CHAPTER 5 - MELTING OF THE SNOWPACK

5-01. INTRODUCTION

5-01.01 General. - Snowmelt is the over-all result of many different processes of heat transfer. The quantity of snowmelt is, moreover, dependent upon the condition of the snowpack itself. As a consequence, the rigorous determination of snowmelt amounts is quite complex and certain simplifying assumptions are used in the practical computation of snowmelt. The relative importance of the various heat-transfer processes involved in the melting of the snowpack vary with time and with locale. Considering solar radiation for example, in the plains area of the United States it is an important direct cause of snowmelt, while it is relatively unimportant except indirectly, in the heavily forested areas of the Pacific Northwest; it is of considerably less importance during the wintertime than during the spring melt season, and decreases in importance with increasing latitude. As a result of this variation in the relative importance of the several heat transfer processes involved in the melting of the snowpack, no single method or index for computing snowmelt is universally applicable to all areas and at all times of the year. In order to intelligently select the best method of computing snowmelt for a given time and area, a complete understanding of the snowmelt process is necessary. In this chapter the heat-transfer processes involved in the melting of the snowpack are first enumerated; a general snowmelt equation is next formulated which relates the net heat transfer to the resultant snowmelt, taking into consideration the thermal condition of the snowpack. Each heat transfer process is then considered separately in some detail; its variation with meteorological conditions, time of year, and place is discussed. The several processes are then summarized and interactions between them pointed out.

5-01.02 Sources of heat energy. - The principal fluxes of heat energy involved in the melting of the snowpack were enumerated by Wilson in his paper, "An outline on the thermodynamics of snowmelt," 38/ and also in Technical Bulletins 2 and 13. These fluxes, with the symbols used subsequently to identify them are:

Absorbed solar radiation (H_{rs})

Net longwave radiation exchange between the snowpack and its environment (H_{rl})

Convective heat transfer (sensible heat) from the air (H_c)

Latent heat of vaporization released by condensate (H_e)

Conduction of heat from underlying ground (H_g)

Heat content of rain water (H_p)

Each of the above is, itself, a function of several components. For example, absorbed solar radiation is the difference between the solar radiation incident on the snowpack and that reflected by it; net longwave radiation is the difference between the longwave radiation emitted by the snowpack and the portion of it radiated back from its environment (i.e., air, trees, and clouds). In the sections which follow each of the foregoing heat fluxes shall be considered separately.

5-01.03 The energy-budget equation. - If all heat fluxes directed toward the snowpack are considered positive and those away from the pack, negative (the sign being included in the term), and these fluxes are then summed, the total must be zero. Thus, considering snowmelt as but another heat transfer process and including the change in thermal quality of the snow itself (see par. 5-01.05),

$$\Sigma H = H_{rs} + H_{rl} + H_c + H_e + H_g + H_p + H_q + H_m = 0 \qquad (5\text{-}1a)$$

where H_q is the change in the energy content of the snowpack and H_m is the heat equivalent of the snowmelt (that is, the quantity of heat involved in the change of state from ice to water--the latent heat of fusion). All other terms are as previously defined in paragraph 5-01.02. From equation 5-1a it follows that

$$H_m = H_{rs} + H_{rl} + H_c + H_e + H_g + H_p + H_q \qquad (5\text{-}1b)$$

considering the heat equivalent of the melt to be positive. In the above equation, H_{rl} is ordinarily negative in the open, H_e and H_q may be either positive or negative, H_c is generally positive, and other terms in the equation are almost always positive.

5-01.04 Units. - The units used in this chapter are a combination of several systems of units. For example, the English units of degrees Fahrenheit and miles per hour are used to express temperatures and wind speeds, while the metric units of calories will ordinarily be used to express heat quanitities. Snowmelt will ordinarily be given in inches depth while unit area will be taken as one square centimeter. This heterogeneous system of units results from an attempt to express the results of this work in the familiar English units such that available data may be used directly in the resulting equations, at the same time taking advantage of some of the numerical simplicities of the metric system. Moreover, some of the basic data are commonly expressed in metric units; for example, radiation quantities are usually given in gram calories per square centimeter. In spite of the different systems of units used in this chapter, the units used to express given quantities are consistent throughout the chapter. The units will be introduced and defined as needed.

5-01.05 Thermal quality of the snowpack. - If snowmelt is defined as the liquid water which leaves the snowpack, the amount of snowmelt resulting from a given quantity of heat energy is dependent upon the thermal quality of the snowpack. While the latent heat of ice is a well established quantity (80 cal/g or 144 Btu/lb), only rarely is snow encountered which consists of pure ice at 32°F. More often, especially during the winter months, its mean temperature is less than 32°F, and some additional heat is required to first raise its temperature to the melting point before melt can begin. On the other hand, during the melt season the snowpack is not only isothermal at 32°F but also contains some free water. That is to say, instead of pure ice to be melted, there is a mixture of ice and water. When the ice matrix is melted, this free water is also released, resulting in a total quantity of water in excess of that required to melt the ice particles themselves as indicated by the latent heat of fusion for water. The actual condition of the snowpack with regard to the amount of water resulting from a given quantity of heat energy is designated as the "thermal quality" of the snowpack. Thermal quality of snow is defined as the ratio of the heat necessary to produce a given amount of water from snow to the amount of heat required to produce the same quantity of melt from pure ice at 32°F. It is usually expressed as a percentage. It may thus be seen that snow at sub-freezing temperatures will have a thermal quality greater than 100 percent, while snow containing free water will have a thermal quality less than 100 percent. This topic is considered in more detail in chapter 8. Methods whereby the thermal quality of the snowpack can be determined are discussed there and typical values of thermal quality given.

5-01.06 Resultant melt. - Since 80 langleys (calories per square centimeter) of heat energy are required to produce one centimeter of water from pure ice at 32°F, 203.2 langleys (2.54 x 80 ly) are required to produce one inch of runoff from a snowpack having a thermal quality of 100 percent. Thus letting $\underline{H_m}$ represent the total heat in langleys supplied to the snowpack, and $\underline{M}$ represent the resultant melt in inches,

$$M = H_m / 203.2 \qquad (5\text{-}2a)$$

for pure ice at 32°F. Letting $\underline{B}$ represent the thermal quality of the snow, 203.2 x B langleys are required to produce one inch of melt from any snowpack having a thermal quality $\underline{B}$. Hence,

$$M = H_m / 203.2B \qquad (5\text{-}2b)$$

for any snowpack. Figure 1 of plate 5-1 illustrates this relationship between heat supply and resultant snowmelt for snowpacks of various thermal qualities.

5-01.07 In connection with the foregoing it should be pointed out that the amount of heat required to ripen* the snowpack is relatively small compared to the amount of heat required to melt the snowpack. An example will serve to illustrate this fact. Consider one gram of snow at an initial temperature of -10°C (14°F). Since the specific heat of ice is approximately 0.5, only 5 calories of heat energy are required to bring its temperature to the melting point. Assuming the free-water-holding capacity of a ripe snowpack to be 3 percent (see chap. 8 for a discussion of free water), an additional 2 calories of heat energy are required to produce sufficient melt water to satisfy this free-water-holding capacity. Thus a total of about 7 calories are required to ripen the gram of snow, while about 78 calories are subsequently required to melt one gram of the resulting water-ice mixture.

