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A Net Radiation Model for Calculating Daily Snowmelt in Open Environments

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Climatological data from three stations in western Canada are used to develop a model for estimating daily net radiation from station latitude, station altitude and cloud cover. Expressions describing: a) the seasonal variation in atmospheric transmissivity; b) the interaction between direct-beam and diffuse short-wave radiation and c) the effect of cloud cover on these components are presented.

The seasonal decrease in albedo of a shallow snowcover is approximated by three line segments. Albedo decreases at an average rate of 0.0061/day during premelt; at an average rate of about 0.071/day during melt and takes on a reasonably constant value of 0.17 when the ground is snowfree.

The exchange of long-wave radiation dominates the radiation balance of a complete snowcover. During these periods it is recommended that the net flux be calculated from observations of ambient air temperature and vapor pressure in standard expressions.

Introduction

Although only approximately one-third of the annual precipitation in the semi-arid region of the Canadian Prairies occurs as snowfall, the shallow snowcover it produces generates 80 % or more of the annual surface runoff from local areas. This runoff serves many beneficial uses: as a domestic and livestock supply, as a wildlife habitat, for recharging soil water reserves, and other purposes; conversely, it may cause flooding, soil erosion and drainage problems. Thus, various agencies concerned with water management have a keen interest in the development of improved methods for predicting the time of melt, the snowmelt rate and the volume of runoff.

Phenomenologically, the energy equation is the physical framework for snowmelt models. In open areas net radiation and sensible heat are the primary fluxes which govern melt. Numerous studies (O'Neill 1972; Granger 1977; Male and Granger 1979, 1981) have demonstrated the importance of net radiation to the energetics of melting prairie snowcovers. The net short-wave component, primarily governs both the timing of melt and the snowmelt rate; the net long-wave flux affects the time of release of snowcover runoff because of its influence on nighttime refreezing and thus on the internal energy status of a snowcover.

This paper focuses on the development of procedures for modelling net radiation from station latitude, station altitude and atmospheric parameters in open environments. Unlike the traditional approach used to calculate the flux of short-wave radiation incident to the earth's surface, that is by moderating the extra-terrestrial flux by cloud cover, the method divides the surface flux into its direct and diffuse components. Relationships for net short-wave, net long-wave and net "all-wave" fluxes are developed and verified against observed data and their application for calculating snowmelt is discussed.

Data

The empirical relationships presented in this paper are based on climatological observations from three stations located in the Prairie Provinces of western Canada: Bad Lake, Edmonton and Winnipeg. These stations monitor from two to four of the four standard radiation fields directly, namely; global, reflected, and diffuse short-wave radiation and net "all-wave" radiation. The instruments used for the measurement of short-wave radiation (Kipp and Zonen pyranometer) and all-wave radiation (Middleton net radiometer) typically have estimated root mean square errors of 4 % when properly used (Latimer 1972). Table 1 provides information on the coordinates and altitude of the stations, the parameters measured and the period of observation of the data included in the analyses.

Table 1 – Information on meteorological stations and observations used in the analysis of daily radiation

Station	Latitude (deg)	Longitude (deg)	Altitude masl	Parameters[a]	Observation Period
Edmonton	53° 33'N	114° 06'W	766	n, RF1,2,4	1977-83
Bad Lake	51° 20'N	108° 24'W	637	n, RF1,2,3,4	1980-85
Winnipeg	49° 54'N	97° 14'W	239	n, RF1,2	1977-83

[a] n = hours of bright sunshine; RF1 = global short-wave; RF2 = diffuse short-wave; RF3 = reflected short-wave; RF4 = net all-wave radiation.

Net Radiation

Net radiation (Q_n) is the sum of the net short-wave (Q_{sn}) and net long-wave (Q_{ln}) fluxes. Since the net long-wave exchange is usually negative, Q_n is expressed as

$$Q_n = Q_{sn} - Q_{ln} \tag{1}$$

Net Short-Wave

Net short-wave radiation is equal to the incident short-wave flux (Q_s) received by the surface less the amount reflected by the surface (Q_r), that is

$$Q_{sn} = Q_s - Q_r \tag{2}$$

Expressing the incident flux as the sum of its direct beam (Q_{drs}) and diffuse (Q_{dfs}) components gives

$$Q_s = Q_{drs} + Q_{dfs} \tag{3}$$

Incident Short-Wave

That portion of the electromagnetic radiation from the sun falling within the range from 0.2 to 4 μm is generally considered short-wave radiation. At a mean earth-sun distance of 149.5×10^6 km the short-wave flux *perpendicular* to the sun's rays is equal to the solar constant, 1.35 kW/m^2. The extra-terrestrial flux incident to a *horizontal* plane at the top of the earth's atmosphere (Q_A) however varies with latitude, season and time of day; but at a fixed geographical position and time it is constant. Q_A can be calculated from the solar constant, the radius vector of the earth's orbit (the distance from the center of the sun expressed in terms of the length of the semi-major axis of the earth's orbit, the sun's zenith distance, and the hour angle measured from solar noon. Values for these parameters are given by List (1968)).

