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SNOWCOVER ABLATION AND RUNOFF

D. H. MALE

Division of Hydrology, University of Saskatchewan, Saskatoon, Saskatchewan

D.M. GRAY

Division of Hydrology, University of Saskatchewan, Saskatoon, Saskatchewan

INTRODUCTION

In many countries snow constitutes a major water resource; its release in the form of melt water can significantly affect agriculture, hydro-electric energy production, urban water supply and flood control. The ablation of a snowcover or the net volumetric decrease in its snow water equivalent is governed by the processes of snowmelt, evaporation and condensation, the vertical and lateral transmission of water within the snowcover and the infiltration of water to the underlying ground. In turn, water yield and streamflow runoff originating from snow are governed by these same processes as well as the storage and the hydraulics of movement of water in channels. In recent years it has become apparent that a better understanding of the physics of the ablation process is central to improving techniques of forecasting the time of melt, the quantity and rate of water released, the volume of water entering the soil and the amount of evaporation.

Regular forecasting of runoff from snowmelt in North America was first attempted at Lake Tahoe, Nevada, in 1909. Engineers of the local power company correlated changes in the lake water levels during the spring with the water content of the snow on Mount Rose (as determined from snow surveys made by Dr. J. E. Church). This correlation allowed the company to regulate releases from the lake to prevent spring flooding and to use the melt water more efficiently for power production. From this early beginning, research into the snow ablation phenomenon has increased significantly. Most studies have been concerned with the prediction of floods and peak discharge rates for

designing hydraulic structures and flood control works, and melt water flow to optimize its use for hydro-electric power generation, recreation and irrigation. Recently, there has been a greater emphasis on studies of the impact of snow on the agricultural industry (crop production) and the environment. The interactions between human activities (for example, deforestation, urban development) and the snowmelt regime are receiving increasing emphasis.

The most comprehensive study of snowmelt published to date is *Snow Hydrology* prepared by the U. S. Army Corps of Engineers (1956). Although the investigations summarized in this report were conducted in mountainous regions in the United States their findings have served as a foundation for many subsequent studies undertaken in other parts of North America. In particular, the results and methods described in this publication form the basis for many of the snowmelt runoff models currently used to forecast streamflow. The major Soviet publication on snowmelt is that by Kuz'min (1961) and deals primarily with conditions on the Russian Steppes. This work contains a thorough discussion of the physical processes influencing snowmelt, and an excellent summary of empirical methods for estimating melt rates suitable for incorporation in streamflow forecasting procedures and for other water management purposes.

PHYSICS OF SNOWMELT

General Considerations

Seasonal snowcovers normally develop from a series of winter storms and are modified by the action of freezing rain, wind and diurnal melting and refreezing at the surface. As a result, both deep snowpacks in the mountains and the shallow covers in regions of low relief develop a characteristic layered structure (Gerdel, 1948; Langham, 1974) with "ice" layers or relatively-impermeable, fine-textured high-density layers alternating with coarse-textured, low density and highly permeable layers. Early in the melt sequence vertical drainage channels develop in the snow contributing further to its heterogeneity. The internal structure significantly influences the retention and movement of melt water through the snow, making a detailed analysis of the transmission process extremely difficult. However, during most of the melt period the total melt water produced is governed by the energy exchanges at the upper and lower snow surfaces.

When the pack is primed to produce melt it is at a temperature of 0°C throughout and its individual snow crystals are coated with a thin film of water; also, small pockets of water may be found in the angles between contacting grains, normally amounting to 3 to 5% of the snow by weight although some investigators have measured values as high as 25% (de Quervain, 1948). Any additional energy input produces melt water which subsequently drains to the ground. When melt rates are at their highest, 20% (by weight) of the pack or more may be liquid water, most of which is in transit through the snow under the influence of gravity.

The amount of energy available for melting snow is determined from the energy equation. This equation is applied to a volume of snow whose upper and lower surfaces are the snow-air and snow-ground interfaces respectively, and may be written as:

$$Q_m = Q_{sn} + Q_{ln} + Q_h + Q_e + Q_g + Q_p - dU/dt, \qquad 9.1$$

where

Q_m = energy flux available for melt,
Q_{sn} = net short-wave radiation flux absorbed by the snow,
Q_{ln} = net long-wave radiation flux at the snow-air interface,
Q_h = convective or sensible heat flux from the air at the snow-air interface,
Q_e = flux of the latent heat (evaporation, sublimation, condensation) at the snow-air interface,
Q_g = flux of heat from the snow-ground interface by conduction,
Q_p = flux of heat from rain, and
dU/dt = rate of change of internal (or stored) energy per unit area of snowcover.

