Cloud Transmissivities for Canada

J. A. DAVIES, M. ABDEL-WAHAB AND J. E. HOWARD

Department of Geography, McMaster University, Hamilton, Ontario, Canada

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ABSTRACT

Transmissivities are determined for different cloud types using nine years of hourly irradiance measurements under overcast skies at six Canadian stations. Values for individual stations and for pooled data using irradiances uncorrected for multiple reflections are similar to values for Blue Hill, Massachusetts but larger than values for Hamburg, West Germany. It is argued that transmissivities used in numerical models which utilize surface observations of cloud layer amounts and types should be determined from irradiances without correction for multiple reflections. This would ensure at least partial compensation for attenuation by undetected cloud above overcast. The superior performance of transmissivities calculated in this manner is demonstrated in numerical model calculations of irradiance. It is also shown that there is no need to replace Blue Hill transmissivities with either the new values for Canada or the values proposed by Atwater and Ball for such models. There is also no indication in the Canadian results that cloud transmissivity varies with cloud amount as suggested by Atwater and Ball. Regional and seasonal variations in the Canadian transmissivities have a negligible effect on calculated irradiance. Irradiance calculations can be simplified with little loss in accuracy using an average transmissivity for each cloud layer; 78, 42 and 32% for high, middle and low cloud, respectively.

1. Introduction

A model that uses surface observations of cloud layer amounts and types has been widely used in North America to estimate solar radiation (Atwater and Ball, 1978; Davies and McKay, 1982; Suckling and Hay, 1978; Powell, 1980). The model estimates spectrally integrated values of global irradiance G. It has the general form

$$G/G_0 = \prod_{i=1}^{n} (1 - C_i + t_iC_i)/(1 - \alpha b), \qquad (1)$$

where G_0 is a theoretical estimate of cloudless sky irradiance, C_i and t_i are cloud amount and transmissivity in cloud layer i, n is the number of cloud layers, and α and b are surface albedo and an atmospheric backscatter coefficient for surface reflected radiation, respectively. The denominator incorporates the effect of multiple reflections between surface and atmosphere.

The importance of cloud transmissivities increases with cloud amount. In the long term, their effects on irradiance estimates depend on cloud amount frequencies. For representative Canadian stations, more than 50% of observed hourly cloud cover in a year exceeds 0.8 and about 40% of the annual solar irradiance is measured under this degree of cloudiness (Table 1). More than 60% of the mean error in irradiance estimates from (1) may be due to errors in cloud transmissivity for cloud amounts greater than 0.7 (Davies *et al.*, 1984).

Models of this type have relied mainly on data for Blue Hill, Massachusetts (Haurwitz, 1948) to calculate cloud transmissivities as functions of optical air mass. There have been few other sources. Vowinckel and Orvig (1962) provided values for polar and arctic locations and for two midlatitude Canadian stations; Dartmouth, Nova Scotia and Edmonton, Alberta. For Edmonton, where data were sufficient to examine seasonal variations, transmissivities for low- and middle-level clouds were larger in winter than summer (Table 2). For both Blue Hill and the Canadian stations, transmissivities for high- and middle-level clouds decreased with increasing air mass. Vowinckel and Orvig also found that transmissivities and the dependence on air mass increased with latitude. Recent determinations of transmissivities for Hamburg, West Germany (Kasten and Czeplak, 1980) and the tropical Atlantic (Atwater and Ball, 1981b) have been used to revise and extend the values of Haurwitz (Atwater and Ball, 1981b). In both studies, transmissivity did not vary with air mass and the large Hamburg data set did not show any seasonal variation. Theoretical calculations for model clouds show that transmissivity varies with solar zenith angle and either cloud optical depth and drop size distribution (Welch *et al.*, 1980) or liquid water content (Stephens, 1978). Transmissivities determined empirically from surface measurements of global radiation vary weakly with zenith angle. The dependence is overshadowed by larger variations of transmissivity with cloud thickness and drop size distribution (Welch *et al.*, 1980). The

TABLE 1. Frequency of occurrence of hourly daylight observations of total cloud amount and measured solar irradiance. Values are for the period 1973–78 for Goose, Montreal, Toronto, Winnipeg and Vancouver (combined).

	Cloud amount										
	0.0	0.1	0.2	0.3	0.4	0.5	0.6	0.7	0.8	0.9	1.0
Total cloud amount	7	9	7	5	4	4	4	6	8	17	29
Solar irradiance (percent of annual irradiance)	10	11	8	7	6	5	6	7	9	15	16

use of constant transmissivity values in (1) has not increased the error in radiation estimates (Howard, 1982).

Results of the four studies (Table 2) show considerable variation of transmissivity within similar climates. In particular, the Hamburg transmissivities are much smaller than the North American values. However, successful use of the Blue Hill transmissivities in radiation calculations throughout North America (Atwater and Ball, 1978; Davies *et al.*, 1975; Davies and McKay, 1982; Powell, 1980; Suckling and Hay, 1977) and Australia (Lyons and Edwards, 1982) does not suggest that regional and seasonal variations are important. Atwater and Ball (1981a) also suggested that transmissivity for a particular cloud type decreases with increasing cloud amount. Radiation estimates for GATE were improved by incorporating this variation into their model. They concluded that this modification would lead to improved results with similar models.

In this paper, several procedures for calculating transmissivities are discussed, values determined for selected Canadian stations are compared, variations with season and cloud cover are examined, and the relative success of different transmissivities in radiation calculations is assessed.