The amount of solar radiation penetrating the atmosphere to be received by the earth's surface depends on the turbidity of the atmosphere, cloud cover, topography (altitude, slope and orientation), and other factors. While passing through the atmosphere, radiation is reflected by clouds, scattered diffusely by air molecules, dust, and other particles and absorbed by ozone, water vapor, carbon dioxide, and nitrogen compounds. The portion absorbed increases the ambient air temperature which in turn increases the amount of long-wave radiation emitted to the earth's surface and to outer space.

The total short-wave flux incident to the earth's surface, Q_s, often referred to as global radiation, is composed of direct-beam and diffuse components, the latter being direct radiation which has been reflected and scattered by clouds or other surfaces and diffused by atmospheric constitutents. As stated above, the common approach used to model Q_s moderates the extraterrestrial short-wave flux, Q_A, by atmospheric transmittancy and cloud cover. Numerous investigators (Penman 1948; de Jong 1973; Brutsaert 1982) have shown that estimates of short-wave radiation incident to a horizontal plane can be obtained by the linear expression

$$Q_s = Q_A \left[a + b \left(\frac{n}{N} \right) \right] \tag{4}$$

in which a and b are empirically-derived coefficients which account for the transmittancy of the atmosphere (see de Jong 1973), n is the actual (measured) number of hours of bright sunshine and N is the maximum possible number of hours of bright sunshine (see List 1968). A ratio of $n/N = 1$ represents a clear-sky condition; $n/N = 0$ designates a completely-overcast sky.

Direct Beam

In snowmelt investigations, where altitude, slope and aspect play an important role in the energy exchange process, it may be advantageous to resolve the incident short-wave flux into its direct-beam and diffuse components. The amount of direct-beam radiation received, which subsequently is either reflected or absorbed by the earth's surface, is strongly affected by topographic facets.

Clear-sky – A useful model describing the direct-beam component of solar radiation incident to a horizontal plane on the earth's surface under clear skies (Q_{dro}) is the following expression proposed by Garnier and Ohmura (1970)

$$Q_{dro} = \frac{I_0}{r^2} \int p^m \cos(X \Lambda S)\, dH \tag{5}$$

where

I_o	– solar constant (1.35 kW/m^2),
r	– radius vector of the earth's orbit (the distance from the center of the sun expressed in terms of the length of the semi-major axis of the earth's orbit,
p	– mean transmissivity of the atmosphere along the zenith path,
m	– optical air mass, the ratio of the distance the sun's rays travel through the atmosphere to the depth of the atmosphere along the zenith path,
$\cos(X \Lambda S)$	– cosine of the angle of incidence of the sun's rays on a slope. (X is the unit normal vector pointing away from the surface and S is the unit vector expressing the sun's position), and
H	– hour angle measured from solar noon. The integral is taken over the duration of sunlight on the slope.

Tables of values and mathematical expressions for estimating the parameters r, m, cos $X \Lambda S$ and H are reported by List (1968), Garnier and Ohmura (1970), Morton, (1983) and Granger and Gray (1990).

The transmissivity (p) represents the fraction of the direct-beam radiation which is allowed to penetrate a clear atmosphere and reach the earth's surface along the zenith path. Its magnitude is therefore affected by the mass of atmosphere lying between the sun and the earth's surface and the amount of water vapor, ozone, dirt, and other impurities it contains. p varies with location and is highest in winter and lowest in summer. Gates (1980) found that the transmissivity often ranges between 0.40 and 0.70, but where the air is clear, such as at high mountain elevations, p = 0.80. For the snowmelt period in the European U.S.S.R. Kuz'min (1972) recommends a value of 0.80 ± 0.05.

When radiation data are available, values for p are obtained by comparing measurements of global radiation on clear days against the corresponding values given by Eq. (5) with different values of "p". That value of "p" which gives the closest agreement between "measured" and "calculated" fluxes is adopted. For the three stations used in the study it was found that the seasonal variation in transmissivity could be described by the expression

$$p = 0.818 - 0.064\sin[(DAY-90)2\pi/365] \tag{6}$$

in which DAY is the Julian day number. Eq. (6) gives a seasonal variation in transmissivity between 0.75 and 0.88. The minimum value occurs on day 181 (June 30) and the maximum occurs on December 31, lagging the solstices by approximately 10 days.

