Appendix D
Meteorological Relationships

D-1. Water Vapor in Air

a. Vapor pressure and saturation. Water vapor present in the atmosphere is measured in terms of the partial pressure exerted by the gas, known as the vapor pressure. As the amount of vapor increases for a given temperature, the pressure increases until it reaches a state of equilibrium with a liquid water surface at that temperature. This is called the saturation vapor pressure. Saturation vapor pressure is specifically related to temperature, as shown in Table D-1. Vapor pressures are commonly measured in terms of millibars of pressure.

Table D1
Saturation Vapor Pressure (mb) Over Water and Over Ice (after Byers 1974)

Temperature, °C	Over Water	Over Ice
-10	2.863	2.597
-5	4.215	4.015
0	6.108	6.108
5	8.719	
10	12.272	
15	17.044	
20	23.373	
25	31.671	
30	42.430	
35	56.236	

b. Relative humidity. The relative humidity is defined as the ratio of the measured water vapor content of the air at a specified temperature to the saturated vapor content at that temperature. It can be computed by the ratio of vapor pressures:

$$RH = \frac{e_a}{e_s} \times 100$$

where

RH = relative humidity, percent

e_a = vapor pressure of the air

e_s = saturated vapor at the temperature of the air

Relative humidity is measured by a sling psychrometer, which contains two thermometers, one in which the bulb is covered with a cloth wetted with distilled water. The dry bulb will indicate the air temperature, and the wet bulb will be cooled below the air temperature by evaporation. The amount of evaporation will depend upon how saturated the air is. Tables are available to relate the difference—the wet bulb depression—to relative humidity.

c. Dew point. The temperature at which the air must be cooled to become saturated is called the dew-point temperature. Since the temperature of the dew point is related to vapor pressure, it is used as a surrogate for vapor pressure in snowmelt equations. Dew point can be computed from relative humidity and air temperature as shown in Figure D-1.

D-2. Solar Radiation

a. Solar constant. The solar constant is defined as the rate of radiant solar energy flux received outside the Earth's atmosphere on a surface normal to the Sun's rays. At the mean distance from the Sun, this value is 1.35 kW/m^2, or 1.94 cal/(cm^2 min) (1.94 ly/min). This value varies about 7 percent during the year primarily because of the changing distance between the Earth and Sun.

b. Incident radiation. The spectral distribution (Planck Curve) of the theoretical radiation emitted by the sun is shown in Figure D-2. Solar radiation (shortwave) radiation generally encompasses the wavelength range of 0.2 to 2.2 µm. Radiation emitted by the atmosphere and Earth (long-wave radiation) has a wavelength range of 6.8 to 100 µm

(1) Solar radiation received at the Earth's surface is actually made up of both direct solar radiation, plus a