2. Determination of transmissivities

True cloud transmissivity can only be determined consistently from airborne radiation measurements above and below cloud. Such measurements have been made rarely and only for short periods. Models require climatological values based on sufficiently long-term data to represent the variation of transmissivity for individual cloud types. Such transmissivities have been determined from surface measurements and observations. Cloud transmissivity can be calculated from Eq. (1) when there is a single cloud layer. It has usually been calculated as the ratio of measured radiation G_m beneath overcast of a single cloud type and cloudless sky irradiance for the same optical air mass. The latter irradiance was obtained as a long-term climatological average of measured values by Haurwitz (1948), Vowinckel and Orvig (1962) and Kasten and Czeplak (1980), while Atwater and Ball (1981b) used calculated values. Depending on the reliability of observed cloud estimates, the use of overcast data eliminates error in specifying cloud amount (Atwater and Ball, 1981a). However, this method unavoidably ignores effects of clouds above the overcast. Therefore, calculated transmissivities, on average, will be smaller than true transmissivities and the difference should be greatest for middle and low clouds.

There is a potentially important difference in the manner in which transmissivities have been calculated. Atwater and Ball calculated transmissivity t' from Eq. (1) with the effect of multiple reflections between ground and atmosphere removed:

$$t' = G_m(1 - \alpha b)/G_0, \tag{2}$$

TABLE 2. Cloud transmissivities from Haurwitz (1948), Vowinckel and Orvig (1962), Kasten and Czeplak (1980) and Atwater and Ball (1981b). Range of values for air masses between 1 and 3 are given for Blue Hill, Dartmouth and Edmonton.

		Vowinckel and Orvig				
			Edmonton			
Cloud type	Haurwitz Blue Hill	Dartmouth	Winter	Summer	Kasten and Czeplak Hamburg	Atwater and Ball
Ci	82–85	69–91	83	91–98	61	90
Cs	71–84	—	—	—	61	90
Ac	47–52	43	64–78	42–50	27	50
As	41	36	53	33–41	27	40
Cb	—	—	—	—	—	22
Cu	—	—	—	—	25	36
Sc	33	42	49	35	25	25
St	25	21–29	38	30	18	15
Ns	15–25	—	—	—	16	15
F	17	—	—	—	—	69

whereas Haurwitz (1948), Vowinckel and Orvig (1962) and Kasten and Czeplak (1980) calculated a transmissivity t'' from irradiances that included multiple reflection effects:

$$t'' = G_m/G_0. \quad (3)$$

Therefore, t'' will be larger than t' for a given cloud type. Because the surface albedo for the tropical ocean is small, values of t' and t'' are similar and the Atwater and Ball transmissivities can be compared with those from the other studies. Over land, especially in Canadian winter conditions, higher surface albedo values would ensure larger differences between t' and t''. Using values of 0.35 and 0.6 for the albedo of cirrus and low cloud (Paltridge and Platt, 1976), t'' values will be larger than t' values by 8–14% in summer (surface albedo of 0.2) and by up to 27–56% in winter when there is snow cover (surface albedo of 0.6).

In an atmosphere with three cloud layers where the lowest (layer 3) is overcast, the total atmospheric transmissivity is given by

$$\frac{G_m}{G_0} = t_3\psi_1\psi_2/(1 - \alpha b) = t'_3/(1 - \alpha b) = t''_3, \quad (4)$$

where $\psi_i = 1 - C_i + t_iC_i$. Hence, $t'_3 < t_3$, t''_3. This inequality also applies to transmissivities determined for overcast in the second layer. For high-level overcast, $t_1 = t'_1$ and both are less than t''_3.

If values of t'_i and t''_i are used in Eq. (1) to calculate atmospheric transmissivity (T', T'') in all sky conditions, then $T' < T$, T'', where T is atmospheric transmissivity calculated using t_i. Therefore, the use of t''_i ensures that, on average, radiation attenuation by cloud above the overcast is compensated to some extent by multiple reflection enhancement. Because the Blue Hill transmissivities have produced irradiance estimates without systematic error, this compensation must be effective and true cloud transmissivities are more closely approximated by t'' than t' estimates. Differences between t' and t'' and their effect on calculated irradiances will be shown later.

3. Data

Hourly integrated values of sunshine and solar radiation (LAT) and hourly cloud observations (LST) for Vancouver, British Columbia; Winnipeg, Manitoba; Toronto, Ontario; Montreal, Quebec; Charlottetown, Prince Edward Island; and Goose, Labrador for 1968–76 were used to calculate transmissivities. The time scales were approximately aligned by assigning each hourly meteorological observation time to the solar time of the nearest midpoint of the radiation integration period.

Two separate data sets were extracted for each cloud type. The first consisted of individual hourly observations of overcast for one cloud layer of one cloud type. This was essentially the selection criterion used by Haurwitz (1948) and Vowinckel and Orvig (1962). The second criterion, derived from the first, consisted only of data with overcast at the start and end of an hour in the cloud data sets. This selection procedure, which was used by Kasten and Czeplak, minimizes the chance of using data where overcast did not persist throughout the radiation integration period. Although about 30% of daylight hours are overcast (Table 1), fewer than 10% satisfy the criteria for the first procedure and fewer than 5% for the second procedure (Table 3). Usable data sets were obtained only for altocumulus Ac, altostratus As, cirrostratus Cs, cirrus Ci, stratus St, stratocumulus Sc and fog F. Small data sets for nimbostratus Ns, stratus fractus Sf and cumulus fractus Cf were combined with St.

Theoretical cloudless sky values of spectrally integrated global irradiance and surface albedo were calculated by the procedures described, and applied to the same stations by Davies and McKay (1982).

TABLE 3. Number of overcast hours for each cloud type selected by the procedures of Haurwitz (1) and Kasten and Czeplak (2).