Cloud Cover – Cloud cover reduces the amount of direct-beam short-wave radiation and this effect is shown in a plot of the ratio of the "observed" direct short-wave flux to the clear-sky direct short-wave flux, Q_{drs}/Q_{dro}, against the sunshine ratio, n/N, in Fig. 1. The scatter in Q_{drs}/Q_{dro} at a given n/N shown in the diagram is attributed to differences in the opacities of cloud types. Unfortunately, cloud type was not recorded and therefore the association could not be tested. Fig. 1 suggests that the relationship between the dimensionless terms is curvilinear and it was assumed the data could be described by the model

$$\frac{Q_{drs}}{Q_{dro}} = a + b\left(\frac{n}{N}\right)^{c} \tag{7}$$

Using the mean range values of the variables, the exponent providing the "bestfit" was calculated to be c = 1.35. The ratio, Q_{drs}/Q_{dro}, was then regressed against the parameter $(n/N)^{1.35}$ and this led to the expression for Q_{drs} equal to

$$Q_{drs}' = Q_{dro}\left[0.024+0.974\left(\frac{n}{N}\right)^{1.35}\right] \tag{8}$$

Eq. (8) suggests that completely-cloudy skies (n/N=0) transmit only 2% of the direct beam radiation; Fig. 1 shows values between 0 and 5%. The mean difference

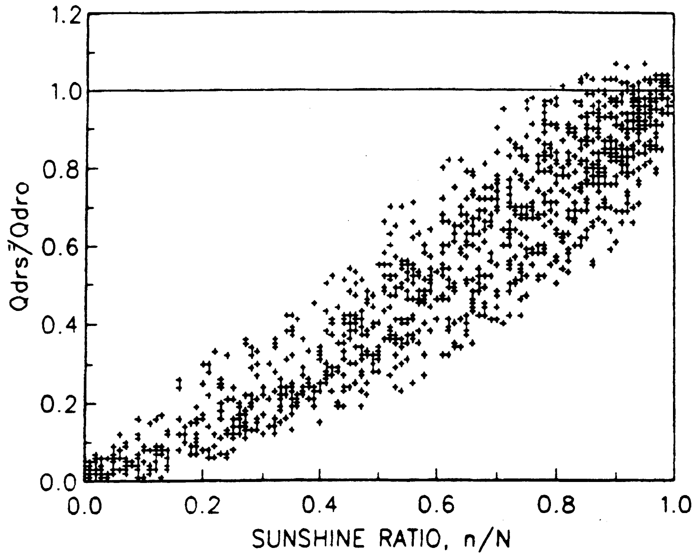

Fig. 1. Ratio of daily direct beam short-wave radiation to the clear-sky, direct beam short-wave flux Q_{drs}/Q_{dro}, plotted against sunshine ratio, n/N for Bad Lake.

between calculated (Eq. (8)) and observed Q_{drs}-values and the standard deviation of the difference were respectively, 0.12 MJ/m^2-d and 1.84 MJ/m^2-d for Bad Lake; −0.08 MJ/m^2-d and 1.76 MJ/m^2-d for Edmonton and 0.32 MJ/m^2-d and 1.64 MJ/m^2-d for Winnipeg.

Diffuse

Clear-sky – Even under clear skies there are usually enough dust particles, water droplets and ice crystals in the atmosphere to cause some diffusion of the incoming direct-beam radiation. Those atmospheric constituents, which tend to reduce the transmissivity for beam radiation, act to increase the scattering and diffusion processes. Generally, the lower the atmospheric transmissivity, the greater the diffuse component; also, the greater the clear-sky direct flux the greater will be the flux of diffuse radiation.

The daily diffuse radiation exhibits a seasonal variation. At Bad Lake the midwinter flux varies between approximately 1 and 3 MJ/m^2-d, while its midsummer range is approximately 4 to 15 MJ/m^2-d. The diffuse flux on clear days ($n/N \geqslant 0.98$) describes the lower envelope curve for the data points. It was postulated that the clear-sky diffuse flux, Q_{dfo}, could be related to those parameters describing the clear-sky, direct beam input and this resulted in the expression

$$Q_{dro} = 3.5\left(\frac{P}{P_0}\right)\cos(X\Lambda S)/p + 0.45\sin[(172-\mathrm{DAY})2\pi/365] \quad (9)$$

in which P/P_0 is the atmospheric pressure ratio, the ratio of the pressure at the surface (P) to the standard pressure at sea level (P_0). Eq. (9) is solved using $\cos(X\Lambda S) = \cos\delta\cos\phi$ (solar noon) in which the declination δ is given by the expression

$$\delta = 0.4093\sin[(DAY-81)2\pi/365] \tag{10}$$

where ϕ is the station latitude in radians (degrees $\times$ 0.0174533).

Comparisons between 91 Q_{dfo}-values by Eq. (9) with corresponding observed values gave a mean difference of -0.15 MJ/m^2-d and standard deviation of the difference equal to 0.31 MJ/m^2-d.

Cloud Cover – In the presence of cloud cover the diffuse flux will always be greater than its clear-sky value. The existence of cloud cover influences diffuse radiation in a compensating manner, however. Increasing cloudiness increases both scattering and diffusion processes, but clouds also block the direct radiation thereby reducing the amount available for scattering and diffusion. The result of these opposing effects is that a maximum for the diffuse radiation will likely occur at some value of the cloud cover index which is less than unity. A comparison of the ratio, Q_{dfs}/Q_{dfo} (Q_{dfo} is calculated by Eq. (9)), with the sunshine ratio, n/N, confirmed this trend. Assuming the effect of seasonal variation to be negligible, the relationship between Q_{dfs}/Q_{dfo} and n/N is described by the quadratic

$$\frac{Q_{dfs}}{Q_{dfo}} = 2.68 + 2.2\left(\frac{n}{N}\right) - 3.85\left(\frac{n}{N}\right)^2 \tag{11}$$

Comparisons between the calculated (Eq. (11)) and observed values of Q_{dfs} showed a mean and standard deviation of the difference of -0.15 MJ/m^2-d, and 1.30 MJ/m^2-d for Bad Lake; -0.08 MJ/m^2-d and 1.24 MJ/m^2-d for Edmonton; and 0.2 MJ/m^2-d and 1.32 MJ/m^2-d for Winnipeg, respectively.

