Atmosphere

Volume 15 Number 4 1977

Canadian Meteorological Society
Société Météorologique du Canada

Atmosphere

Volume 15 Number 4 1977

Contents

ISSN 0004-6973

Canadian Meteorological Society
Société Météorologique du Canada

EDITORIAL

This issue of *Atmosphere* is the last in its present format. *Atmosphere* will be replaced by ATMOSPHERE-OCEAN at the beginning of volume 16. In addition to the change of name, which is the result of our Society having formally incorporated the Canadian oceanographic community, we are taking this opportunity to make certain physical and administrative changes arising out of the continuing development of our journal.

The new Editor in liaison with Council will institute a number of modifications in the administrative structure and composition of the Editorial Board from that at present (see inside front cover). These entail a formal recognition of the special responsibilities of certain members of the Board. In the future there will be an Editor, a Deputy Editor, a Technical Editor, a Book Review Editor and a number of Associate Editors. The criteria governing these appointments will be those of expertise in branches of meteorology and oceanography, balanced regional and linguistic representation, and to a lesser degree ease of communication within the Board, and with the printers.

In response to the expressed wishes of the expanded readership the physical appearance of the journal will be modified and we think improved. The cover will be re-designed to visually express the combined interests of our Society in the atmosphere and the oceans. We may also make certain alterations to the page size, the binding, the typesetting format and the inclusion of abstracts in both English and French. The final decisions will, however, have to be strongly influenced by the costs of production.

On the other hand, the editorial policy will remain unchanged. ATMOSPHERE-OCEAN will be the medium of our Society for the publication of the results of original research, survey articles, essays, book reviews, notes and correspondence in all fields of the atmospheric and oceanographic sciences. Foreign contributions in either English or French will continue to be welcomed and it is hoped that Canadians will use ATMOSPHERE-OCEAN as a primary vehicle for the dissemination of their work to the international scientific community.

T.R. Oke

Comparison of Drag Coefficients Over Water Measured Directly and Determined by Wind Profile

Amin H. Meshal*

Atlantic Oceanography Laboratory, Bedford Institute of Oceanography Dartmouth, N.S., Canada

[Manuscript received 12 August 1976; in revised form 26 April 1977]

ABSTRACT

Wind profiles under stable conditions have been observed at heights between 0.5 and 7 m above a water surface. Simultaneous direct measurements of wind velocity fluctuations have been taken by a sonic anemometer. The profiles were closely log-linear in the range of stabilities observed. Our data are best fitted taking von Kármán's constant to be 0.38 ± 0.03 rather than 0.40 as conventionally used. Comparison between the measured and profile-derived drag coefficients shows good agreement. The drag coefficient decreased with increasing stability and did not show clear dependence on wind speed. Wind speeds measured simultaneously by cup and sonic anemometers at the same height were compared. There was no over-speeding of the cup anemometers. The ratio of wind speed measured by cup anemometer to that measured by sonic anemometer was 0.98.

1 Introduction

This paper presents analysis and results of data that include direct measurements of wind fluctuations as well as wind profiles in the atmospheric boundary layer over a water surface. The data were collected under stable conditions during the summer of 1975 (from 15 August to 2 September) at a site in Bedford Basin, N.S., Canada.

The purpose of this paper is to examine the application of the log-linear law to wind profiles in the observed range of stability, to compare the drag coefficient calculated from the profile with that determined directly, and to examine the effect of the variation of stability and wind speed on the drag coefficient. The data were also used to compare wind speed measured simultaneously by cup and sonic anemometers at the same level for estimating the frequently reported overspeeding of the cup anemometer. The sonic anemometer is ideal for this purpose due to its high accuracy, its linearity and because it is an absolute instrument.

*On leave of absence from the Institute of Oceanography and Fisheries, Egypt.

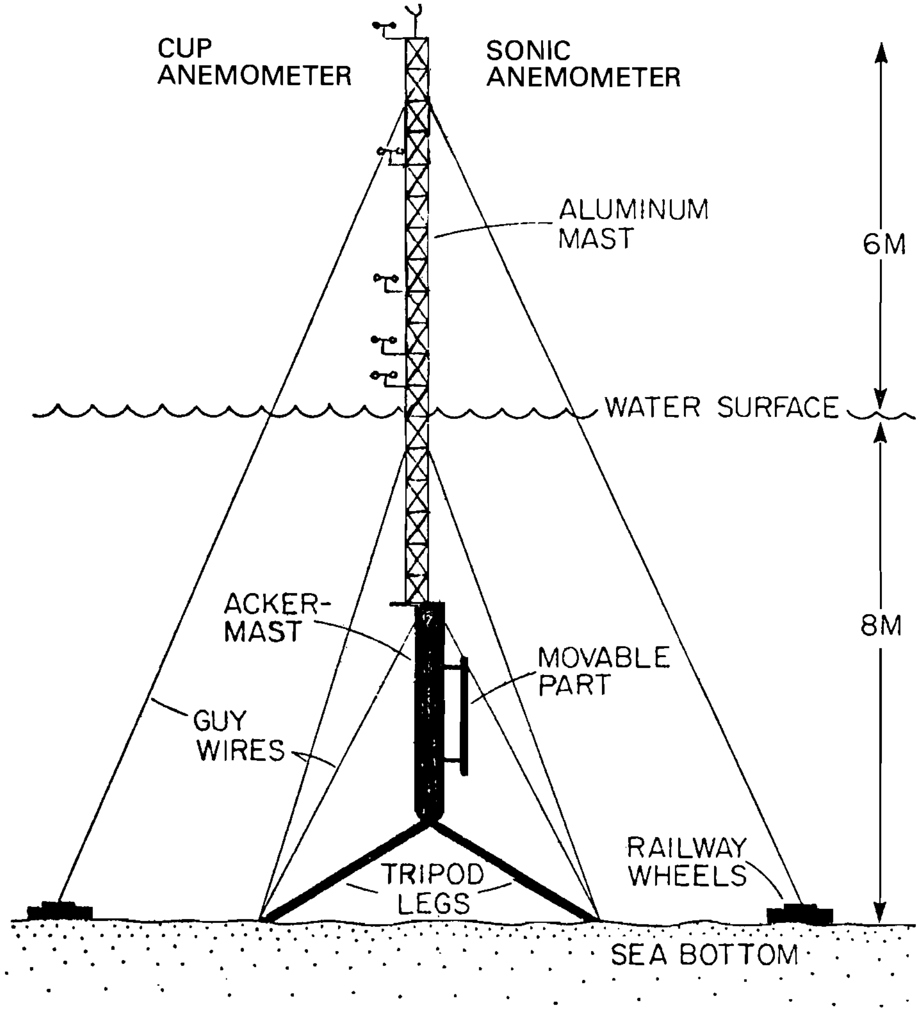

Fig. 1 The tower and the attached mast.

2 Experimental details

The experiment was carried out in Bedford Basin at a site 155 m northwest of the jetty of Bedford Institute of Oceanography (BIO) and 120 m from the nearest shore. At this site a fetch over water of 3.6 to 6.0 km is available for wind coming from azimuths 280 to 330° The depth of water at the lowest tide is 8 m and the tidal range varies from 1.8 to 2.5 m.

The tower (Fig. 1) consists of a steel mast fitted to tripod legs and supported by three guy wires. An aluminum mast of the square lattice type was fixed to the top of the tower.

A Kaijo Denki PAT 311-1 sonic anemometer-thermometer (Mitsuta, 1966) and a microbead thermistor manufactured by Victory Engineering Corporation (model E41A401C) were mounted near the top of the aluminum mast (Fig. 1). A resistance wire wave staff (Taiani, 1971) was used to measure the wave height. Five cup anemometers (Meshal, 1976) were mounted on the mast with vertical separations of 0.5, 1, 2, and 2 m. Anemometers were interchanged frequently so as to reduce errors from calibration.

Signals from the sensors on the mast were transmitted by underwater cables to a recording and monitoring base located on the BIO jetty. For the cup anemometers the accumulated total number of pulses occurring in the previous 30 s were digitally recorded on magnetic tape using a data logger built in the Metrology Electronic Shop, BIO. Smith's (1974a) system was used for recording the signals from the sonic anemometer and the thermistor. The output from the sonic anemometer, the thermistor and the wave staff were digitized by the use of an A-D computer program (Smith, 1974b).

Details concerning the calibration of the sensors have been reported elsewhere (Meshal, 1976).

3 Analysis and results

a *Dimensionless Wind Shear* ϕ_m

The dimensionless wind shear ϕ_m can be defined as (Businger et al., 1971)

$$\phi_m = \frac{kz}{u_*} \cdot \frac{\partial U}{\partial z} \cdot \tag{1}$$

Values of ϕ_m were calculated from (1) by introducing the approximation, suggested by Panofsky (1965), that the derivative of U with respect to z can be given by finite differences as

$$\frac{\partial U}{\partial z} \approx \frac{U_2 - U_1}{\sqrt{z_1 z_2}\,\ln(z_2/z_1)}. \tag{2}$$

The approximation is restricted to logarithmic profiles but for non-neutral profiles an error of about 3% is expected (Paulson, 1967). Computations of ϕ_m from Eq. (2) were first made by taking von Kármán's constant equal to the widely used value of 0.40. The calculated values of ϕ_m were then plotted against z/L (Fig. 2) and the best straight line fit was found by the least squares method. From the plot, $\phi_m(0)$ at neutral stability ($z/L = 0$) was found to be 1.06 rather than 1.0 as supposed. This may be due to the approximation introduced in (2). However, if von Kármán's constant k is taken as 0.377 instead of 0.40, $\phi_m(0)$ will become 1.0. Accordingly, it was decided to use k as 0.38 in all our calculations here with an expected error of $\pm$ 8%. Values of k ranging from 0.34 to 0.41 have been reported in the literature (e.g., Tennekes, 1968; Dyer and Hicks, 1970; Businger et al., 1971; Webb, 1970). A regression line of ϕ_m on z/L has the form

$$\phi_m = 1.0 + 2.5\frac{z}{L} \tag{3}$$

with correlation coefficient of 0.86 and standard error of estimate of 0.09.

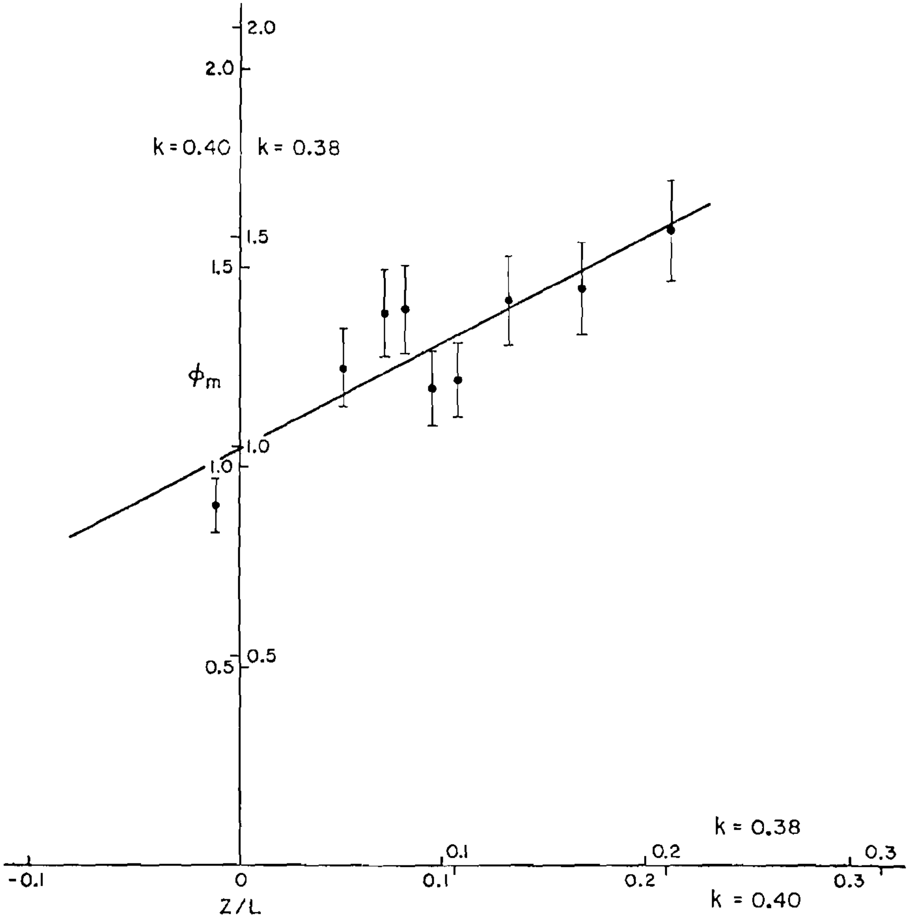

Fig. 2 The relation between ϕ_m and z/L when computations are made with $k = 0.40$ and 0.38.

Businger et al. (1971) found that $\phi_m(0)$ was about 1.15 and they concluded that $\phi_m(0)$ would be 1.0 if k is 0.35 instead of 0.40. They showed that under stable conditions ϕ_m varies linearly with z/L according to the form

$$\phi_m = 1.0 + 4.7\, z/L. \tag{4}$$

Comparison between the two relationships shows that the effect of stability on ϕ_m is stronger in (4) than in (3). However, the two equations become more consistent if we take into consideration the remarks given by Businger et al. (1971) on their equation. They stated that (4) is a good overall fit, but its slope near neutrality changes rapidly and is about 50% greater than indicated by the observations. They estimated the slope to be 3.0 for nearly neutral conditions. Eq. (3) is obtained from nine observations which lie in the region where the slope of ϕ_m changes rapidly.

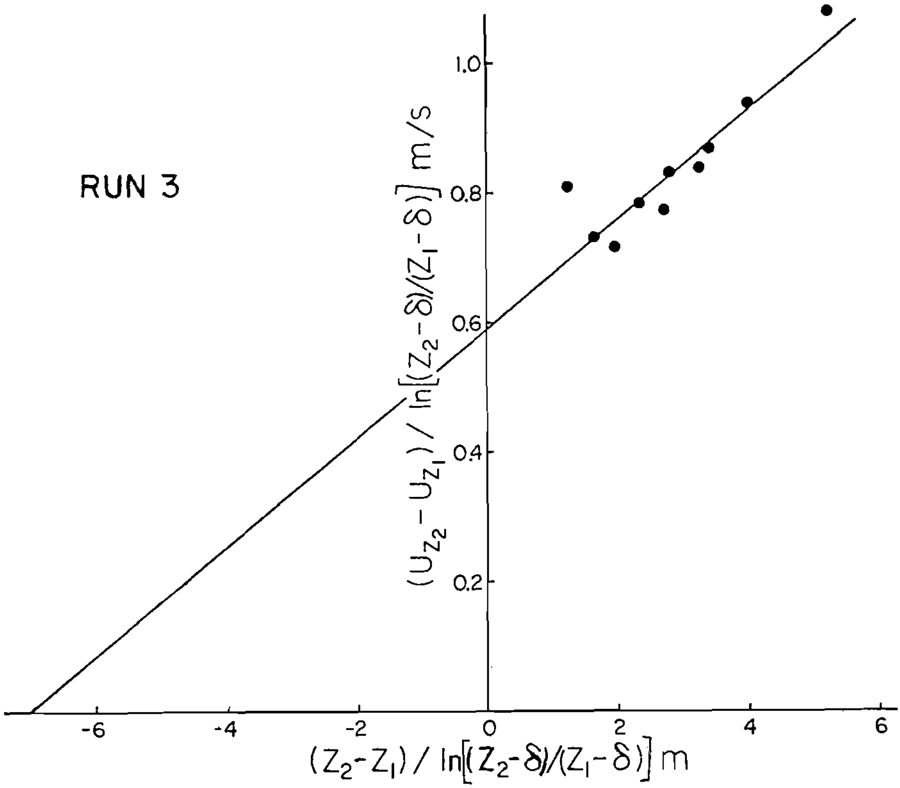

Fig. 3 An example of the log-linear analysis of wind profile.

b *Log-Linear Law*

The log-linear method for analysing wind profiles has been shown to be valid in a small range of unstable and a wide range of stable conditions (Webb, 1970). Since the present data were taken in that range of stability (z/L from -0.01 to $+3.2$) the log-linear law was chosen and the results were found to be in good agreement with simultaneous direct measurements. The influence of thermal stratification on the wind profile is expressed to the first approximation in the Monin-Obukhov (1954) log-linear law as

$$\frac{\partial U}{\partial z} = \frac{u_*}{kz}\left(1 + \alpha\frac{z}{L}\right). \tag{5}$$

Following Webb (1970) (5) is integrated between any two heights z_1 and z_2 and then divided by ln (z_2/z_1) to give

$$\frac{U_2 - U_1}{\ln (z_2/z_1)} = \frac{u_*}{k}\left[1 + \frac{\alpha}{L}.\frac{z_2 - z_1}{\ln (z_2/z_1)}\right]. \tag{6}$$

A zero displacement δ is taken to be half the observed rms wave height so that the effective level of the anemometers is $(z-\delta)$. Putting and

$$\left.\begin{aligned} Y &= \frac{U_2 - U_1}{\ln[(z_2 - \delta)/(z_1 - \delta)]} \\ X &= \frac{z_2 - z_1}{\ln[(z_2 - \delta)/(z_1 - \delta)]} \end{aligned}\right\}, \tag{7}$$

Eq. (6) can be written as

$$Y = \frac{u_*}{k}\left(1 + \frac{\alpha}{L}X\right). \tag{8}$$

The values of Y and X were calculated for each run from every available pair of heights using (7). Y was plotted against X and the best straight line was fitted by the least squares method. The plots are illustrated by an example in Fig. (3). It is evident from (8) that the line intercepts the vertical axis ($X = 0$) at u_*/k and hence u_* can be estimated. The distribution of points in the plots suggests that u_* is better determined by that method than α, since small changes in the slope of the regression line produce larger differences in α/L than in u_*. From (1) and (5) α can be calculated from

$$\phi_m = 1 + \alpha\frac{z}{L}. \tag{9}$$

U_{10} was estimated from (6) and values of U_{10}, u_*, ϕ_m, α, and C_{10}, determined from the profile measurements, are given in Table 1. The average value of α is 3.8 ± 2.9 which is smaller than the value (5.2 ± 0.5) reported by Webb (1970) for stable conditions.

c *Drag Coefficient Measured Directly and Estimated from Wind Profiles*

The average drag coefficient determined from sonic anemometer measurements (Table 1) is $10^3 C_{10} = 1.2 \pm 0.4$ (standard deviation). This average is estimated from 17 runs after excluding the very low values ($<0.3 \times 10^{-3}$) of runs 15 and 19. These two runs are also excluded in the comparison between the sonic- and the profile-derived drag coefficients. The mean drag coefficient estimated from wind-profile data (11 runs, Table 1) is $10^3 C_{10} = 1.1 \pm 0.4$ which is in good agreement with the direct measurements. Comparison between the directly measured and the profile-derived drag coefficient is based on nine simultaneous runs (Table 1). The mean difference between the two sets of results is 10.8% with standard deviation of 0.13×10^{-3}. A regression line of C_{10} calculated from the profile and that determined by the sonic anemometer gives a correlation coefficient of 0.84 with standard error of estimate of 0.20. The mean value of the drag coefficient obtained from the present experiment lies within the range $(1.0 \text{ to } 1.5) \times 10^{-3}$ generally reported over the sea (Smith, 1967; Weiler and Burling, 1967; Hasse, 1968; Smith, 1973).