	Vancouver		Winnipeg		Toronto		Montreal		Charlottetown		Goose	
Cloud type	1	2	1	2	1	2	1	2	1	2	1	2
Ac	49	8	218	49	191	27	212	42	147	25	52	9
As	25	8	170	46	92	11	97	12	62	13	81	17
Cs	53	10	164	47	119	28	98	24	36	7	113	34
Ci	57	12	162	28	92	19	39	0	18	5	97	19
Cu	0	0	16	0	64	5	3	0	0	0	0	0
St, Ns, Sf, Cf	143	31	499	212	167	52	141	50	254	63	69	24
Sc	325	87	1545	638	1094	381	1069	320	1273	481	756	266
F	117	64	69	25	252	132	59	29	287	134	34	11
Total	769	220	2843	1045	2071	655	1718	477	2077	728	1202	380
Percent of all hours	2	1	9	3	6	2	5	1	8	3	4	1
Percent of overcast hours	9	3	31	11	23	7	19	6	23	8	10	3

Atmospheric backscatter for (2) was generally equated with cloud albedo, except at Montreal and Toronto where the cloud component was augmented by an aerosol component of 0.02. Values of 0.6 and 0.55 were used for low- and middle-cloud albedo. Since the transparency of cirriform cloud can vary widely, high-cloud albedo was allowed to vary linearly between 0.15 for thin cloud and 0.35 for thick cloud as a function of measured bright sunshine duration. The two extreme conditions were associated with maximum and zero sunshine duration, respectively.

4. Results

a. *Average transmissivities*

Transmissivities were computed for each hourly radiation measurement under overcast. To avoid unrealistic transmissivity values at large zenith angles, only data for air mass values $\leqslant 5$ have been used. Means and standard deviations for each station and for pooled data from the six stations are presented for four sets of transmissivities in Table 4. Sets H1 and K1 were calculated from irradiances selected by the Haurwitz (H1) and Kasten and Czeplak (K1) procedures, while sets H2 and K2 are corresponding results for irradiances without the multiple reflection component. Mean transmittances calculated from fewer than 20 hours at individual stations have been omitted in Table 4 although data for these hours were used to obtain the pooled results.

With the exception of Ac and F, transmissivities in H1 for the pooled data agree well with Blue Hill values. However, there are geographical variations. Except for F, transmissivities for eastern Canada

TABLE 4. Means and standard deviations (in parentheses) of transmissivities. Asterisks indicate means calculated from less than 50 hours of data. Dashes indicate less than 20 hours of data.

Cloud type	Blue Hill	Vancouver	Winnipeg	Toronto	Montreal	Charlottetown	Goose	Pooled
H1. By Haurwitz' method (uncorrected irradiances):								
Ci	82–85	83 (12)	79 (15)	88 (10)	84 (12)	83 (15)	84 (3)	83 (13)
Cs	75–84	75 (14)	76 (15)	76 (14)	76 (13)	69 (16)	75 (17)	75 (15)
Ac	47–52	25* (11)	46 (21)	38 (19)	36 (16)	37 (19)	42 (17)	39 (16)
As	41	29* (12)	52 (17)	47 (19)	42 (16)	45 (19)	46 (17)	46 (17)
Cu	—	—	—	28 (19)	—	—	—	28 (19)
Sc	31–35	18 (12)	41 (20)	34 (18)	25 (16)	37 (20)	36 (17)	34 (18)
St/Ns	25	25 (16)	39 (17)	27 (14)	23 (17)	30 (17)	33 (13)	32 (16)
F	17	44 (21)	38 (19)	27 (18)	42 (23)	28 (17)	29 (12)	32 (18)
K1. By Kasten and Czeplak's method (uncorrected irradiances):								
Ci	61	—	78* (15)	—	—	—	—	84 (12)
Cs	61	—	75* (11)	79* (10)	76* (14)	—	75* (18)	75 (14)
Ac	27	—	35* (8)	33* (21)	36* (14)	28* (16)	—	33 (17)
As	27	—	50* (17)	—	—	—	—	45 (16)
Cu	25	—	—	—	—	—	—	—
Sc	25	15 (9)	34 (17)	30 (15)	23 (14)	31 (15)	30 (13)	30 (15)
St/Ns	16–18	13* (9)	34 (14)	21 (10)	17 (11)	22 (13)	24 (13)	29 (12)
F	—	36 (16)	25* (9)	26 (15)	44* (24)	24 (13)	—	28 (15)
H2. By Haurwitz' method (corrected irradiances):								
Ci	82–85	80 (13)	73 (16)	83 (9)	78 (13)	80 (15)	77 (15)	77 (14)
Cs	71–84	72* (14)	69 (16)	70 (14)	69 (13)	63 (13)	76 (18)	69 (15)
Ac	47–52	22* (10)	36 (16)	32 (17)	29 (13)	31 (16)	33 (14)	32 (15)
As	41	24* (11)	38 (13)	36 (15)	32 (12)	33 (13)	33 (12)	34 (13)
Cu	—	—	—	23 (15)	—	—	—	23 (15)
Sc	31–35	16 (11)	31 (15)	27 (14)	20 (12)	30 (16)	28 (14)	27 (14)
St/Ns	25	21 (15)	31 (13)	21 (10)	18 (13)	25 (14)	26 (10)	26 (13)
F	17	37 (18)	31 (15)	23 (16)	35 (20)	24 (15)	24 (10)	27 (16)
K2. By Kasten and Czeplak's method (corrected irradiances):								
Ci	61	—	70* (14)	—	—	—	—	78 (12)
Cs	61	—	68* (13)	74* (10)	69* (13)	—	68* (19)	69 (14)
Ac	27	—	27* (12)	28* (18)	29* (9)	24* (13)	—	27 (13)
As	27	—	36* (12)	—	—	—	—	33 (11)
Cu	25	—	—	—	—	—	—	—
Sc	25	13 (8)	25 (11)	24 (11)	18 (11)	24 (12)	22 (9)	23 (11)
St/Ns	16–18	14* (8)	23 (9)	17 (8)	14 (9)	19 (11)	18 (9)	22 (9)
F	—	32 (14)	21* (6)	22 (13)	37* (22)	21 (11)	—	24 (13)

(Goose, Charlottetown, Montreal and Toronto) and Blue Hill are similar, while those for Winnipeg are larger for low and middle-level cloud and those for Vancouver are smaller for middle cloud and Sc.

