributed to El Chichon for reasons mentioned above.

In 1983, the decrease in a and the increase in both b and α_s resulted in a slight decrease in ρ. Thus, the decrease in a, coupled with the increase in primary backscatter, is approximately balanced by the increase in the trapping of radiation.

In accordance with ρ, the decrease in $^a/b$ implies that the system is slightly less efficient at gaining radiation in 1983.

This section has shown that the aerosol produced by El Chichon affected the net solar radiation balance of the surface-lower atmosphere system only slightly.

Table 1 Median, First and Third Quartiles of a, b, $^{a}/b$ and ρ

Each cell contains three numbers which represent the categories in Fig. 4 to 7 numbered in ascending order. The first number is the median category, and the second and third numbers are the first and third quartiles, respectively.

	GOOSE BAY		TORONTO		WINNIPEG	
	1977-1982	1983	1977-1982	1983	1977-1982	1983
a	3,3,4	4,3,5	3,2,4	3,2,4	3,2,5	3,2,4
b	1,1,2	4,2,5	1,1,2	2,2,4	1,1,2	3,2,4
$^{a}/b$	3,2,4	2,2,3	3,3,4	2,2,3	4,3,4	3,2,3
R	4,3,6	6,5,6	4,3,6	6,5,7	4,2,6	6,5,7
ρ	2,2,3	3,2,3	2,2,2	2,2,2	2,2,2	2,2,2

Table 2 Aerosol Properties and Energy Balance Parameters

	GOOSE BAY			TORONTO			WINNIPEG		
	1977 - 1982	1983	Δ%	1977 - 1982	1983	Δ%	1977 - 1982	1983	Δ%
a	0.051 ± 0.03	0.073 ± 0.03	43	0.059 ± 0.03	0.064 ± 0.04	8	0.065 ± 0.04	0.058 ± 0.02	-11
b	0.012 ± 0.01	0.038 ± 0.02	217	0.012 ± 0.01	0.026 ± 0.02	117	0.012 ± 0.01	0.026 ± 0.02	117
f	0.035	0.107	206	0.048	0.124	158	0.040	0.085	113
a/b	12.1 ± 26.0	2.6 ± 1.9	-79	15.8 ± 49.3	2.99 ± 49.3	-81	16.9 ± 398.7	3.30 ± 2.44	-80
a/b	4.25 ± 0.64	1.92 ± 0.18	-55	4.92 ± 0.66	2.46 ± 0.34	-50	5.42 ± 0.79	2.23 ± 0.33	-59
R	0.359 ± 0.18	0.490 ± 0.10	36	0.401 ± 0.18	0.579 ± 0.13	44	0.340 ± 0.17	0.504 ± .14	48
ρ	1.198 ± 0.12	1.324 ± 0.12	11	1.124 ± 0.10	1.101 ± 0.12	-2	1.184 ± 0.18	1.145 ± 0.13	-3
α_s	0.44	0.51	16	0.26	0.30	15	0.32	0.37	16
# of observ	217	92		314	121		916	87	