5-01.08 Nocturnal snow crusts. - During the snowmelt season, it is usual for net flux of heat from all sources to be positive (toward the snowpack) during the day and negative (away from the snowpack) during the night. This diurnal change is especially noticeable in areas of little or no forest cover and during periods of clear weather. It results from the net longwave radiation loss during the nighttime exceeding the gains of heat resulting from convection and condensation. (During the daytime, the solar radiation combined with the greater convective transfer resulting from higher air temperatures usually far exceeds the net longwave loss.) This nocturnal loss of heat energy from the snowpack results in the formation of a crust on the surface of the snowpack. The free water within the pack is refrozen and the snow cooled to some temperature below freezing. The effect is usually confined to the top layers of the pack. In spite of the fact that this deficit may amount to some 80 langleys during a single night, such crusts seldom exceed 10 inches. They are generally much thinner, being around 6 inches in thickness and representing heat deficits of 20 - 40 calories. The nocturnal snow crust represents a heat-energy deficit that must be subtracted from the subsequent day's net gain in determining daily snowmelt amounts. It is sometimes referred to and expressed as a "negative melt" quantity, in which case it is expressed in units of inches of melt rather than langleys.

5-01.09 Data. - Much of what follows is based on snow lysimeter studies of snowmelt made at CSSL (see Res. Notes 17 and 25); however, what is presented herein is generally applicable to any snow-covered area. Reference is made to several publications which contain general data pertinent to the energy-budget approach to snowmelt, and

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*For the purposes of this chapter a ripe snowpack is defined as one which is isothermal at 32°F and has all of its free-water-holding capacity satisfied. (Free water includes only that water permanently held within the snowpack; that is, water held by adsorption and capillarity. It does not include water in the process of percolating through the pack or water impounded in the pack as a result of poor drainage conditions.)

which were consulted in the preparation of this report. Sverdrup's, "The eddy conductivity of the air over a smooth snow field," 34/ gives much valuable data from carefully made meteorological observations over a barren snow field. It also presents the general energy-budget approach to snowmelt along with detailed theoretical background. John Hopkins University, Publications in Climatology, 18/ 36/ likewise present much detailed data on vertical air temperature, humidity, and wind-speed gradients near the ground, and on radiative heat transfer near the ground. These data are generally for snowfree conditions although some data are included for measurements made over snow. These publications also include some of the most recent developments in the theory of heat transfer near the ground. Geological Survey Circular 229, "Water-loss Investigations: Volume 1 - Lake Hefner Studies, Technical Report," 2/ and Sverdrup's Oceanography for Meteorologists 35/ both contain a general reivew of the theory of the energy-budget approach and also include considerable data obtained over water surfaces. Brunt's Physical and Dynamical Meteorology, 6/ Sutton's Micrometeorology, 33/ and Geiger's The Climate Near the Ground 13/ are three excellent texts dealing with the questions pertinent to this report which were referred to frequently in its preparation.

5-02. RADIATION THEORY

5-02.01 Planck's law. - All bodies radiate energy, the intensity of the radiation being a function of the temperature of the radiating body; moreover, the spectral distribution of the radiation is also a function of the temperature of the radiating body. Generally speaking, the higher the temperature, the greater the intensity of the total radiation emitted and the shorter the wave length of the maximum intensity. The spectral distribution of the energy of a radiating black body is given by Planck's Law,

$$E_\lambda = \frac{C_1}{\lambda^5\left(e^{C_2/\lambda T} - 1\right)} \qquad (5\text{-}3)$$

where E_λ is the intensity of the emitted radiation of wave length λ, T is the temperature of the radiating body and C_1 and C_2 are constants. Figure 2 of plate 5-1 and figure 1 of plate 5-3 show the theoretical distribution of radiation intensities in the spectrum of a black body in accordance with equation 5-3. Figure 2 is for a body at a temperature of 6000 degrees K (approximate sun temperature) and figure 1 for a body at 273 degrees K (temperature of a melting snowpack). This is the general expression for radiant energy. Other expressions may be derived from it as given in the following two paragraphs.

5-02.02 Wien's law. - For any given temperature, E_λ is zero for $\lambda = 0$ and for $\lambda = \infty$; for some intermediate value of λ, E_λ has its maximum value. This wave length of maximum intensity of radiation can be determined from equation 5-3. Differentiating and equating to zero, gives,

$$\lambda_m T = \text{constant} \tag{5-4}$$

where λ_m is the wave length at which E_λ is a maximum. Equation 5-4 is know as Wien's Law. The value of the constant is usually taken as 2940 for wave lengths expressed in microns (equals 10^{-6}m) and temperatures in degrees K. Thus the wave lengths of maximum intensity are 0.49μ and 10.8μ for the temperatures of 6000 and 273 degrees K, respectively.

5-02.03 Stefan's law. - The total energy emitted in all wave lengths (per unit time and area) by a black body may be determined by integrating equation 5-3. Thus the total radiation in all wave lengths, E, from a black body at a temperature T is,

$$E = \sigma T^4 \tag{5-5}$$

where,

$$\sigma = \int_0^\infty \frac{C_1}{(\lambda T)^5 \left(e^{C_2/\lambda T} - 1\right)} d(\lambda T)$$

Equation 5-5 is known as Stefan's law and σ as Stefan's constant. The value of σ is 0.826 x 10^{-10} (ly/min) / $(^{o}K)^4$. For a radiating body other than a black body, its radiation relative to the radiation of a black body is expressed by a ratio known as its emissivity. Figure 2 of plate 5-3 is a graphical presentation of Stefan's law, and it also shows radiation intensities corresponding to the various temperatures for emissivities less than unity (black-body emissivity).

5-02-04 Solar and terrestrial radiation. - Only a very small portion of the entire electromagnetic spectrum (which ranges from cosmic rays and the emissions of radioactive substances with wave lengths of the order $10^{-6}\mu$ to low-frequency radio waves having wave lengths of the order 10^4 m) is involved in the radiation melt of the snowpack: the radiation between about 0.15μ to 80μ. This radiation is divided into two general categories: solar and terrestrial. Solar (or shortwave) radiation is included in the range from about 0.15 to 4μ which encompasses the visible spectrum (0.4 to 0.7μ). It has its maximum intensity in the visible spectrum at about 0.5μ. It also extends into the ultraviolet and the infrared. Figure 2, plate 5-1 shows the theoretical distribution of intensity of solar radiation at the different wave lengths, and also shows the limits of the visible spectrum. It may be seen that roughly half of the solar radiation lies within the range 0.4 to 0.7μ, that is, half the total solar radiation is in the form of visible radiation or light. Terrestrial (or longwave) radiation is generally included in the range 3μ to 80μ: it has its maximum intensity in the infrared at around 11μ. Figure 1, plate 5-3 shows the spectral distribution of radiation intensity from a black body at 273°K (32°F), as given by Planck's law (equation 5-3). From Stefan's law, the total energy radiated per unit time and area is found to be 0.459 ly/min or 27.5 ly/hour. In the following two sections, both of the foregoing types of radiation involved in the melting of the snowpack will be considered separately; their variations with cloud and tree cover will be discussed and methods presented by which they may be estimated in the absence of measurements.