Albedo

For practical applications, hydrologists are most interested in the integrated reflectance by a surface for light of all wavelengths falling within the short-wave radiation band. Reflected radiation is calculated using the surface albedo (A) defined as

$$A = \frac{\text{Reflected short-wave}}{\text{Incident short-wave}} = \frac{Q_r}{Q_s} = \frac{\int_{\lambda_1}^{\lambda_2} r(\lambda)I(\lambda)d(\lambda)}{\int_{\lambda_1}^{\lambda_2} I(\lambda)d(\lambda)} \tag{12}$$

where

$r(\lambda)$ – reflectivity at wavelength λ,
$I(\lambda)$ = intensity of monochromatic radiation of wavelength λ, and
λ_1, λ_2, = limits of integration which are determined primarily by the characteristics of the sensor of the measuring instrument.

The reflective properties of snow vary widely depending on wetness, impurity content, particle size, density and composition, surface roughness and the spectral composition and direction of the illuminating beam (see Mellor 1966; Kondratyev 1969; Manz 1974). Although albedo depends primarily on the reflection, scattering and absorption processes occurring within a shallow surface layer of snow, some of the radiation incident to the surface penetrates to deeper depths. Giddings and LaChappelle (1961) found the reflectivity of snow for radiation with a wavelength of 0.6 μm becomes relatively independent of depth when the snow depth exceeds 10-20 mm. O'Neill and Gray (1973) suggested that the underlying ground becomes important when the snow depth is less than 60-80 mm and Kung *et al.* (1964) found a marked change in the ratio of surface albedo to that of bare ground for snowcovers as deep as 130 mm. Because of radiation impinging on the ground surface the albedo-decay of a shallow Prairie snowcover during ablation is more rapid and the shape of the curve differs appreciably from that of a deep, mountain snow pack.

Modelling Albedo

Gray and Landine (1987) examined the albedo-depletion of seasonal prairie snowcovers between Feb. 1 and the end of ablation over several years of record and found that the decrease could be divided into three periods (Fig. 2):

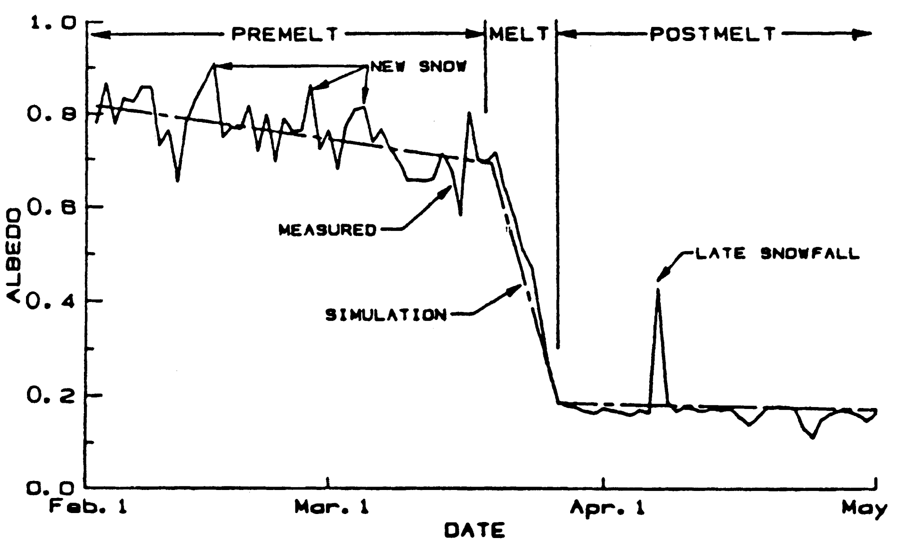

Fig. 2. Schematic of the decrease in albedo of a Prairie snow-cover during premelt, melt and postmelt periods.

Premelt – The period extending from Feb. 1 up to the start of active melt. During this period, except for increases produced by snowfall and decreases caused by periodic melt events, albedo decreases at a relatively-low, but constant rate, due to metamorphic processes. Rates of depletion ranged from 0.004 to 0.009/day with an average of 0.0061/day.