Concerning the effect of wind speed on the drag coefficient, the results presented here did not show any clear dependence of C_{10} on wind speed. For nearly the same range of wind speed (5–10 m/s), Smith (1967, 1974a) did

TABLE 1. Summary of data from sonic and profile measurements over Bedford Basin

					Sonic anemometer data									Profile data					
Run	Date 1975	Starting time GMT	Duration min	Ht z m	U_{10} m/s	U_{10} deg	u_* cm/s	C_{10} $\times 10^3$	$10^4\langle Tw\rangle$ °C·m/s	T °C	$10^3 z_0$ cm	z/L	Gust factor	U_{10} m/s	u_* cm/s	C_{10} $\times 10^3$	α	ϕ_m	δ m
	Aug.																		
1	15	1652	40	6.5	8.18	300	25.4	0.96	−133	21.3	2.6	0.067	1.54	8.01	25.3	1.00	4.8	1.32	0.13
2	15	1750	22	6.0	7.64	300	25.3	1.10	−198	21.0	5.7	0.093	1.41	—	—	—	—	—	0.13
3	19	1905	43	6.0	6.82	290	27.4	1.61	−275	20.7	46.6	0.102	1.60	6.53	22.5	1.19	1.6	1.16	—
4	19	2245	42	5.5	5.14	300	17.4	1.14	−119	20.7	7.2	0.160	1.58	5.02	16.7	1.11	2.4	1.38	0
5	20	1207	41	5.5	6.03	310	23.6	1.53	+23	17.0	35.8	−0.012	1.73	6.22	22.3	1.29	11.7	0.86	—
6	20	1705	28	7.0	7.79	300	28.9	1.38	−334	21.5	21.1	0.123	1.67	8.09	31.1	1.48	2.8	1.35	—
7	20	2145	43	5.5	8.45	290	33.5	1.57	−298	20.7	41.3	0.059	1.62	—	—	—	—	—	0.23
8	21	1435	43	5.0	9.09	280	29.3	1.04	−244	18.7	4.0	0.062	1.59	—	—	—	—	—	0.24
9	21	1652	30	5.75	8.68	290	36.1	1.73	−368	19.1	67.3	0.077	1.71	9.14	38.5	1.77	4.3	1.33	0.21
10	21	1810	21	7.2	9.19	280	34.1	1.38	−393	19.6	21.1	0.090	1.58	9.52	32.2	1.14	1.6	1.14	0.20
11	21	2230	34	6.0	6.06	280	22.1	1.33	−211	18.2	17.6	0.149	1.64	—	—	—	—	—	0.16
12	23	1430	35	6.4	8.82	290	25.8	0.86	−100	16.3	1.2	0.048	2.04	8.47	26.8	1.00	4.0	1.19	0.19
13	23	1535	43	6.8	9.40	280	34.3	1.33	−163	16.6	17.5	0.035	1.51	—	—	—	—	—	0.20
14	28	1430	42	6.25	7.59	320	22.4	0.87	−297	19.3	1.3	0.210	1.47	7.18	20.6	0.82	2.5	1.52	0.18
15	28	1530	42	6.10	6.51	330	7.1	0.12*	−144	20.2	0	3.18	1.54	6.78	14.9	0.48	1.2	4.97	0.16
16	28	1645	43	6.0	8.43	330	23.2	0.76	−368	19.8	0.5	0.224	1.61	—	—	—	—	—	0.32
17	28	1750	43	6.5	7.84	330	17.8	0.52	−396	21.9	0.02	0.577	1.46	—	—	—	—	—	0.18
18	28	1845	42	6.75	6.57	320	15.4	0.55	−300	22.6	0.04	0.694	1.60	—	—	—	—	—	0.15
19	29	1855	42	6.6	5.47	320	—	−0.28*	−150	20.5	0	—	1.80	—	—	—	—	—	0
	Sept.																		
20	2	1340	42	7.0	5.63	320	9.1	0.26*	−39	17.2	0	0.464	1.26	5.29	13.5	0.65	4.4	3.04	—
mean								1.16					1.60			1.08	3.8	1.75	
standard deviation								0.37					0.16			0.36	2.9	1.20	

*not included in calculating the mean

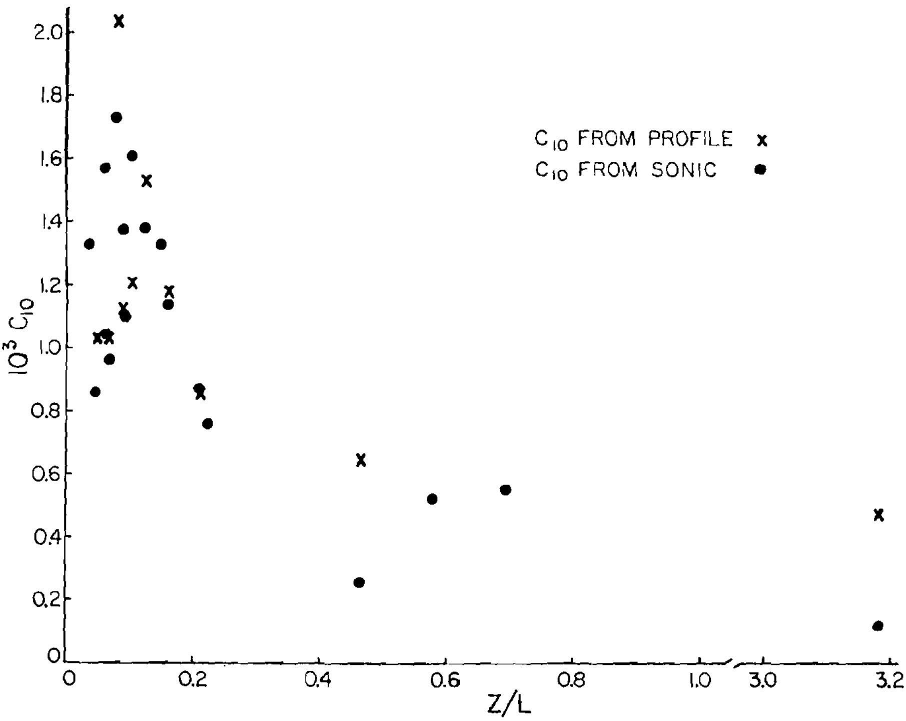

Fig. 4. Drag coefficient as a function of z/L.

not find significant variation of the drag coefficient with wind speed. However, for the very limited range of wind speed reported here one cannot really say much about the dependence of the drag coefficient on wind speed.

The effect of stability variation on the drag coefficient for the present data is illustrated in Fig. 4. Starting from near-neutral stability C_{10} increases with increasing stability up to $z/L = +0.1$ after which value there was a remarkable decrease of C_{10} in the direction of increasing stability. DeLeonibus (1966, 1971) detected a dependence of the drag coefficient on atmospheric stability and his data showed that the drag coefficient was decreasing with increasing stability.

d *Gust Factor*

The gust factor G was determined from sonic anemometer data using the formula

$$G = 1 + (u_{max}/U_z). \tag{10}$$

The calculated values of G vary from 1.26 to 2.04 (Table 1). The results are similar to that of Davis and Newstein (1968) and Monahan and Armendariz (1971) who reported gust factors in the range from 1.0 to 2.2. The mean value of the gust factor (Table 1) $G = 1.60 \pm 0.16$ is higher than that reported

TABLE 2. Comparison between wind speeds measured by cup and sonic anemometers

Run No.	Height z m	Wind speed (m/s) measured by Sonic anemometer U_s	Cup anemometer U_c	Differences $U_s - U_c$ m/s	$\frac{U_s - U_c}{U_s} \times 100$
1	6.50	7.90	7.61	+0.29	+3.7
3	6.00	6.47	6.17	+0.30	+4.6
4	5.50	4.88	4.61	+0.27	+5.5
5	5.50	5.68	5.93	−0.25	−4.4
6	7.00	7.53	7.68	−0.15	−2.0
9	7.75	8.45	8.33	+0.12	+1.4
10	7.20	8.91	9.19	−0.28	−3.1
12	6.40	8.53	8.08	+0.45	+5.3
14	6.25	7.33	6.76	+0.57	+7.8
15	6.10	6.42	6.70	−0.28	−4.4
19	7.0	5.55	5.21	+0.34	+6.1

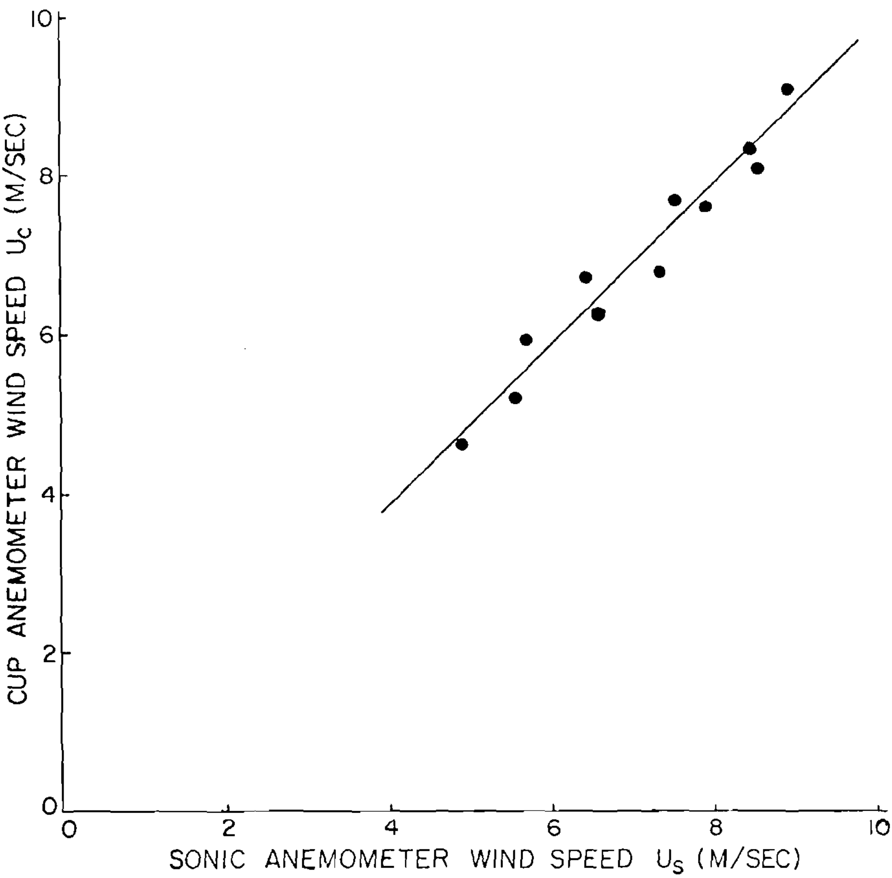

Fig. 5 Comparison between wind speeds measured by cup and sonic anemometers.

by Smith (1973) over the Atlantic Ocean (1.29 ± 0.04), and by Smith (1974a) over Lake Ontario (1.30 ± 0.10). The present data did not show any dependence of the gust factor on stability variations or on wind direction.

e *Comparison Between Wind Speeds Measured by Cup and Sonic Anemometers*

The comparison is based on 11 simultaneous runs (Table 2) with the two anemometers placed at the same height. Wind speeds measured by the cup and sonic anemometers, U_c and U_s, respectively, are plotted in Fig. 5 and the best fit straight line is obtained by the least squares method. The correlation coefficient is 0.97 with standard error of estimate of 0.3. The most striking aspect of this plot is that the cup anemometer did not overestimate the wind speed relative to the sonic anemometer measurements. Table 2 shows that, of the 11 measurements taken, the differences $(U_s - U_c)$ are positive in seven and negative in the other four. The percentage difference is 1.9 ± 4.6. The average ratio of wind speeds measured by cup and sonic anemometer U_c/U_s is 0.98 ± 0.05 and there is no obvious effect of stability or of wind direction on this ratio.

This result contradicts the finding of Izumi and Barad (1970) who reported cup anemometer overspeeding of about 10%. Their measurements were taken on land where the roughness length is several times larger than that on the water. Busch and Kristensen (1976) showed that the overspeeding of the cup anemometer is a function of several parameters among which are the ratio between the distance constant of the anemometer and the roughness length and the ratio between the height of measurement and the roughness length. The overspeeding of our cup anemometers would be less than 0.3% according to their figures, which is smaller than our calibration error.

4 Conclusions

The wind profile and the direct measurements of wind velocity fluctuations taken over sea water and under stable conditions yield the following main conclusions:

(i) The profile analysis showed that the dimensionless wind shear ϕ_m is equal to 1.0 for neutral stability when von Kármán's constant is taken as 0.38.

(ii) The log-linear law used in analyzing the wind profile remained valid over the range of stabilities observed.

(iii) The drag coefficient calculated from the profile is in good agreement with that measured by the sonic anemometer, with a mean difference of 10.8%. The drag coefficient decreased with increasing stability.

(iv) The mean gust factor was 1.6 ± 0.2 and there was no clear effect of stability or of wind direction on this factor.

(v) The cup anemometer did not overestimate the wind speed; the average wind speed measured by the cup anemometer was 0.98 of that of the sonic anemometer. Further investigation is needed to verify this conclusion.

Acknowledgments

My deep gratitude goes to Dr S.D. Smith and Dr F. Dobson of the Air-Sea Interaction Group, Bedford Institute of Oceanography, Dartmouth, N.S., Canada, for providing guidance and offering every possible assistance during the course of this work. The valuable assistance during the field experiment of R. Anderson, D. Hendsbee, K. Stewart, E. Banke, and D. Harvey is sincerely acknowledged. The new circuit for the cup anemometer was designed by K. George and B. Allen of the Metrology Division, Bedford Institute of Oceanography. This study was conducted while the author was holding a postdoctorate fellowship granted by the National Research Council of Canada.

Notation

$C_{10} = \tau/\rho U_{10}^2 = \langle -uw \rangle / U_{10}^2 = u_*^2/U_{10}^2$ wind drag coefficient at 10-m height
$G = 1 + (u_{max}/U_z)$ gust factor
g = acceleration of gravity
$k = 0.38$, von Kármán's constant
$L = -Tu_*^3/gk\langle tw \rangle)$ the Monin–Obukhov length
T = absolute air temperature
t = temperature fluctuation
U = mean wind speed, subscript indicates the height of measurement (m)
u,w = longitudinal and vertical components of wind velocity
$u_* = (\tau/\rho)^{1/2} = \langle -uw \rangle^{1/2}$ friction velocity
z = vertical coordinate or height of measurement above water surface
z_0 = roughness length
α = numerical constant
δ = zero plane displacement.
ρ = density of air
$\tau = -\rho\langle uw \rangle$ wind stress on sea surface
$\phi_m = kz/u_* \cdot \partial U/\partial z$ dimensionless wind shear

Angle brackets denote average over a data run.

References

BUSCH, N.E. and L. KRISTENSEN. 1976. Cup anemometer overspeeding. RISC Report No. 339, Danish Atomic Energy Commission, 20 pp.

BUSINGER, J.A.; J.C. WYNGAARD; Y. IZUMI; and E.F. BRADLEY. 1971. Flux-profile relationships in the atmospheric surface layer. *J. Atmos. Sci.* **28**: 181–189.

DAVIS, F.K. and NEWSTEIN, H. 1968. The variation of gust factors with mean wind speed and with height. *J. Appl. Meteorol.* **7**: 372–378.

DE LEONIBUS, P.S. 1966. Momentum flux observations from an ocean tower. Paper presented at Second Int. Oceanogr. Congr., Moscow, May 30–June 9, 1966.