5-03. SOLAR RADIATION

5-03.01 The solar constant. - Of the tremendous quantity of radiant energy emitted by the sun, only an infinitesimally small portion is intercepted by the earth and its atmosphere. Yet this small portion is the ultimate source of all the earth's energy. The amount of solar energy intercepted by the earth varies slightly with seasons due to the varying distance between the earth and sun, and is thought to have small day-to-day variations due to changes in the solar output; however, these variations are quite small. The intensity of incident radiation on the earth is given by the solar constant which is defined thus: the intensity of solar radiation received on a unit area of a plane normal to the incident radiation at the outer limit of the earth's atmosphere with the earth at its mean distance from the sun. The value of the solar constant is generally taken to be 1.94 langleys per minute which is based on the 1913 Smithsonian standard scale. 29/ Until recently, it was thought this value was too high; a solar constant of 1.90 ly/min being the best available figure. 11/ More recently there has been evidence that these values are too low, a value of 2.00 ly/min being offered. 24/ These differences arise from the fact that solar radiation, measured at or near the earth's surface, is less than that incident at the outer limit of the earth's atmosphere even during clear weather. Estimates must be made of the portion absorbed and reflected by the atmosphere to arrive at the solar constant. This is particularly true in the ultraviolet where the radiation is largely absorbed. Recently observations of the solar spectrum have been made at high altitudes from rockets, and estimates of the solar constant have been made using these data. 24/ However, even those observations leave portions of the spectrum to be estimated. In this report the solar constant will be taken as 1.94 ly/min.

5-03.02 Insolation. - Of direct concern to the study of snowmelt is the amount of solar radiation incident on a horizontal surface. This is termed insolation. The daily amount of insolation received at the outer limit of the earth's atmosphere (or at the earth's surface in the absence of an atmosphere) may be calculated from the solar constant for any given latitude and time of year by taking into consideration: (1) the distance between the earth and sun, (2) the angle of incidence of the sun's rays, and (3) the duration of sunlight. Figure 3, plate 5-1 shows daily insolation amounts as a function of latitude and time of year, between latitudes 20° and 70° N.

5-03.03 Transparency of atmosphere. - The portion of the insolation given by figure 3 of plate 5-1 which actually reaches the earth's surface depends upon the transparency of the atmosphere and the optical airmass through which it must pass.* Some of the incident solar radiation is reflected, some scattered, and some absorbed by the atmosphere. In the absence of clouds these amounts are relatively small and quite constant barring unusual atmospheric conditions such as dust storms. The variations that occur are chiefly a result of variations in the amount of water vapor and dust in the air. The effects of water vapor and dust on the absorption and scattering of solar radiation has been extensively investigated; however, a detailed consideration of these effects is beyond the scope of this report. Reference is made to Technical Bulletin 5 and Research Note 3 and an article by Klein 27/ for summaries of work done on this subject. For purposes of this report it is sufficient to point out that the average daily insolation received at the earth's surface with clear skies may be determined for any given locality by plotting daily totals of measured insolation (see par. 5-03.09) as a function of time of year, and drawing a curve which envelopes these values, being guided by the curve giving the values of daily insolation received at the outer limits of the earth's atmosphere for the latitude of the particular site. This is done for CSSL (lat. 39° 22' N) as shown by figure 4 of plate 5-1. It is to be pointed out that, generally speaking, the higher the elevation of the station the greater is the atmospheric transmission, other things being equal. Also, the less the zenith distance, the greater is the atmospheric transmission. These follow from the fact that the smaller the optical airmass, the greater is the transmission for any given condition of the atmosphere. (Atmospheric transmission coefficient is defined as the ratio of the insolation received at the earth's surface with a cloudless sky, I_c, to the insolation received at the outer limit of the earth's atmosphere, $\overline{I_o}$. Actually, for the same airmass, the transmission is somewhat greater in the winter than in the summer because of the usually clearer air that prevails during the winter. It may be noted in figure 4 of plate 5-1 that the atmospheric transmission is greatest in the summer and least in the winter in consequence of the differences in optical airmasses. The atmospheric transmission coefficient varies from about 80 percent at time of the winter solstice to about 85 percent at the time of the summer solstice.

5-03.04 Atmospheric transmission coefficients are based on the total insolation received at the earth's surface and, as such, include both the direct solar beam and the diffuse sky radiation (scattered light reaching the earth's surface). They include an amount of diffuse sky radiation based on normal conditions. One factor that has a pronounced effect on the amount of diffuse sky radiation reaching the earth's surface is the albedo or reflectivity of the surface itself. Since a

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*Optical airmass is very closely given by the secant of the zenith (distance) angle for sea-level locations. For elevated stations, the secant must be multiplied by the ratio of the average station pressure to standard sea level (p/p_o).

portion of the reflected bea is also scattered back to the earth's surface, the greater the reflectivity of the surface, the greater is the diffuse sky radiation, other things being equal. Since for ordinary conditions of bare ground, the albedo is quite small and constant, this is a minor effect. However, over snow surfaces, the generally high albedo and its large range (40 to 80 percent) makes this effect of considerable importance. Thus the amount of insolation received at the earth's surface increases with the albedo of the snowpack, other things being equal. Figure 4 of plate 5-1 includes the average result of this effect at CSSL. During the winter, the higher average albedoes resulting from new-fallen snow tend to increase the amount of insolation reaching the earth's surface. During the spring, the lower albedoes of the older snow decrease the amount of diffuse sky radiation, while during the summer months, the bare ground makes this effect practically nil. Thus the decrease in the atmospheric transmission coefficient during the winter months due to the greater optical airmass is somewhat offset, for snow covered areas, by the increase in diffuse sky radiation (in addition to the clearer air of winter which also tends to increase the atmospheric transmission coefficient.)

5-03.05 Effect of clouds. - By far the largest variations in the portion of solar radiation transmitted by the atmosphere are caused by clouds. Yet this variation is also one of the most difficult to evaluate. The transmitted radiation varies with type, height, density, and amount of clouds. Several investigations have been made which relate the ratio of the insolation actually received at the earth's surface ($\underline{I}$) to the average insolation received at the earth's surface with cloudless skies ($\underline{I_c}$), to the amount of cloud cover ($\underline{N}$). Thus,

$$I/I_c = 1 - k'N \qquad (5\text{-}6a)$$

For $\underline{N}$ expressed in tenths of sky cover an average value for $\underline{k'}$ of 0.71 is given by Sverdrup (p. 51). 35/ However, there must necessarily be a considerable variation in this value from place to place, season to season and for different types and densities of clouds. It may be seen that for an overcast sky ($N = 1$), this value of $\underline{k'}$ makes, $I = 0.29I_c$. Recently a value of 0.54 has been suggested by Newmann 32/ which results in $I = 0.46I_c$ for an overcast sky. Haurwitz 21/ gives the I/I_c ratio in the form,

$$I/I_c = 1 - (1 - k)\,N \qquad (5\text{-}6b)$$

where $\underline{k}$ is equal to $1 - k'$ and has a certain physical significance: it is the ratio of the insolation received with overcast skies to the insolation received with cloudless skies. This is the more usual form of the equation used to relate the ratio I/I_c to the amount of cloud cover, $\underline{N}$. Values of $\underline{k}$ have been determined by Haurwitz 21/ for different types of clouds. Figure 5 of plate 5-1 shows variations in the value of $\underline{k}$ with cloud type and height (after Haurwitz) and also the resultant variation in the percentage of insolation transmitted with various cloud heights and amounts using the values of k when substituted in equation 5-6b.