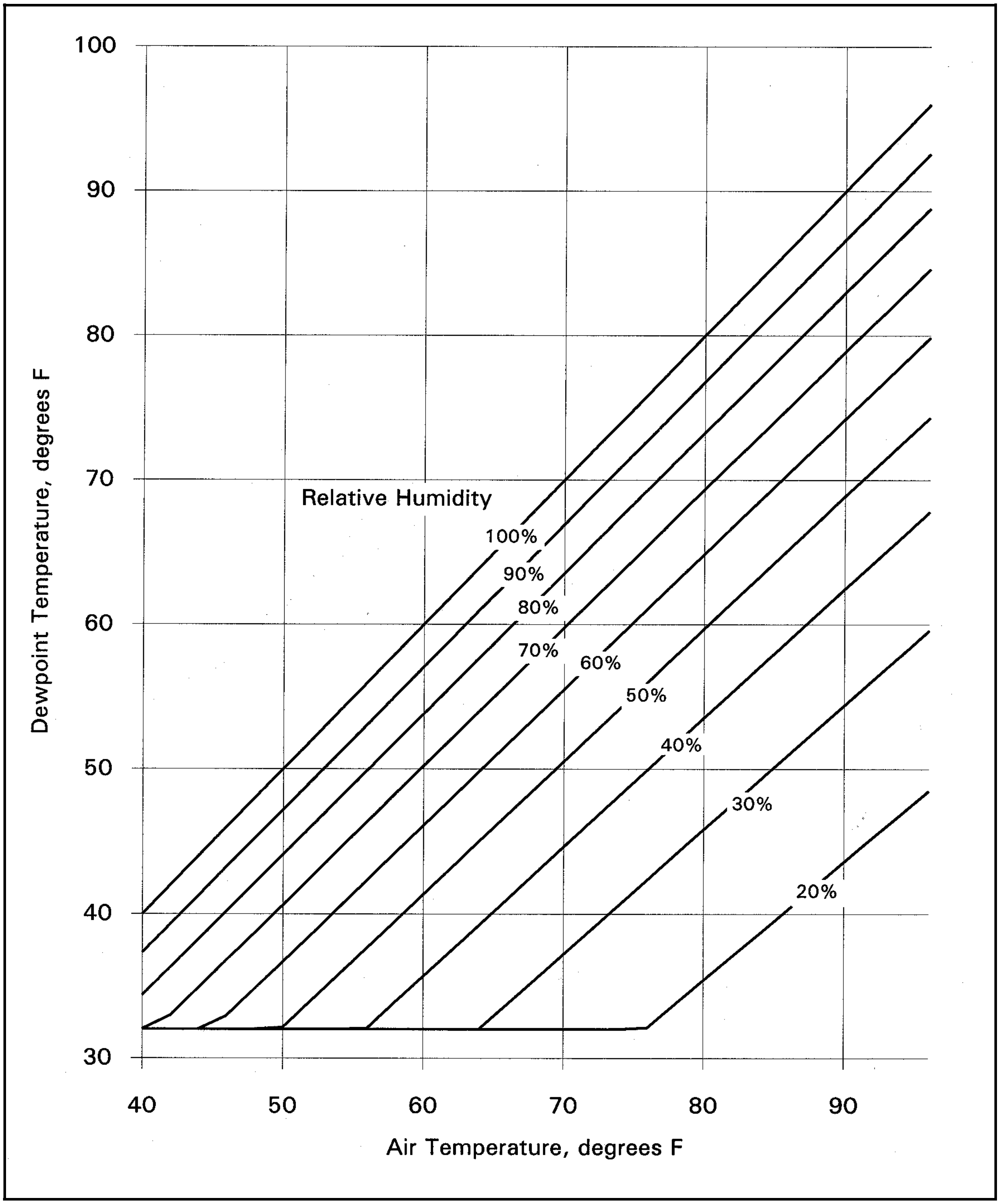


Figure D-1. Dew-point temperature as a function of air temperature and relative humidity

mall component that is scattered by the atmosphere (diffuse or sky radiation). The rate at which the total is received on a horizontal surface is termed insolation. This is expressed as a flux per unit area (flux density), such as watts per square meter or megaJoules per square meter per day. An older convention, used in *Snow Hydrology*, is g-cal/(cm^2 min), or langleys (ly) per minute, where a langley is equivalent to 1 g-cal/cm^2. Another term used to express flux density is irradiance. Table D-2 summarizes the comparisons

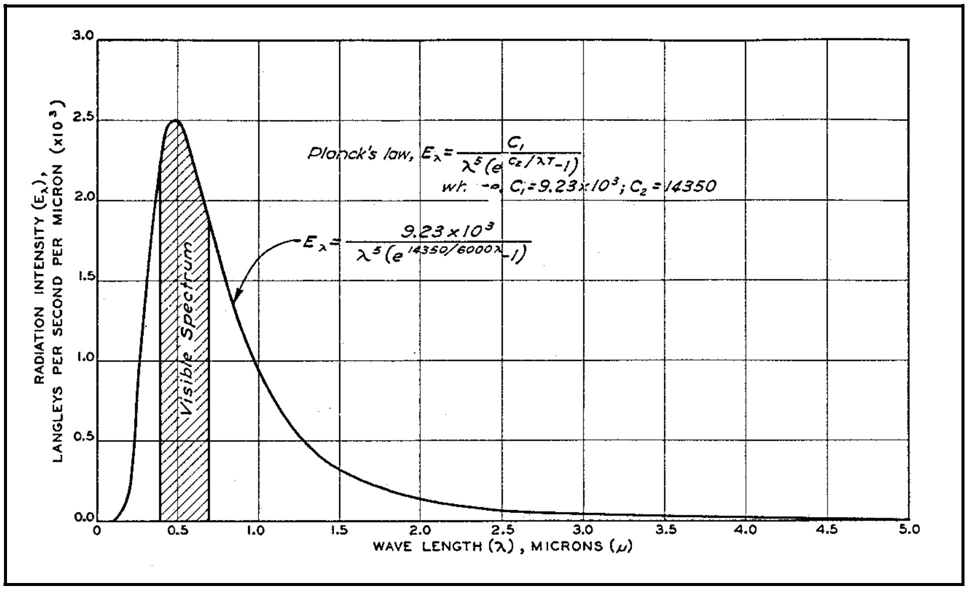


Figure D-2. Spectral distribution (Planck Curve) of the Sun's radiation (Figure 2, Plate 5-1, *Snow Hydrology*)

among three common conventions for expressing insolation. Table D-3 contains typical values of daily insolation at 45° north latitude for conditions outside the atmosphere, and for the Earth's surface assuming a cloudless sky at the maximum (spring equinox) and minimum (winter equinox) sun angles.