Melt – During melt the decay of albedo is accelerated by changes in the optical properties of the snowcover and the reflection of short-wave radiation penetrating to the ground surface. The general shape of the curve during continuous ablation is "Ogee" in which a period of rapid melt is preceded and followed by 1 to 2 days when the rate of change in albedo is smaller. Examination of albedo-depletion curves for twelve periods of continuous melt showed that the decrease could be approximated by the expression

$$A(t) \equiv A_i - 0.071t \tag{13}$$

in which A_i is the albedo of the snow surface at the start of "active" melt and t is the number of days after the start of "continuous" depletion. Normally, melt spans only 4 to 7 days.

Postmelt – Following the disappearance of the seasonal snowcover the albedo of the ground surface takes on a relatively-constant value of 0.17. The decrease in albedo of late-occurring snows is more rapid than the rate of decay of the seasonal snowcover and can be approximated by the expression

$$A(t) \equiv A_i - 0.196t \tag{14}$$

Net Long-Wave

Long-wave radiation emitted by a body (Q_l) is a function of its surface temperature, and is described by the familiar relationship

$$Q_l = \varepsilon\sigma T_s^4 \tag{15}$$

where ϵ is the emissivity, σ is the Stefan-Boltzmann constant (4.899×10^{-9} MJ/m^2-d), and T_s is the absolute temperature of the surface. The net long-wave flux, Q_{ln}, is composed of downward radiation emitted by the atmosphere, $Q_{l\downarrow}$, and the upward flux emitted by the surface, $Q_{l\uparrow}$. It is important to include the exchange in snowmelt calculations in open environments, especially where there is strong diurnal cycling of temperature, because of its influence on the internal energy content of the shallow snowcover. Before significant amounts of water are released, the negative energy deficit must be satisfied and the snowcover brought to isothermal conditions at 0° C.

Clear Sky

Atmospheric Flux – The atmospheric flux of long-wave radiation is emitted by ozone, carbon dioxide, dirt, water vapor and other contaminants from all atmospheric levels; under clear skies the major portion originates from the lowest 100 m layer. Since the largest flux is presumed to originate from the water vapor, many investigators have succeeded in correlating the atmospheric clear-sky, long-wave flux ($Q_{l\downarrow o}$) with air temperature and vapor pressure. The most widely quoted of these relationships is the Brunt (1932) formula

$$Q_{l\downarrow 0} = \sigma T_a^4 (a + b\sqrt{e_a}) \tag{16}$$

where T_a is the absolute air temperature (K), e_a is the actual vapor pressure of the air (mb), and a and b are regression coefficients whose magnitudes change with location and climate. Kondratyev (1969, p. 572) lists values for the coefficients for various locations in the world which show "a" ranging from 0.30 to 0.66 and "b" ranging from 0.039 to 0.127.

Direct comparisons of Brunt's equation with measured data over snow are scarce. Gray and Landine (1987) found $a = 0.58$ and $b = 0.09$ for clear days over non-melting snow surfaces in an open prairie environment. Male and Granger (1979) compared measured values of $Q_{l\downarrow o}$ with those calculated with the mean daily air temperature and vapor pressure over a melting snowcover and concluded that Brunt's formula is a poor estimator of this flux. They attributed the poor association to fog, haze and ice crystals which are commonly present at night above a snow cover which has experienced melting during the day.

A modified form of the Brunt equation which may be used to estimate $Q_{l\downarrow o}$ over snow is the empirical expression presented by Satterlund (1979)

$$Q_{l\downarrow 0} = 1.08\sigma T_a^4 [1 - \exp(-e_a^{(T_a/2016)})] \tag{17}$$

in which T_a, the air temperature at screen height, and e_a, the vapor pressure at screen height are in K and mb, respectively. Satterlund shows superior estimates of $Q_{l\downarrow o}$ by Eq. (17), than those obtained by other relationships at temperatures below freezing.

Surface – Since the emissivity of a smooth snow surface varies within a relatively narrow range (dirty snow – 0.97; fresh snow – 0.99), $Q_{l\uparrow o}$ is often calculated directly by Eq. (15). The accuracy of this estimate depends strongly on the measure of surface temperature, and therefore it is usually more reliable under melting conditions than under non-melt situations.

Net exchange – The net long-wave exchange under clear skies can be obtained by summing the downward- and upward-directed fluxes calculated by the procedures outlined above. That is

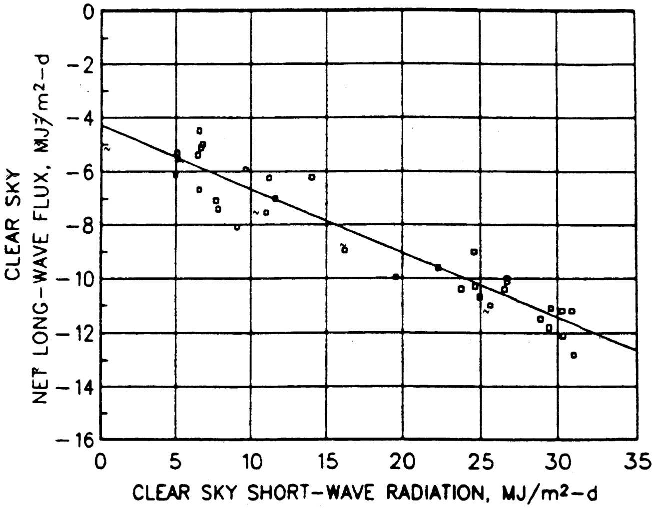

Fig. 3. Relationship between daily net long-wave and net short-wave fluxes on clear days at Bad Lake.