———. 1971. Momentum flux wave spectra observations from an ocean tower. *J. Geophys. Res.* **76**: 6506–6527.

DYER, A.J. and B.B. HICKS. 1970. Flux gradient relationship in the constant flux layer. *Quart. J. Roy. Meteorol. Soc.* **96**: 715–721.

HASSE, L. 1968. On the determination of vertical transport of momentum and heat in the atmospheric boundary-layer at sea. In *Hamburger Geophysicalische*

Einzelschriften, Heft 11, 70 pp.

IZUMI, Y. and M.L. BARAD. 1970. Wind speeds as measured by cup and sonic anemometers and influenced by tower structure. *J. Appl. Meteorol.* **9**: 851–856.

MESHAL, A.H. 1976. Drag coefficient measured directly and derived from wind profile. Rep. BI-R-76-5, Bedford Institute of Oceanography, Dartmouth, N.S., 21 pp.

MITSUTA, Y. 1966. Sonic anemometer-thermometer for general use. *J. Meteorol. Soc., Japan*, **44**: 12–23.

MONAHAN, H.H. and M. ARMENDARIZ. 1971. Gust factor variations with height and atmospheric stability. *J. Geophys. Res.* **76**: 5807–5818.

MONIN, A.S. and A.M. OBUKHOV. 1954. Basic relationships of turbulent mixing in the surface layer of the atmosphere. *Akad. Nauk. SSSR Trud. Geofiz. Inst.* **24**: 163–187.

PANOFSKY, H.A. 1965. Re-analysis of Swinbank's Kerang observations: Flux of heat and momentum in the planetary boundary-layer. Rep. Dept. Meteorol., Pennsylvania Univ., 66–76.

PAULSON, C.A. 1967. Profiles of wind speed, temperature and humidity over the sea. Unpubl. PH.D. thesis, University of Washington, 116 pp.

SMITH, S.D. 1967. Thrust-anemometer measurements of wind-velocity spectra and Reynolds stress over a coastal inlet. *J. Mar. Res.* **25**: 239–262.

———. 1973. Thrust-anemometer measurements over the sea re-examined. Rept. BI-R-73-1, Bedford Institute of Oceanography, Dartmouth, N.S., 23 pp.

———. 1974a. Eddy flux measurements over Lake Ontario. *Boundary-Layer Meteorol.* **6**: 235–255.

———. 1974b. Program A to D for analog-to-digital conversion and processing of time series data. Computer Note Series/BI-C-74-1, Bedford Institute of Oceanography, Dartmouth, N.S., 67 pp.

TAIANI, P.M. 1971. Nova Scotia Research Foundation, Drawing E-1-1, Flexible Wave Sensor, Dartmouth, N.S.

TENNEKES, H. 1968. Outline of a second-order theory of turbulent pipe flow. *AIAA Journal*, **6**: 1735–1740.

WEBB, E.K. 1970. Profile-relationships: the log-linear range, and extension to strong stability. *Quart. J. Roy. Meteorol. Soc.* **96**: 67–90.

WEILER, H.S. and R.W. BURLING. 1967. Direct measurements of stress and spectra of turbulence in the boundary-layer over the sea. *J. Atmos. Sci.* **24**: 653–664.

The Energy Budget of the Sable Island Ocean Region

E. Vowinckel and Svenn Orvig,
Department of Meteorology, McGill University, Montreal
[Manuscript received 24 June 1976; in revised form 30 June 1977]

ABSTRACT

The heat exchange between ocean and atmosphere over cold water is studied by calculating all terms in the energy balance twice each day for the year 1971 for the Sable Island region.

The atmospheric long-wave radiation is relatively constant because of frequent overcast and low clouds. The surface long-wave balance is markedly negative in winter but slightly positive for a short time in summer, due to strong advection of warm moist air over the cold water. In winter, the turbulent fluxes are directed upwards and are strong, the upward fluxes beginning after the middle of August and lasting until mid-March. The maximum daily values of latent heat flux are 400 to 500 ly day^{-1} (194 to 242 W m^{-2}), about a third or a quarter of the magnitude over the warmer Gulf Stream water. The summer fluxes are fairly constant and directed downward.

The water of the Labrador Current in the Sable Island region warms substantially from March to September and conversely cools intensely in the period November–January.

A comparison of the energy exchange for a current and for water without motion shows that the surface temperatures would be similar in summer, and the temperature drop would be about equal until November. From that time on, the surface temperature would level off for a water body with no current, but in actual conditions the surface temperature continues to drop to a late winter minimum of about 1°C.

Atmospheric advection of latent heat was calculated by assuming that the daily precipitation was always caused first by condensation of all locally evaporated water with any remainder being supplied by water-vapour advection. The main cause for atmospheric heating in the Sable Island area was found to be condensation of imported water vapour. The region is, in summer, a marked sink for atmospheric heat and water content. For water it remains a sink even in winter. For sensible heat it becomes a source from November to March. The warming of the atmosphere is caused by release of latent heat of advected water vapour in the period February–August. During the months September–January the heat sources are both water-vapour advection and surface turbulent terms.

1 Introduction

The Northern Hemisphere continental east coasts are affected by two opposite ocean currents – warm water from the south and cold from the north.

In the western Atlantic the cold stream includes the Scotian Shelf water at Sable Island. The cold current is still clearly discernible directly off shore past Cape Cod and even to Cape Hatteras.

It is of interest to study the heat exchange between ocean and atmosphere over cold water, to ascertain the amount of heat transfer to the atmosphere under conditions of winter-time cold continental air advection, and to examine the degree of transformation of the cold water by atmospheric (including radiative) processes. While it is known that vertical heat exchange may be very large over warm ocean currents, it is not well known how cold water areas compare. Also, the cold waters off Canada's east coast cause prevalent spring and summer fog and a study of the energy exchange under such conditions would seem to be of interest.

2 Data and method

Since no ship or island station is located in the ideal position for the proposed study, namely in the southern part of the Grand Banks, there was a choice between using observations from Weather Ship B or from Sable Island. Weather Ship B was located at 56°30′N, 51°W until the summer of 1975 and regularly reported sea-surface temperatures. It did not observe precipitation. Sable Island (43°56′N, 60°01′W) has reliable precipitation records but no sea-surface temperatures and, obviously, is not as much a part of the oceanic environment as a ship. However, the island lies in a more desirable position than the ship and is small. Its location is in a region of frequent, strong, cyclonic activity and much cloud. It is therefore argued that no significant diurnal effects or development of land-sea breeze should be expected. Also, the observations of interest are taken at 00 and 12 GMT, i.e., at 20 and 08 LST, away from the time of maximum insolation.

It was therefore decided to use twice-daily synoptic and radiosonde data from Sable Island, although the lack of simultaneous sea-surface temperatures might be troublesome. We are grateful to Dr S.D. Smith of the Bedford Institute of Oceanography who obtained for us 39 water-temperature charts for the period September 1970 to late May 1972. Such sea-surface temperature charts, based on ships' reports are compiled by the METOC Centre, Maritime Command, Halifax. They are probably reliable to about ±3°C.

A smooth curve was drawn through the interpolated values for Sable Island as shown in Fig. 1 which also shows, for comparison, long-term monthly mean temperatures, taken from the Meteorological Office Atlas (1948). Daily surface-temperature values were extracted from Fig. 1 for the period 1970–72. It is not possible to assess the errors caused by this method, but it is an advantage that the sea-surface temperature is rather conservative. The meteorological data were kindly supplied by the Canadian Atmospheric Environment Service. In the following the observations at 12 GMT (08 LST) have been regarded as representative for daytime and those at 00 GMT (20 LST) for the night. Other radiosonde observations are, unfortunately, not available.

The calculations of the various energy-budget terms were performed according to the energy-budget programmes designed and described by Vowinckel

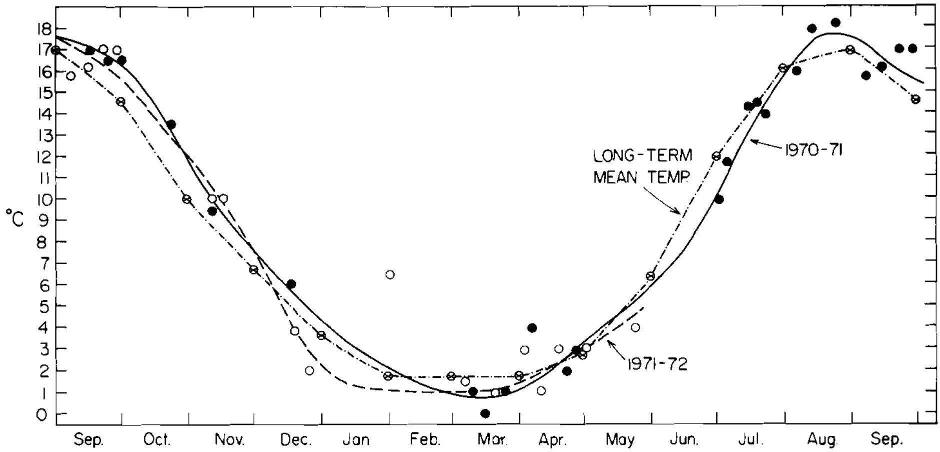

Fig. 1 Sea-surface temperatures, Sable Island.

TABLE 1. Absorption of short-wave radiation in various layers of the sea

Layer in metres from the surface:	0–1	1–3	3–5	5–10	10–20	20–30	30–40	40–50	50–60
Absorption (per cent):	40	33	7	7	7	3	1	1	1

and Orvig (1972, 1973, 1974, 1976). Once a thermal profile in the water is given as an initialization, calculations may proceed with no further sea-surface temperature information as it may be generated by a programme subroutine. This needs a statement of the water depth. Initially, 200 m was assumed since no seasonal temperature variation is observed below this depth. A depth of 100 m was subsequently tested, but only small differences resulted in the calculated flux values so the following discussion is based on a water depth of 200 m.

Values for short-wave radiation absorption in water were adapted from Laevastu (1960) and are listed in Table 1. For the dynamic mixing, infinite fetch was assumed. The sea water density was not corrected for salinity, i.e. the water was regarded as either fresh or isohaline.

The calculations were performed both with observed sea-surface temperatures (from Fig. 1) and without specification of such temperatures. In the latter case a water body of 200 m was prescribed and the programme generated its own surface temperature, as mentioned above, using the same set of meteorological data. This condition implies that no ocean advection is allowed, i.e. it is similar to a lake. There can be no doubt that such conditions would change the weather, both because of different air-mass modification and because of some change in the atmospheric circulation. The authors have a programme available for assessing the change in air-mass modification, but no possibility exists of obtaining numerically the weather changes caused by (probably slight) alterations in circulation. It was, therefore, judged preferable to retain the meteorological observations unchanged. The result will most likely be an overestimate of the fluxes, but the magnitude has not been estimated.

To permit effects of horizontal water motion, the surface temperature was re-set each day to actual values. The results of such "ocean" calculations are compared to the "lake" runs in order to illustrate the influence of ocean advection. In the case of using observed sea-surface temperatures, it was noted above that they are probably not reliable to better than ±3 °C. The calculated fluxes will, obviously, be sensitive to errors in the ocean surface boundary conditions. In the case of long-wave radiation from the surface, which is the largest term in the surface energy balance (Table 2), an error of 1 °C in surface temperature will alter the daily value by 10 cal cm^{-2} (4.8 W m^{-2}). An error of 3 °C will therefore cause an error in calculated long-wave radiation of 4.5% in March and 3.6% in August-September.

The following terms have been calculated:

SGA: Short-wave (solar) radiation absorbed at the surface.
RLD: Long-wave radiation down (atmospheric back radiation).
RLU: Long-wave radiation from the surface (terrestrial radiation).
QS: Sensible heat flux at the surface, positive when directed upward.
QE: Latent heat flux at the surface, sign as for QS.
S300: Incoming solar radiation at 300 mb (the "top" of the troposphere).
SAA: Solar radiation absorbed in the atmosphere.
U150: Upward long-wave radiation flux at 150 mb.

In addition, atmospheric advection of latent and sensible heat was calculated, as described in the later discussion of the atmospheric heat budget.

3 The surface heat budget

a *Radiation*

Fig. 2 shows the monthly means of the various surface energy-budget terms. They are also included in Table 2. The absorbed solar radiation (SGA) displays the normal annual pattern; compared to the atmospheric back radiation (RLD) it is relatively small in all months. This condition is particularly well expressed in the region of Sable Island for a number of reasons. First, since there is frequent overcast and low clouds, the back radiation originates at low elevation, i.e. at relatively high temperature. Even under clear sky conditions the air has high moisture content and RLD originates near the surface. Second, the high moisture content and frequent cloudiness reduces SGA, thereby further enhancing the relative importance of long-wave radiation in the total incoming radiation at the surface.

The seasonal variation in RLD is noticeably less than that of SGA. This is generally observed in middle and high latitudes but is more pronounced at a maritime location such as Sable Island. The long-wave radiation emitted from the surface, RLU, is a function of the surface temperature. Fig. 2 and Table 2 show the long-wave radiation balance (RLD-RLU). As a normal condition at the earth's surface the long-wave balance is negative. In continental climates,

TABLE 2. Monthly mean values of surface energy budget terms (ly day^{-1})

1971	SGA	RLD	RLU	Long-wave balance	QS	QE	Net surface balance
JAN.	132.9	599.5	683.7	−84.2	219.4	169.1	−339.8
FEB.	220.4	594.1	666.9	−72.8	118.9	89.9	−61.2
MAR.	305.2	608.6	660.9	−52.3	4.5	40.1	+208.3
APR.	374.7	646.5	672.2	−25.7	−22.0	4.9	+366.1
MAY	404.1	683.9	696.4	−12.5	−23.0	−9.5	+424.1
JUN.	471.4	703.0	730.9	−27.9	−21.3	−7.4	+472.2
JUL.	401.6	787.4	783.1	+4.3	−21.1	−20.5	+447.5
AUG.	389.8	785.2	825.5	−40.3	−15.5	12.5	+352.5
SEP.	336.2	754.8	825.7	−70.9	11.6	106.8	+146.9
OCT.	233.3	720.6	797.1	−76.5	56.2	136.4	−35.8
NOV.	127.9	684.3	757.9	−73.6	137.9	181.8	−265.4
DEC.	113.6	624.0	708.4	−84.4	181.9	169.1	−321.8

(1 ly day^{-1} = 0.48 W m^{-2}. 1 W m^{-2} = 2.07 ly day^{-1}.)

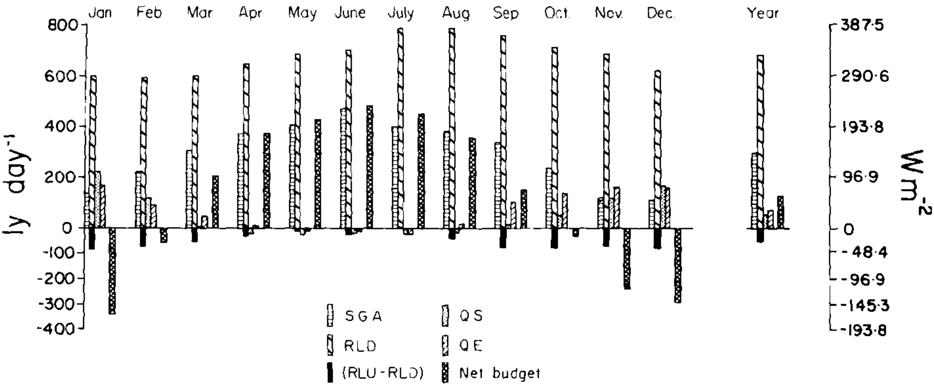

Fig. 2 Surface energy-budget terms. Monthly mean and annual average values for 1971.

without major ground heat storage, the negative long-wave balance has its maximum value in summer, with maximum short-wave radiation income and highest surface temperature. The usual continental small negative balance in winter may on occasion be replaced by a positive long-wave radiation balance, such as occurs with a strong surface inversion, particularly if accompanied by high moisture content and clouds.

Conditions are opposite over water, as its high conductive capacity permits long periods of positive or negative energy budgets before the surface temperature adjusts. Accordingly, the long-wave balance is generally strongly negative in winter and slightly negative in summer. These conditions may be altered in areas near land or oceanic boundaries where advection of different air masses may occur. Fig. 2 shows that at Sable Island the winter maximum negative long-wave balance is clearly developed. Towards summer the deficit becomes less and in July the long-wave balance is actually slightly positive (Table 2). This is only possible with cases of strong advection of warm, moist air with a low ceiling.

The slight positive long-wave balance in mid-summer is remarkable and a unique feature, probably restricted to mid-latitude cold currents where they are sharply delineated from warm water and land masses. One might predict similar conditions over cold water along continental west coasts such as the California and Canary Currents. The air masses originating over the subtropical continents would certainly be sufficiently warm. However, they lack the moisture to form the required low cloud cover. Once the moisture has been acquired, the temperature inversion has been weakened to minimize the effect. The phenomenon seems to be restricted to the east coast cold currents.