The net long-wave radiation and convective heat transfer processes are operative *at* the snow-air interface whereas the short-wave radiation exchange is strongest at the surface although limited amounts penetrate into the pack. The ground heat flux, which is usually small, may produce small amounts of melt near the snow-ground interface. Water is released from this lower layer when the snow reaches 0°C and is holding the maximum liquid. Rain may penetrate considerable depths into the snow resulting in a mass transport of heat which is distributed more uniformly throughout the pack than the heat obtained from other sources. However, melt water is generated primarily at the snow-air interface.

The daily amount of melt produced by a given value of Q_m ($kJ/m^2 \cdot d$) (see Eq. 9.1) may be calculated by the expression

$$M = Q_m/(\rho h_f B), \qquad 9.2$$

in which

M = snowmelt water equivalent (cm/d),
h_f = latent heat of fusion, (kJ/kg),
ρ = density of water, (kg/m^3), and

B = thermal quality or the fraction of ice in a unit mass of wet snow.

For normal melt conditions $h_f = 333.5$ kJ/kg and $\rho = 1000\ kg/m^3$. Thus, Eq. 9.2 reduces to

$$M = Q_m/(3335\ B). \qquad 9.3$$

As mentioned previously, a melting snowpack generally will retain 3 to 5% water (by weight) against free drainage, corresponding to a thermal quality between 0.95 and 0.97.

Table 9.1 indicates the magnitude of the various energy fluxes during the melt period for clear days in Saskatchewan at a site with no vegetation protruding above the snow surface. The wide range of possible values for each of the fluxes is evident. The presence of a cloud cover would produce an even greater variation. Relative changes in the fluxes can be attributed to changing wind conditions, relative humidity, air temperature and time of year, factors considered in detail below. Table 9.1 also shows the dominant size of the net radiation Q_n during snowmelt in open areas.

Table 9.1
SELECTED DAILY ENERGY FLUX TRANSFER (kJ/m^2)[a] DURING THE MELT PERIOD IN THE ABSENCE OF VEGETATION (BAD LAKE, SASKATCHEWAN).

Date (Day/Mon/Yr)	Q_{sn}	Q_{ln}	Q_n[b]	Q_h	Q_e	Q_g
11-4/75	8090	-6320	1770	186	-855	- 45
12-4/75	9620	-8480	1140	782	26	- 22
14-4/75	12290	-9430	2860	13	-395	- 4
17-3/76	4630	-4500	130	1830	-555	64
27-3/76	7200	-7720	- 520	1517	-208	-237
28-3/76	7790	-7120	670	70	-201	-111
29-3/76	9070	-7660	1410	532	- 60	-180
30-3/76	9290	-6040	3250	827	140	-270

[a] positive values indicate an energy gain by the snow.
[b] the daily net radiation flux transfer: $Q_n = Q_{sn} + Q_{ln}$.

Short-wave (Solar) Radiation

Incident radiation

The radiation which influences snowmelt is electromagnetic radiation emitted from a medium by virtue of its temperature and falls in the wavelength range from about 0.2 to 100 μm. Within this range two types of radiation are distinguished. Short-wave radiation (emitted by the sun) is generally considered as that portion falling within the range from 0.2 to 2.2 μm; long-wave radiation (emitted by the atmosphere and the earth) lies between 6.8 and 100 μm.

Radiation from the sun is short-wave radiation falling within a very narrow wavelength band with maximum intensity at 0.47 μm. At the top of the earth's atmosphere extra-terrestrial solar radiation incident on a surface perpendicular to the sun's rays at a mean earth-sun distance of 149.5×10^6 km is equal to the solar constant, 1.365 kW/m^2. This value varies by about 7% during a year primarily because of the changing distance between the earth and the sun. The amount of solar radiation penetrating the earth's atmosphere to be received at the surface varies widely depending on latitude, season, time of day, topography (slope and orientation), vegetation, cloud cover and atmospheric turbidity. 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 absorbed energy increases the temperature of the air which in turn increases the amount of long-wave radiation emitted to the earth's surface and to outer space.

Short-wave radiation reaching the surface of the earth has two components: a direct beam component along the sun's rays and a diffuse component scattered by the atmosphere but with the greatest flux coming from the direction of the sun. The daily amount of direct beam radiation is a complex function of the factors mentioned above. Garnier and Ohmura (1970) have presented a useful method for analyzing the interaction between these factors. They express the direct clear day radiation I_d falling upon a slope as

$$I_d = (I_o/r^2) \int p^m \cos(X \Lambda S) dH, \qquad 9.4$$

where

I_o = solar constant,
r = radius vector of the earth's orbit, (the distance from the centre 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; this is a measure of the fraction of solar radiation which reaches the earth's surface without being scattered or absorbed,

m = optical air mass which is 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 the slope, (X is a unit normal vector pointing away from the surface and S is a unit vector expressing the sun's position), and
H = hour angle measured from solar noon, the integral being taken over the duration of sunlight on the slope.