Table D2
Conversion Factors for Insolation Units

	ly/day cal/(cm²·day)	mJ/(m² day)	W/m²
1 ly/day =	1	0.04186	0.4844
1 mJ/(m²·day)=	23.89	1	11.57
1 W/m² =	2.064	0.0864	1

Table D-3
Typical Daily Insolation Values

For Latitude 45° N	Langleys	mJ/m²	W/m²
Top of atmosphere, 21 June	990	41	480
Top of atmosphere, 20 Dec	250	11	120
Earth's surface,[1] 21 June	750	31	360
Earth's surface, 20 Dec	200	8	97

[1] For a horizontal surface and a clear day.

(2) Insolation magnitude depends upon the solar constant, the angle of the Sun's rays (a function of season and latitude), and the amount of depletion in the atmosphere. Depletion results from absorption by gas molecules, dust, smoke, etc., and cloud particles. Clouds have by far the greatest effect in reducing the amount of radiation energy received on Earth. Figure D-3 shows the daily insolation amounts outside of the atmosphere, before attenuation by the atmosphere. The effect of atmospheric influences under cloudless skies is shown on Figure D-4, which is based upon measurements at the Central Sierra Snow Laboratory.

(3) The effect of clouds on solar radiation received can be quite pronounced and highly variable. Two factors, the amount of cloud cover (percent of sky covered) and the cloud height, are involved. Figure D-5 illustrates the effect of cloud height and cover.

(4) Another determinant for solar radiation falling upon a surface is the slope of the surface itself. In the northern hemisphere, it is obvious that a south-facing slope will receive more solar radiation than a north-facing slope of the same magnitude. This effect is more pronounced in the winter. Figure D-6 illustrates the effect of slope on incident solar radiation for latitude 46° 30' N.

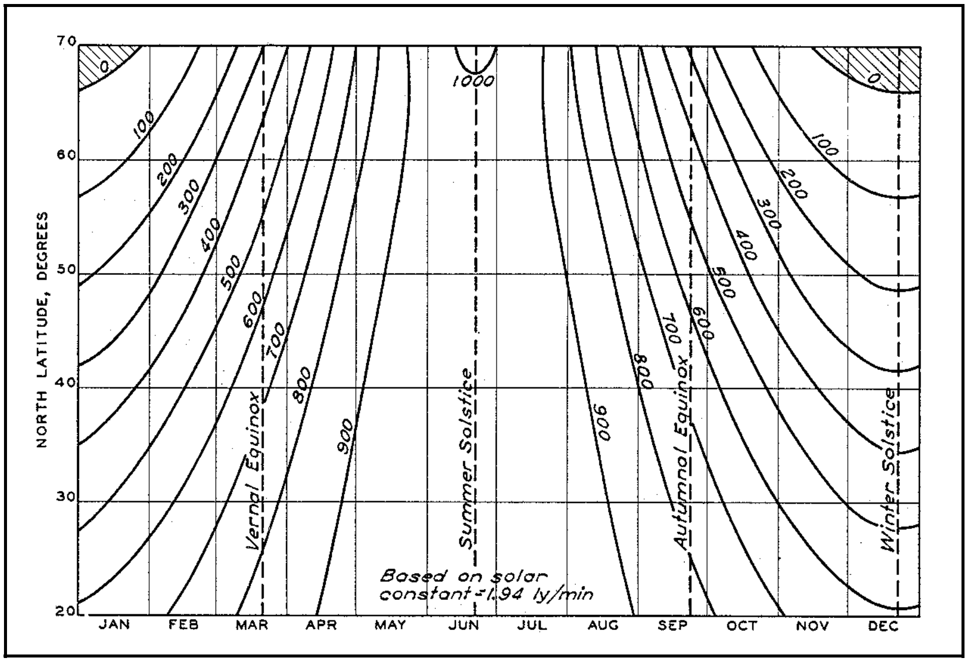


Figure D-3. Seasonal and latitudinal variation of solar radiation outside the Earth's atmosphere (Figure 3, Plate 5-1, *Snow Hydrology*)