Daily values of long-wave radiation balance show sharp short-term variations at all times of the year, indicating that fluctuations in the synoptic conditions (e.g. cloudiness) are similar in all seasons although the temperature level is higher in summer. Generally, the main characteristics of the local energy budget are climatically determined, and synoptic events only cause

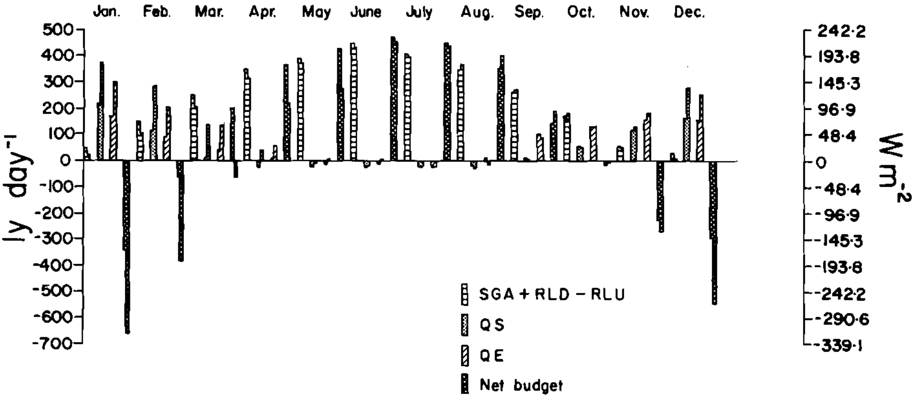

Fig. 3 Surface energy-budget terms for "lake" and "ocean". Monthly mean values for 1971. Right-hand column shows "lake" condition (no current), left-hand column shows "ocean" condition (with advection).

variations on this general theme. Occasionally the synoptic situation is such that very warm, moist air is advected over the cold water in winter, resulting in a positive long-wave radiation balance as high as on individual summer days.

Considering the total radiation balance (including SGA) it is apparent that all monthly means are positive, although there are individual days with negative balance not caused by especially low values of solar radiation but rather by a sharp reduction in RLD, caused by advection of very cold air.

The influence of the cold current is largely eliminated in the calculations if the surface temperature is not forced to the observed values but left to self adjustment. In this case, both RLD and SGA remain unchanged but the terrestrial radiation (RLU) varies and so does, therefore, the radiation balance. The comparison between the two conditions is shown for each month in Fig. 3. Apart from minor exceptions, in late summer and autumn, the radiation budget is always less positive for the non-advective "lake" condition. This indicates that, especially in winter, the "lake" temperatures would be higher. Thus is seen the first indication of the substantial withdrawal of energy from the area via ocean export. This will be discussed in more detail below.

From the discussion so far it is concluded that there is energy available at the surface, for use other than radiation, in every month of the year. The calculations reported here indicated that the average day would have 243 cal cm^{-2} (118 W m^{-2}) to be used in turbulent fluxes or ocean current export.

b *Turbulent Fluxes*

The monthly mean values of latent heat flux (QE) and sensible heat flux (QS) are also shown in Fig. 2 and Table 2. These turbulent terms are shown positive when directed upward, i.e. when they are heat loss terms for the surface. The winter months are characterized by strong upward fluxes. From December to February the sensible heat flux is greater than the latent heat flux because the air temperature drops faster than the water temperature. This has the conse-

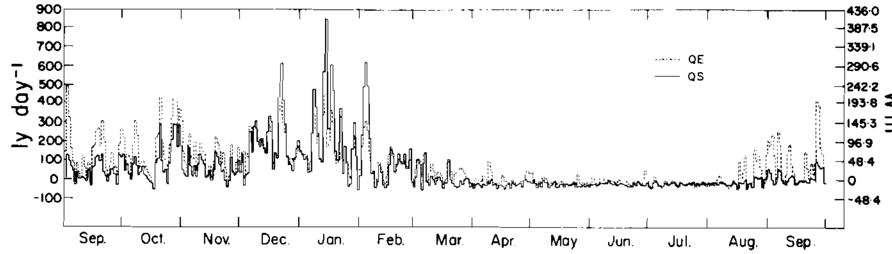

Fig. 4 Daily values of turbulent terms, September 1970-September 1971.

quence that the increase in the temperature gradient, which governs QS, is greater than the change in the moisture gradient caused by reduced saturation mixing ratio at low air temperatures. The changeover from downward to upward fluxes takes place in autumn in individual daily values (Fig. 4). Upward fluxes begin to appear after the middle of August when cool air advection is mainly shown by QE which reacts sharply to the decreased air moisture content. As the season progresses, both air and water temperatures drop but the air-sea temperature difference increases, and QS becomes greater than QE from mid-December. By mid-March transition to summer conditions is apparent (Fig. 4).

The maximum QE values reach 400 to 500 ly day^{-1} (194 to 242 W m^{-2}), which are much less than the highest values found over the warm water of the Gulf Stream. There, average daily values in January are 600–750 cal cm^{-2} (290–363 W m^{-2}) (Budyko, 1963) and maximum daily values over 1500 cal cm^{-2} (725 W m^{-2}) may occur. The peak daily values of QS at Sable Island are of similar magnitude to QE, reaching 600 to 900 cal cm^{-2} (290 to 436 W m^{-2}).

The monthly mean values of the turbulent terms (Fig. 2) show a remarkable summer reversal, as QS is directed towards the surface from April to August and QE is downward from May to July. These summer fluxes are also particularly stable (Fig. 4). It is only rarely that this area experiences a summer air mass which is colder than the water. With such stable conditions in the near-surface air layer, saturation will be restricted to a shallow thickness and frequent summer fog occurs. The mean frequency of fog at Sable Island is 15% in July, i.e. visibility is less than 1 km. The fog season is at its height in that month, and Osborne (1976) reported that Sable Island had conditions below Visual Flight Rules limits 40% of the time in July (ceiling 1,000 feet, visibility 3 miles).

c *Heating and Cooling of the Water*

If the monthly mean values of the surface energy budget are considered (Fig. 2 and Table 2), it is found that the net annual balance is +117 ly day^{-1} (56 W m^{-2}) in spite of the fact that the average annual combined heat loss by the turbulent terms is 124 ly day^{-1} (60 W m^{-2}). The cause is the radiative surplus. This large amount of energy goes into storage, i.e. it is used to increase the water temperature and is carried away with the ocean current. Such net annual values, when quoted, give a distorted picture for an area where the water is moving, because no balancing of positive and negative periods is possible—cooling of the water at one time cannot be compensated by warming at another. The net drift velocity over the Scotian Shelf is about 10 km day^{-1} to the southwest, and a parcel of water will reside on the shelf for 2 to 3 months at most.

If the heating during May, June and July is considered, the average input is 447 ly day^{-1} (Table 2). Confining this heating to the upper 20 m of ocean, the average temperature increase for an advecting water column would be 0.2°C day^{-1}. Combining this figure with the drift speed, a net horizontal temperature gradient of 2°C is obtained over 100 km. This gradient is a little high but not totally inconsistent with observed surface gradients in summer.

The warming of the upper 20 m of ocean water, being 0.2°C day^{-1}, would amount to about 12°C in two months. This is not inconsistent with the observation of Hachey (1938).

The appropriate time interval for heat-budget calculations over the ocean will depend on the current speed and the areal extent for which the observed parameters can be regarded as representative. Monthly means are used in the following discussion. If the monthly values are considered individually (Fig. 2) it is seen that substantial heating, essentially by short-wave radiation, takes place in summer from March to September, with the maximum at the time of highest radiation in June-July. The winter months November-January are characterized by intense cooling caused by strong upward turbulent fluxes from the surface.

The summer heating is relatively independent of the sea-surface temperature, except for its influence on the long-wave balance. The rate of heating of a cold ocean current is independent of the degree of "coldness" with respect to surrounding water masses. The determining factor is the cloud cover which regulates the radiation budget.

The heat gained by radiation at the surface is spread vertically downward into the water by wind-created turbulence. A comparison between "ocean" conditions (with current advection) and "lake" conditions (no horizontal motion) is shown in Fig. 5, which gives observed sea-surface temperatures ("ocean") and calculated surface temperatures ("lake") for the same days. It is apparent that there is little difference in summer. This may indicate that the surface water of the current in the Sable Island area then is in thermal equilibrium, in the direction of the current, and that one should perhaps not speak of a "cold current." If it is assumed that the current speed is constant, this means that the surface heat balance must be uniform over a fairly extended area of the Labrador Current. A map of sea-surface isotherms will, of course, show the current because of the extreme warmth of the Gulf Stream, which is maintained throughout the year. The cold current must still be present at depth as is apparent from the subsequent development in autumn, which is discussed in the following.

The summer period is, as seen, characterized by substantial radiational heating and southward export of the heated water, but during winter the

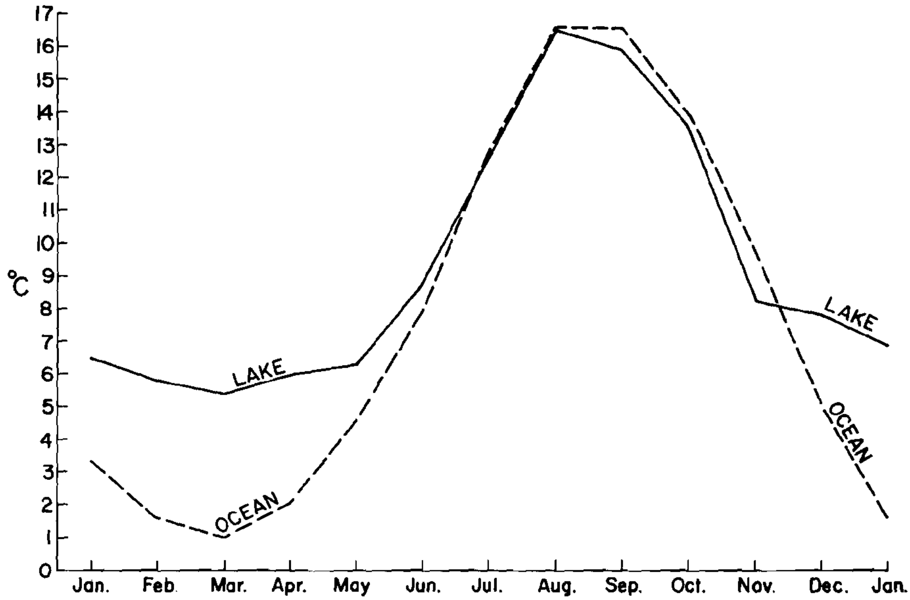

Fig. 5 Water surface temperatures for "lake" and "ocean" conditions. Values for the 15th day of each month, January 1971-January 1972.

surface heat budget is determined by the turbulent terms which effect very large extraction of energy (Fig. 2 and Table 2). In comparison with the radiative fluxes, these terms are influenced much more strongly by synoptic events, and the highest daily heat loss values in winter occur in synoptically determined short intervals, with the main bulk of the cooling taking place in a few brief events. Daily surface heat balance values of -800 to -1200 cal cm $^{-2}$ (387–580 W m^{-2}) may occur.

Comparing the calculated turbulent terms under real ("ocean") conditions and without ocean advection ("lake") as shown in Fig. 3, it is apparent that winter-time turbulence would be even more intense under "lake" conditions. These terms are determined by wind speed, air temperature and air moisture content, all kept unchanged in the comparison, and by sea-surface temperature which is different and therefore the cause of the calculated dissimilarity. In early autumn the drop in surface temperature is about equal under the two conditions (Fig. 5), but the "lake" temperature then tends to level off from November onwards, due to thermal mixing involving progressively deeper water masses, and the calculated winter minimum is just above $+5\,°C$. Under actual conditions the surface temperature continues its steep drop to a late winter value close to $+1\,°C$.

It can be concluded that, in reality, the heat content of the water transformed in summer becomes exhausted in November, and the continuing decrease in sea-surface temperature is caused by arrival of progressively colder water from the northeast, in the Labrador Current.

The strongly negative energy balance at the sea surface in winter indicates that the southward flowing water masses are then still cooling in the region of Sable Island. The actual heat loss would, of course, be even greater if the current were not already so cold. This can be seen by comparing the current's net surface energy balance with that of the warmer "lake" for November-February (Fig. 3). In March the "lake" would still be cooling, while the real current shows a positive net heat balance, i.e. the beginning of the summer heating cycle.

4 The atmospheric heat budget

Additional terms appear in the atmospheric heat and water budgets: short-wave radiation absorbed (SAA); long-wave radiation to space at the top of the atmosphere—here taken to be at 150 mb (U150); a term containing the release of latent heat of condensation (measured by the amount of precipitation); terms measuring the advection of sensible heat and of water vapour as well as the change of atmospheric storage of sensible heat and water vapour. In writing heat and water balance equations for the atmosphere, all of these terms must be included and, in addition, certain of the terms discussed under the surface heat budget: the long-wave fluxes (RLD and RLU) and the sensible and latent heat fluxes (QS and QE).

It is apparent that SGA (solar radiation absorbed at the surface) does not enter into the atmospheric balance, and QE disappears from the sensible heat budget while the amount of precipitation appears in this budget for the atmosphere. The precipitation falling on the surface does not involve energy transformation there and appears only in the surface water budget.

Table 3 gives the monthly mean values of the atmospheric energy-budget terms. These indicate (as do the calculated daily values, which are not shown) that two clearly different seasons can be distinguished during the year. Most of the twelve months, including the summer, are characterized by atmospheric energy loss. The actual daily heat loss varies with the weather pattern in a complicated manner.

The months November to January experience frequent cases of positive atmospheric balance and, during the peak cyclonic periods, the daily values may vary greatly. The reason for the high positive mid-winter values lies in the magnitude of the sensible heat flux term (QS). Table 3 includes the monthly means of the atmospheric radiative budget terms and it is apparent that the radiation balance remains negative throughout the year and that it is fairly constant, in spite of much less available short-wave radiation (S300) in winter. This is caused by the relatively high winter surface temperature and, accordingly, rather large RLU values and thus smaller negative atmospheric long-wave balance at that time. This compensates for the decrease in SAA. The atmospheric radiation budget is dominated by the long-wave terms, while the surface radiation is dominated by the short-wave term rather than by the net long-wave flux.

The change to the winter condition of positive atmospheric heat budget is

TABLE 3. Monthly mean values of atmospheric energy-budget terms (ly day^{-1})

1971	S300	SAA	Surface long-wave balance	U150	Atmospheric long-wave balance	Atmospheric radiation balance	QS	QE	Net atmospheric balance
JAN.	305.0	48.7	84.2	399.5	−315.3	−266.6	219.4	169.1	+121.9
FEB.	431.5	59.8	72.8	353.5	−280.7	−220.9	118.9	89.9	−12.1
MAR.	616.0	84.6	52.3	369.2	−316.9	−232.3	4.5	40.1	−187.7
APR.	806.3	114.8	25.7	373.8	−348.1	−233.3	−22.0	4.9	−250.4
MAY	942.5	147.2	12.5	403.0	−390.5	−243.3	−23.0	−9.5	−275.8
JUN.	998.6	156.2	27.9	413.8	−385.9	−229.7	−21.3	−7.4	−258.4
JUL.	967.7	171.3	−4.3	422.0	−426.3	−255.0	−21.1	−20.5	−296.6
AUG.	853.6	153.4	40.3	434.3	−394.0	−240.6	−15.5	12.5	−243.6
SEP.	679.7	114.7	70.9	436.6	−365.7	−251.0	11.6	106.8	−132.6
OCT.	498.8	83.9	76.5	429.9	−353.4	−269.5	56.2	136.4	−76.9
NOV.	350.5	67.1	73.6	353.5	−279.9	−212.8	137.9	181.8	+106.9
DEC.	271.2	47.7	84.4	340.1	−255.7	−208.0	181.9	169.1	+143.0

(1 ly day^{-1} = 0.48 W m^{-2}. 1 W m^{-2} = 2.07 ly day^{-1}.)

thus the result of the large surface sensible heat flux at that time. All areas experiencing the flow of continental arctic air over open water will have high QS values. It is not necessary that the water should be particularly warm, although it would increase the flux somewhat. The main requirement is very cold air.

There is one other, quite different, geographical area with high QS values—namely the subtropical deserts in summer. Most of the available absorbed solar radiation at the surface is then transferred directly to the atmosphere via sensible heat flux. Atmospheric heat gain areas are therefore found off the east coasts of cold continents in winter, and over subtropical deserts in summer.

Table 3 shows the atmospheric balance, including the latent heat exchange with the surface. In order to assess the actual, sensible, heat gain or loss by the atmosphere, it is necessary to include the amount of heat gained by condensation, measured by the amount of precipitation. There are two sources of water vapour: local evaporation, and advection (import from outside the area). In the following it is assumed that all evaporated water vapour is first condensed and the remainder of the precipitation is supplied by imported moisture.

The heat and moisture contents of the atmosphere were ascertained by two radiosonde ascents each day throughout the year 1971, and the daily precipitation was used to calculate the 24-h latent heat release. As indicated above, it was thus possible to calculate daily values of the various terms in the atmospheric sensible heat budget. Table 4 shows monthly mean values of evaporation (QE), advected latent heat, advected sensible heat, total gain from condensation and resulting atmospheric sensible heat gain. The relationships between values in Tables 3 and 4 are explained in the Appendix.

From Table 4 it appears that, of the terms shown there, the main cause for atmospheric heating is the condensation of imported water vapour. It is the major contributor from February to June, and the largest factor on an average for the year.

Considering the monthly mean values of advected sensible heat and advected latent heat (Table 4) it is seen that October to March had opposite signs for the two terms while April to September had import of both forms of heat. The Sable Island region in summer is thus a marked sink for the atmosphere for both heat and water content. For water it remains a sink even in winter. For sensible heat it becomes a source from November to March.

It is interesting to attempt to relate these findings to the usual intensification of cyclones along the east coast and to consider whether this is restricted solely to the Gulf Stream area, or whether the causes are purely dynamical and that the surface influences are unimportant. From the last four columns in Table 3 and from Table 4 it seems that the radiative and turbulent terms are insufficient, at least from February to August, to produce substantial changes in the atmospheric energy content over the cold water region investigated. Striking increases in energy input can then be experienced only through the release of the latent heat of the advected water vapour. During the months September-January the heat sources are both advected water vapour and surface turbulent terms.