Differences in transmissivities between sets H1 and K1 are small for pooled data though larger for some cloud types at individual stations, especially Winnipeg. The Hamburg transmissivities are not well matched for high and middle cloud except for Ac at Charlottetown. The differences between the Blue Hill and Hamburg values cannot be accounted for by differences in methods of data selection but must reflect differences in climate. Closer examination of Canadian high-cloud values suggests that the large differences for high cloud may be due to an additional influence. The distribution of pooled transmissivities for Cs in K1 is bimodal; 70% represent either thick or thin cloud as characterized by bright sunshine measurements of 0 and 100%, respectively (Fig. 1). The corresponding mean transmissivities of 63 and 87% compare reasonably with the Hamburg value of 61% and the upper limit of 84% for Blue Hill. The mean value for thin Ci (100% sunshine) for pooled data is 90% but there were insufficient data to calculate a worthwhile mean value for thick Ci (0% sunshine). Denser cirriform cloud over Hamburg than over both Canada and the northeastern United States (the latter between 1938 and 1945) is consistent with greater amounts of cirrus over Western Europe at the eastern end of the North Atlantic flight route due to aircraft condensation trails (Freeman and Liou, 1979).

In previous work transmissivities for Cu have been determined only for Hamburg and the tropical Atlantic. Overcasts of this cloud type were only reported at Toronto. The H1 and H2 transmissivities of 28 and 23% are similar to the Hamburg value of 25%

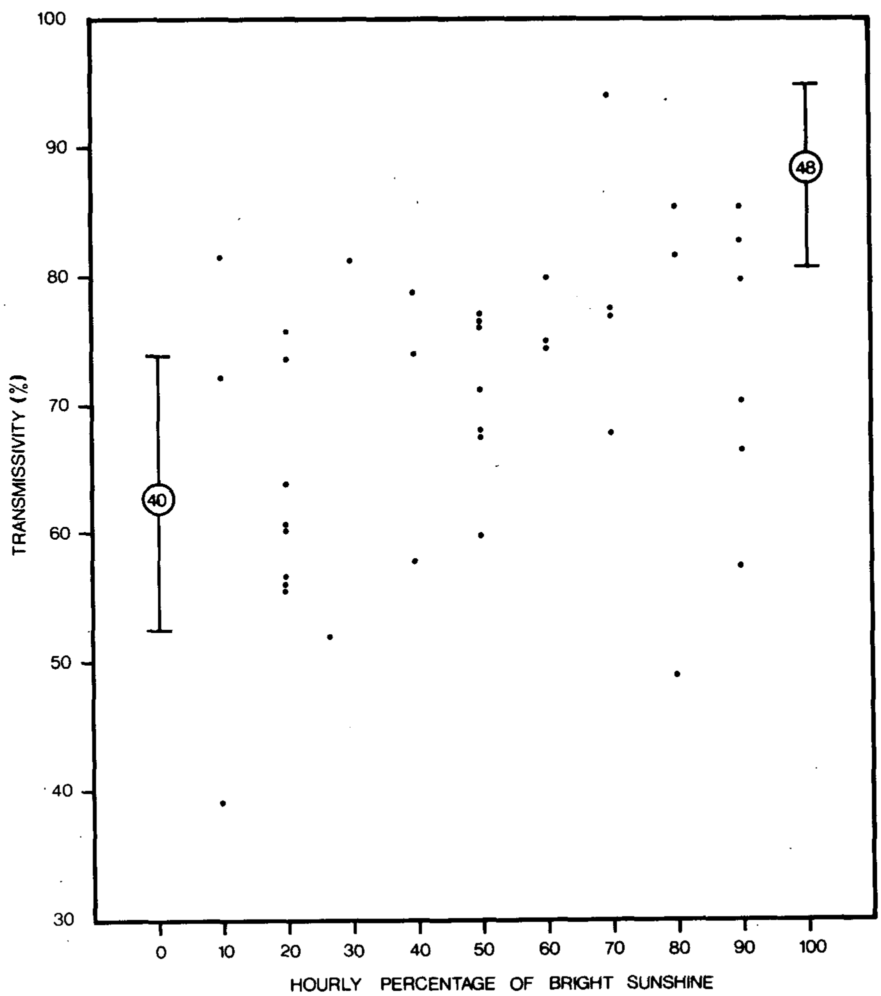

FIG. 1. Distribution of transmissivity (set K1 pooled values) for Cs according to hourly percentage of measured bright sunshine. Mean transmissivity and standard deviation for the number of values indicated are plotted for 0 and 100% sunshine. Individual transmissivities are plotted for intermediate sunshine values.

TABLE 5. Values of t''/t' for different cloud types.

	t''/t'							
	Ci	Cs	Ac	As	Cu	Sc	St	F
H1/H2	1.08	1.10	1.22	1.35	1.22	1.26	1.24	1.19
K1/K2	1.08	1.09	1.22	1.41	—	1.30	1.32	1.17

but smaller than the value of 36% advocated by Atwater and Ball (1981b).

Standard deviations for transmissivities are generally large. Since values do not increase for middle and low clouds, fluctuations in thickness and optical properties of the overcast cloud type must be more important than the effect of variable cloud amount and type above. The use of sequential overcast hours reduces standard deviations of pooled results by 1–4%, and correction for multiple reflection effects leads to a further reduction of similar magnitude.