5-03.06 Another approach which has been used to estimate the depletion of insolation by clouds is to relate the ratio I/I_c to the percentage of possible sunshine (S), as determined by a sunshine recorder, by an equation of the form,

$$I/I_c = a + b\ S \tag{5-7a}$$

The coefficients a and b may be evaluated statistically. Fritz and MacDonald 12/ determined values of 0.35 and 0.61 using monthly data for the United States. A more usual form of the equation used to relate the ratio to sunshine amounts is,

$$I/I_c = k'' + (1-k'')\ S \tag{5-7b}$$

Here k" is the value of the ratio (fraction of cloudless sky insolation) on a day with zero recorded sunshine. Values of k" in equation 5-7b for the United States are given by Kimball. 26/ The most generally used value of k" is 0.22; however, it too is subject to variation with all of the aforementioned factors plus possible differences in the settings of the sunshine recorders. A general summary of this approach to the estimation of insolation--i.e. from sunshine data--is to be found in a paper by Hamon and others. 19/ This paper gives a graphical means of computing insolation from percent possible sunshine, with the additional parameters of latitude and time of year. (See pl. 6-1 for a reproduction of this graph.) The linear relationship between the insolation ratio, I/I_c, and the percent sunshine, S, implicit in equations 5-7 is not used in this paper; rather an empirical curvilinear relationship is derived. Equations 5-6 and 5-7 are for total insolation received at the earth's surface; hence they include diffuse sky radiation. As was previously mentioned, the quantity of diffuse sky radiation relative to the direct solar beam, is affected by the albedo of the surface. This effect is even more pronounced with cloudy skies than it is with clear skies; not only is the ratio of diffuse sky radiation to direct radiation increased by the presence of clouds, but the light reflected from the earth's surface is strongly re-reflected by the clouds. Thus, over snow-covered areas, greater values of the ratio I/I_c are to be expected than are found over non-snow-covered areas for a given type and amount of cloud cover. The actual values of the constants are also dependent, for snow-covered areas, upon the albedo of the pack and forest cover, in addition to all of the aforementioned variables. In view of these complexities, the previously given values of the constants should be used as general guides only in the determination of insolation amounts. Moreover, they apply, strictly, only to long-term averages of data and may be considerably in error for a given day. Only by actual measurement at the site can the true amount of daily insolation be accurately determined.

5-03.07 Effect of slope. - It is obvious that outside of the tropics (northern hemisphere), the radiation incident on south-facing slopes exceeds that on north-facing slopes. For moderate slopes during the springtime, as a result of the high solar altitude, the effect of slope is slight. During the winter the effect is more pronounced. At any given instant, the radiation on a sloping surface relative to the radiation received on a horizontal surface may be determined from the geometry of the situation (the slope and its aspect in conjunction with the solar altitude and azimuth). However, if daily totals are to be determined, the problem is more complex. Such a determination involves the integration of the solar path relative to the sloping surface. It, of course, varies with the time of year as a result of the changing solar path and is different for every slope and slope aspect. In addition, there is the variability in the transmission of radiation with solar altitude in consequence of the differences in optical airmass through which radiation must pass. Then too since, diffuse sky radiation is the same for all slopes and aspects, this constant factor must be included in all computations. All these effects have been included in some analyses made by Hoeck 22/ of daily totals of radiation on slopes of 25 degrees for a latitude of 46° 30' N. and an elevation of 1577 meters. Data are given for both south- and north-facing slopes (it is shown that radiation on east- and west-facing slopes is the same as that on a horizontal surface). These data are given in figure 6 of plate 1. This figure also presents values of total solar radiation (direct plus diffuse sky radiation). The above cited paper by Hoeck is an excellent and detailed treatise on the role of radiation (both longwave and short-wave) in the melting of the snowpack. Reference is also made to Geiger 13/ (p.224) and a paper by Thelkeld and Jordan 37/ for more detailed information on evaluation of incident radiation on sloping surfaces.

5-03.08 Effect of forest cover. - As is the case with cloud cover, the determination of the effect of forest cover on the amount of insolation reaching the ground is somewhat inexact. The transmission percentage varies with the density, type and condition of the trees. Deciduous trees, of course, show marked variations in the transmission ratio with the season; the analysis of transmission variation for this type of forest is most difficult. Fortunately, however, the principal type of forest in the snow-covered areas of the western United States is coniferous. Figure 1 of plate 5-2 gives an average curve for the transmission percentage of insolation through coniferous forest canopies of various densities. It is presented only as a general guide; actual quantities for a particular area may vary considerably from the curve. For the data in the figure, the canopy cover is defined as the horizontal projection of tree crown area. This relationship is based on snow laboratory data, and includes cloudy- as well as clear-weather data. There is some variation in the ratio with the relative amounts of direct and diffuse sky radiation, hence the ratio would also vary with degree of cloudiness and solar altitude. This subject is considered in some detail by Miller 31/ who examined the data from the snow laboratories along with other pertinent data. It will not be dwelt on further here. Research Notes 5, 9, and 12 also deal with this question in more detail.

5-03.09 Measurement. - The preceding paragraphs describe the causes of variations in insolation incident on the snow surface and give relationships by which insolation amounts may be estimated in the absence of actual measurements. They were derived from actual measurements at pyrheliometer sites for application to areas where no such instruments are located. Of course, the most accurate and sometimes the only practical method of determination of insolation is to be had from actual measurements. In this country, measurements of insolation are made almost exclusively by means of Eppley pyrheliometers. 20/ This instrument consists of an evacuated bulb in the center of which is a disk having a white center and concentric bands of black and white. A thermopile, having its alternate junctions in the black and in the white, produces an emf in proportion to the temperature difference between the rings, and hence to the radiation incident upon them. This emf may then be recorded on a recording potentiometer, suitably calibrated to give radiation intensity. Because of the glass envelope, longwave radiation is excluded (as is also a portion of the solar radiation, both due to reflection at the air-glass interface, and to absorption of certain wave lengths by the glass). Errors result from the degree of heating of the bulb in different ambient temperatures; however, they are quite small and are generally ignored. There are currently, in the United States, some seventy pyrheliometer stations at which insolation is continuously recorded. These data are published monthly in the U. S. Weather Bureau's National Summary of Climatological Data. (Prior to January, 1950 they were published in the Monthly Weather Review.) Continuous measurements of insolation were made at UCSL and CSSL throughout the period of operation of these laboratories (see chap. 2). These data are published in the Hydrometeorological Logs for these laboratories.