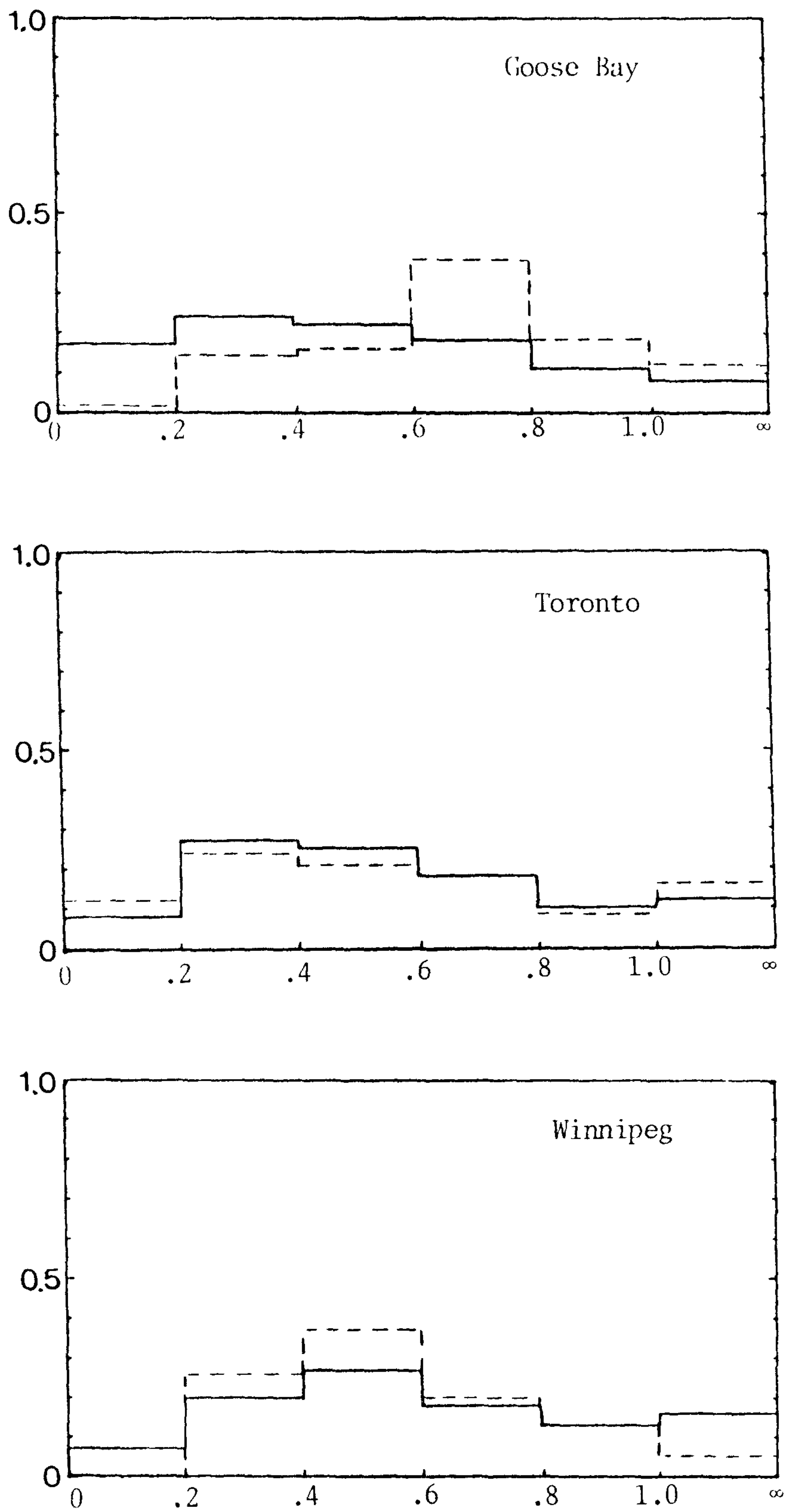

Fig. 4 Relative Frequency of a ——— 1977 - 1982

----- 1983

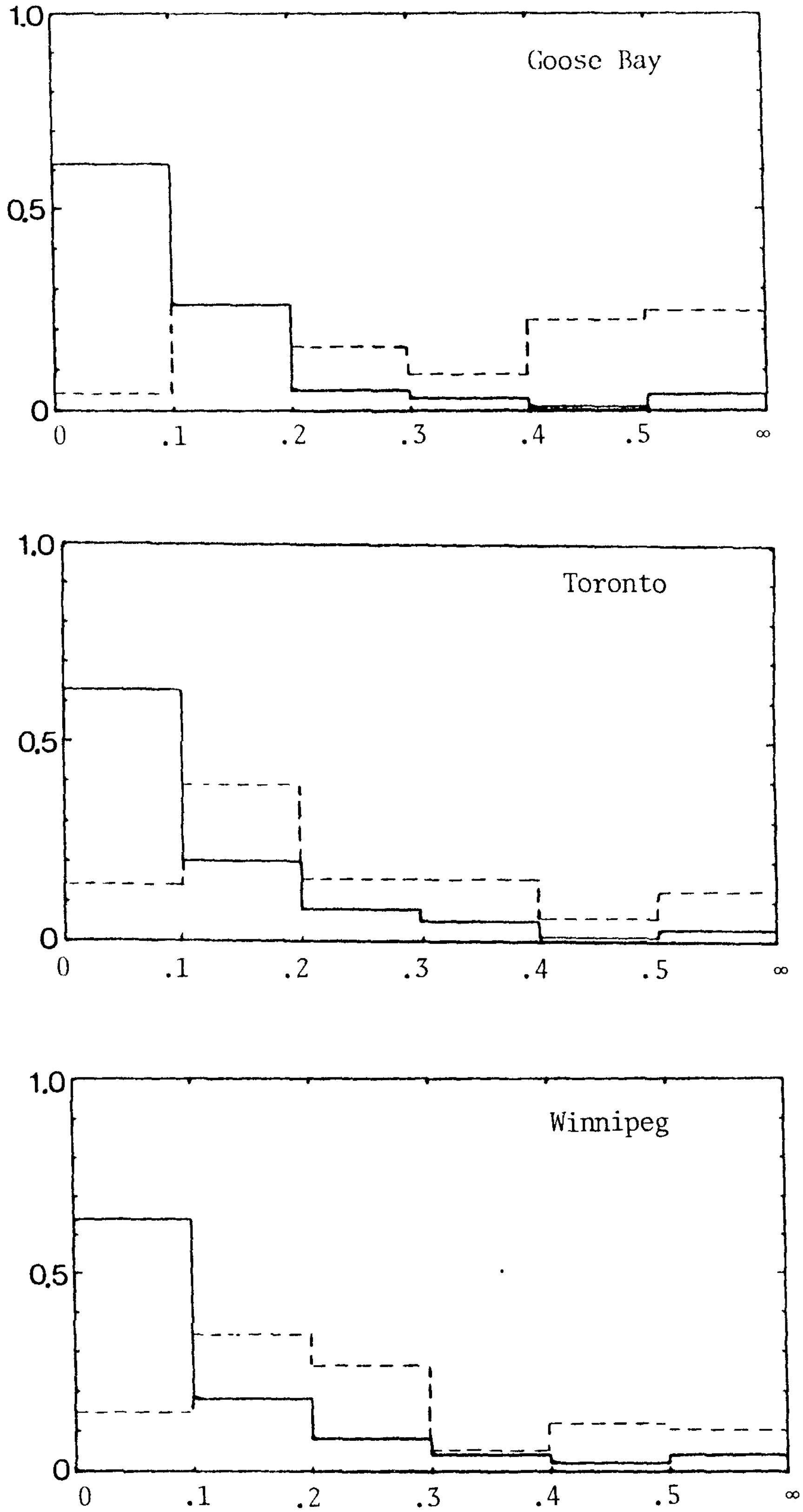

Fig. 5 Relative Frequency of b ——— 1977 - 1982

------ 1983

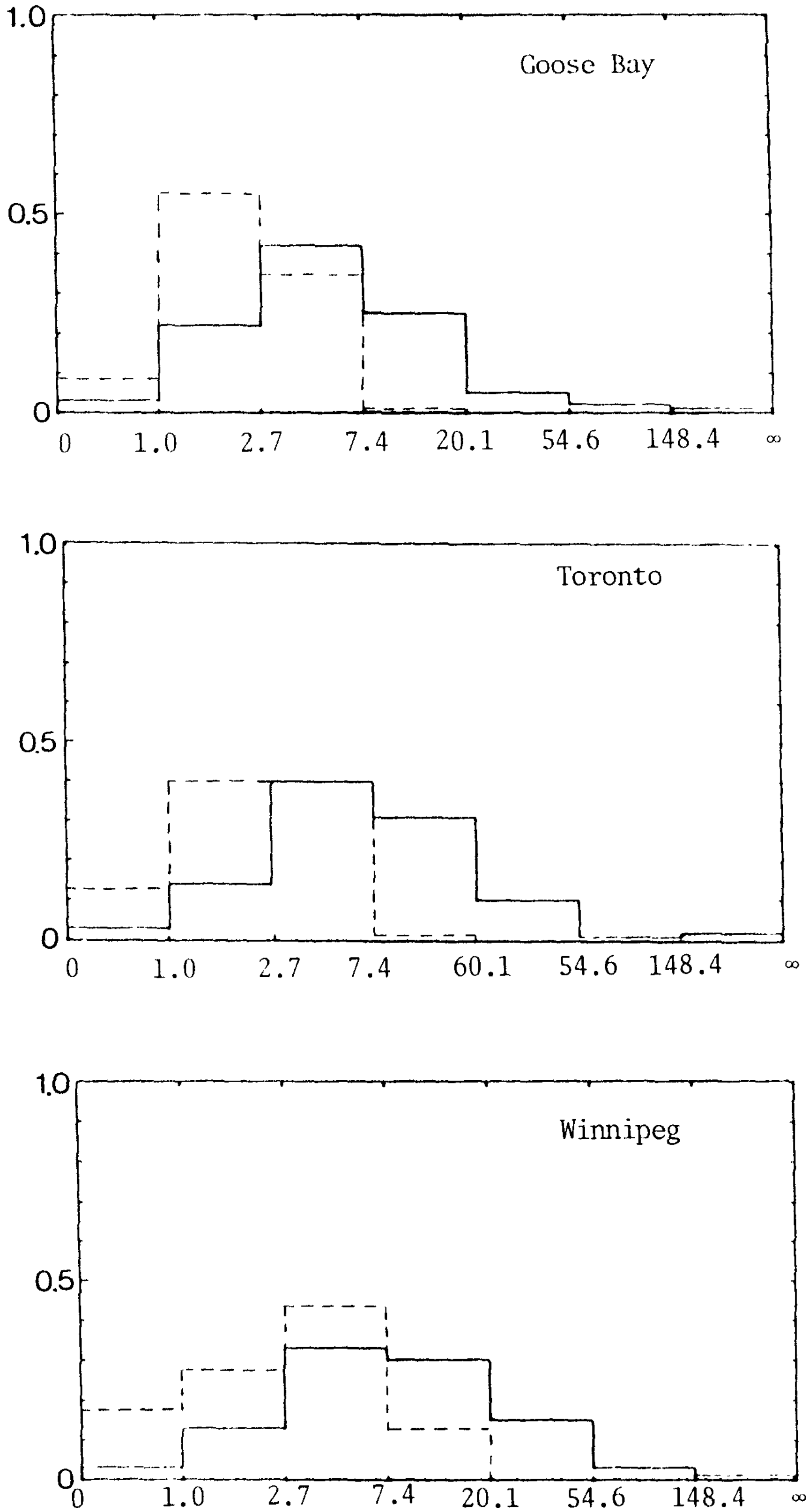

Fig. 6 Relative Frequency of $^a/b$ ——— 1977 - 1982

------ 1983

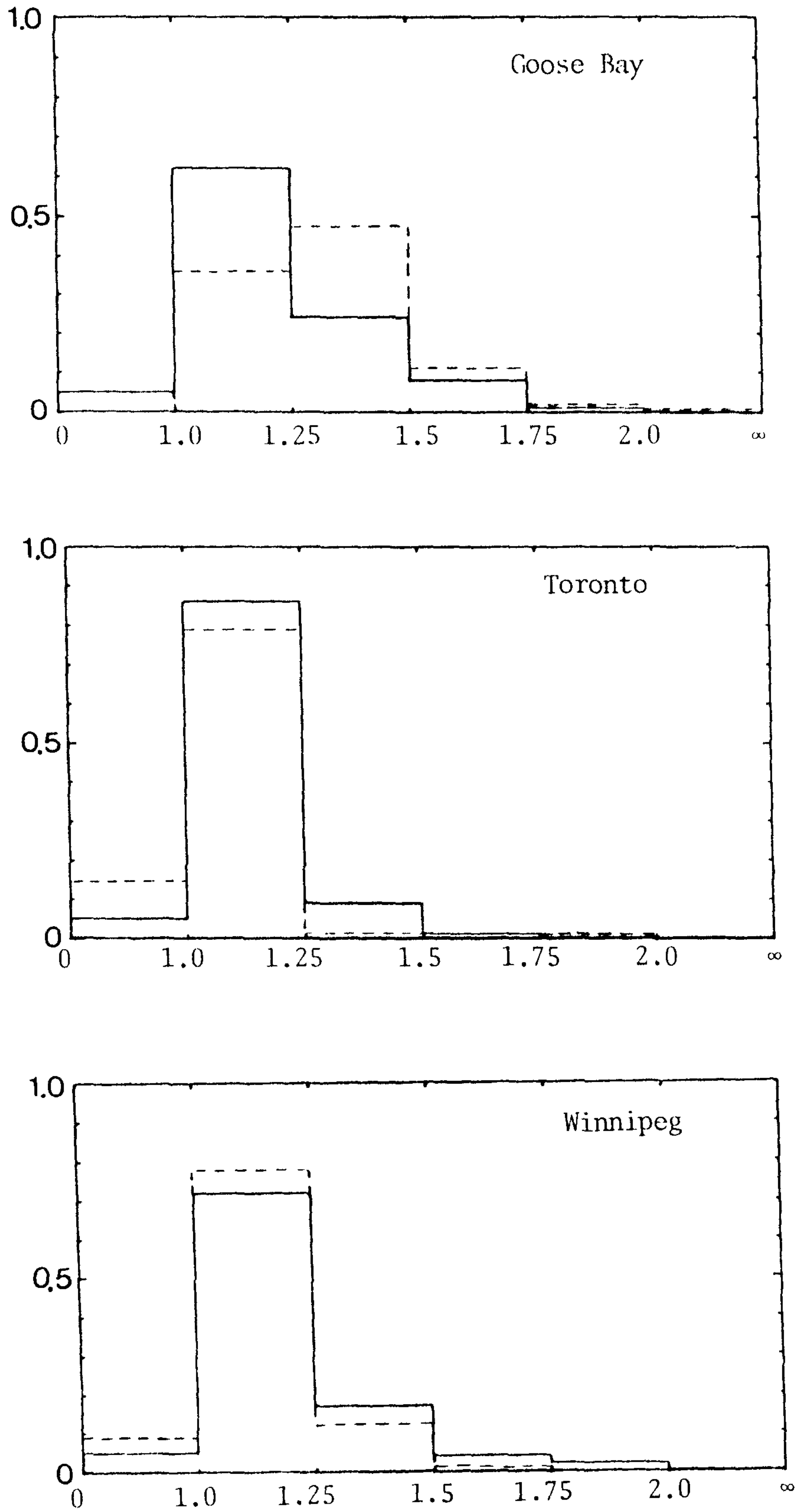

Fig. 7 Relative Frequency of ρ ——— 1977 - 1982

------ 1983

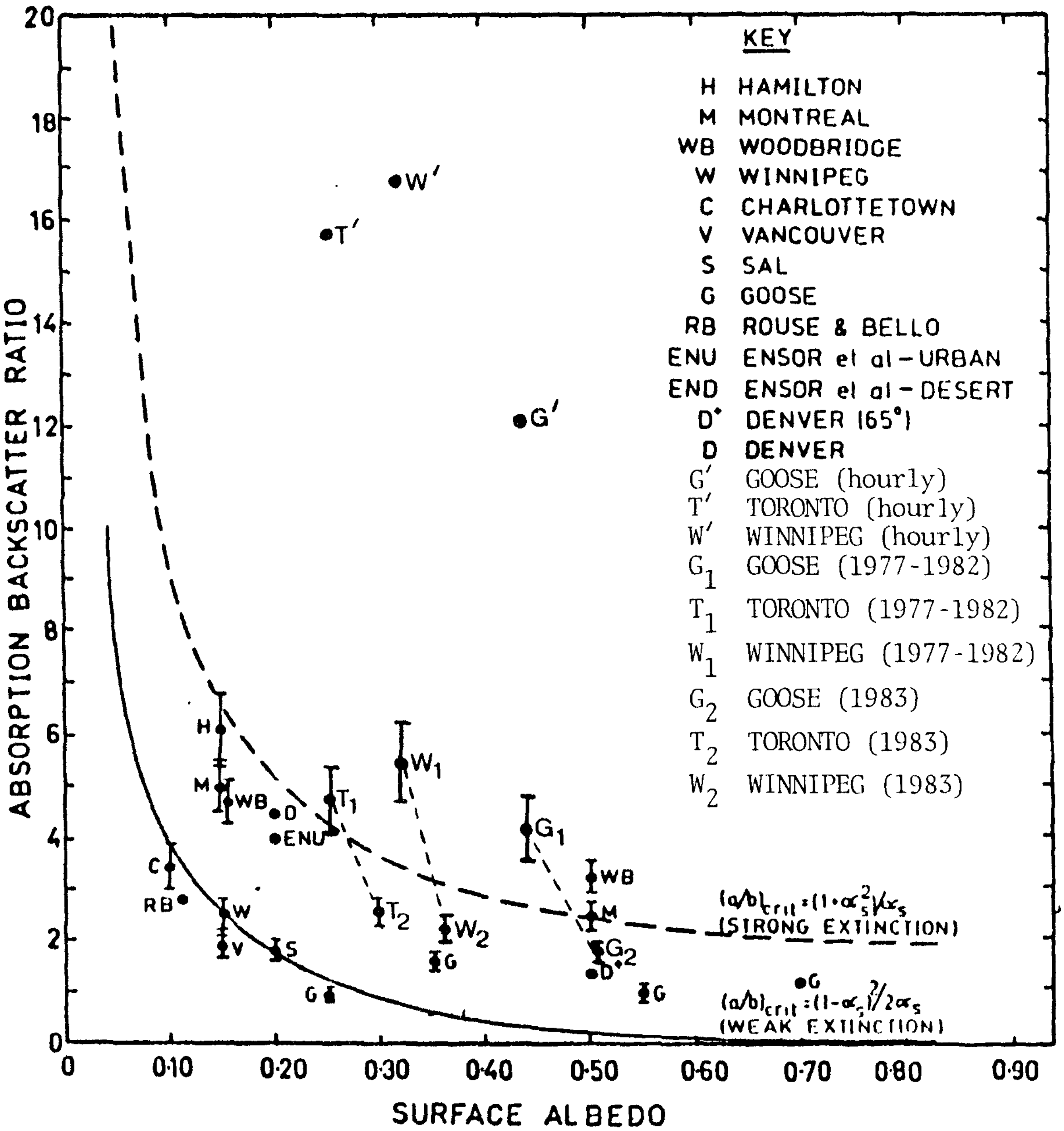

Fig. 8 **Absorption to backscatter ratio as a function of surface albedo.**

PART 5

CONCLUSIONS