The removal of multiple reflection effects from irradiances reduces mean values of calculated transmissivities by 4–12%. Values of t''/t' are similar using pooled results from either sets H1 and H2 or K1 and K2 (Table 5). If t'' approximates true transmissivity, values of t''/t' for low and middle clouds (1.17–1.41) imply average transmissivities of about 70–85% for unseen cloud fields above the overcast.

b. Seasonal variation of transmissivity

Only Vowinckel and Orvig (1962) have shown seasonal variations in transmissivity. They attributed larger transmissivities for low and middle cloud at Edmonton in winter to thinner cloud due to weaker convection. To determine seasonal variation, transmissivities in sets H1 and H2 were grouped into cool (C) season (November–March) and warm (W) season (April–October) categories (Table 6).

In both seasons transmissivities are generally larger at Edmonton than at the other stations. Values for Winnipeg, the closest station to Edmonton, are simialr in winter but smaller in summer. These differences are unlikely to be due to differences in the calculation of transmissivities. Although Vowinckel and Orvig used measured rather than calculated cloudless sky irradiances, differences in transmissivities by the two methods are less than 3%.

With the exception of F and all low cloud at Vancouver, transmissivities for low and middle cloud are larger in winter. This would seem to support the conclusion of Vowinckel and Orvig. However, seasonal differences for low and middle cloud decrease when multiple reflection effects are removed (set H1/H2). In the case of the eastern stations, most of the seasonal differences can be attributed to these effects. Thus, the Vowinckel and Orvig hypothesis is not supported. Differences in winter transmissivities between sets H1 and H2 are greater at Winnipeg than at the other stations. This may be due to thinner cloud in a drier winter atmosphere.

Although at most stations cloud transmissivities vary seasonally, it will be shown later that model estimates of irradiance do not show seasonal bias.

c. Model performance with revised transmissivities

Results of irradiance calculations for Winnipeg and Montreal for 1976 using transmissivities from Tables

TABLE 6. Seasonal variation in mean cloud transmissivity for data sets A and C. Values for Edmonton are from Vowinckel and Orvig (1982); C and W denote cool and warm seasons, respectively. Asterisks indicate means calculated from less than 50 hours of data. Dashes indicate less than 20 hours of data.

Cloud type	Data set	Vancouver		Edmonton		Winnipeg		Toronto		Montreal		Charlottetown		Goose		Pooled	
		C	W	C	W	C	W	C	W	C	W	C	W	C	W	C	W
Cs	H1	—	74*	—	—	73	78	76*	76	79	72*	—	68*	74	76	75	75
	H2	—	71*	—		64	74	68*	71	69	68	—	65*	64	71	66	71
Ci	H1	—	81*	81	92	79	80	—	88	—	87	—	—	86*	83*	81	84
	H2	—	81*	—	—	69	77	—	83	—	83	—	—	76*	79*	72	80
Ac	H1	—	25*	78	46	56	30	46	36	47	31	43*	32	—	35*	48	32
	H2	—	22*	—	—	39	26	35	32	34	28	32*	28	—	28*	35	28
As	H1	—	—	54	39	52	37	49	42*	46	—	47*	36*	47	34*	48	37
	H2	—	—	—	—	36	32	36	36*	32	—	33*	30*	31	27*	33	32
Sc	H1	16	22	49	36	48	30	38	30	28	23	38	32	38	32	37	29
	H2	14	19	—	—	33	26	29	26	22	20	29	28	26	26	27	25
St	H1	18	33	36	32	41	26	24	20	21	18	27	26	38*	23	29	25
	H2	15	29	—	—	29	22	19	18	16	16	21	23	26*	19	22	22
F	H1	43	44*	—	—	31*	38*	24	34	29*	52*	25	27	—	27*	29	33
	H2	36	39*	—	—	24*	33*	20	30	24*	45*	21	24	—	23*	24	29

TABLE 7. Mean bias error (MBE) and root-mean-square error (RMSE) values for Winnipeg and Montreal in 1976 using the cloud transmissivity results of Tables 4 and 6. AH1(SEA) refers to seasonal transmissivities H1 (SEA). Upper values are in MJ m^{-2} day^{-1} and lower values in percent of mean measured daily irradiance G_m for net AH1 in Table 6.

		Station transmissivities					Pooled transmissivities				
Station	G_m	H1	K1	H2	K2	AH1(SEA)	H1	K1	H2	K2	AH1(SEA)
MBE:											
Winnipeg	14.40	0.17 1.2	−0.42 −2.9	−0.91 −6.3	−1.49 −10.3	−0.09 −0.7	−0.2 −1.4	−0.39 −2.7	−1.08 −7.5	−1.24 −8.6	−0.41 −2.9
Montreal	11.95	−0.84 −7.0	−0.98 −8.2	−0.98 −8.2	−1.77 −14.8	−0.92 −7.7	−0.33 −2.8	−0.69 −5.7	−1.28 −10.8	1.59 −13.3	−0.69 −5.8
RMSE:											
Winnipeg		1.78 12.4	1.77 12.3	1.99 13.8	2.42 16.8	1.78 12.4	1.7 11.8	1.78 12.4	2.10 14.6	2.25 15.6	1.77 12.3
Montreal		2.17 18.1	2.22 18.6	2.22 18.6	2.71 22.7	2.24 18.8	2.04 17.1	2.17 17.6	2.38 19.9	2.58 21.6	2.19 18.4

4 and 6 are given in Table 7. Mean bias error (MBE) and root-mean-square error (RMSE) values are presented. Transmissivities from sets H2 and K2 systematically underestimate irradiance, thus supporting the argument in Section 2. The best results were obtained with transmissivities from sets H1 and K1 computed from pooled data. This suggests that the overcast sky data sets for individual stations were not sufficiently representative and that transmissivity did not vary significantly over the range of climate conditions represented in this study. It is also clear that separate transmissivities for the cool and warm seasons are unnecessary. The absence of seasonal bias in calculated irradiances is apparent in plots of measured and calculated mean daily irradiance for Winnipeg in 1976 (Fig. 2). Differences are not statistically significant. Calculations were made with set H1 transmissivities for pooled data. The Winnipeg results are representative of plots for other years and other stations.