5-03.10 Albedo of the snowpack. - The albedo, or reflectivity, of the snowpack varies over a considerable range: new-fallen snow may reflect 80 percent or more of the incident insolation, while a ripe, granular snowpack may reflect as little as 40 percent. Consequently, the albedo of the snow has an important role in the melting of the snowpack. The albedo is primarily a function of the condition of the surface layers of the snowpack. Before taking up methods by which the albedo of the snow surface may be estimated, it is well first to consider briefly the manner in which albedo of the snow is commonly measured and also to consider the manner in which this reflection takes place and its effects upon the measurement of reflected radiation. Albedo of the snowpack is commonly measured (in this country) by means of two Eppley pyrheliometers as was done at the snow laboratories. The two pyrheliometer bulbs are first compared in direct sunlight to assure identical calibration. One of these is mounted in the normal position (see par. 5-03.09) and measures the insolation received. The other is inverted and measures the shortwave radiation reflected by the snowpack. Albedo of the snow is generally considered to be the inverse ratio of these two quantities. Two considerations that are of importance to the measurement of the albedo of snow are: (1) the diffuse and (2) the spectral character of the albedo. If the albedo of the snow varies with the angle of incidence of the radiation, then, even for identical snow conditions, different albedoes will result from annual and diurnal changes in the sun's altitude. If the albedo of the snow is different for different wave lengths of solar radiation, then the measurements of reflected and incident radiation intensities may not be directly comparable. Each of these two considerations will be examined separately in the following paragraphs.

5-03.11 Snow is generally considered to be a good diffuse reflector; that is to say, the intensity of reflected light is independent of the angle of the incident beam. The intensity of reflected light should thus conform to the cosine of the angle of the reflected beam. While this is nearly true for small angles of incidence, the cosine law does not hold strictly for large angles of incidence, higher albedoes being associated with larger angles of incidence. This variation may also be partially the result of changes in the structure of the snow itself. In the morning and in the evening a crust may occur on the snow surface. During midday, when the melt rate is at a maximum, the higher concentration of liquid water in the top layers of the snowpack undoubtedly decreases the albedo. However, even an immature, non-melting snowpack exhibits the same characteristic, although, it is generally believed to be to a lesser degree. This effect has also been investigated by Hubley. 23/ In practice, this diurnal variation in albedo is obviated by use of a mean daily albedo. This specular quality of the snow also has a seasonal effect on the albedo of the snowpack as a result of the annual change in the sun's altitude. Thus slightly higher albedoes would be expected to result during the winter than during the spring for the same snow conditions in consequence of the smaller solar altitudes.

5-03.12 As previously mentioned, the albedo of the snow is usually determined from the readings of a pair of Eppley pyrheliometers, one in the normal position measuring incident radiation on a horizontal surface and the other inverted to measure reflected radiation. Since Eppley bulbs are calibrated in direct sunlight and since varying amounts of incident radiation are reflected and absorbed by the glass envelope of the bulb, depending upon the wave length of the radiation, the resulting calibration is for the spectral distribution of the radiant energy of the incident solar radiation only. Should the spectral distribution of the reflected rays differ significantly from the incident, the calibration would no longer strictly hold. Spectral measurements of the albedo of snow have shown it to be quite constant through the visible range, as is obvious from the dazzling whiteness of the snow surface. Progressing into the near infrared the reflecting power decreases rapidly. This is illustrated in figure 2 of plate 5-2 which shows the spectral reflectivity for a ripe, melting snowpack. (Data for this figure are from SIPRE Report 4. 30/) However, since the transmission of the glass envelope of the Eppley bulb and the intensity of the radiation itself both decrease with increasing wave lengths in the near infrared, the longer wave lengths of solar radiation have relatively little effect upon the calibration of the pyrheliometer and consequently, the lesser albedoes in this portion of the solar spectrum are quite unimportant.

5-03.13 Continuous measurements of incident and reflected radiation over a snowpack were made at CSSL during the period of snow cover for the years 1946 through 1954. These measurements were made using the two-Eppley pyrheliometer method previously described. Several studies have been made, using these data, to determine the variation of albedo with time, and with accumulated heat supply (as determined by radiation and temperature indexes). (Reference is made to Tech. Bull. 6 and to Res. Note 1.) The results of these studies are summarized in figures 3 and 4 of plate 5-2 which show the variation of albedo with time and with accumulated heat supply index. Different curves are given for different times of the year in figure 4. It may be noted that during the melt season, the high albedoes of new-fallen snow quickly decrease to the albedo of the older snow. There is a lower limit of about 40 percent. These curves are for an uncontaminated snow surface; snow naturally contaminated by forest litter, dust, etc., or artifically contaminated would, of course, exhibit lower albedoes. The curves are general ones derived from the study of many separate occurrences. Individual situations may vary significantly from the average values represented by the curves. For example, when an old, melting snowpack having an albedo of, say, 50 percent is covered by a light snow fall, its albedo may increase to 80 percent and then return to 50 percent in a day or two when the thin cover of newly-fallen snow is melted. The curves presented, however, are good general guides. They may be entered at any point corresponding to the current albedo of the snow, and estimates of the future albedo of the snowpack made therefrom. A more complete discussion of the albedo of the snowpack can be found in the previously referenced report by Miller.31/

5-03.14 Absorption of radiation by the snowpack. - The difference between the solar radiation incident on the snow surface and that reflected by it is the solar radiation absorbed by the snowpack. This absorption occurs not only at the surface as with more opaque materials, but, because of the translucent nature of snow, extends to some depth within the pack. For deep, ripe snowpacks, this penetration is of little practical concern; the heat absorbed by the snow would result in the same quantity of melt were it all absorbed in the top surface or should it penetrate to a depth of a foot or so. However, for shallow snowpacks the penetration of radiation may result in a measurable quantity being transmitted by the snowpack to the underlying surface. As a result of the usual low albedo of this surface (rock or soil), most of this transmitted radiation is absorbed. Most of the heat energy may be returned to the snowpack by conduction and/or longwave radiation, producing about the same melt as would have occurred in a deeper snowpack. On the other hand, in the case of frozen ground, some of this heat may be absorbed by the ground without any melt resulting at the ground level.

5-03.15 For snowpacks having a homogeneous structure, the penetration of solar radiation into the snowpack may be expressed by a logarithmic law,

$$I_d = I_a e^{-kd} \qquad (5\text{-}8)$$

where I_a is the intensity of radiation transmitted through the snow surface, I_d is the intensity at depth d beneath the surface, and k is the extinction* coefficient for snow. In experiments made of the absorption of solar radiation by snow, the value of k has been found to vary, its value being chiefly dependent upon the density of the snowpack; the higher the density the greater is the penetration. For depths expressed in centimeters, in equation 5-8, the following values of k have been experimentally determined (Tech. Rpt. 8, Int. Rpt. 1):

Snow Density (percent)	Extinction Coefficient (k)
26.1	0.280
32.2	0.184
39.7 } 44.8 }	0.106

Figure 5 of plate 5-2 illustrates the penetration of radiation in the snowpack for different densities of snow using the coefficients given above. From this figure it may be seen that, for ripe, high density snow, about 4 percent of the radiation absorbed by the snow penetrates as far as one foot; for lesser densities of snow the radiation penetration is less. A further discussion of this topic may be found in Technical Report No. 8 (Interim Report No. 1) and in a paper by Gerdel.14/

5-03.16 The albedo of snow is mainly determined by the character of the snow surface; sub-surface conditions have little effect on the albedo. When an old surface of low-albedo snow is buried by an appreciable layer of new-fallen snow, the snowpack exhibits the albedo of the new surface. When the old surface is reexposed by the melting of the new layers, its albedo becomes as it was previously. Similarly the albedo of shallow snow (as low as 6 inches) is thought to be little affected by the type and condition of ground beneath the snow. Measurements of albedo, however, usually show a marked decline in albedo as the snow cover becomes thin. This results mostly from the patchiness of the shallow snow, the reflected radiation being the result of reflection both from bare ground and snow. Under these conditions, melt is usually accelerated because of the greater portion of solar radiation absorbed by the combined ground and snow surfaces, even though little if any more radiation is directly absorbed by the snow itself. The warming of the ground by solar radiation accelerates melt both by conduction of heat from the ground, and indirectly, by warming the air passing over it. This results in a rapid edge melting of snow patches.