5 Summary and conclusions

Twice-daily synoptic and radiosonde data from Sable Island for the year 1971 indicate that during that year the solar radiation was relatively small compared to atmospheric back radiation. This is generally the case in the region, due to frequent overcast and low clouds. The surface long-wave radiation balance was strongly negative in winter and for a brief period, in July, it was slightly positive because of strong advection of warm, moist air with low ceiling.

All months showed a positive mean total radiation balance, which was used in turbulent fluxes or ocean current export.

The winter months were characterized by strong upward turbulent fluxes. By mid-March transition to summer conditions was apparent, and the downward fluxes lasted until after the middle of August when cool air advection began to appear. The maximum fluxes of latent heat reached values of 500 ly day^{-1} (242 W m^{-2}) which is about one-third of the highest values found over the warm Gulf Stream.

The turbulent fluxes indicate that it is only rarely that the area experiences a summer air mass which is colder than the water. Such stable conditions cause frequent summer fog.

The calculated available heat during the summer months would, if it were confined to the upper 20 m of ocean, increase the temperature of the column by 0.2 °C day^{-1}. This would give a warming of about 12 °C in two months and, if the water is advected at 10 km day^{-1}, create a horizontal temperature gradient of 2 °C in 100 km.

An assessment of the effect of horizontal water motion on the sea-surface temperature indicated that this was negligible in summer and that the surface heat balance may be uniform over a fairly extended area. The heat content of the water transformed in summer became exhausted in November, and the continuing decrease in sea-surface temperature is caused by arrival of colder water from the northeast.

The atmospheric heat budget was mostly characterized by energy loss, even in the summer. During the peak cyclonic periods the daily values varied greatly. In winter there were frequent cases of positive atmospheric heat balance, caused by large surface sensible heat flux, although it appears that the main cause for atmospheric heating is condensation, and imported water vapour is more important than local evaporation. Of the various terms, water-vapour advection is the major contributor from February to June, and the largest factor on the average for the year. The Sable Island region seems to be a sink for advected water vapour all throughout the year. Striking heating of the atmosphere can only be accomplished, from February to August, by the release of latent heat of advected water vapour. During the period September-January, the heat sources are both advected water vapour and surface turbulent terms.

Appendix

In Table 4 will be found QE values from Table 3. The last column in Table 3 gives the energy balance without consideration of atmospheric advection in or

TABLE 4. Monthly mean values of atmospheric sensible heat budget terms (ly day^{-1})

1971	QE	Advected latent heat	Total gain from condensation	Advected sensible heat	Atmospheric heat gain
JAN.	+169	+126	295	−248	+47
FEB.	+90	+192	282	−180	+102
MAR.	+40	+215	255	−27	+228
APR.	+5	+218	223	+32	+255
MAY	−10	+173	163	+103	+266
JUN.	−7	+177	170	+81	+251
JUL.	−21	+137	116	+160	+276
AUG.	+13	+62	75	+182	+257
SEP.	+107	+74	181	+59	+240
OCT.	+136	−76	60	+153	+213
NOV.	+182	+159	341	−266	+75
DEC.	+169	+130	299	−273	+26
Average for the YEAR:	+73	+131	204	−17	+187

(1 ly day^{-1} = 0.48 W m^{-2}. 1 W m^{-2} = 2.07 ly day^{-1}.)

out of the area. Therefore, if the monthly net value for February (−12), for example, is added to the (positive) advected latent heat from Table 4 (+192), the result should be the sensible heat available for export (−180). The atmospheric latent heat advection is based on 24-h precipitation amounts, less QE.

In Table 4 total gain from condensation (February: 282) less sensible heat export (−180) gives +102 as the atmospheric heat gain by these processes. This heat is used together with gain by QS (119—see February in Table 3), for a total of 221 ly day^{-1}, to account for the February atmospheric radiation balance (−221, see Table 3).

References

BUDYKO, M.I. (ed.) 1963. Atlas of the heat balance of the earth. Mezhduvedom. Geofiz. Kom. pri Prizid. Akad. Nauk SSSR. Moscow, 69 pp.

HACHEY, H.B. 1938. The origin of the cold water layer of the Scotian Shelf. *Proc. and Trans. Roy. Soc. Can.* Sect. III, **32**: 29–42.

LAEVASTU, T. 1960. Factors affecting the temperature of the surface layer of the sea. *Soc. Sci. Fennica*, **XXV**, Helsinki, 136 pp.

METEOROLOGICAL OFFICE, AIR MINISTRY. 1948. Monthly Meteorological Charts of the Atlantic Ocean. London. His Majesty's Stationery Office, M.O. 483, 122 pp.

OSBORNE, A.H. 1976. Techniques for the prediction of sea fog over the Scotian Shelf. Env. Canada, Tech. Memo. 825, 18 pp.

VOWINCKEL, E. and S. ORVIG. 1972. EBBA—An energy budget programme. McGill Univ., Arctic Meteorology Research Group, Sci. Rep. No. 105, 50 pp. plus Appendix. (Available from authors on request.)

——— and ———. 1973. The heat and water budgets of a beaver pond. *Atmosphere*, **11**: 166–178.

——— and ———. 1974. Répercussions météorologiques découlant de modifications naturelles ou délibérées de la surface—principes généraux et prospectives, région de la Baie de James. *Cahiers de Géographie de Québec*, **18**: 445–461.

——— and ———. 1976. Experiments with the surface heat and water budgets of the Canadian Middle North. *Tellus*, **28**: 442–450.

A Cloud Layer–Sunshine Model for Estimating Direct, Diffuse and Total Solar Radiation

P.W. Suckling* and J.E. Hay
Department of Geography,
University of British Columbia, Vancouver
[Manuscript received 12 May 1977; in revised form 26 September 1977]

ABSTRACT

A previous study (Suckling and Hay, 1976a) described a method for calculating hourly values of the direct and diffuse solar radiation for cloudless sky conditions. This paper presents an extension which incorporates the effects of clouds through the use of hourly values of cloud amount and type for up to four layers and hourly bright sunshine totals. The latter data provide a more accurate measure of the length of time the direct radiation of the sun is not attenuated by cloud. On an average, the cloud layer–sunshine (CLS) model estimated daily total solar radiation at five Canadian locations to within ±15 per cent of the measured values. This was an improvement over an earlier model (Davies et al., 1975) based on cloud data alone, but the relative advantage, as well as the overall errors themselves, were diminished as the averaging period was increased to five and ten days. The CLS model has the additional advantage of calculating the separate direct and diffuse components of the total solar radiation.

1 Introduction

Suckling and Hay (1976a), presented a model which calculated hourly and daily totals of solar radiation, including its direct and diffuse components, for cloudless days. This paper describes how the effects of cloud have been incorporated into the model.

Clouds have the most variable and often the greatest attenuating effect on the transmission of solar radiation through the atmosphere. In order to calculate the total solar radiation at the surface ($K\downarrow$) with cloud present, some researchers have derived statistical relationships between the ratio of actual $K\downarrow$ to cloudless sky solar radiation ($K\downarrow_0$) and the ratio of measured to total possible number of hours of bright sunshine (Ångström, 1924; Kimball, 1927; Bennett, 1965; Exell, 1976) or the total amount of cloud (Kimball, 1928; Neumann, 1954; Mateer, 1963; Kimura and Stephenson, 1969; Thompson, 1976). Their formulae do not consider either the variations in the intensity of solar radiation which can occur irrespective of the value obtained from a sunshine recorder or the effects of different cloud types and cloud layers. These

*Present address: Department of Geography, Brandon University, Brandon, Man.

models do perform reasonably well for monthly mean data but the empirical coefficients in the relationships vary spatially and temporally (Bennett, 1965; Kimura and Stephenson, 1969; Thompson, 1976).

The effects of different cloud types have been included in the solar radiation studies of London (1957), Monteith (1962), Lumb (1964), Lettau and Lettau (1969), Hay (1970a), Atwater and Brown (1974), Manabe and Strickler (1964) and Davies et al. (1975). The last three studies also showed the need to consider the effects of cloud layers. These models generally succeed in estimating $K\downarrow$ with greater accuracy than the more empirical relationships. However, none considers the direct ($S\downarrow$) and diffuse ($D\downarrow$) components of solar radiation separately. Some applications of solar radiation information, such as those in the area of solar energy, require knowledge of $S\downarrow$ and $D\downarrow$ (Duffie and Beckman, 1974; Garnier, 1975; Hay, 1976).

In order to accomplish this task of calculating the direct and diffuse components and hence the total solar radiation under cloudy conditions, a more refined model will be presented. In addition to the cloud-layer data used in the model proposed by Davies et al. (1975), the present approach also requires a knowledge of the reported hourly bright sunshine amounts to overcome inadequacies in the way cloud data alone can be used to specify the portion of time cloud obscures the sun. Any consequent improvement in the overall performance of the model will, however, be achieved at the expense of a predictive capability. This may be an unwarranted restriction for some workers (e.g. general circulation modellers).

2 Description of the cloud layer–sunshine (CLS) model

The initial inputs into the model are the cloudless sky values of the total ($K\downarrow_0$), direct ($S\downarrow_0$) and diffuse ($D\downarrow_0$) solar radiation calculated by the method of Suckling and Hay (1976a). When computing the attenuation by cloud, hourly values of cloud amount and cloud type by layer and bright sunshine amounts are required. Such data are available from many of the stations operated by the Canadian Atmospheric Environment Service (see Section 4).

a *Direct Solar Radiation Component*

In order to calculate the portion of the hour during which direct solar radiation is received, an effective cloud amount (n_e) is calculated from:

$$n_e = \frac{An_s + B(1 - s)}{A + B} \tag{1}$$

where n_s is the sum of the cloud amounts as seen from the surface but excluding cirriform, s is the total bright sunshine amount expressed as a fraction of an hour and A and B are weighting factors for the cloud and sunshine measurements. Both cloud amount and sunshine values are used to overcome inadequacies of utilizing only one of these parameters. If one uses cloud only, then a random distribution of the cloud must be assumed. The sole use of sunshine amounts is also inadequate, since sunshine records suffer from "over-burn" and the intensity required to "burn" the card in the record is not sufficient until the sun is 5° above the horizon (Brooks and Brooks, 1947). In equation (1) it is assumed that the use of both variables (weighted according to the empirical coefficients A and B) will lead to an overall reduced error in the calculated transmissivity for direct solar radiation. This procedure also assumes that sufficient direct solar radiation passes through cirriform clouds to register on the sunshine records. The direct component of solar radiation uncorrected for cirriform attenuation ($S{\downarrow}_x$) is then calculated as:

$$S{\downarrow}_x = (1 - n_e)S{\downarrow}_0. \tag{2}$$

With the inclusion of attenuation by cirrus the direct solar radiation becomes:

$$S{\downarrow} = [(1 - n_e) - n_{Ci}(1 - t_{Ci})]S{\downarrow}_0 \tag{3}$$

where n_{Ci} is the reported amount of cirriform cloud (i.e. $n_{Ci} = n\text{-}n_s$, where n is the total cloud amount as reported by a ground-based observer) and t_{Ci} is the transmissivity of the cirriform cloud. In the case of the direct component, it is appropriate to use the amount of cirriform visible to a ground-based observer.

b *Diffuse Solar Radiation Component*

The diffuse component of solar radiation is initially calculated in three parts for the cloudless, the cirriform and the remaining portions of the sky. For the cloudless portion of the sky, the diffuse solar radiation ($D{\downarrow}_{cs}$) is given by:

$$D{\downarrow}_{cs} = (1 - n)D{\downarrow}_0. \tag{4}$$

For the portion of the sky covered solely by cirriform cloud, the diffuse solar radiation ($D{\downarrow}_{Ci}$) is given by:

$$D{\downarrow}_{Ci} = n_{Ci}t_{Ci}D{\downarrow}_0 \tag{5}$$

where t_{Ci} is again the cirriform transmissivity. In equations (4) and (5), only $D{\downarrow}_0$ is used, since the incident direct solar radiation ($S{\downarrow}_0$) for these portions of the sky was considered in the calculation of $S{\downarrow}$.

For the remaining cloudy portions of the sky (i.e. excluding portions with only cirriform cloud), the diffuse solar radiation ($D{\downarrow}_{cy}$) is calculated by a layered procedure similar to that in the model of Davies et al. (1975). Since cloud layer amounts are reported from surface-based observations, a correction must be applied to upper layer amounts to negate the effect of obscuring by the layers below. The assumption has been made that the fraction of the unobstructed sky covered by an upper cloud is representative of the total proportion of the sky covered by that cloud layer (Davies et al., 1975). Thus, the corrected amount of cloud in a layer becomes:

$$n_i = n_i'/(1 - n_r), \tag{6}$$

where n_i' is the reported cloud layer amount and n_r is the simple sum of reported cloud amount for the layers beneath. In the example shown in Fig. 1 where $n_1' = 0.5$, $n_2' = 0.3$ and $n_3' = 0.1$, $n_1 = n_1' = 0.5$ while the corrected amount

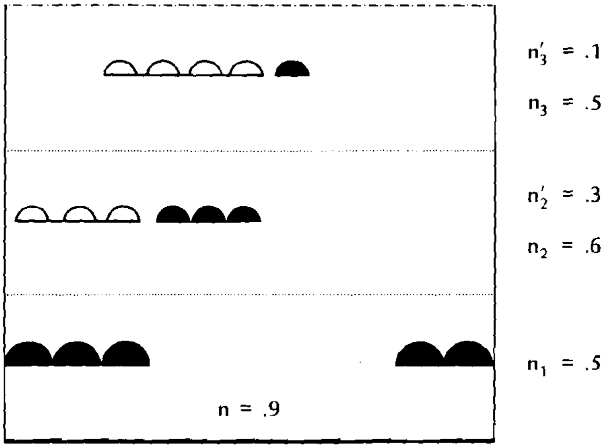

Fig. 1 Example of the correction for cloud layer amounts for layers obstructed by lower cloud.

TABLE 1. Cloud transmissivities (%) for solar radiation from various sources

	Cloud Type									
Source	Fog	Ns	St	Sc	Cu	Cb	As	Ac	Cs	Ci
Haurwitz (from List, 1966)	17–19	15–25	24–25	29–35	—	—	41	45–52	65–84	80–85
Houghton (1954)	—	—	25	—	20	—	48	48	—	78
London (1957)	—	17–25	27–41	—	23	20	44–50	—	—	75–84
Vowinckel and Orvig (1962)										
Polar Ocean:	—	—	30–70	40–68	—	—	50–72	45–88	—	65–91
Arctic Coast:	—	—	31–60	38–67	—	—	50–78	57–84	—	77–100
Edmonton:	—	—	28–50	31–60	—	31–47	25–60	33–78	—	79–112
Dartmouth:	—	—	20–29	38–44	—	—	28–38	38–44	—	20–91
Makarevsky (Kondratyev, 1969)	—	—	—	—	—	—	—	10–35	46–73	62–84
Drummond and Hickey (1971)	—	—	—	—	—	—	45	45	40–80	40–80
Vonder Haar and Cox (1972)	—	—	35–50	35–50	15–25	—	25–55	25–55	—	—
Reynolds et al. (1975)	—	—	—	27–46	10–14	3	—	—	27–40	27–40
Liou (1976)	—	3	25–49	—	10–23	3	14–28	—	—	—

for the second layer becomes $n_2 = 0.3/(1–0.5) = 0.6$ and for the third layer, $n_3 = 0.1/(1–0.8) = 0.5$. If n_r reaches 1.0, higher layers are completely obscured and must be neglected. Additional assumptions could be used to infer the character of this obscured cloud but they do not appear justified at this time. Few measurements or estimates of the cloud-type transmissivities (t_i) are available while those reported vary widely, as shown in Table 1. Variations are due to the angle of incidence of the solar beam, (usually expressed in terms of the optical air mass, m), the optical depth of the cloud, cloud height and cloud composition. Many of these variables are not routinely measured. The work of Haurwitz (1948) allows for a variable transmissivity dependent upon optical air mass:

$$t_i = \left(\frac{1}{K{\downarrow}_0}\right)\left(\frac{a}{m}\right) e^{-bm} \tag{7}$$

TABLE 2. Values of the coefficients used in the Haurwitz (1948) cloud-transmission functions (a has units: $\mathrm{kJ\,m^{-2}\,h^{-1}}$)

	Cloud Type							
	Fog	Ns	St	Sc	As	Ac	Cs	Ci
a	645.3	469.3	997.2	1453.9	1634.1	2199.8	3649.5	3444.2
b	0.028	−0.167	0.159	0.104	0.063	0.112	0.148	0.079

TABLE 3. Cloud-type assignments

Haurwitz Cloud Type	Additional Types Assigned to Same Category
Fog	Obstruction
Ns	Cb
St	Fs
Sc	Cu, Fc, Cu++
As	—
Ac	Acc
Cs	Cc
Ci	—

where a and b are coefficients obtained by Haurwitz by least-squares procedures from his observed data. Haurwitz's values of a and b for different cloud types are given in Table 2. Following the works of Nunez et al. (1971), Atwater and Brown (1974) and Davies et al. (1975), the Haurwitz transmission functions will be used in this study. Values for t_{Ci} in equations (3) and (5) were also calculated from equation (7). The Atmospheric Environment Service of Canada reports 16 different cloud-type categories, which must therefore be reduced to the 8 different categories used by Haurwitz. Following the descriptions of cloud types given in the *Manual of Standard Procedures for Surface Weather Observing and Reporting* (Department of Transport, 1970) and the work of London (1957) and Hay (1970a), the cloud types were assigned as shown in Table 3.