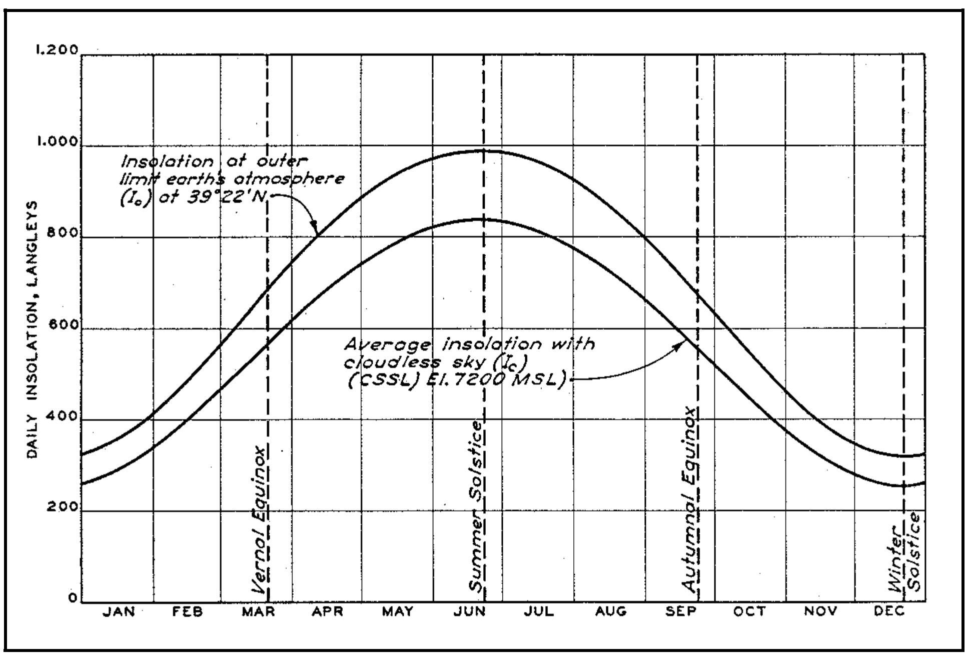


Figure D-4. Seasonal variation in insolation at the Central Sierra Snow Laboratory, showing atmospheric depletion (Figure 4, Plate 5-1, *Snow Hydrology*)

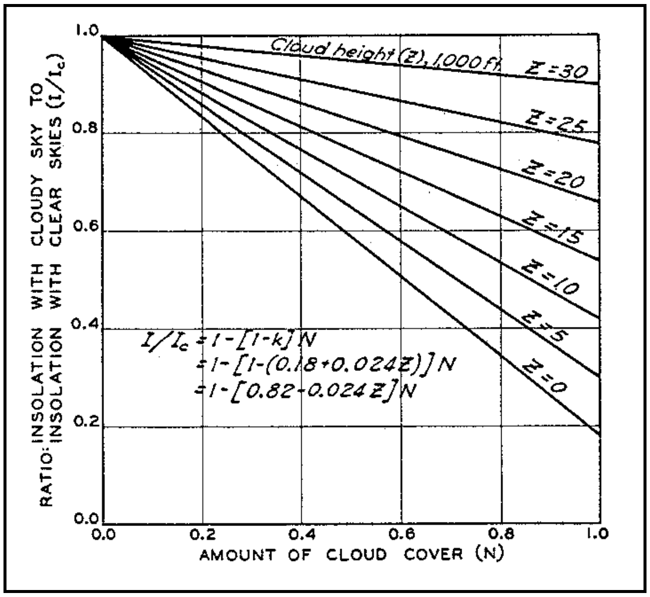


Figure D-5. Variation of insolation with cloud height and amount of cloud cover (Figure 5, Plate 5-1, *Snow Hydrology*)

(5) Forest cover also plays an important part in the amount of solar energy that reaches the snow surface. For only coniferous forests, the transmission percentage varies with the season, because of variation in the shading effect of the trees with the solar altitude. The determination of the amount of sunshine transmitted through the forest is at best approximate. Figure D-7 shows a mean transmission curve for daily insolation amounts, expressed in terms of forest canopy density. In the generalized snowmelt equations, the transmission coefficient and forest density are combined into a single factor *F*, which is termed the effective forest cover.

(6) One way of expressing the effect of cloud cover is in terms of percentage of possible sunshine. With this as a variable, a practical nomograph has been developed to estimate daily insolation as a function of latitude and season. This is shown in Figure D-8.