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*The term "extinction coefficient" is used rather than "absorption coefficient" inasmuch as some of the decrease in the intensity of the incident beam with depth results from internal reflections; not all the radiation which penetrated the snow surface is absorbed.

5-04. TERRESTRIAL RADIATION

5-04.01 Radiation emitted by the snowpack. - Snow is very nearly a perfect black body with respect to longwave radiation, 10/ absorbing all such radiation incident upon it and emitting the maximum possible radiation in accordance with Stefan's law (equation 5-5). While this may seem somewhat strange, considering the high albedo of snow with respect to shortwave radiation, particularly in the visible spectrum, a consideration of the character of the snow surface indicates the reason for this phenomenon. Since the snow surface is composed of small grains of ice having many facets, when considered microscopically, it is extremely rough. The multi-faceted crystals and the interstices between the crystals quite effectively trap incoming longwave radiation, and, conversely, are an efficient emitting surface. Since snow radiates as a black body in accordance with Stefan's law, the longwave radiation emitted by the snowpack may be readily calculated. The intensity of black-body radiation for different temperatures is given in figure 2, plate 5-3. Since the temperature of snow is limited to a maximum of 32°F, the maximum intensity of radiation that may be emitted by it is 0.459 ly/min (equals 27.5 ly/hr). Radiation intensities corresponding to emissivities other than unity are also given by the figure (these will be discussed subsequently).

5-04.02 Back radiation to the snowpack. - Back radiation to the snowpack is the integrated result of radiation from: (1) the earth's atmosphere, (2) clouds, and (3) forest cover. Each of these radiative fluxes will be considered separately in the paragraphs which follow; the combined effect will then be discussed. Considering first the back radiation to the snowpack in an unforested area under conditions of clear skies, it is pointed out that the earth's atmosphere, unlike the snowpack, is not a black or even a gray body*. Rather it absorbs and emits radiation to varying degrees dependent upon the wave length. At certain wave lengths the air absorbs and radiates almost as a black body; at others it is practically transparent to radiation. Only two of the gases of the atmosphere, carbon dioxide and water vapor, have any appreciable effect on absorption in the longwave portion of the spectrum. Since the proportion of carbon dioxide in the atmosphere is practically constant, its effect in absorbing and radiating longwave radiation may be considered as fixed. The over-all effect of carbon dioxide is also less important to radiation exchange, both by virture of its lesser quantity and also by virtue of its lesser absorption bands. Carbon dioxide has a narrow, intense absorption band centered at 14.7μ and extending from 12μ to 16.3μ. The amount of water vapor in the atmosphere, however, exhibits wide variations. As a result it is the controlling variable in the amount of back radiation from the atmosphere with clear skies. The absorption

*A gray body is one which at a given temperature, emits a fixed proportion of the black body radiation at that temperature in all wave lengths.

spectrum for water vapor is quite complex and extensive. For certain wave lengths in the longwave spectrum almost no absorption takes place, while other wave lengths are almost totally absorbed by the existing water vapor in the atmosphere. Other wave lengths are absorbed to varying degrees. Thus there are "windows" through which some of longwave radiation emitted by the pack (and other terrestrial surfaces) may escape to space.

5-04.03 The back radiation to the snow surface from the earth's atmosphere is the result of radiation from all levels of the atmosphere. It is dependent upon the moisture content and temperature distribution of the entire atmosphere. A method has been advanced whereby the temperature and moisture distribution throughout the troposphere may be utilized in determining downward longwave radiation at the earth's surface;9/ however, the method is quite complex and requires that upper air sounding be made. Since the layers of the atmosphere nearest the earth's surface ordinarily have the greatest moisture content and the highest temperatures, they have the greatest influence on the downward longwave radiation. The temperature and moisture content of the upper atmosphere has comparatively little variation, and as a result, its contribution to downward longwave radiation is, like that of carbon dioxide, fairly constant. As a consequence, several investigators have found that estimates of downward longwave radiation from the earth's atmosphere can be had from surface air temperature and vapor pressure alone. For example, Brunt6/ has proposed an equation wherein the ratio of the back radiation from the earth's atmosphere (R_d), to the theoretical black-body radiation computed using surface air temperature (σT_a^4) was correlated with the square root of the surface vapor pressure (e_a). that is,

$$R_d/\sigma T_a^4 = a + b\sqrt{e_a} \qquad (5\text{-}9)$$

Other investigators have determined values of a and b in Brunt's equation for a variety of locations, some of which are given in the table below (as listed by Goss and Brooks17/):

Investigator	Place	a	b	Corr. Coeff.	Range of e_a (mb)
Goss & Brooks	California	0.66	0.039	0.89	4 - 22
Angstrom	California	0.50	0.032	0.30	------
Ramanathan & Desai	India	0.47	0.061	0.92	8 - 18
Eckel	Austria	0.47	0.063	0.89	------
Dines	England	0.52	0.065	0.97	7 - 14
Asklof	Sweden	0.43	0.082	0.83	2 - 8
Angstrom	Algeria	0.48	0.058	0.73	5 - 15
Boutaric	France	0.60	0.042	----	3 - 11
Anderson	Oklahoma	0.68	0.036	0.92	3 - 30

Another form of empirical equation, advanced by Angstrom, 4/ relates the ratio to the vapor pressure of the air, as follows:

$$R_d/\sigma T_a^{4} = a - be^{-k\,e_a} \qquad (5\text{-}10)$$

where values of e is the base for Naperian logarithms and e_a is the vapor pressure in millibars. Values of the constants a, b, and k, are given in the following table (from Anderson 2/):

Investigator	Place	a	b	k
Angstrom	Sweden	0.806	0.236	0.115
Kimball	Virginia	0.80	0.326	0.154
Eckel	Austria	0.71	0.24	0.163
Raman	India	0.79	0.273	0.112
Anderson	Oklahoma	1.107	0.405	0.022

A straight line relationship between the ratio and the vapor pressure of the air has also been derived from measurements made at Lake Hefner.2/ Thus,

$$R_d/\sigma T_a^{4} = a + b\,e_a \qquad (5\text{-}11)$$

For e_a expressed in millibars in the above equation, values of 0.740 and 0.0049 are given for a and b. The differences that exist between the foregoing equations are small, especially when one considers the large scatter that exists in the actual observations. Figure 3 of plate 5-3 shows the variation of the ratio, $R_d/\sigma T_a^4$, with vapor pressure, e_a, as given by equations 5-9, 5-10, and 5-11, using values of the coefficients and exponents given for the Lake Hefner study (Anderson, in the above tables). The values of the ratio, as determined by the Lake Hefner coefficients, are generally greater than those found by the other investigators yet they agree most closely with the constant ratio found for the CSSL, within its limited range (see following paragraph). It will be noted that all three of the equations give quite similar results.