$$Q_{lno} \equiv Q_{l\downarrow o} - Q_{l\uparrow o} \quad (18)$$

The discussions above suggest that the traditional approach of estimating the clear sky net long-wave flux, $Q_{l\updownarrow o}$, from the air temperature and vapor pressure may not provide reliable, accurate estimates of the term. In nature, the atmospheric conditions on which the empirical relationships are based are not often the same because even under clear-sky conditions the temperature and humidity lapse rates may vary widely.

For a natural surface, the fluctuations in surface and atmospheric temperatures (and to a large extent the variations in atmospheric humidity) are driven by the energy supplied to the surface, thus an alternative approach for estimating the net long-wave flux might attempt to develop a correlation with the incident short-wave flux. Fig. 3 shows values of the net long-wave radiation, Q_{lno}, plotted against the corresponding values of the incident short-wave flux, $Q_{so} = Q_{dro} + Q_{dfo}$, as calculated with Eqs. (5) and (9), for clear days at Bad Lake. The best-fit line describing the data is given by

$$Q_{lno} = -4.25 - 0.240 Q_{so} \quad (19)$$

The regression gives a correlation coefficient of 0.904, with a standard error of the estimate for Q_{lno} of 0.79 MJ/m^2-d.

Cloud Cover

In general, cloud cover increases the flux of atmospheric radiation and causes a decrease in the upward flux by reducing the incident short-wave flux, decreasing the surface temperature and increasing the atmospheric absorption of thermal energy radiated by the earth's surface. In 1948, Penman suggested that the net long-wave flux, Q_{ln} could be approximated by a linear relationship with the net clear-sky long-wave flux, Q_{lno} and the degree of cloudiness. For the observations at Bad Lake the best-fit line is defined by the equation

$$\frac{Q_{ln}}{Q_{ln0}} = 0.25 + 0.75\left(\frac{n}{N}\right) \tag{20}$$

A comparison of daily net long-wave radiation calculated by Eq. (20) and the observed values for the Bad Lake data gave a mean difference between the calculated and observed values of 0.16 MJ/m^2-d, with a standard deviation of the differences of 1.25 MJ/m^2-d. Because there was no evidence of a seasonal effect on the magnitude of the variance, it is presumed that the relatively large value is due mainly to differences in cloud type and height and to the fact that the sunshine ratio indexes cloud cover only during the daylight hours. The absence of night-time observations of cloud cover constitutes the largest, single drawback to this approach. Another limitation of the method is that it appears to impose upper and lower limits to the range of calculated values of Q_{ln} between -1.5 and 12 MJ/m^2-d.

Models

Comparison with Observed Data

Fluxes of daily net "all-wave" radiation, using Eqs. (8) and (11) to estimate the short-wave incident flux and Eq. (20) to estimate the net long-wave flux are compared to their corresponding observed values for a grass surface during the snow-free season in Fig. 4 and for periods when the surface is snow-covered (excluding the melt period) in Fig. 5. Fig. 4 shows reasonably good agreement between calculated and observed values (mean difference of 0.40 MJ/m^2-d and a standard deviation of the difference equal to 1.30 MJ/m^2-d). When the ground is snowcovered the calculated values appear to be limited to the range from -4 to 0 MJ/m^2-d while the observed values ranged from -5 to 2 MJ/m^2-d. The inferior agreement between modelled and observed net radiation over snow is attributed to the poor association between net "all-wave" and net "short-wave" fluxes. Gray and Landine (1987) found Q_n and Q_{sn} correlated only with $r = 0.08$ during premelt. During the period, net radiation is often negative (see Fig. 5) due to low fluxes of solar radiation, high albedo, low ambient temperatures, low number of sunshine hours and clear skies. Thus, for periods of snowcover it is strongly suggested that either the Brunt equation (Eq. (16)) or the Satterlund equation (Eq. (17)) be used to calculate Q_{ln}. Gray and Landine (1987) provide the following forms of these expressions for the prairies of western Canada

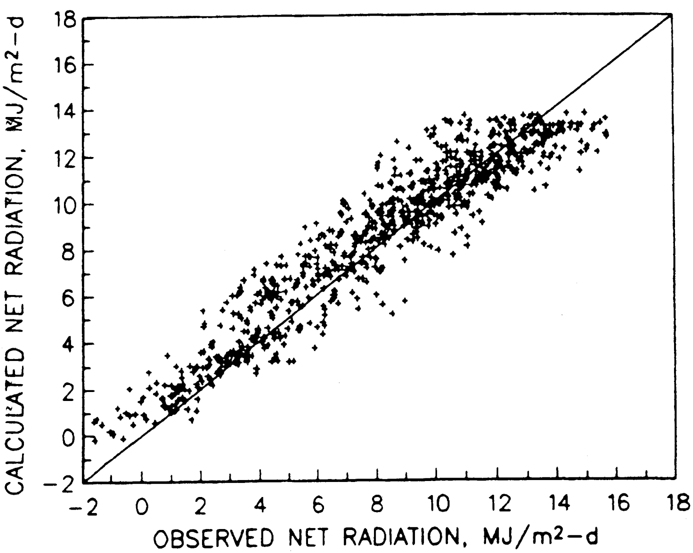

Fig. 4. Comparison of the calculated and observed daily net all-wave radiation for snow-free periods at Bad Lake.