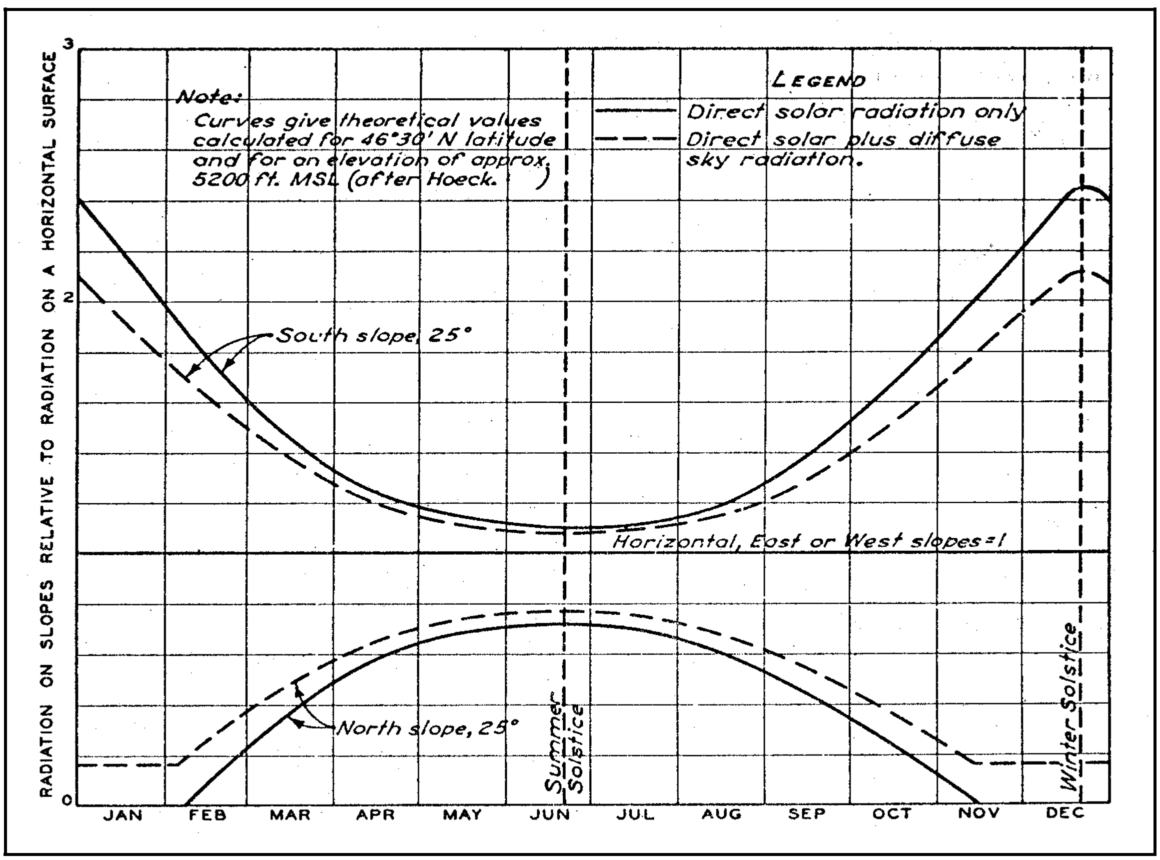


Figure D-6. Seasonal variation— radiation on slopes versus radiation on a horizontal surface (Figure 5, Plate 5-1, *Snow Hydrology*)

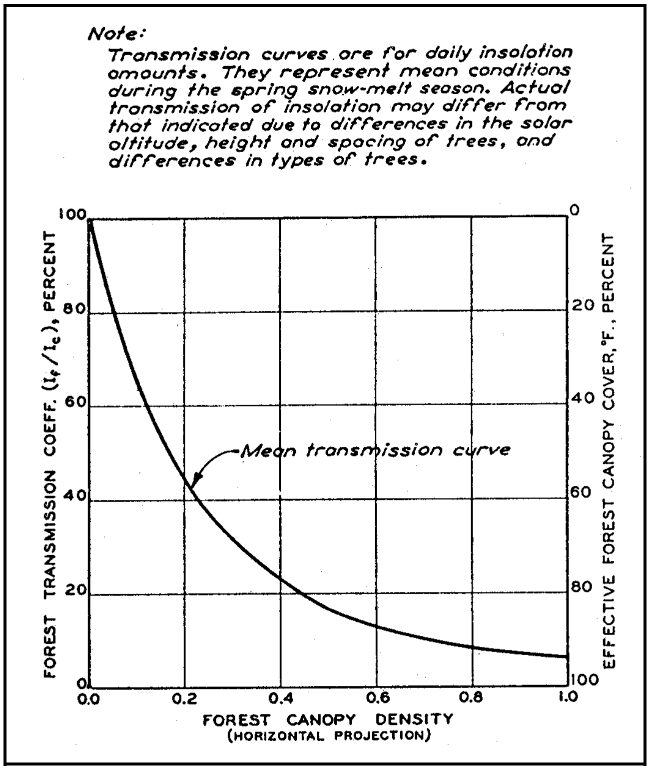


Figure D-7. Transmission of solar energy by a forest canopy (Figure 1, Plate 5-2, *Snow Hydrology*)