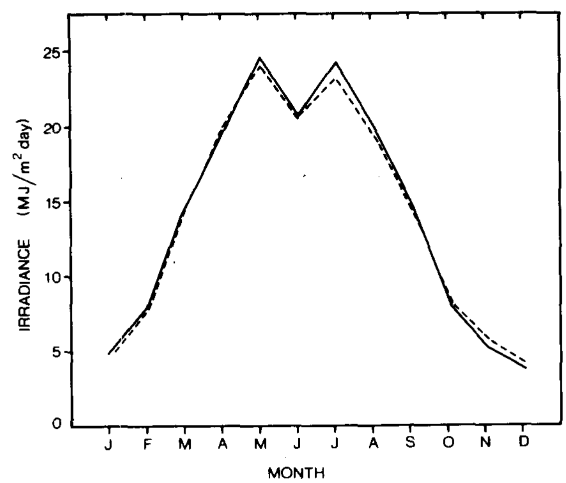

FIG. 2. Variation of measured (solid line) and calculated (dashed line) mean daily irradiance for Winnipeg in 1976.

Irradiances estimated with the "best" transmissivities from this study were compared with estimates using previous transmissivities. The following sets of transmissivities were used:

- H1(S): Station values for H1 (Table 4).
- H1(P): Pooled values for H1 (Table 4).
- H1(P) Cu: Pooled values for H1 with Cu transmissivity determined from partial cloud cover as discussed by Atwater and Ball (1981b). Three years of hourly data (1974–76) for Winnipeg, Toronto, Montreal and Goose were used in the analysis. Mean values of 47, 45, 44 and 46% were obtained for these stations, respectively. An average of 46% was used for Cu in these calculations.
- 3T: A single transmissivity for each cloud layer. Using the pooled values for H1 in Table 4, transmissivities of 78, 42 and 32% were obtained for high, middle and low cloud.
- AB1 Atwater and Ball's transmissivities (Table 2) raised to the power of the layer cloud amount as they recommend.
- AB2 Atwater and Ball's transmissivities held constant with cloud amount.
- KC: Kasten and Czeplak's transmissivities (Table 2).
- BH1: Blue Hill transmissivities expressed as exponential functions of optical air mass (Davies and McKay, 1982).
- BH2: Haurwitz' transmissivities (Table 2).

Results (Table 8) show that recent evaluations of transmissivities have not improved upon the effectiveness of the Blue Hill transmissivities. Kasten and Czeplak's transmissivities are inappropriate in Canada

TABLE 8. Model performance with cloud transmissivities from this study, Haurwitz (1948), Atwater and Ball (1981) and Kasten and Czeplak (1980). Upper values are in MJ m^{-2} day^{-1} and lower values in percent of mean measured daily irradiance G_m.

Station	G_m	H1(S)	H1(P)	H1(P)Cu	3T	AB1	AB2	KC	BH1	BH2
MBE:										
Vancouver	10.5	−0.79	−0.29	−0.19	−0.33	1.82	0.05	−1.88	−0.04	−0.42
		−7.6	−2.8	−1.8	−3.1	17.4	0.4	−17.9	−0.4	−4.0
Winnipeg	14.40	0.17	−0.2	0.02	−0.27	1.74	0.43	−1.96	−0.04	−0.41
		1.2	−1.4	0.1	−1.9	12.1	3.0	−13.6	−0.3	−2.8
Toronto	13.42	−0.80	−0.87	−0.53	−0.90	0.83	−0.56	−2.54	−0.62	−0.68
		−5.9	−6.5	−3.9	−6.7	6.2	−4.1	−18.9	−4.6	−5.1
Montreal	11.95	−0.84	−0.33	−0.21	−0.4	1.42	−0.04	−2.02	−0.08	−0.14
		−7.0	−2.8	−1.8	−3.3	11.9	−0.3	−16.9	0.7	−1.1
Charlottetown	12.67	0.07	0	0.12	−0.09	1.23	−0.10	−1.76	0.01	0.06
		0.05	0	1.0	−0.7	9.7	−0.8	−13.9	0.1	0.5
Goose	10.75	−0.28	−0.04	0.28	−0.02	1.94	0.33	−1.75	0.19	0.23
		−2.6	−0.4	2.6	−0.2	1.81	3.1	−16.3	1.8	2.1
RMSE:										
Vancouver		2.21	1.93	1.86	1.91	2.71	1.98	3.05	1.79	1.88
		21.1	18.4	17.7	18.2	25.9	18.8	29.1	17.1	18.0
Winnipeg		1.78	1.7	1.51	1.66	2.93	2.19	2.87	1.74	1.86
		12.4	11.8	10.5	11.5	20.4	15.2	19.9	12.1	13.0
Toronto		2.4	2.45	2.15	2.43	2.39	2.46	3.47	2.28	2.29
		17.9	18.3	16.0	18.1	17.80	18.4	25.9	17.0	17.1
Montreal		2.17	2.04	1.92	2.05	2.56	2.24	2.98	1.91	1.91
		18.1	17.1	16.1	17.2	21.4	18.7	24.9	16.0	16.0
Charlottetown		2.13	2.2	2.12	2.20	2.89	2.76	2.93	2.13	2.19
		16.8	17.3	16.8	17.4	22.8	21.8	23.1	17.1	17.3
Goose		2.2	2.22	2.04	2.19	3.26	2.63	3.09	2.1	2.15
		20.5	20.6	18.9	20.5	30.4	24.5	28.7	19.6	20.1

and the Atwater and Ball values, and their suggested variation with cloud amount, offer no improvement. The use of Cu transmissivity from the partial cloud analysis improves results slightly while the use of fixed transmissivities for individual cloud layers essentially reproduced results from pooled transmissivities. Differences between the results from the two sets of Blue Hill transmissivities indicate little advantage in allowing transmissivities to vary with optical air mass.

d. Consideration of the Atwater and Ball hypothesis

Atwater and Ball (1981b) found during Phase 3 of GATE that the use of transmissivities based on Blue Hill data in (1) underestimated hourly solar irradiance for cloud cover by between 0.4 and 0.8. The systematic error was reduced by allowing transmissivity to vary with cloud amount. Davies *et al.* (1975) noted similar underestimation in southern Ontario using data collected for IFYGL. Incorrect cloud transmissivities and neglected contributions to irradiance from reflections from sides of clouds were suggested as possible explanations.