If cirriform clouds are present, then the amount for such layers in the "cloudy portion" of the sky is calculated as:

$$n_{i(Ci)} = n_i - n_{Ci} \tag{8}$$

where $n_{i(Ci)}$ is the "cloudy portion" layer amount for a cirriform layer and n_i is the total amount in the cirriform layer calculated from equation (6). The Haurwitz transmissivities (t_i) from equation (7) are used for the various cloud types to calculate individual layer transmissivities (ψ_i) for the "cloudy portion" of the sky where:

$$\psi_i = 1 - (1 - t_i)\frac{n_i}{n_s}. \tag{9}$$

The diffuse solar radiation through the "cloudy portion" is then given by:

$$D\downarrow_{cy} = n_s K\downarrow_0 \prod_{i=1}^{j} \psi_i \tag{10}$$

where j is the number of layers. Whereas the calculations of $D\downarrow_{cs}$ and $D\downarrow_{Ci}$ used $D\downarrow_0$, those for $D\downarrow_{cy}$ utilize $K\downarrow_0$ since both direct and diffuse solar radiation are incident upon the top of these clouds and transmitted only as diffuse solar radiation.

Multiple reflection is then added as a fourth part of the diffuse solar radiation ($D\downarrow_{mr}$). Following Hay (1970a) only the first-order component is used. Therefore,

$$D\downarrow_{mr} = (S\downarrow + D\downarrow_{cs} + D\downarrow_{Ci} + D\downarrow_{cy})\alpha(\alpha_c n_s + \alpha_{Ci} n_{Ci}) \tag{11}$$

where α is the surface albedo and α_c and α_{Ci} are the non-cirriform and cirriform cloud-base albedoes, respectively. The cloud-base albedo is split into two parts since the cirriform albedo differs considerably from that of other clouds (London, 1957; Canover, 1965; Liou, 1976). Following Davies et al. (1975) and Hay (1976), α_c was assigned the constant value of 0.6 since the albedoes of non-cirriform clouds are all moderately high. London (1957) had non-cirriform cloud albedoes ranging from 0.5 to 0.7 although other studies have shown an even greater variation (Canover, 1965; Drummond and Hickey, 1971; Reynolds et al., 1975; Liou, 1976). A value of 0.2 was assigned for α_{Ci} (London, 1957; Drummond and Hickey, 1971) which is considerably lower than that for other clouds. In this study α was assumed to be 0.2 for no snow cover and 0.8 when snow was present (Sellers, 1965).

The diffuse component of solar radiation is therefore given by:

$$D\downarrow = D\downarrow_{cs} + D\downarrow_{Ci} + D\downarrow_{cy} + D\downarrow_{mr}. \tag{12}$$

Finally, the total solar radiation is given by the sum of the direct and diffuse components. Thus,

$$K\downarrow = \Big[[(1 - n_e) - n_{Ci}(1 - t_{Ci})]S\downarrow_0 + [(1 - n) + n_{Ci}t_{Ci}]D\downarrow_0 + n_s K\downarrow_0 \prod_{i=1}^{j} \psi_i \Big][1 + \alpha(\alpha_c n_s + \alpha_{Ci} n_{Ci})]. \tag{13}$$

3 Performance of the model

For the three-year period 1968–70, solar radiation and hourly meteorological data, including cloud amount and types in up to four layers and bright sunshine amounts, were obtained from the Atmospheric Environment Service of Canada for the following five locations: Goose, NFLD, Edmonton, ALTA, Summerland, B.C., Vancouver, B.C. and Sandspit, B.C. Values for the extra-terrestrial radiation were calculated using the solar climate calculator routine of Furnival et al. (1970) and the cloudless sky solar radiation values were obtained by the method of Suckling and Hay (1976a). For Summerland, the hourly meteorological data from Penticton were used.

TABLE 4. Performance of the CLS model for estimation of $K\downarrow$ on a daily basis for 1968–70. (Unit for mean $K\downarrow$ and RMSE: MJ m^{-2} day^{-1})

Location	A	B	Number of days	Mean $K\downarrow$ Measured	Mean $K\downarrow$ Modelled	RMSE	RMSE (%)	Measured/ modelled
Goose	1	1	1065	10.31	10.63	1.78	17.2	0.970
	2	1	1065	10.31	10.40	1.64	15.9	0.992
	1	2	1065	10.31	10.88	2.06	20.0	0.948
Edmonton	1	1	1035	12.11	13.23	2.32	19.2	0.915
	2	1	1035	12.11	12.93	2.02	16.7	0.937
	1	2	1035	12.11	13.54	2.74	22.7	0.895
Summerland	1	1	885	12.06	12.38	1.38	11.5	0.975
	2	1	885	12.06	12.22	1.25	10.4	0.987
	1	2	885	12.06	12.54	1.63	13.5	0.962
Vancouver	1	1	1070	11.91	11.87	1.27	10.6	1.003
	2	1	1070	11.91	11.69	1.36	11.4	1.019
	1	2	1070	11.91	12.05	1.33	11.2	0.988
Sandspit	1	1	1089	9.83	9.25	1.45	14.7	1.062
	2	1	1089	9.83	9.08	1.64	16.6	1.082
	1	2	1089	9.83	9.43	1.40	14.2	1.043

TABLE 5. Comparison of the performances of the Layer and CLS models for 1968–70 – Root Mean Square Errors (%)

Location	Model: Layer	CLS 1,1	CLS 2,1	CLS 1,2
(a) Daily Basis				
Goose	19.8	17.2	15.9	20.0
Edmonton	18.5	19.2	16.7	22.7
Summerland	12.2	11.5	10.4	13.5
Vancouver	16.0	10.6	11.4	11.2
Sandspit	22.9	14.7	16.6	14.2
Average	17.9	14.6	14.2	16.3
(b) Five-Day Daily Means				
Goose	10.3	10.3	8.4	
Edmonton	10.5	14.7	11.6	
Summerland	6.7	7.7	6.3	
Vancouver	9.4	5.7	6.3	
Sandspit	15.6	9.2	11.3	
Average	10.5	9.5	8.8	
(c) Ten-Day Daily Means				
Goose	7.8	8.4	6.3	
Edmonton	8.7	14.1	10.8	
Summerland	4.9	7.0	5.5	
Vancouver	7.5	4.4	5.0	
Sandspit	14.2	8.3	10.4	
Average	8.6	8.4	7.6	

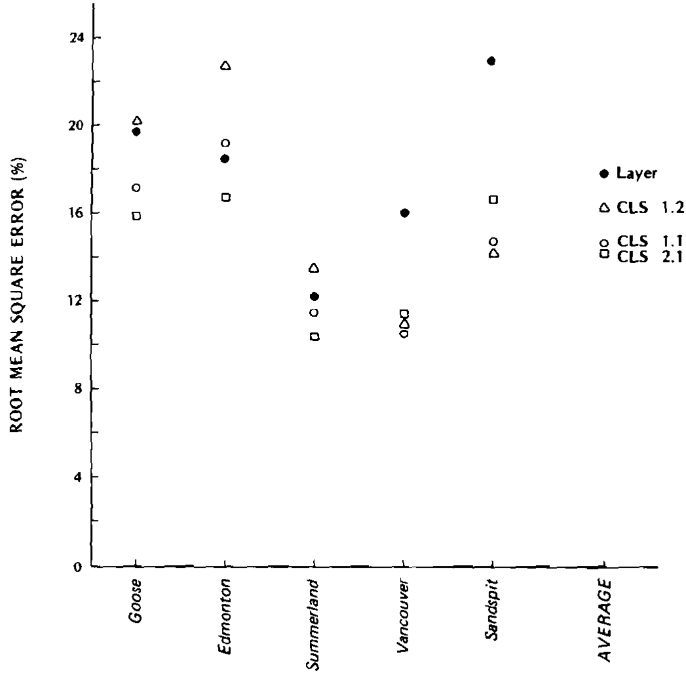

Fig. 2 Root mean square errors (%) on a daily basis for the layer and CLS solar radiation models during 1968–70.

Three versions of the model were tested with the empirical weighting coefficients in equation (1) being given arbitrary values of: (i) $A = B = 1$; (ii) $A = 2$ and $B = 1$; and (iii) $A = 1$ and $B = 2$. The results for the estimation of $K\downarrow$ on a daily basis are given in Table 4. Of the three versions, the first two (which will be referred to as the CLS 1, 1 and CLS 2, 1 models) performed much better than the third (referred to as the CLS 1, 2 model). When the results from all five stations were averaged, the three models had average root mean square error values of 14.6%, 14.2% and 16.3%, respectively. The Davies et al. (1975) Layer model was also used to calculate $K\downarrow$ for these five Canadian locations. Performances of the two approaches are compared in Table 5(a) and Fig. 2. The CLS 2, 1 model estimated solar radiation better than the Layer model at each of the five locations and by 3.7% on average (root mean square error of 14.2% compared to 17.9%). For the west coast stations of Vancouver and Sandspit, the CLS 1, 1 and CLS 1, 2 models showed even more improvement. The performance of the CLS 1, 1 model for estimating $K\downarrow$ at Vancouver is shown in Fig. 3 for a 10% random sample of days. Systematic errors do occur with the CLS model, these being over-estimations at Goose, Edmonton and Summerland (for the CLS 1, 1 and CLS 2, 1 models) and under-estimations at Vancouver and Sandspit.

The central limit theorem would suggest that the calculation of solar radia-

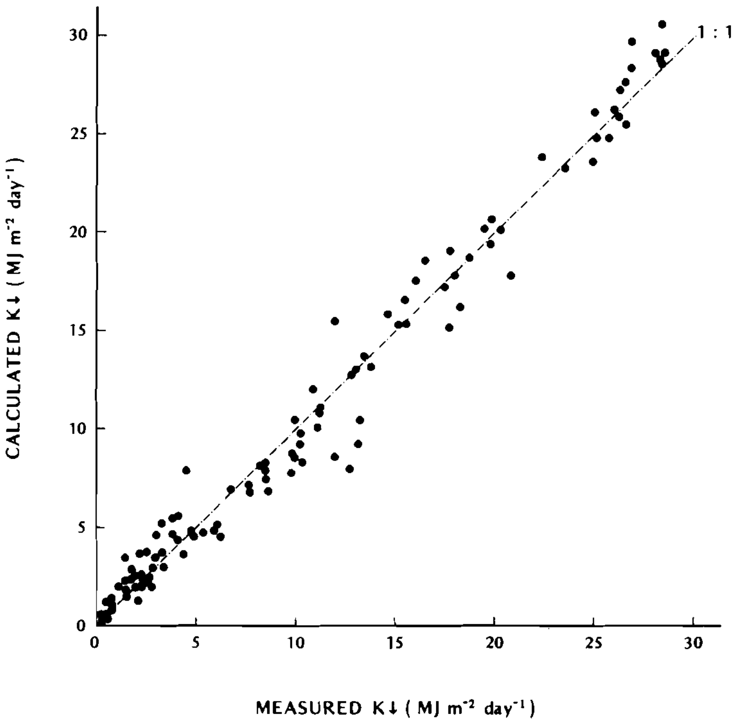

Fig. 3 Performance of the CLS 1, 1 model for estimating $K\downarrow$ at Vancouver for a 10% random sample of days during 1968–70. (Unit: MJ m^{-2} day^{-1}.)

tion fluxes over longer periods should produce better agreement between the estimated and measured values. The results of the Davies et al. (1975) Layer model and the CLS 1, 1 and CLS 2, 1 models were analysed for five-day and ten-day daily means (see Table 5(b) and 5(c), respectively). Compared to the daily $K\downarrow$ performances (Table 5(a)), it can be seen that in all cases, improvement resulted when values were averaged over longer periods. From Table 5 it can also be seen that the advantage of using the CLS model over the Layer model diminishes as the averaging period is lengthened. Compared to the CLS 2,1 model, the Layer model becomes superior for Edmonton for five-day daily means and for both Summerland and Edmonton for ten-day daily means. For several of the locations, both models are approaching or are within ±5% accuracy when modelled values are averaged over ten days.

The effectiveness of the CLS model for estimating the separate direct and diffuse components of solar radiation can be analysed only for Goose since diffuse solar radiation measurements are available at this station. Table 6 summarizes these results. Although the relative root mean square errors for $S\downarrow$ and $D\downarrow$ (28.7% and 30.4%, respectively) are much higher than for $K\downarrow$ (15.3%) for the CLS 2, 1 model, the corresponding absolute errors are in a similar range (1.43 and 1.79 MJ m^{-2} day^{-1} compared to 1.67 MJ m^{-2} day^{-1}). The systematic errors, indicated by the ratio of average measured to

TABLE 6. Performance of the CLS model for $S\downarrow$ and $D\downarrow$ estimation at Goose on a daily basis for 1968–70. (Unit for Mean and RMSE: MJ m^{-2} day^{-1})

	A	B	Number days	Mean		RMSE		Measured/ modelled
				Measured	Modelled		(%)	
K↓	1	1	938	10.88	11.20	1.82	16.7	0.972
S↓	1	1	938	4.99	5.62	1.55	31.1	0.889
D↓	1	1	938	5.89	5.59	1.79	30.4	1.055
K↓	2	1	938	10.88	10.96	1.67	15.3	0.993
S↓	2	1	938	4.99	5.40	1.43	28.7	0.925
D↓	2	1	938	5.89	5.56	1.79	30.4	1.059
K↓	1	2	938	10.88	11.46	2.11	19.4	0.949
S↓	1	2	938	4.99	5.84	1.82	36.5	0.855
D↓	1	2	938	5.89	5.62	1.78	30.3	1.047

average calculated flux, show that the direct and diffuse components are, on the average, over- and under-estimated, respectively.

The average model estimation errors for daily totals of $K\downarrow$ less than ± 2 MJ m^{-2} day^{-1} indicated in Table 4 can be compared to the errors which occur when solar radiation data are obtained by extrapolation from the measuring stations. Suckling and Hay (1976b) have shown that extrapolations of measured solar radiation over distances greater than 50 km in British Columbia and Alberta produce average errors for $K\downarrow$ estimation greater than 2 MJ m^{-2} day^{-1}. Therefore, it is advantageous to use the CLS model for estimating $K\downarrow$ for any location greater than 50 km from an existing solar radiation measuring station. Even shorter distances will yield similar errors when local factors are particularly well developed. Under such cases and, if the representative meteorological data are available, it would be more appropriate to calculate numerically rather than simply to extrapolate the solar radiation data.

4 Utility of the model in Canada

As of January 1, 1976, there were 51 stations measuring total solar radiation ($K\downarrow$) in Canada (Atmospheric Environment Service, 1976). The CLS model can supplement this solar radiation network by calculating $K\downarrow$ for locations where hourly meteorological (including cloud and bright sunshine) data are available. Values of precipitable water for the cloudless sky solar radiation calculations would have to be assigned using nearby radiosonde stations, meteorological grid-point data (Suckling, 1977) or statistical relationships between precipitable water and surface vapour pressure (e.g. Monteith, 1961; Idso, 1969; Hay, 1970b). The Canadian stations where solar radiation is measured or where there is a potential to model solar radiation data were determined from climatological data inventory catalogues (Atmospheric Environment Service, 1976). The results of this survey are summarized in Table 7 and Fig. 4. A total of 134 locations (51 measured and 83 modelled) could have solar radiation data. This is 2.6 times the number of measuring stations. The increase is small in the northern territories while a three-fold increase in the number of locations is possible in southern Canada.

TABLE 7. Number of solar radiation stations in Canada (Measuring and Potential modelling as of January 1, 1976)

Region	Number of stations for Measuring	Modelling	Total	Fractional Increase
East Coast Provinces	7	17	24	3.4
Quebec	6	11	17	2.8
Ontario	6	13	19	3.2
Prairie Provinces	8	23	31	3.9
British Columbia	9	14	23	2.6
The North	15	5	20	1.3
Canada (excl. North)	36	78	114	3.2
Canada	51	83	134	2.6

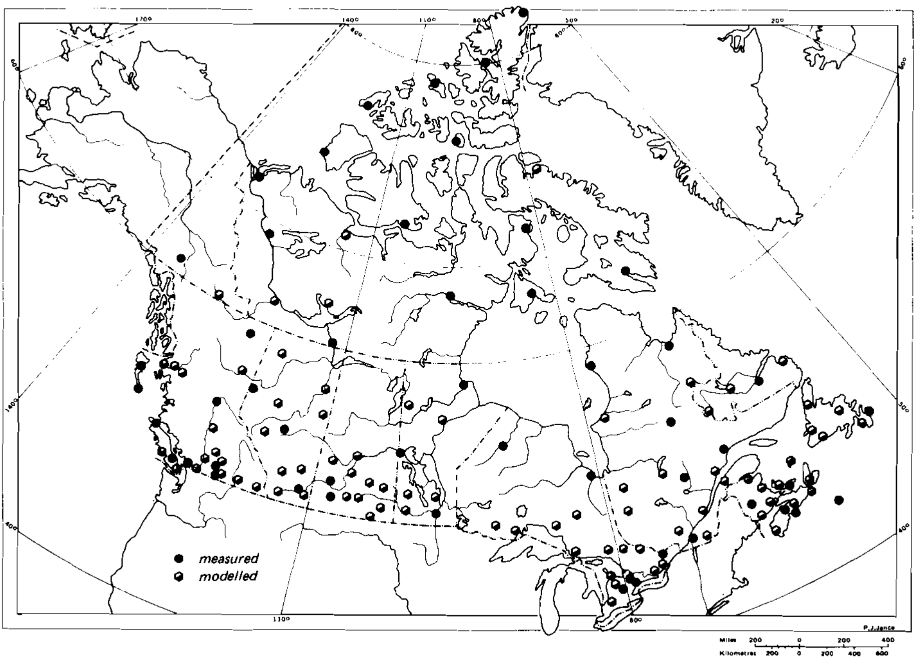

Fig. 4 Available solar radiation network in Canada as of January 1, 1976: ● measured; ◑ modelled.