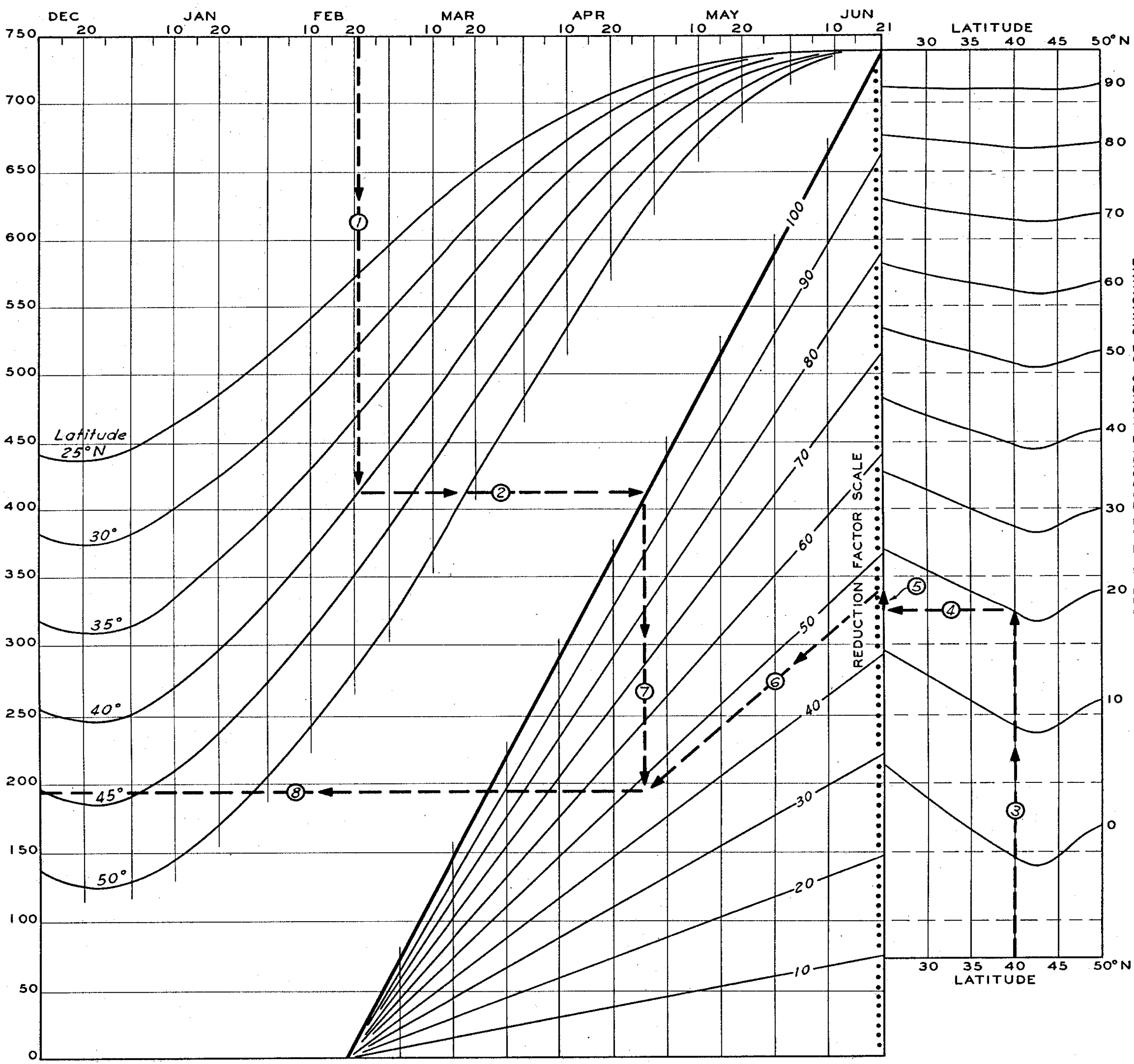
D-3. Long-wave Radiation

Long-wave or thermal radiation, emitted by the sky and Earth, encompasses wavelengths from about 6.8 to 50 µm. Figure D-9 is the spectral distribution of radiation intensity for a blackbody at 0 °C, which is approximately equivalent to melting snow. Since snow is nearly a blackbody, the outgoing long-wave radiation is essentially a constant, computed by Stephen's law as 0.45 ly/min. Back-radiation (towards the Earth) is emitted by the atmosphere, clouds, and forest cover and is a complex phenomenon that must be computed experimentally. The net long-wave radiation is equal to outgoing radiation flux less back-radiation.

a. Net radiation from clear skies. Radiation from the atmosphere can be expressed in terms of the temperature of the air and its moisture content, the latter measured by vapor pressure of the air. Figure D-10, based upon experimental evidence, illustrates the net radiation associated with open clear skies. This shows that most of the time there is an outgoing flux of radiation under clear skies—the air temperature must be 69 °F for a gain to the snowpack to occur.

b. Net radiation with cloud cover. Figure D-11 is a curve representing the theoretical net exchange under overcast skies, which are assumed to be radiating as a blackbody. This curve further illustrates the effect of cloudy skies in reducing the radiation loss that would occur for the same temperature under clear skies.

c. Net radiation with forest cover. The presence of a forest canopy is a somewhat similar situation to that of cloud cover with regard to net radiation exchange with the snowpack. The canopy, if a solid cover, absorbs and emits all possible radiation, acting at the temperature of the tree leaves, which is approximately the ambient air temperature. This effect is illustrated in Figure D-12.