Recently, extensive calculations with (1), using Blue Hill transmissivities (H1), have been made for Goose, Montreal and Winnipeg. Mean hourly values of measured and calculated irradiance for the 1968–76 period are shown as a function of total cloud amount in Fig. 3. The model underestimates, as in the GATE and IFYGL studies, but there is seasonal variation (Fig. 4). At Goose and Winnipeg, underestimation is strongest in summer, weaker in fall and spring and is replaced by overestimation in winter. On the basis of the seasonal transmissivity results, the model, using average cloud transmissivities, would be expected to overestimate in summer and underestimate in winter. Further, since the model underestimates direct beam irradiance in a similar manner (Figs. 3 and 4) it is unlikely that Atwater and Ball's hypothesis applies.

The systematic differences in both global and direct beam irradiances could be due to small systematic errors in cloud amount. The direct beam underesti-

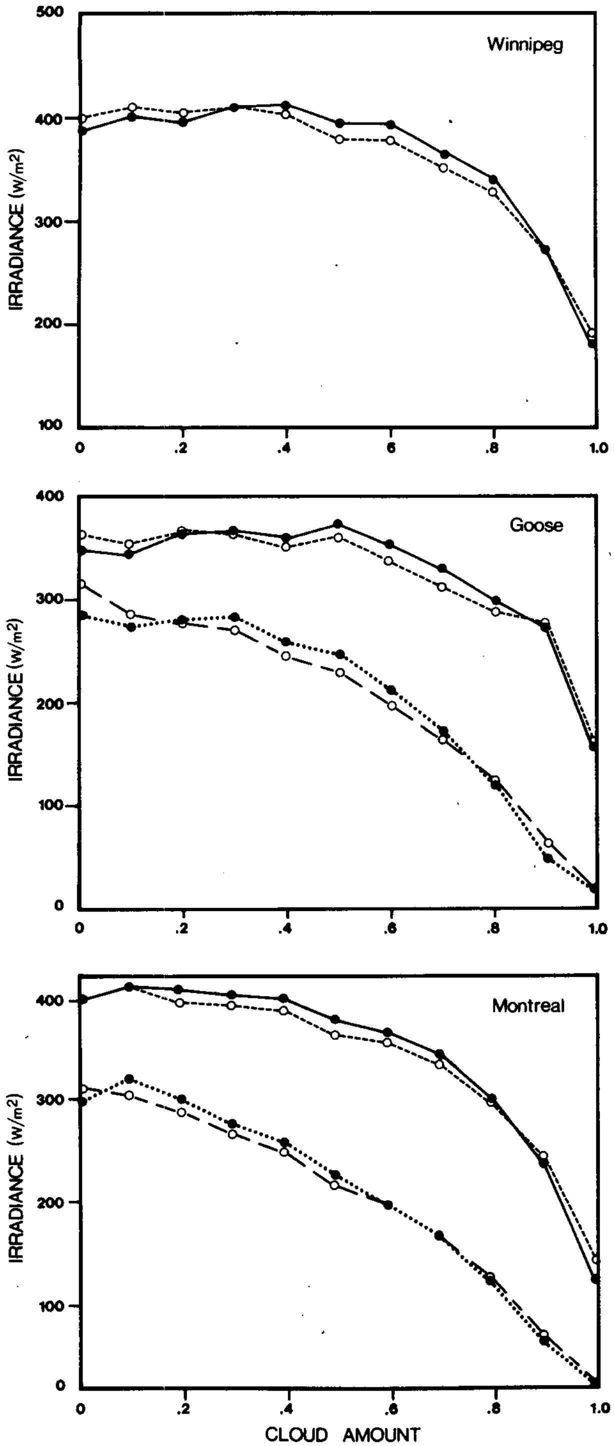

FIG. 3. Variation of mean (1968–76) measured (dots) and calculated (open circles) global and direct beam irradiance with cloud amount.

mates in Fig. 3 can be produced by cloud overestimates of less than 0.05, which is only one third of the average cloud cover overestimate by ground-based observers (Hoyt, 1977). The model calculations of direct beam irradiance I were made with

$$I = I_0(1 - C), \tag{5}$$

where C is total cloud opacity which is usually less than cloud amount. If total cloud amount is used in (5), then irradiance is underestimated by about 23%.

The following explanation is suggested for the seasonal variation in differences between measured and calculated direct beam and global irradiances shown in Figs. 3 and 4. Neglect of aerosol attenuation at Goose and Winnipeg introduced a tendency for both irradiances to be overestimated. This is apparent in winter but is more than offset by larger cloud cover overestimates in summer when, as a result of stronger convection, greater expanses of cloud sides are visible to the observer. At Montreal this seasonal variation is modified by aerosol effects; the most important of these is the transmission function k^m where m is optical air mass. The calculations used a constant value for k (0.91) which was probably not suitable in all seasons. True values are likely to be smaller in summer and larger in winter (Davies and Uboegbulam, 1983). If k was too large in summer by ≤ 0.05, the effect of cloud cover overestimation (as shown for Goose) would be offset. If it was too small in winter by the same magnitude, irradiances would be underestimated by the amounts shown in Fig. 4.