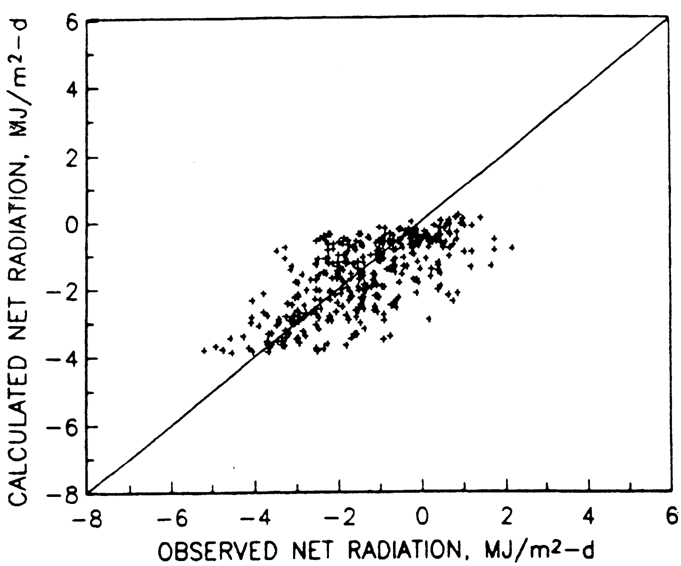

Fig. 5. Comparison of the calculated and observed daily net all-wave radiation over snow, excluding the melt periods for Bad Lake (albedo was assumed equal to 0.7).

Brunt:

$$Q_{ln} \equiv -0.85 + 0.97\{\sigma T_a^4(-0.39+0.093\sqrt{e_a})[0.26+0.81(\frac{n}{N})]\} \quad (21)$$

Satterlund:

$$Q_{ln} \equiv -0.13 + 0.97\{\sigma T_a^4[1.08(1-\exp(-e_a^{(T_a/2016)}))-0.97] \times [0.25+0.83(\frac{n}{N})]\} \quad (22)$$

in which Q_{ln} is in MJ/m^2-d with T_a in K and e_a in mb. Eq. (21) has a correlation coefficient of 0.82 and a standard error of estimate of 1.09 MJ/m^2-d; the corresponding statistics for Eq. (22) are 0.83 and 1.07 MJ/m^2-d respectively. Because Q_{ln} is usually small during periods when the ground is covered with snow and the error in estimating the term is not systematic, the expressions have been applied successfully in modelling changes in the internal energy content of a snowcover (Gray and Landine 1988). Fig. 6 plots daily values of Q_n estimated by Eqs. (9) (12) and (21) (the Brunt expression) against the flux observed over snow on 319 days during the winter and spring months, December through March. The association between the calculated and observed values is strong; the mean difference is 0.36 MJ/m^2-d and the standard deviation of the difference is 1.22 MJ/m^2-d.

Correlation of Net All-Wave and Net Short-Wave Radiation

Several authors (Davies 1965, 1967; Davies and Idso 1979; Gray and Landine 1987) have demonstrated that over natural surfaces plots of net "all-wave" radiation (Q_n) and net short-wave radiation (Q_{sn}) are remarkably linear with high correlation coefficients during periods when the incident short-wave radiation is the dominant flux. Fig. 7 shows the daily net all-wave radiation plotted against the corresponding net short-wave flux for snow-free periods at Bad Lake. The best-fit linear regression is described by the expression

$$Q_n \equiv -2.24 + 0.651Q_{sn} \quad (23)$$

The comparison involved 795 data pairs and produced a correlation coefficient of 0.87 and a standard error of the estimate of 1.47 MJ/m^2-d. The mean difference and standard deviation of the differences between daily net radiation calculated by Eq. (23) and their corresponding observed values were calculated to be -0.05 MJ/m^2-d and 1.46 MJ/m^2-d. These values differ only slightly from the corresponding statistics of 0.40 MJ/m^2-d and 1.30 MJ/m^2-d obtained when the Q_n values by Eq. (1) were compared with the observed data.