Values for I_o, r and m are given by List (1968).

Figure 9.1, which is calculated using Eq. 9.4, shows the annual variation in daily values of solar radiation received by a horizontal surface at several latitudes; p is assumed to be 1.0, implying that all the energy reaches the surface. The influence of transmissivity is illustrated in Fig. 9.2. By multiplying the data in Fig. 9.1 by the ratios of Fig. 9.2, the daily radiation values received by a horizontal surface can be calculated for any atmospheric transmissivity. For example, at 50° N latitude, if p is 0.7, the radiation would be $18 \times 0.405 = 7.29$ MJ/m^2 on March 1, 13.9 MJ/m^2 on April 1, about twice the value of a month earlier. The time of year obviously is an important factor governing the solar radiation flux incident on the earth's surface. Since this flux is a major component in the total energy flux of the snow, the time of year also has an important influence on the melt rate — a fact well known to hydrologists involved in flood forecasting. As a rule, the longer the spring melt is delayed the greater the danger of flooding. This is due partly to increases in the radiative flux and partly to the increased probability of rain.

The transmissivity p is the highest in winter and lowest in summer because the atmosphere contains more water vapor during summer. It also varies somewhat with latitude, increasing northwards. Kondratyev (1969) reports a mean annual value for p at Pavlovsk, USSR, of 0.745 with a deviation of ± 0.05 during the years 1906-1936. Williams et al. (1972) suggest average monthly values ranging from 0.92 to 0.50 for four locations in middle latitudes (Pavlovsk, USSR; Benson, U.K.; Paris, France; Tacubaya, Mexico). Kuz'min (1961) recommends the use of 0.80 ± 0.05 for the snowmelt period within the European USSR.

It is a common observation that snow on a south-facing slope melts faster than snow on a north-facing slope, the reason being that the orientation of the slope affects the amount of direct beam solar radiation the area receives. The effect of slope and orientation is illustrated in Fig. 9.3. The results are symmetric about a north-south line; as might be expected the influence of orientation diminishes towards the summer solstice. Even on a 10° slope the effect of orientation can be significant; e.g., at 50° N on April 1, a south-facing

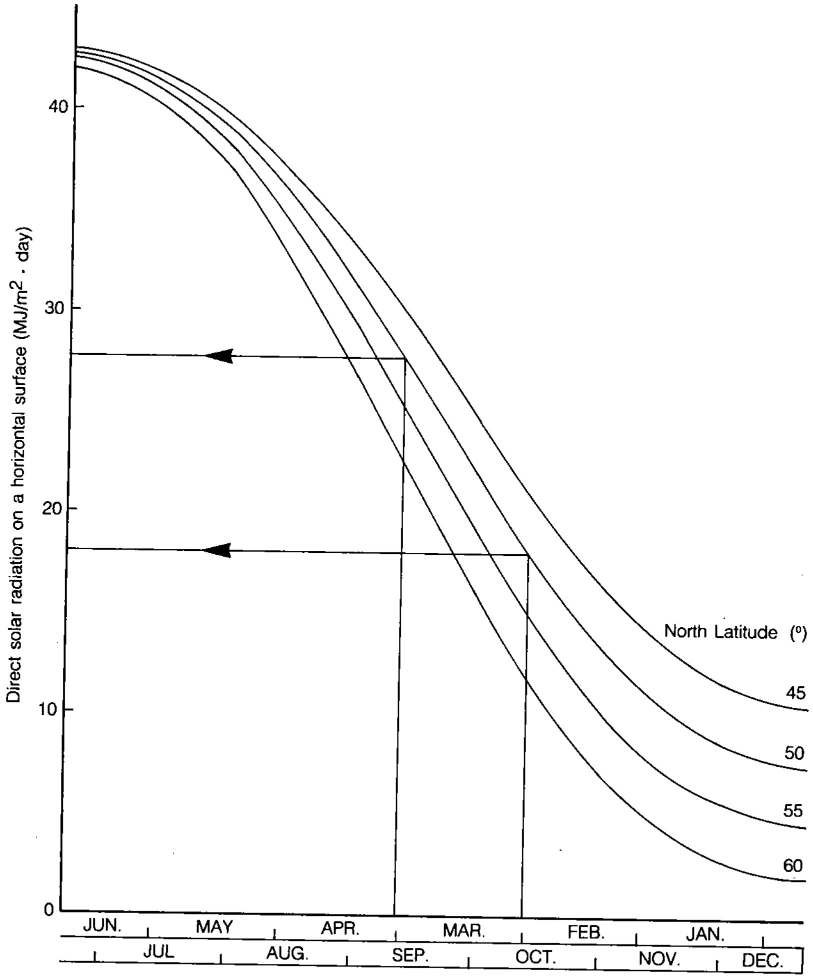

Fig. 9.1 Daily values of direct solar radiation received on a horizontal surface — transmissivity p = 1.