5-04.04 Over extensive snowfields, wide variations of vapor pressure of the air are not ordinarily encountered. The vapor pressure has a strong tendency to remain close to that of the snow surface since the snowpack is both a sink and a source for vapor pressures greater or less than that of the snow. For air over a melting snowpack, the tendency is thus toward a vapor pressure of 6.11 millibars (the saturated vapor pressure at 32°F); the range of observed vapor pressures is usually between 3.0 and 9.0 millibars. Measurements of back radiation made over snow at CSSL, with vapor pressures within this range, indicate the ratio, $R_d/\sigma T_a^4$, to be quite constant and independent of the vapor pressure of the air for this limited range. The value of this constant ratio of

0.757 is also shown in figure 3 of plate 5-3 for the limited range of vapor pressures for which it holds. It will be noted that within this range, the values given by equations 5-9, 5-10, and 5-11 also show little change. The use of a constant ratio is tantamount to making the coefficient b in equations 5-9, 5-10, and 5-11 equal to zero, the constant ratio then being the value of the constant, a.

5-04.05 Net radiation with clear skies. - Using the constant ratio of the preceding paragraph for $R_d/\sigma T_a^4$ (= 0.757), the net longwave radiation exchange over a melting snow surface (R_u = 0.459 ly/min) for clear-weather conditions is given in figure 4, plate 5-3 as a function of air temperature. (Values are also given for ratios equal to 0.80, 0.85 and unity.) From this figure it may be seen that with clear skies the air temperature must exceed 69°F in order for a net gain of longwave radiation by the snowpack to result. This is, of course, strictly true only for conditions of a melting snowpack and for vapor pressures near the saturated vapor pressure for melting snow (6.11 mb). Yet it is generally true for clear weather radiation exchange over the snowpack during the snowmelt season since these conditions usually prevail. Snow surface temperatures are usually not too different from 32°F, and the differences that do occur result in a relatively small change in the emitted longwave radiation. For example, a snow surface temperature of 20°F results in an emitted radiation of 24.9 ly/hr in contrast to the 27.5 ly/hr emitted by the melting snowpack (see fig. 2 plate 5-3); this is 90.5 percent of the black-body radiation at 32°F. Also, over a melting snow surface, the vapor pressure of the air usually remains fairly close to that of the snow surface (i.e., 6.11 mb) as was previously discussed.

5-04.06 Radiation from clouds. - So far the discussion has been restricted to longwave radiation exchange between the snowpack and the atmosphere during clear weather only. In the presence of clouds the foregoing relationships do not hold, since clouds have a dominant effect on longwave radiation. The absorption spectrum for liquid water is quite similar in pattern to that for water vapor; however, the magnitude of the absorption is much greater for liquid water. For example at a wave length of 10 microns, which is in the middle of the transparent band for water vapor and also in the portion of the spectrum where the absorption is least for liquid water, 0.1 mm of liquid water transmitts only 1 percent of the incident radiation. Since even relatively thin clouds contain more precipitable water than this, all clouds are considered to be black bodies with respect to longwave radiation. Under overcast conditions, the net longwave radiation exchange between the snowpack and the atmosphere may be considered to be the net longwave radiation exchange between two black bodies having temperatures corresponding to the snow surface temperature and the cloud base temperature. That is,

$$R = \sigma(T_c^4 - T_s^4) \tag{5-12}$$

where R is the net longwave radiation and T_c denotes cloud base temperature. The net radiation exchange for overcast conditions over a melting snow surface is given by the curve labeled "(black-body)" in figure 4 of plate 5-3. It may thus be seen that in situations where the cloud base temperature is greater than the snow surface temperature, there will be no loss of heat energy from the snowpack by longwave radiation, but rather a net gain will result.

5-04.07 Net longwave radiation loss from the snowpack under conditions of partial cloud cover may be estimated as follows. Angstrom 4/ found that since the net longwave radiation loss from the snowpack is inversely proportional to the amount of cloud cover, the net longwave radiation loss with cloudy skies (R) may be roughly approximated by an equation of the form

$$R = R_c\ (1 - kN) \qquad (5\text{-}13)$$

where R_c is the net longwave radiation loss with clear skies and N is the portion of sky covered by clouds. An average value of k of 0.9 has been suggested by Angstrom; however, the value of the coefficient, k, has been shown to vary with type and height of the clouds among other things. Meinander (as quoted by Geiger 13/) found the following k values for the different cloud types:

Cloud type	k
Low thick clouds (Ac, Sc, Ns, St)	0.76
High thinner clouds (Ac, As, Cs)	0.52
Thin cirrus veils	0.26

Phillips (as quoted by Geiger 13/) found the following values of k as a function of cloud height:

Ceiling		k
km	1000 ft	
1.5	4.92	0.87
2	6.56	0.83
3	9.84	0.74
5	16.40	0.62
8	26.24	0.45

From this it would seem the average k of 0.9 must thus be best applied to low-level clouds. From the above table, the approximate relationship, $k = 1 - 0.024z$, where z is the height of the cloud base in thousands of feet, may be derived (see fig. 5, plate 5-3). Substituting this relationship in equation 5-13, gives the equation,

$$R = R_c \left[1-(1-0.024z)N\right] \qquad (5\text{-}14)$$

which is illustrated in figure 5 of plate 5-3. Using Meinander's k values, the relationship is illustrated in figure 3 of plate 5-6. Also shown in this figure is an average relationship between the degree of cloudiness and the longwave loss ratio as found by Lauscher (as quoted by Hoeck 22/). This latter relationship reflects average conditions wherein the tendency is for lower, thicker clouds to be associated with higher degrees of cloudiness. Obviously these relationships are only approximations of net longwave radiation under cloudy conditions. The value of k in equation 5-13 is dependent upon other factors such as thickness and density, as well as cloud type and height. Moreover, the value of the ratio, R/R_c varies with the distribution of the partial cloud covers, N. The assumed linear relationship with each of these variables (z and N) is not absolutely valid. Then too, the equation represents only average conditions. A normal decrease in cloud base temperature with elevation is implied; however, in any given instance, the actual temperature may vary from the normal. For these reasons, the relationship of figure 5, plate 5-3 (equation 5-14) is of little practical value in estimating back radiation from partially cloudy skies. It is given, rather, to illustrate the relative effects of clouds on the back radiation from the sky. Because of the many complexities involved in its computation, it is necessary that back radiation from the sky be measured if accurate results are to be obtained (see par. 5-04.11).