The sum effect of irradiance underestimates for cloud amounts between 0.3 and 0.7 is not large since less than 25% of hourly cloud observations and 30% of the annual irradiance lie within this interval (Table 1). The overall accuracy of model estimates is determined by performance in less cloudy and more cloudy conditions. The model performs well in these conditions (Figs. 3 and 4). It follows that the Blue Hill transmissivities are satisfactory in very cloudy conditions.

These results do not indicate a need to allow cloud transmissivity to vary with cloud amount. The small underestimation of irradiance (less than 30 W m^{-2}) is more likely to be the result of overestimating cloud amount than underestimating transmissivity. Also, few of the differences between measured and calculated values in Figs. 3 and 4 are statistically significant.

5. Conclusions

This study shows that:

1) There is no need to replace the Blue Hill transmissivities with the values of Atwater and Ball (1981b) or to let cloud transmissivity vary with cloud amount as they suggested.

2) Both regional and seasonal variations in transmissivity have negligible effects on calculated irradiance and can be neglected for Canada, at least.

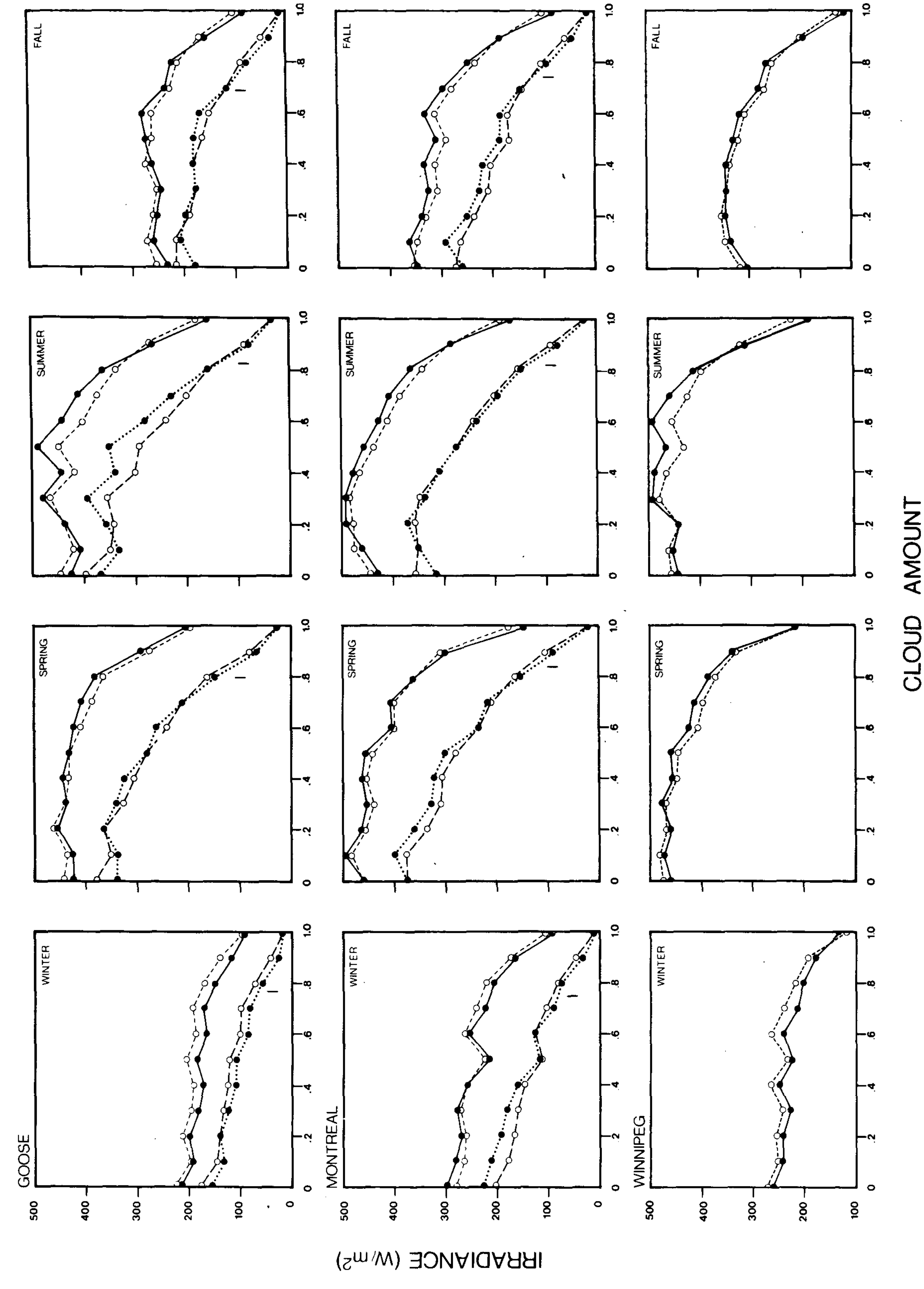

FIG. 4. Seasonal variation of mean (1968–76) measured (dots) and calculated (open circles) global and direct beam irradiance (incident on a horizontal surface) with cloud amount for Goose, Montreal and Winnipeg. Direct beam irradiance is indicated by I.

3) Transmissivities from irradiances with the multiple reflection component removed systematically underestimate irradiance.

4) Transmissivities using all overcast hours and hours with overcast at the start and end are similar.

5) Irradiance calculations can be simplified with little loss in accuracy by grouping cloud type transmissivities into single values for each cloud layer.

The performance of the radiation model is characterized by mean bias errors less than 5% of the mean measured irradiance and daily rms error (RMSE) values between 10 and 20%. The RMSE and the standard deviation of the differences between calculated and measured irradiance are virtually identical and can be considered to measure nonsystematic error (Davies *et al.*, 1984). Large uncertainties in cloud transmissivities and layer amounts, which determine the RMSE values, may prohibit significant reduction in the error statistic. However, results from this model are at least as good as those from recent satellite models (Gautier *et al.*, 1980; Diak and Gautier, 1983;) and regression models.

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