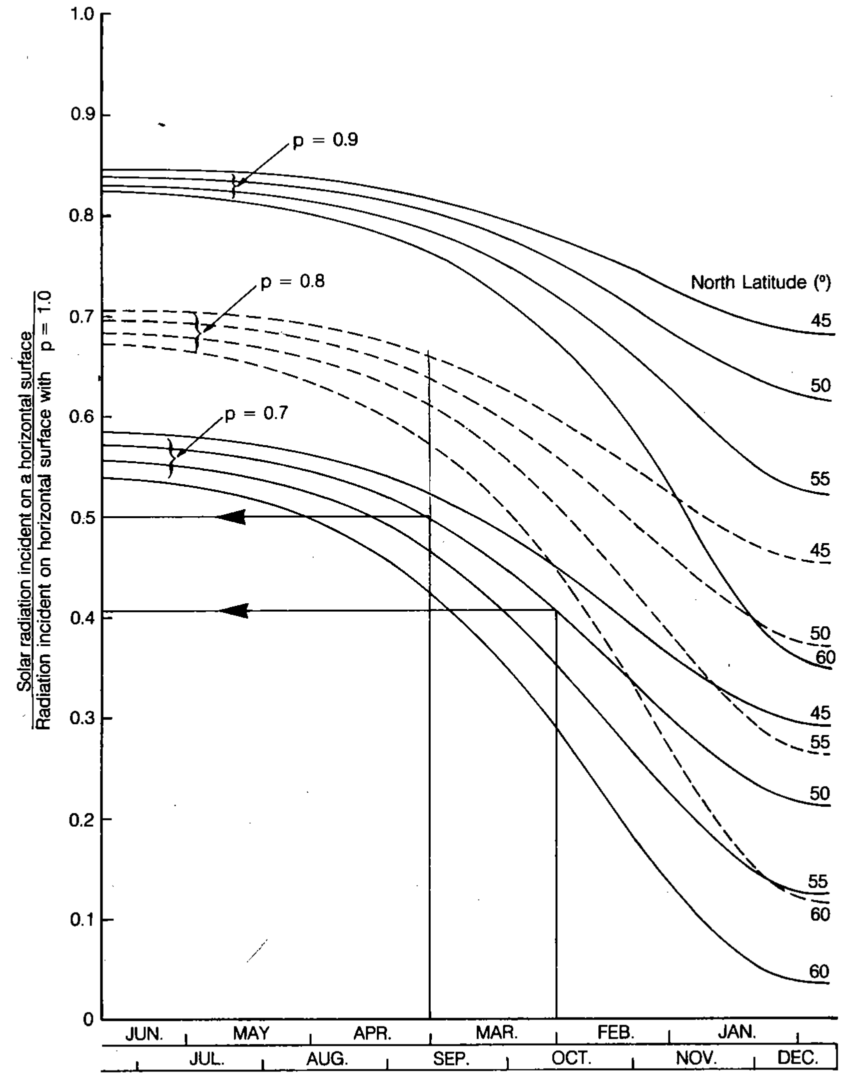

Fig. 9.2 Influence of transmissivity (p) on daily solar radiation received on a horizontal surface.

slope receives approximately 40 percent more direct beam radiation than a north-facing slope (see Fig. 9.3a). The ratios in Fig. 9.3 are calculated assuming p = 1.0; and decrease slowly as p decreases, becoming about 5 percent less for p = 0.7.