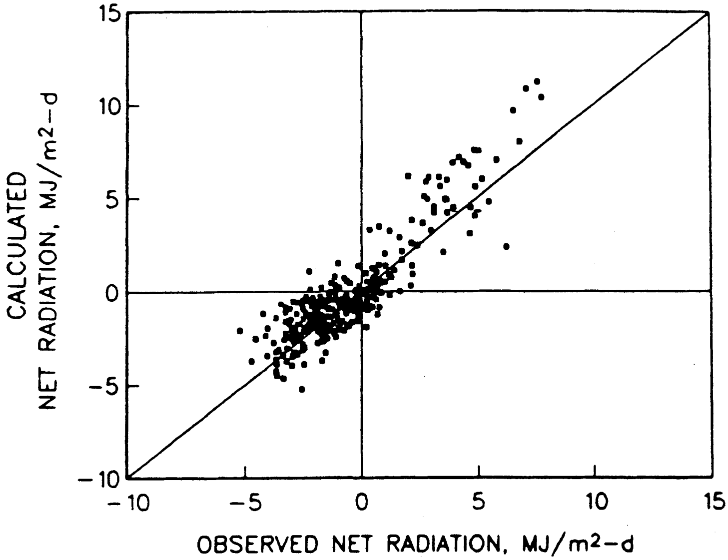

Fig. 6. Calculated daily net radiation using the Brunt espression (Eq. (21)) for the net long-wave exchange plotted against observed daily net radiation over snowcover during winter and spring.

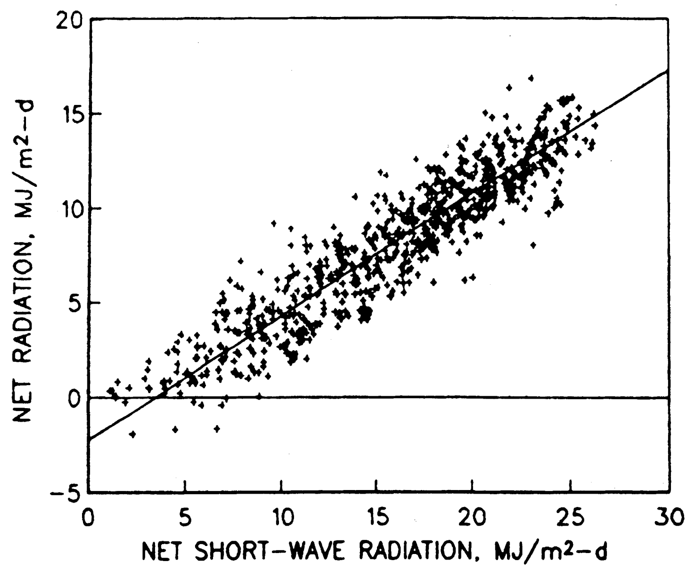

Fig. 7. Daily net radiation versus daily net short-wave radiation for Bad Lake.

Summary

Climatological observations from three stations in western Canada are used to develop models for calculating daily net all-wave radiation at the earth's surface in open environments. The procedure divides the net all-wave radiation into its long- and short-wave parts in the customary fashion, however, the incoming short-wave radiation is presented as the sum of its direct and diffuse components. For all components a clear-sky value is first calculated and this value is then adjusted in the presence of cloud cover.

For the short-wave component, the clear-sky direct-beam radiation, calculated by the method of Garnier and Ohmura (1970) (Eq. (5)), is moderated in the presence of cloud to obtain the direct-beam, short-wave component (Eq. (8)). The diffuse component is expressed as a function of the clear-sky, direct-beam flux and the degree of cloudiness (Eqs. (9) and (11)).

Expressions reported by Gray and Landine (1987) represent a convenient approach to modelling snowcover albedo. The seasonal decrease in albedo is approximated by three line segments describing the periods of premelt, melt and postmelt. Albedo decreases at an average rate of 0.0061/day during premelt, at an average rate of about 0.071/day during melt and takes on a reasonably constant value of 0.17 when the ground is snowfree.

Two methods are presented for calculating the net long-wave exchange. The first represents a new approach for estimating the net long-wave exchange from the incident clear-sky short-wave flux and the sunshine ratio (Eqs. (19) and (20). The second is an adaptation of the Brunt formula (Eq. (21)).

Comparisons between calculated and observed daily net all-wave radiation fluxes show that the model, using the first approach for calculating the net long-wave exchange, performs well over snowfree surfaces, with no apparent seasonal bias; the mean difference between calculated and observed values is 0.40 MJ/m^2-d with a standard deviation of 1.30 MJ/m^2-d. This model is not recommended for computing the exchange over snow when the net short-wave flux is small relative to the long-wave exchange and the net and short-wave fluxes are poorly correlated. Under these conditions improved estimates of net radiation are obtained using estimates of net long-wave radiation from formulations such as that of Brunt (1932) which incorporate temperature and humidity effects.

Over snow-free surfaces, when incident short-wave radiation is the dominant flux, net all-wave radiation is highly correlated with net short-wave radiation. An empirical expression describing this association is presented. Comparisons between daily net radiation, calculated by the regression, and observed values gave a mean difference of -0.05 MJ/m^2-d and a standard deviation of the difference of 1.46 MJ/m^2-d. These statistics do not differ appreciably from the corresponding values found when daily net radiation, calculated as the sum of the direct and diffuse components, is compared with observed values.

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