Notes:

1. *The sample shown by dashed lines with arrows estimates the daily total insolation at a station at latitude 40°N, on February 21 with 20 percent possible sunshine: Step 1 — enter the graph at February 21 and move downward to the curve labelled 40°N. Step 2 — thence move horizontally to establish a point at the intersection with the heavy line. Step 3 — enter lower right of graph at 40°N, proceeding upward to 20 percent of possible hours of sunshine. Step 4 — thence move horizontally toward the left to the vertical line through 25°N. Step 5 — here the seasonal correction (+2) from the table at the right is added on the reduction factor scale. Step 6 — thence move downward toward the convergence point. Step 7 — from the reference point of step 2 move downward to an intersection with the line extended in step 6. Step 8 — from the point of intersection of these two lines move horizontally to the insolation scale to read the estimated value, 195 langleys per day.*
2. *For use between June 21 and December 21, the curves are symetrical about June 21.*

SEASONAL CORRECTION TO REDUCTION FACTOR

MONTH	PERCENT OF POSSIBLE SUNSHINE							
	0	10	20	30	40	50	60	70
JAN	+4	+3	+3	+2	+2	+2	+1	+1
FEB	+3	+3	+2	+2	+2	+1	+1	+1
MAR	−1	−1	−1	−1	−1	0	0	0
APR	−2	−2	−1	−1	−1	−1	−1	0
MAY	−4	−3	−3	−2	−2	−2	−1	−1
JUN	−5	−4	−4	−3	−2	−2	−2	−1
JUL	−5	−4	−3	−3	−2	−2	−2	−1
AUG	−4	−3	−3	−2	−2	−2	−1	−1
SEP	−2	−2	−1	−1	−1	−1	−1	−1
OCT	0	0	0	0	0	0	0	0
NOV	+2	+2	+1	+1	+1	+1	+1	0
DEC	+4	+3	+3	+2	+2	+2	+1	+1

Figure D-8. Nomograph for estimating insolation as a function of latitude, date, and duration of sunshine (Figure 3, Plate 6-1, *Snow Hydrology*)

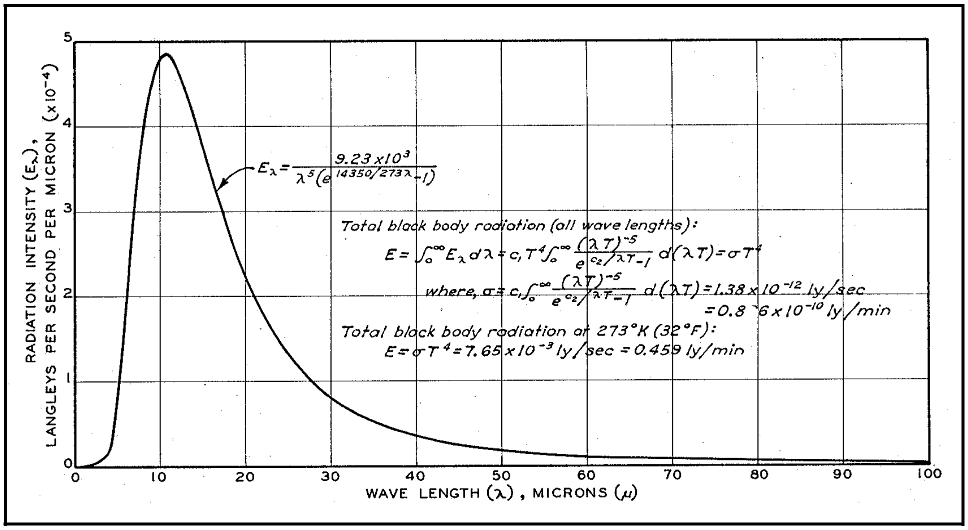


Figure D-9. Theoretical spectral distribution for a snow surface at 0 °C (Figure 1, Plate 5-3, *Snow Hydrology*)