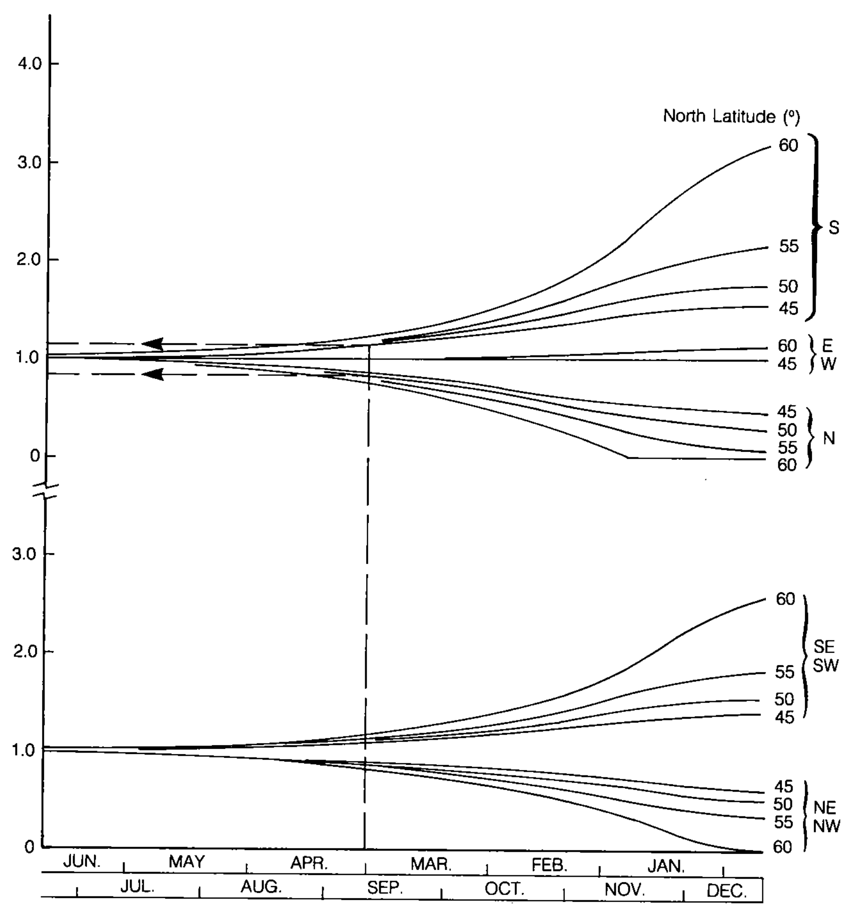

Fig. 9.3a Ratio of direct daily solar radiation received by a 10° slope to that received by a horizontal surface.

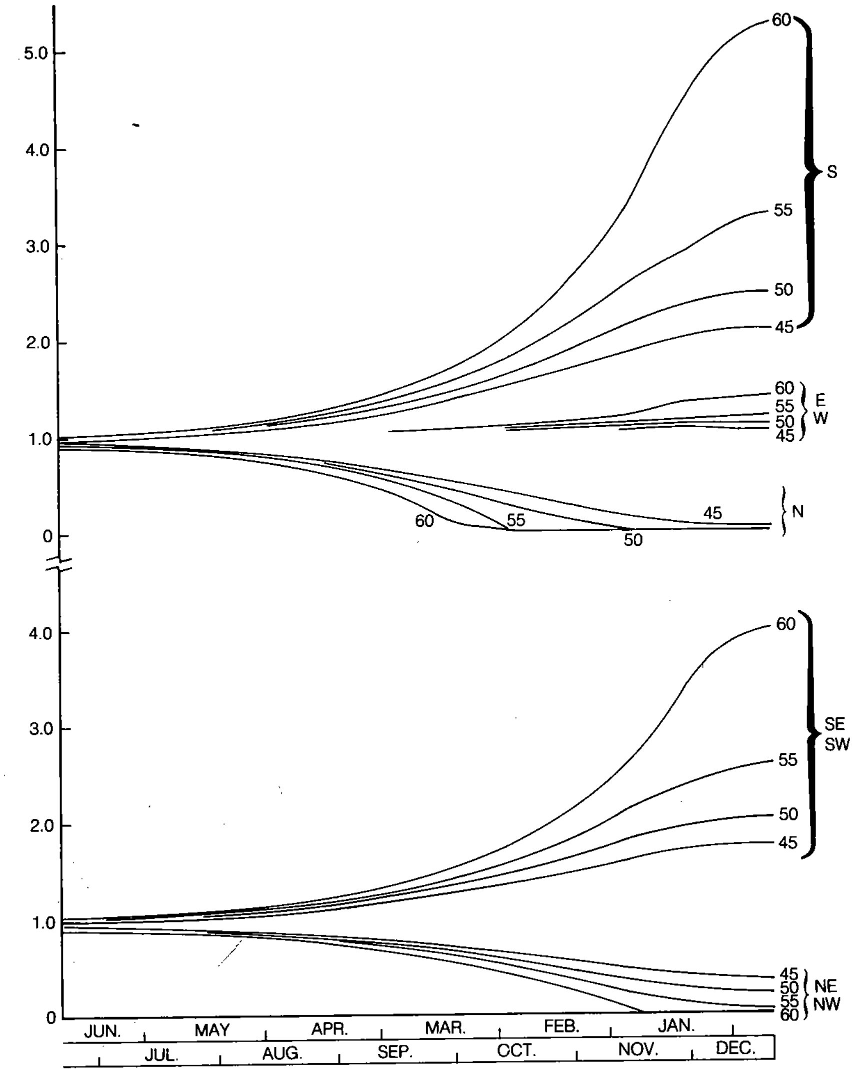

Fig. 9.3b Ratio of direct daily solar radiation received by a 20° slope to that received by a horizontal surface.