The use of the multiple reflection method has been shown to be invaluable for estimating the effect aerosols have on the solar radiation balance of the surface-lower atmosphere system. Although Robinson's method is a good approximation of a, b, f and $^{a}/b$ for low surface albedo, it significantly overestimates b and f and underestimates a and a_T for high surface albedo. Also, Robinson's method significantly underestimates ρ for all surface albedos used in this study.

In the absence of protracted tropospheric and stratospheric aerosol anomalies, the efficiency of the surface-lower atmosphere system at all three locations to gain solar radiation exceeded the efficiency of a clean atmosphere by 10 - 20%. This implies that the aerosol which typically occurs on clear days at Winnipeg, Toronto and Goose Bay tends to warm the system. However, dominant variable governing ρ is the surface albedo with aerosol backscatter and absorption having a secondary affect, particularly at large α_s.

The effect of El Chichon aerosol on the net solar radiation balance of the surface-lower atmosphere system was almost negligible. However, the vertical distribution of absorbed energy was altered, as increased total attenuation caused more radiation to be absorbed aloft with less being absorbed at the surface (Hansen et al, 1978).

At low surface albedo, increases in a and b decrease the efficiency of the system to gain radiation, while at large albedos the opposite is true.

Therefore, the results of Parts 4.1 and 4.2 complement one another in that multiple reflections of radiation between the surface and aerosol significantly influence the net solar radiation balance of the surface-lower atmosphere system.

APPENDIX

The purpose of this section is to show the statistically significant and insignificant differences between Robinson's method and the multiple reflection method. All tests are two-tailed t-tests at $\alpha = 0.05$ with 95 degrees of freedom.

Table 1A Statistical Comparison between Robinson's method and the Multiple Reflection Method

Statistical Difference exists between:	Statistical Difference does not exist between:
a_T and a ; a_T and A ; R and R_R — for $\alpha_s = 0.6$ ρ and ρ_R for all α_s	$^a/_b$ and $^A/_B$ a_T and a ; R and R_a — for $\alpha_s = 0.2$

Note that R_R is the aerosol recovery rate and ρ_R is the aerosol sensitivity parameter as defined by Robinson's method,

$$R_R = \frac{F}{F + A + B} \tag{1A}$$

$$\rho_R = \frac{(1 - \alpha_s)(1 - A - B) + A}{B + \alpha_s(1 - A - B)} \quad \frac{\alpha_s}{1 - \alpha_s} \tag{2A}$$

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