Historically, solar radiation has generally been measured for a relatively short period of time when compared to the much longer cloud and sunshine records (Atmospheric Environment Service, 1972). With the CLS model the solar radiation record for many locations could be extended back in time thereby providing a more reliable data set from which to determine averages, variabilities and trends.

There are only four Canadian stations (Goose, Toronto, Montreal and

Resolute, N.W.T.) which currently and routinely measure diffuse solar radiation (Atmospheric Environment Service, 1976). Since the CLS model calculates the direct and diffuse components separately, this model has great potential for increasing our spatial knowledge of these component fluxes.

5 Conclusion

A previous study (Suckling and Hay, 1976a) described a method for calculating hourly values of the direct and diffuse solar radiation for cloudless sky conditions. In this study, the effects of cloud have been incorporated through the use of hourly values of cloud amount and type for up to four layers and hourly bright sunshine totals. The latter data provide a more accurate measure of the length of time the direct radiation of the sun is not attenuated by cloud. On an average, this cloud layer-sunshine (CLS) model estimated daily total solar radiation at five Canadian locations during the period 1968–70 to within ±15 per cent of the measured values. This was an improvement over an earlier model (Davies et al., 1975) based on cloud data alone, but the relative advantage, as well as the overall errors themselves, diminished as the averaging period was increased to five and ten days. The CLS model, however, has the additional advantage of calculating the separate direct and diffuse component fluxes of the total solar radiation.

Analysis of the available solar radiation and meteorological data in Canada shows that the CLS model has a potential to almost triple the number of locations producing total solar radiation data and for many locations can extend the historical records of solar radiation backwards in time.

Acknowledgements

The financial assistance and cooperation of the Canadian Atmospheric Environment Service and the National Research Council of Canada are gratefully acknowledged.

References

ÅNGSTRÖM, Å. 1924. Solar and terrestrial radiation. *Quart. J. Roy. Meteor. Soc.* **50**: 121–125.

ATMOSPHERIC ENVIRONMENT SERVICE. 1972. *Climatological Data Archive Inventory Summary*. Environment Canada, Downsview, Ont.

———. 1976. *Climatological Station Data Catalogue*. Environment Canada, Downsview, Ont.

ATWATER, M.A. and P.S. BROWN, JR. 1974. Numerical computations of the latitudinal variation of solar radiation for an atmosphere of varying opacity. *J. Appl. Meteor.* **13**: 289–297.

BENNETT, I. 1965. Monthly maps of mean daily insolation for the United States. *Solar Energy* **9**: 145–158.

BROOKS, C.F. and E.S. BROOKS. 1947. Sunshine recorders: A comparative study of the burning-glass and thermometric systems. *J. Meteor.* **4**: 105–115.

CANOVER, J.H. 1965. Cloud and terrestrial albedo determinations from Tiros satellite pictures. *J. Appl. Meteor.* **4**: 378–386.

DAVIES, J.A.; W. SCHERTZER; and M. NUNEZ. 1975. Estimating global solar radiation. *Boundary-Layer Meteor.* **9**: 33–52.

DEPARTMENT OF TRANSPORT. 1970. *Manual*

of Standard Procedures for Surface Weather Observing and Reporting. Meteorological Branch, Toronto.

DRUMMOND, A.J. and J.R. HICKEY. 1971. Large-scale reflection and absorption of solar radiation by clouds as influencing earth radiation budget: New aircraft measurement. Preprints, *International Conference on Weather Modification.* Canberra, Australian Academy of Science and American Meteorological Society, pp. 267–276.

DUFFIE, J.A. and W.A. BECKMAN. 1974. *Solar Thermal Energy Processes.* Wiley, New York, 386 pp.

EXELL, R.H.B. 1976. The solar radiation climate of Thailand. *Solar Energy* **18**: 349–354.

FURNIVAL, G.M.; W.E. REIFSNYDER; E. WYLER; and T.G. SICCAMA. 1970. Solar climate calculator. In *Proceedings of the Third Forest Microclimate Symposium*, Canadian Forestry Service, pp. 181–187.

GARNIER, B.J. 1975. A note on the climatology of solar energy potential. In *The Potential of Solar Energy for Canada.* Conference of the Solar Energy Society of Canada Inc. (R.K. Swartman, ed.), Ottawa, pp. I 43–46.

HAURWITZ, B. 1948. Insolation in relation to cloud type. *J. Meteor.* **5**: 110–113.

HAY, J.E. 1970a. *Aspects of the Heat and Moisture Balance of Canada.* PH.D. thesis, University of London, 2 vols., 322 pp.

———. 1970b. Precipitable water over Canada. I Computation. *Atmosphere* **8**: 128–143.

———. 1976. A revised method for determining the direct and diffuse components of the total short-wave radiation. *Atmosphere* **14**: 278–287.

HOUGHTON, H.G. 1954. On the annual heat balance of the Northern Hemisphere. *J. Meteor.* **11**: 1–9.

IDSO, S.B. 1969. Atmospheric attenuation of solar radiation. *J. Atmos. Sci.* **26**: 1088–1095.

KIMBALL, H.H. 1927. Measurements of solar radiation intensity and determinations of its depletion by the atmosphere with bibliography of pyrheliometric measurements. *Mon. Wea. Rev.* **55**: 155–169.

———. 1928. Amount of solar radiation that reaches the surface of the earth on the land and sea and methods by which it is measured. *Mon. Wea. Rev.* **56**: 393–398.

KIMURA, K. and D.G. STEPHENSON. 1969. *Solar Radiation on Cloudy Days.* Research Paper No. 418, Division of Building Research, National Research Council of Canada, Ottawa, 8 pp.

KONDRATYEV, K. YA. 1969. *Radiation in the Atmosphere.* Academic Press, New York, 912 pp.

LETTAU, H. and K. LETTAU. 1969. Shortwave radiation climatonomy. *Tellus* **21**: 208–222.

LIOU, K.-N. 1976. On the absorption, reflection and transmission of solar radiation in cloudy atmospheres. *J. Atmos. Sci.* **33**: 798–805.

LIST, R.J. 1966. *Smithsonian Meteorological Tables.* 6th revised edition, The Smithsonian Institution, Washington, 527 pp.

LONDON, J. 1957. *A Study of the Atmospheric Heat Balance.* Final Report No. AF. 19(122)–165, Department of Meteorology and Oceanography, New York University, 75 pp.

LUMB, F.E. 1964. The influence of cloud on hourly amounts of total solar radiation at the sea surface. *Quart. J. Roy. Meteor. Soc.* **90**: 43–56.

MANABE, S. and R.F. STRICKLER. 1964. Thermal equilibrium of the atmosphere with a convective adjustment. *J. Atmos. Sci.* **21**: 361–385.

MATEER, C.L. 1963. On the relationship between global radiation and cloudiness at Ocean Station P. *Arch. Meteor. Geophy. Bioklim.* **B 12**: 482–490.

MONTEITH, J.L. 1961. An empirical method for estimating long-wave radiation exchanges in the British Isles. *Quart. J. Roy. Meteor. Soc.* **87**: 171–179.

———. 1962. Attention of solar radiation: a climatological study. *Quart. J. Roy. Meteor. Soc.* **88**: 508–521.

NEUMANN, J. 1954. Insolation in relation to cloud amount. *Mon. Wea. Rev.* **82**: 317–319.

NUNEZ, M.; J.A. DAVIES; and P.J. ROBINSON. 1971. *Solar Radiation and Albedo at a Lake Ontario Tower Site.* Third report, No. HO81276, Department of Geography, McMaster University, Hamilton, Ont. 82 pp.

REYNOLDS, D.W.; T.H. VONDER HAAR; and S.K. COX. 1975. The effect of solar radiation absorption in the tropical troposphere. *J. Appl. Meteor.* **14**: 433–444.

SELLERS, W.D. 1965. *Physical Climatology.* The University of Chicago Press, Chicago, 272 pp.

SUCKLING, P.W. 1977. *A Solar Radiation Model and an Analysis of Synoptic Solar Radiation Regimes in British Columbia.* PH.D. thesis, Department of Geography, University of British Columbia, 271 pp.

SUCKLING, P.W. and J.E. HAY. 1976a. Modelling direct, diffuse and total solar radiation for cloudless days. *Atmosphere* **14**: 298–308.

——— and ———. 1976b. The spatial variability of daily values of solar radiation for British Columbia and Alberta, Canada. *Climatol. Bull.*, McGill University, **20**: 1–7.

THOMPSON, E.S. 1976. Computation of solar radiation from sky cover. *Water Resources Research* **12**: 859–865.

VONDER HAAR, T.H. and S.K. COX. 1972. Simultaneous measurements of solar radiation from aircraft and satellites during BOMEX. In preprints, Conference on Atmospheric Radiation. Fort Collins, Colo., Amer. Meteor. Soc., pp. 128–133.

VOWINCKEL, E. and S. ORVIG. 1962. Relation between solar radiation and cloud type in the Arctic. *J. Appl. Meteor.* **1**: 552–559.

VEGETATION AND THE ATMOSPHERE, volumes I and II. J.L. Monteith, editor. Academic Press, 1976. Vol. I, 278 pp., $26.00; Vol. II, 439 pp., $38.75.

These two volumes attempt to summarize the micrometerological work, done mainly during the last 20 years, which has a direct bearing on plant ecology. Thus they are aimed not only at meteorologists but also at biologists, hydrologists, foresters, and agronomers. They are advertised as advanced textbooks for senior students and research workers. Given the wealth of material accumulated in recent years and the piecemeal approach of most researchers, such an overview is desirable.

Volume I presents the basic principles of micrometeorological research. Volume II contains case studies, attempting a synopsis of available data for selected types of vegetation. Each chapter is written by a different author. The international (SI) system of units (with identical symbols) is used throughout both volumes.

Volume I: The introduction on micrometeorology and ecology (Elston and Monteith) outlines the essential differences between the physical and plant-physiological approach and the need for synthesis between the two if further progress is to be achieved. The discussion of radiation (Ross) gives a good account of the complexity of radiative transfer in vegetation. The section on momentum, mass, and heat exchange by Thom is perhaps too heavily centred on his own work and that of his colleagues in the U.K. Emphasis is on traditional one-dimensional concepts, almost without discussion of what he calls "non-characteristic profile shapes" (with vanishing or negative velocity gradients). More recent studies have shown these to be not all that uncommon even in extended crops, let alone in vegetation that does not live up to the "normal" fetch requirements. The hydrological cycle section (Rutter) is a well-balanced overview, commendably interpreting data in terms of the geographical or climatological conditions under which they were obtained. Chamberlain's work on the movement of particles is a thorough summary of field and laboratory experiments with analysis of the differences between the two. Discussion of mathematical models by Waggoner occasionally leaves the reader with a suspicion that the author jumps too optimistically from the proven success of very limited models to anticipated success of more complex ones. Successful predictions of temperature, humidity, and CO_2 profiles are presented, but we do not learn how consistently such agreement could be obtained. The chapter on instrumentation (Szeicz) could be very valuable, providing exactly those almost-obvious experimental hints that too often get overlooked in the heat of battle.

Volume II: The chapters on cereals (Denmead), maize and rice (Uchijima), sugar beet and potatoes (Brown), sunflower (Saugier), and cotton (Stanhill) illustrate the degree of sophistication of experimental work done on commercially important crops. All authors run through the same cycle of radiation, transfer processes, and CO_2 exchange (with biomass production), with individual additions and deletions (very few give IR emissivities; only one discusses transpiration suppressants), so that one might ask if a more straightforward analysis could have been achieved if the editor had predigested the information and classified the relative similarities and differences between different vegetative systems. Yet some compensation for the occasional duplication results, through a growing appreciation of the diversity of this field, stemming from the bias and preoccupations of the different authors, which would have been lost in a more homogenized presentation. It is not clear, however, why rice and maize were combined in the same chapter, a forced marriage producing some strange offsprings (such as reflection coefficient for rice and profiles of net radiation for maize in the same figure). Most authors are short on discussion of experimental accuracy. The notable exception is Saugier who is also the only one to provide a welcome summarizing conclusion.

Townsville Stylo (Rose, Begg, and Torssell) is an interesting ecological management problem; the description of possible effects of intermittent cloud cover and wind may not agree with the one-dimensional steady-state picture, an aspect largely ignored by the previous authors. The review on coniferous forests (Jarvis, James, and Landsberg) is very extensive and thorough, but one wishes for a summary and outlook. By comparison, there is a very meager, unsatisfactory chapter on deciduous forests (Rauner), which quotes largely from the author's book on the subject, and has 14 references (compared with 160 on coniferous forests), of which only 5 are by non-Russian authors. The paper on CO_2 exchange and turbulence in a Costa Rica forest (Allan and Lemon) is an instructive demonstration of a working model on biomass production, seen close enough to reveal "warts and all." The section on citrus orchards (Kalma and Fuchs) has some interesting reading, but would have profited by some extrapolation (where to go next?) and even speculation. The sections on swamps (Linacre), grasslands (Ripley and Redmann), and tundra (Lewis and Callaghan) weigh in less heavily in terms of data because of lesser economic significance, but are of particular interest to the Canadian reader.

Both volumes are readable compendia of a wealth of information, never long winded, ideal for looking up order-of-magnitude figures, or for quickly reviewing concepts on the full spectrum of agrometeorological phenomena. They are not textbooks insofar as concepts are often summarized rather than derived (a lot of ... "has been shown to be" ...), but full referencing details can be obtained from the literature. They will therefore complement rather than replace textbooks on micrometeorology. The referencing is generally fair and comprehensive, with the exception of one or two authors who succumb to the temptation to view their field too much from the focal point of their own work without warning the reader of this bias. Some references (particularly Russian, to a lesser degree European and Japanese ones) are not readily available to the average reader; they will however provide a useful indication of work done outside the normal field of vision of the North American scientist. The editor must be faulted for some of the reservations expressed in this discussion. There is a relatively high number of minor errors and misprints which dot particularly Volume I with very uneven density, such as missing or improperly listed references and abbreviations, and small errors or omissions in formulas or figures. They are not too difficult to spot and hopefully will be corrected in future printings.

In terms of their goals, it appears that these books are ideal for researchers already working in one of the disciplines of agrometeorology who do not normally see their field in the overall perspective. For such workers, they should be highly recommended as a good source of reference material. They will also appeal to any meteorologist who cares to extend his view of the atmosphere to include its lower boundary. They will easily convince the micrometeorologist to pay more attention to plant-physiological processes. Whether or not (in style and language) it encourages the biologist to draw more heavily on the physicist's contribution is another question that the meteorologically trained reader cannot easily decide.

Peter H. Schuepp
Department of Agricultural Chemistry and Physics
Macdonald Campus of McGill University

ATMOSPHERIC SCIENCE. John M. Wallace and Peter V. Hobbs. Academic Press, New York, 1977, 467 pp., $ U.S. 17.95.

In the Preface, the authors explain that this textbook is intended to support several university courses in atmospheric sciences: introductory survey courses at the junior or senior undergraduate level or the first year of graduate study; the undergraduate physical meteorology course; and the undergraduate synoptic laboratory. Reflecting these different uses, the book begins with an appealing survey chapter and follows with chapters on thermodynamics, synoptic-scale disturbances, cloud and precipitation physics, radiation, the global energy balance, dynamics, and the general circulation, in that order. Thus embodying a fairly even mix of physical and dynamical meteorology, it is perhaps more closely akin to Hess's very serviceable *Introduction to Theoretical Meteorology* than to any other book in current use, though there is obviously some overlap with the topics in Byers's venerable *General Meteorology*. Entirely up-to-date, written with great clarity, and judiciously balancing detail against breadth of coverage, this new book deserves to take first position among the survey texts.

Atmospheric Science is pitched at an intermediate level and is therefore slightly more advanced than the texts of Byers and Hess. For instance, the Helmholtz and Gibbs potentials are introduced in thermodynamics, and the distinction between radiance (intensity) and irradiance (flux) is carefully drawn in radiation. SI units are used throughout for the first time in a meteorology text. Also as an indication of the book's timeliness, sections are included on atmospheric aerosols (a concise 15-page review), profile-inversion from satellite measurements, and weather modification.

The book is attractively produced and contains a liberal number of half-tone prints. Cloud photographs and satellite images are used extensively and effectively to enliven the narrative. An especially nice touch is the inclusion of biographical footnotes on the many scientists who have contributed to meteorology. Ranging from Richard Assman to C.T.R. Wilson, these entries include such outstanding scientists as Newton, Poisson, and Kelvin, and lesser known but remarkable individuals such as Luke Howard and Christopher Buys-Ballot. Every chapter contains problems and answers, usually in profusion (52 in thermodynamics, 37 in radiation). Most are listed at the end of each chapter but some are used in the text to illustrate a point or introduce a related concept. Also attractive from the instructor's viewpoint is the thoughtful arrangement of material within chapters which permits some of the more difficult concepts to be omitted with no loss in continuity.