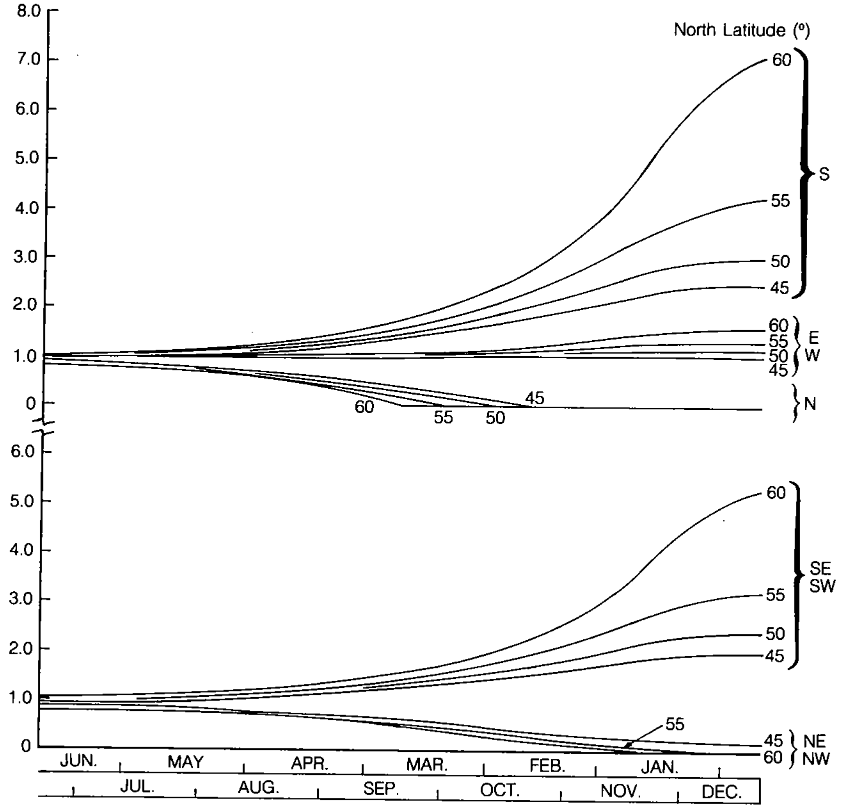

Fig. 9.3c Ratio of direct daily solar radiation received by a 30° slope to that received by a horizontal surface.

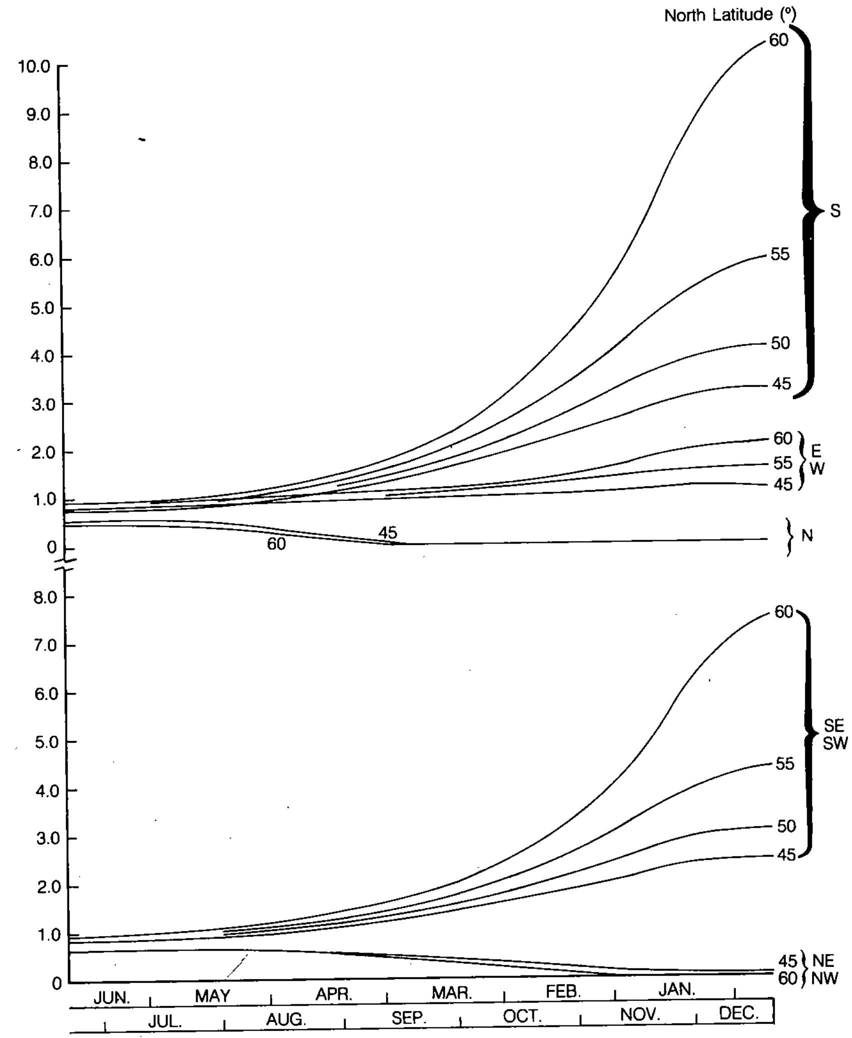

Fig. 9.3d Ratio of direct daily solar radiation received by a 50° slope to that received by a horizontal surface.