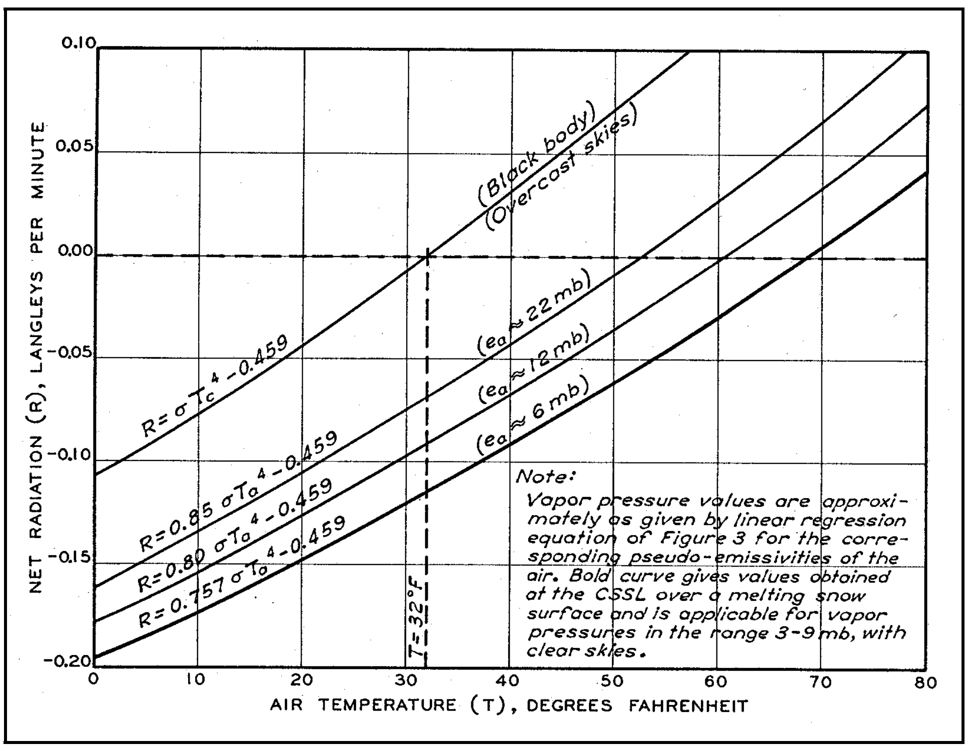


Figure D-10. Net long-wave radiation exchange between the snowpack and the atmosphere, clear skies (Figure 4, Plate 5-3, *Snow Hydrology*)

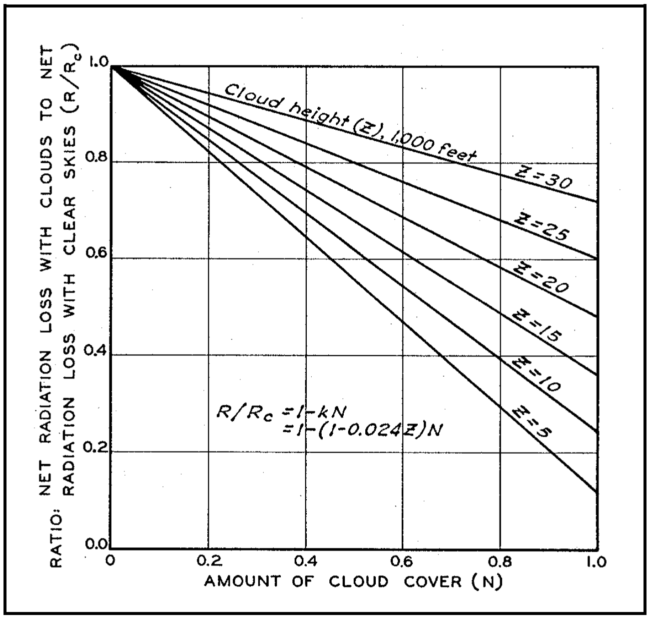


Figure D-11. Variation in net long-wave radiation loss with cloud height and amount (Figure 5, Plate 5-3, *Snow Hydrology*)

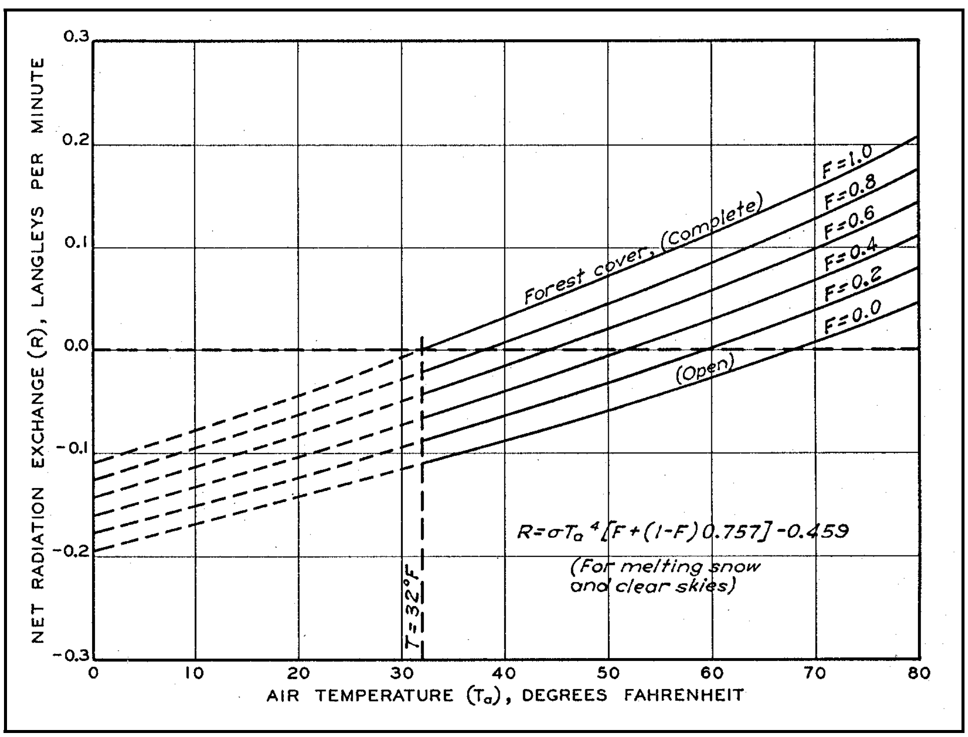


Figure D-12. Net long-wave radiation exchange in forested areas (Figure 6, Plate 5-3, *Snow Hydrology*)