In addition to being well suited for the classroom, this reasonably priced book should enjoy considerable use as a reference on general meteorology.

R.R. Rogers
McGill University
Montreal

LASER MONITORING OF THE ATMOSPHERE. Volume 14 in Topics in Applied Physics. Edited by E.D. Hinkley, Springer-Verlag, Berlin, Heidelberg, New York, 1976, 380 pp., $8.00.

Optical techniques have long been employed for studies of the atmosphere. The advent of laser sources has led to substantial improvements in many of these techniques and has introduced a number of new methods of atmospheric observation with greatly improved spatial and temporal resolution. As a result of these new capabilities, the role of lasers in atmospheric monitoring has steadily increased in the last decade. There is already available a sizeable body of information on the demonstrated capabilities of laser systems based on the optical properties of scattering, fluorescence, absorption, and emission. The aim of this book is to present in a comprehensive and tutorial manner these fundamental techniques of laser diagnostics of the atmosphere.

Each chapter covers a different application area and is written by one or more experts in the particular field.

The chapter headings and authors are:

1. Introduction. By E.D. Hinkley
2. Remote Sensing for Air Quality Management. By S.H. Melfi
3. Laser-Light Transmission Through the Atmosphere. By V.E. Zuev
4. Lidar Measurement of Particles and Gases by Elastic Backscattering and Differential Absorption. By R.T.H. Collis and P.B. Russell

5. Detection of Atoms and Molecules by Raman Scattering and Resonance Fluorescence. By H. Inaba
6. Techniques for Detection of Molecular Pollutants by Absorption of Laser Radiation. By E.D. Hinkley, R.T. Ku and P.L. Kelley
7. Laser Heterodyne Detection Techniques. By R.T. Menzies

Although each chapter is written by separate authors the editor has been successful in achieving a good continuity between the chapters. There is considerable cross-referencing and a uniform set of symbols is provided in Chapter 1. Chapter 2 summarizes briefly the structure and properties of the atmosphere and indicates how laser techniques can be applied to their measurement. Emphasis is placed on air quality and pollution monitoring requirements. Chapter 3 provides a basic overview of the fundamental properties of light transmission through the atmosphere and pays special attention to the quantitative aspects relating to laser beam propagation. In Chapter 4 the lidar (laser radar) measurement technique is described in considerable detail. Lidar measurements involving both scattering and absorption processes are presented and the chapter is liberally provided with descriptions of experimental equipment and observational results. In Chapter 5 the material covers the applications of Raman scattering and resonance fluorescence for the measurements of atmospheric properties utilizing the lidar mode of operation. This chapter contains a substantial amount of quantitative information (based on both theory and experiment) on the capabilities of these techniques for atmospheric monitoring.

Chapter 6 deals primarily with the use of tunable lasers for absorption measurements in the atmosphere. A summary is given of the various tunable lasers available. The associated detection techniques, particularly in the infra-red, are highlighted. The final chapter deals with the laser heterodyne technique in which the laser is employed as a local oscillator in a coherent radiometer system to enhance the signal-to-noise characteristics of the system. The material contains a rather complete overview of the essential advantages and limitations of this method.

I can recommend this book as an extremely useful addition to the library of anyone interested in atmospheric diagnostics. Its tutorial approach, with sufficient coverage of the background and fundamental material, would make it very useful reading for any beginner in the field. In addition the expert already well aware of these activities will find that this book presents a good compilation of recent information. The book contains close to seven hundred references which are reasonably complete to about 1975. These references are grouped appropriately at the end of each chapter and in addition a number of references are organized under topical headings at the end of the book.

A.I. Carswell
Department of Physics
York University

Sea Ice Processes and Models

A symposium discussing these topics was held in Seattle on September 6–9, 1977. The occasion was the impending wind-up of the Arctic Ice Dynamics Joint Experiment (AIDJEX). This project, mainly American although with substantial Canadian input, included, as the experimental basis for the understanding of sea ice dynamics on the Polar Ocean, preliminary pilot programs in 1970–1972, and the main experiment from March 1975 to May 1976 in the northern Beaufort Sea.

The main output of the project was a model of dynamics and thermodynamics of polar sea ice. The model incorporated a momentum balance equation describing air and water stress upon the ice, Coriolis forces, sea surface tilt forces and internal ice stresses; constitutive laws relating ice stress and strain; a yield strength relationship determined from the rate of work done upon the ice; and an ice-thickness distribution equation including both thermodynamic and mechanical processes. The model has been coded as a finite-difference numerical scheme and has been tested in preliminary fashion using data gathered in the pilot experiments. Shore boundary effects have also been taken into account. The main problems concern calculations of the internal ice stress, the optimum form of which is still under debate.

In the Arctic Ocean, as a rule, the wind stress drives the sea ice sheet which in turn drives the water. The meteorological program of AIDJEX had as its main purpose to relate the air stress upon the ice surface to the easily measured parameter, the geostrophic wind. This was done with some success upon the open Arctic Ocean, but greater difficulties can be foreseen in the narrow channels of the Canadian Arctic Archipelago. During these meteorological experiments a good deal of very useful boundary-layer data was gathered under Arctic conditions. In peripheral studies heat-flux measurements were made over narrow leads. The water stress on the bottom of the ice sheet was measured (by groups from McGill) and many other oceanographic data were gathered. Perhaps most interesting were observations of vigorous eddies with a horizontal size of about 10 km and a vertical scale of 100 m found in the stable layer 100-300 m below the bottom of the ice. A few papers on remote sensing were included in the AIDJEX reports.

About half the time at the symposium was occupied by reports on sea-ice studies not directly involved in AIDJEX. Several models of sea-ice dynamics and thermodynamics of various complexities were offered as competition for the AIDJEX model. Some individuals who have actually forecast sea-ice conditions for a few months ahead reported that their techniques were analog, relating ice conditions in the area of interest to the atmospheric pressure distribution occurring some months previously. Since no practical methods exist to measure sea-ice thicknesses over large areas there was considerable interest in a paper which reported near-simultaneous measurements of the ice-bottom depths by sonar from a British nuclear submarine on her way to the North Pole and the ice-top elevations by laser from a Canadian Argus in flight overhead. One paper described differences in January climates, as computed by the GISS general circulation model, corresponding to minimum and maximum extent of Arctic sea ice. Results for maximum ice included colder temperatures over the north Pacific Ocean, over the North Atlantic Ocean and across North America; larger poleward temperature gradients; larger eddy sensible-heat transfer; and in general a slightly more energetic circulation.

All in all a very stimulating symposium. Of greatest interest to Canadian meteorologists might be the large bank of boundary-layer data for Arctic conditions, and possibly some of the ideas behind the ice dynamics models.

E.R. Walker
Institute of Ocean Sciences
Patricia Bay, B.C.

CALL FOR PAPERS – TWELFTH ANNUAL CONGRESS

The Twelfth Annual Congress and Annual General Meeting of the Canadian Meteorological and Oceanographic Society will be held at The University of Western Ontario May 31-June 2, 1978. The theme of the opening session will be entitled "Energy, the Meteorologist and the Oceanographer" as presented by invited and contributed papers. Subsequent sessions will deal with contributed papers on meteorology and oceanography, including sessions of common interest according to the papers submitted.

Titles and definitive abstracts (less than 300 words) should reach the program committee by *1 February, 1978*. Papers on meteorology should be sent to Prof. D.R. Hay, Department of Physics, The University of Western Ontario, London, Ont. N6A 3K7. Papers on oceanography should be sent to Mr Farrell M. Boyce, Canada Centre for Inland Waters, 867 Lakeshore Road, Burlington, Ont. L7R 4A6.

CALL FOR NOMINATIONS – 1977 AWARDS

Nominations are requested from members and Centres for the 1977 CMOS Awards to be presented at the 1978 Annual Meeting. Four awards are open for competition: 1) the President's Prize for an outstanding contribution in the field of meteorology by a member of the Society; 2) the Dr Andrew Thomson Prize in Applied Meteorology, for an outstanding contribution by a member in that field; 3) the Graduate Student Prize, for a contribution of special merit, and 4) the Rube Hornstein Prize in Operational Meteorology, for outstanding service in providing operational meteorological service. The awards will be made on the basis of contributions during the 1977 calendar year, except 4) which may be awarded also for work performed over a period of years.

Nominations are also requested for the award of citations to individuals or groups in Canada who have made some outstanding contribution in helping to alleviate pollution problems or in developing environmental ethics.

All nominations should reach the corresponding secretary not later than *1 February 1978*.

NOTICE OF FEES INCREASE

The following fees for 1978 were approved at the Eleventh Annual General Meeting:

General Member	\$25.00
Student Member	\$ 5.00
Sustaining Member	\$60.00 (min.)
Institutional Subscription	\$25.00

APPEL DE TEXTES - DOUZIEME CONGRES ANNUEL

Le douzième congrès annuel et l'assemblée générale annuelle de la Société canadienne de météorologie et d'océanographie auront lieu à l'Université Western Ontario du 31 mai au 2 juin 1978. Les présentations de la session d'ouverture traiteront du thème "L'énergie, le météorologiste et l'océanographe" tandis que les autres sessions seront consacrées aux présentations ayant trait à divers aspects de la météorologie et l'océanographie, incluant des sessions d'intérêt commun, selon les papiers soumis.

Les titres ainsi que les sommaires définitifs (maximum 300 mots) devront parvenir au comité du programme d'ici le *1er février 1978*. Les textes en météorologie devront être envoyés au professeur D.R. Hay, Département de Physique, l'Université Western Ontario, London, Ontario, N6A 3K7, tandis que ceux traitant d'océanographie devront être expédiés à M. Farrell M. Boyce. Centre Canadien des Eaux Intérieures, 867 Lakeshore Road, Burlington, Ontario, L7R 4A6.

APPEL AUX CANDIDATURES POUR LES PRIX ET CITATIONS 1977

On demande aux membres et aux centres locaux de soumettre leurs nominations aux candidatures pour les prix de la Société pour l'année 1977. Il y a quatre prix: 1) le prix du président pour un travail exceptionnel en météorologie par un membre de la Société, 2) le prix de météorologie appliquée Dr Andrew Thomson pour un travail exceptionnel dans ce domaine par un membre, 3) le prix aux étudiants gradués et 4) le prix de météorologie opérationnelle Rube Hornstein pour un travail exceptionnel dans l'exploitation des services météorologiques.

Tous les prix seront attribués pour un travail qui a été effectué durant l'année 1977, à l'exception de 4) qui peut aussi être attribué pour un travail effectué durant une période couvrant plusieurs années.

On demande aussi des nominations de candidats susceptibles de se voir décerner une citation par la Société. Une citation peut être décernée à un individu ou à un groupe qui a apporté une contribution exceptionnelle à la solution des problèmes de la pollution, à l'amélioration de l'environnement ou au développement d'une éthique écologique.

Toutes les nominations recueillies par le secrétaire-correspondant avant *le 1er février 1978* seront remises aux comités des récompenses et des citations, selon le cas.

AVIS D'AUGMENTATION DES COTISATIONS

Les cotisations suivantes pour 1978 ont été approuvées au cours de la onzième assemblée générale annuelle:

Membre	$25.00
Membre étudiant	$ 5.00
Membre de soutien	$60.00 (min.)
Groupe	$25.00

Page Numbers of Issues

Figures in italics refer to 11th Annual Congress Issue

Section A: Index of Contributions by Author

Section B: CMS Subject Index

Section C: General Subject Index

INFORMATION FOR AUTHORS

Editorial policy. *Atmosphere* is a medium for the publication of the results of original research, survey articles, essays and book reviews in all fields of atmospheric science. It is published quarterly by the CMS with the aid of a grant from the Canadian Government. Articles may be in either English or French. Contributors need not be members of the CMS nor need they be Canadian; foreign contributions are welcomed. All contributions will be subject to a critical review before acceptance. Because of space limitations articles should not exceed 16 printed pages and preferably should be shorter.

Manuscripts should be submitted to: *Atmosphere*, Dept. of Geography, University of British Columbia, 2075 Wesbrook Mall, Vancouver, British Columbia, V6T 1W5. Three copies should be submitted, typewritten with double spacing and wide margins. Heading and sub-headings should be clearly designated. A concise, relevant and substantial abstract is required.

Tables should be prepared on separate sheets, each with concise headings.

Figures should be provided in the form of three copies of an original which should be retained by the author for later revision if required. A list of legends should be typed separately. Labelling should be made in generous size so that characters after reduction are easy to read. Line drawings should be drafted with India ink at least twice the final size on white paper or tracing cloth. Photographs (halftones) should be glossy prints at least twice the final size.

Units. The International System (SI) of metric units is preferred. Units should be abbreviated only if accompanied by numerals, e.g., "10 m," but "several metres."

Footnotes to the text should be avoided.

Literature citations should be indicated in the text by author and date. The list of references should be arranged alphabetically by author, and chronologically for each author, if necessary.

RENSEIGNEMENTS POUR LES AUTEURS

Politique éditoriale. *Atmosphère* est un organe de publication de résultats de recherche originale, d'articles sommaires, d'essais et de critiques dans tous les domaines des sciences de l'atmosphère. Il est publié par la SMC à l'aide d'une subvention accordée par le gouvernement canadien. Les articles peuvent être en anglais ou en français. Il n'est pas nécessaire que les auteurs soient membre de la SMC; les contributions étrangères sont les bienvenues. A cause des limitations d'espace les articles ne doivent pas dépasser 16 pages dans le format final. Tout article sera soumis à un critique indépendant avant d'être accepté.

Les manuscrits doivent être envoyés à: *Atmosphère*. Dép. de géographie, Université de la Columbie-Britannique, 2075 Wesbrook Mall, Vancouver, La Columbie-Britannique, V6T 1W5. Ils doivent être soumis en trois exemplaires dactylographiés à doubles interlignes avec de larges marges. Les titres et sous-titres doivent être clairement indiqués. Chaque article doit comporter un résumé qui soit concis, pertinent et substantiel.

Les tableaux doivent être préparés et présentés séparément accompagnés d'un titre concis et d'un numéro.

Les graphiques doivent être présentés en trois copies dont les originaux devraient être conservés par l'auteur au cas où ils seraient nécessaire de les reviser. Une liste des légendes des graphiques doit être dactylographiée séparément. L'étiquettage doit être de grand format de façon à ce qu'il soit facilement lisible après réduction du format. Le traçage des lignes doit s'effectuer au moyen d'encre de chine en doublant, au moins, le format final, le tout sur papier blanc ou sur papier à calquer et identifié adéquatement. Les photographies (demi-teintes) devraient être présentées sur épreuves glacées au double du format final.

Les unités. Le Système International (SI) d'unités métriques est préférable. Les unités devraient être abrégées seulement lorsqu'elles sont accompagnées de nombres, ex: "10m," mais "plusieurs mètres."

Les notes de renvoie au texte doivent être évitées.

Les citations littéraires doivent être indiquées dans le texte selon l'auteur et la date. La liste des références doit être présentée dans l'ordre alphabétique, par auteur et, si nécessaire, dans l'ordre chronologique pour chaque auteur.

The Canadian Meteorological Society / La Société météorologique du Canada

The Canadian Meteorological Society came into being on January 1, 1967, replacing the Canadian Branch of the Royal Meteorological Society, which had been established in 1940. The Society exists for the advancement of Meteorology, and membership is open to persons and organizations having an interest in Meteorology. At nine local centres of the Society, meetings are held on subjects of meteorological interest. *Atmosphere* as the scientific journal of the CMS is distributed free to all members. Each spring an annual congress is convened to serve as the National Meteorological Congress.

Correspondence regarding Society affairs and CMS membership should be directed to the Corresponding Secretary, Canadian Meteorological Society, c/o Dept. of Geography, Simon Fraser University, Burnaby, B.C., V5A 1S6.

There are three types of membership – Member, Student Member and Sustaining Member. For 1978 the dues are $25.00, $5.00 and $60.00 (min.), respectively. The annual Institutional subscription rate for *Atmosphere* is $25.00.

Correspondence relating to institutional subscriptions should be directed to the University of Toronto Press, Journals Department, 5201 Dufferin St., Downsview, Ontario, Canada, M3H 5T8. Cheques should be made payable to the University of Toronto Press.

La Société météorologique du Canada a été fondée le 1er janvier 1967, en replacement de la Division canadienne de la Société royale de météorologie, établie en 1940. Cette société existe pour le progrès de la météorologie et toute personne ou organisation qui s'intéresse à la météorologie peut en faire partie. Aux neuf centres locaux de la Société, on peut y faire des conférences sur divers sujets d'intérêt météorologique. *Atmosphère*, la revue scientifique de la SMC, est distribuée gratuitement à tous les membres. A chaque printemps, la Société organise un congrès qui sert de Congrès national de météorologie.

Toute correspondance concernant les activités de la Société et les souscriptions à la SMC devrait être adressée au Secrétaire-correspondant, Société météorologique du Canada, Département de Géographie, L'Université Simon Fraser, Burnaby, B.C., V5A 1S6.

Il y a trois types de membres: Membre, Membre-étudiant, et Membre de soutien. La cotisation pour 1978 est de $25.00, $5.00 et $60.00 (min.) respectivement. Les institutions peuvent souscrire à *Atmosphère* au coût de $25.00 par année.

La correspondance concernant les souscriptions des institutions doit être envoyée aux Presses de l'Université de Toronto, Département des périodiques, 5201 rue Dufferin, Downsview, Ontario, Canada, M3H 5T8. Les chèques doivent être payables aux Presses de l'Université de Toronto.

Council / Conseil d'administration 1977–78

ATMOSPHERE