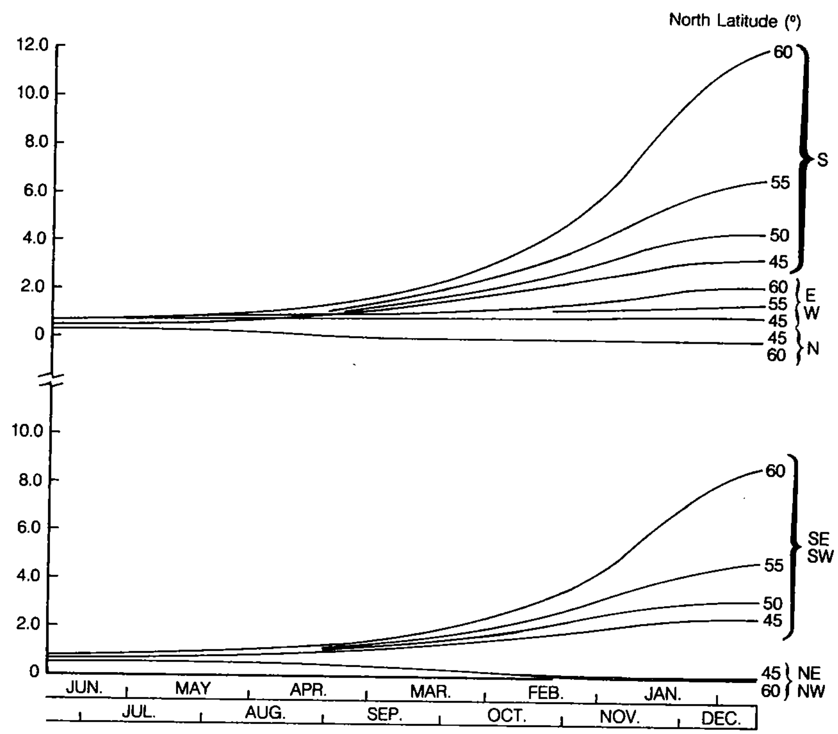

Fig. 9.3e Ratio of direct daily solar radiation received by a 70° slope to that received by a horizontal surface.

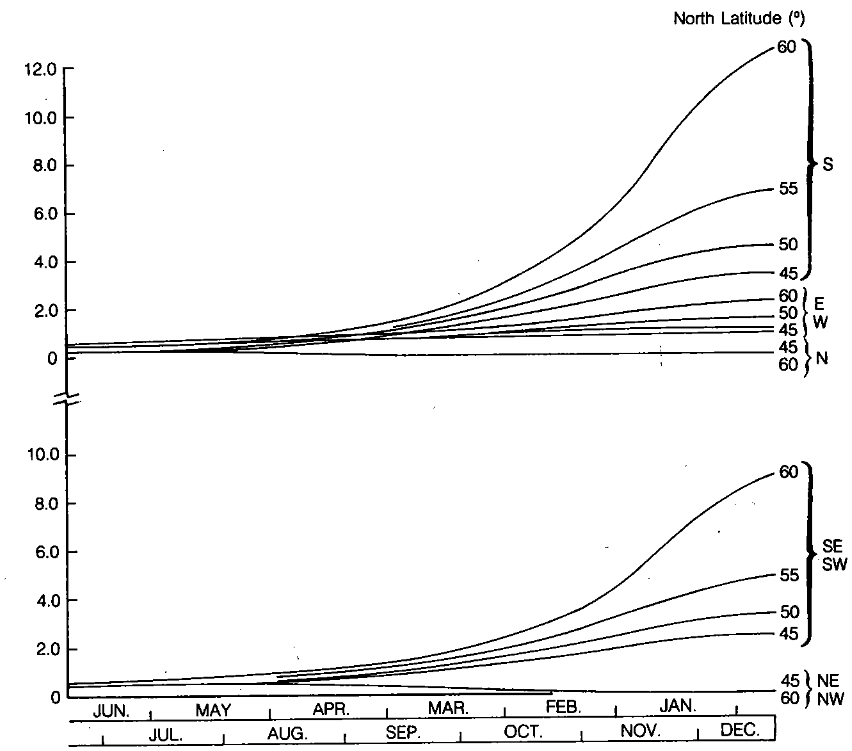

Fig. 9.3f Ratio of direct daily solar radiation received by a 90° slope to that received by a horizontal surface.

In recent years models of solar radiation transfer have been developed which more closely approximate the physical processes occurring in the atmosphere (Suckling and Hay, 1976; Dozier, 1979). In these models separate transmission functions are defined for the attenuation of radiation resulting from absorption by ozone, water vapor, oxygen, carbon dioxide, methane and nitrous oxide, and for scattering by atmospheric molecules (Rayleigh scattering) and aerosols. Each function depends on the optical path length and the attenuation coefficient for each atmospheric constituent.

In mountainous terrain shadows play an important role in determining the amount of direct solar radiation reaching a given point. Methods for considerating this effect have been developed by Williams et al. (1972) and Dozier (1979). The region being analyzed is divided into a uniform grid. By considering the elevation at each grid point and searching along the line of the sun's azimuth for points high enough to block the sun, the grid points in shadow at any hour angle may be found.

Forest canopy further complicates computation of the surface net radiation flux. The forest canopy is a heterogeneous, anisotropic medium that absorbs, scatters and reflects the direct beam solar radiation and emits long-wave radiation. The amount of direct radiation received by the snow surface has been reduced by the shading of the trunks and canopy. To calculate the reduced amount the incoming direct radiation I_d is usually multiplied by the factor (1-V); where V is a beam-shading function depending on the solar angle. V effectively averages the surface in the shade with those in the sun. Dozier (1979) describes a complete solar radiation model which includes a shading function.

The diffuse component of the short-wave radiation reaching the earth's surface is less amenable to mathematical computation and is often ignored in energy balance estimates; the direct beam component is then taken as an index of the total short-wave flux. However, the diffuse component may be the significant energy flux: on clear, bright days it may be 10% of the short-wave radiation received while on partly cloudy days it may be more than 50% and on overcast days it is 100%.

Under clear sky conditions the amount of diffuse radiation depends on the altitude of the sun above the horizon, the atmospheric transmittance and the amount of incident radiation reflected by the underlying surface. Kuz'min (1961) reports that the increase in scattered radiation over snow relative to that over bare ground ranges from 65% (solar elevation $\sim 0°$) to 12% (solar elevations: 36 to 55°). Various methods have been suggested for calculating the diffuse flux under cloudless sky conditions (Kuz'min, 1961; List, 1968; Kondratyev, 1969) but to obtain precise energy balance data over snow, direct measurements are normally made.

The simplest expression for calculating D_o, the diffuse radiation flux on a

horizontal surface under cloudless conditions is given by List (1968, p. 420 attributed to Fritz):

$$D_o = 0.5\,[(1 - a_w - a_o)I_t - I_d], \qquad 9.5$$

where a_w = radiation absorbed by water vapor (assumed to be 7%),
a_o = radiation absorbed by ozone (assumed to be 2%),
I_t = extra-terrestrial radiation on a horizontal surface,

$$I_t = (I_o/r^2) \int \cos z_s \, dH,$$

where r is the radius vector of the earth (see Eq. 9.4) and z_s is the sun's zenith distance (values of I_t are tabulated by List (1968, p. 418)) and,
I_d = direct radiation reaching a horizontal surface of the earth (see Eq. 9.4).

The factor 0.5 applied in Eq. 9.5 expresses the assumption that half of the direct solar beam is scattered toward the surface and half is scattered away from it. Other empirical methods of calculating D_o are described by Liu and Jordan (1960) and Stanhill (1966). To use these procedures some estimate or measurement of the total (or global) short-wave flux (direct and diffuse components) on a horizontal surface is required. Models have also been developed to calculate the diffuse component by taking the scattering process into account. Dozier (1979) presents a computing scheme in which the amount of radiation initially scattered out of the beam is determined by wavelength-dependent coefficients for aerosol and Rayleigh scattering and for ozone and water vapor absorption. This type of model is complex requiring estimates of the ozone and water vapor distributions with altitude and the change in air pressure with altitude.

An important source of diffuse radiation to the snow surface is the back-scattered short-wave radiation, i.e., that portion of the incoming radiation reflected at the snow surface which is again scattered and reflected downward by the atmosphere. Backscattering models have been developed by Hay (1976) and Dozier (1979). The critical parameter in model calculations is the reflectance or albedo of the snow surface, considered below.

For a sloping surface the diffuse component can be calculated from the expression of Kondratyev (1969), assuming that the diffuse radiation field is isotropic; that is, it has the same properties in all directions:

$$D = D_o \cos^2 (\theta/2), \qquad 9.6$$

where θ is the angle of inclination of the slope. In mountainous regions any slope is likely to have a sky dome that is restricted by the surrounding topography. The fraction by which the downward scattered radiation is reduced can be calculated by assuming that the radiation intensity is constant from all parts of the sky dome, i.e., isotropic. For a specific location the