5-04.08 To summarize the effect of cloud height and amount in the net exchange of longwave radiation over the snowpack, and give a general expression which also holds approximately for clear skies, the relationships of the foregoing paragraphs may be combined into one general equation. Since R_c may be approximated by the equation,

$R_c = 0.757\sigma T_a^4 - \sigma T_s^4$ it follows that,

$$R = \left[0.757\sigma T_a^4 - \sigma T_s^4\right]\left[1-(1-0.024z)N\right] \tag{5-15}$$

Equation 5-15 is a general expression for the net longwave radiation exchange over a melting snowpack for an unforested area. The sign convention followed here is as elsewhere in this chapter: Fluxes of heat energy directed toward the snowpack are considered positive and those directed away, negative. Again the reader is cautioned to remember this relationship is only an illustrative one; practical determination of longwave radiation under conditions of partial cloud cover must be based upon actual measurements. Quite often several layers of clouds are involved, in varying amounts, such that computation of their net effect is practically impossible.

5-04.09 Radiation from forest canopy. - Somewhat analogous to the radiation from clouds is the radiation from the forest canopy. A solid canopy also approximates a black body in the longwave portion of the spectrum, absorbing and emitting all possible radiation. The effective leaf temperature can conveniently be considered to be the same as the ambient air temperature.* Thus the net longwave radiation heat exchange between a solid forest canopy and the snowpack (R_f) may be expressed as

$$R_f = \sigma(T_a^4 - T_s^4) \tag{5-16}$$

since both the snowpack and the tree canopy are, effectively, black bodies, and the effective temperature of the tree leaves is taken as air temperature. For forest conditions other than 100 percent cover, the situation is more complex. Assuming a melting snowpack (σT_s^4 is constant), and a constant value for the ratio $R_d/\sigma T_a^4$, it is possible to illustrate the effect of a varying forest cover in longwave radiation exchange. Since net longwave radiation exchange for the forested areas is given by equation 5-16, and radiation in the open may be expressed as $R_c = (0.757\sigma T_a^4 - \sigma T_s^4)$ these two terms can be weighted in accordance with the amount of forest cover and combined to arrive at the over-all net exchange in the forest. Letting F represent the degree of forest cover (solid canopy equals unity),

$$\begin{aligned} R &= FR_f + (1-F)\ R_c \\ &= \sigma T_a^4 \left[F + (1-F)\ 0.757\right] - \sigma T_s^4 \end{aligned} \tag{5-17}$$

This variation in R with forest cover is illustrated in figure 6 of plate 5-3 for net exchange over a melting snow surface. The foregoing (eq. 5-17) is, of course, an oversimplification of the problem; ramifications are considered in Research Note 12 and in a paper by Miller.31/

5-04.10 The relationship given by figure 6 of plate 5-3 (eq. 5-17) is, as is the case with the relationship for cloud cover, only a relative one used to illustrate the role of forest cover in the flux of longwave radiation. The type of trees, degree of maturity, spacing, etc., all affect the relationship. In addition, the sun's altitude, the wind speed, and degree of cloud cover, affect the approximation of using air temperature as the effective canopy temperature. Still another important source of variation in the above equation is the method by which the degree of forest cover is estimated. This is a quite subjective quantity, and estimates for the same site may vary considerably

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*Actually the leaves heated by direct solar radiation may be somewhat higher than the ambient air temperature; they transfer heat to the air by longwave radiation and by convection to the air passing over them. However, the leaves which face the snowpack and hence have the dominant role in radiative heat transfer between the trees and the snowpack, are generally shaded by the tree crown and hence are very nearly at air temperature. There is a radiative heat transfer that occurs within the foliage itself.

with the method used and even for a given method, with the observer. In view of the complexities, the accurate determination of instantaneous or daily amounts of net longwave radiation exchange is dependent upon measurements. This will be discussed in the paragraph which follows.

5-04.11 Measurement. - In this country most current measurements of longwave radiation are made using Gier-Dunkle Radiometers. These instruments are non-selective absorbers of radiation, that is, they are sensitive to both shortwave and longwave radiation. Two varieties are made: one, a total hemispherical radiometer, measures the total hemispherical irradiation upon a plane surface; the other, a net-exchange radiometer, measures the net radiative heat transfer across the plane of the meter surface. Reference is made to papers by Gier and Dunkle 16/ and by Dunkle and others 8/ for a description of the construction and operation of these instruments. Basically they consist of a 4-inch-square flat plate which serves as a heat-flow meter. Both surfaces of the plate are blackened to absorb, non-selectively, a high percentage of the radiation incident upon them. A silver-constantan thermopile inside the plate has its alternate junctions in the upper and lower surfaces and thus produces an emf portional to the temperature difference between the surfaces and hence to the difference in irradiation falling upon the surfaces. The flat plate is mounted in an air blast from a blower which keeps the convective losses from the two surfaces approximately equal. Thus the net radiometer measures the difference in radiation falling on the two surfaces. The total hemispherical radiometer has one of its plates shielded by a highly reflecting plate which is also located in the air blast and thus remains at about the same temperature as the shielded surface. Since these meters measure both shortwave and longwave radiation, they are ideal for the measurement of total radiative heat transfer between the snowpack and its environment. Thus the net-exchange radiometer integrates into a single measurement all radiative fluxes to and from the snowpack. For analytical purposes, however, it is sometimes desirable to determine the longwave component separately. One method of doing this is to make the measurements at night when shortwave radiation is non-existent. This method has commonly been employed in the past, which accounts for the fact that longwave radiation is often referred to as "nocturnal radiation." Nighttime measurements, however, usually restrict the range of air temperatures to the lower values; during the daytime, air temperatures, and hence the back radiation, are usually higher than at night. In order to determine the back radiation during daylight hours it is necessary that measurements be made of the incident shortwave radiation (see para. 5-03.09) simultaneously with the allwave radiation measurements; this former quantity can then be subtracted from the latter to arrive at the longwave component. This method, however, gives rise to considerable error. Since the longwave component is but a small difference between two much larger quantities, any small errors in either of them is magnified many-fold in the final result. Because of this source of error, other investigations, for example the Lake Hefner study, 2/ have led to the conclusion that the measurement of longwave radiation during the day is impractical.

Nevertheless, the nocturnal measurements still offer a reliable means of determining longwave radiation, and the radiometers are well suited to the determination of the total radiative heat transfer, day or night. Measurements of longwave radiation exchange by Gier-Dunkle radiometers were made at CSSL during several years of its operation. Several studies have been made of the data collected. Reference is made to Research Notes 6, 7, and 11 and Technical Bulletin 12. Another study of net longwave radiation exchange is given in Technical Bulletin 7.

5-05. RADIATION SUMMARY

5-05.01 General. - So far the shortwave and longwave components or the total radiative flux have been considered separately. The effects of atmospheric water vapor, clouds, forest cover, etc., on each were individually discussed. In this section the effects of those factors on the combined allwave radiative flux will be considered. Illustrative examples are given for conditions as they exist at CSSL (approx. 40° N and 7200 ft msl) and for the north slopes of the Alps (approx. 46° N and 5200 ft msl). For the Alps, the variation of radiation throughout the year is shown. For CSSL example two situations will be examined: one for the winter and the other for the spring melt season. The winter situation is included to illustrate the importance of the seasonal change in daily insolation amounts in the melting of the snowpack and the dominant effect of radiative heat transfer in controlling snowmelt. February 15 was selected to typify winter conditions and May 20 to typify spring melt conditions. On those dates the daily amounts of insolation received with clear skies at CSSL are 400 langleys and 800 langleys, respectively. This marked change in insolation coupled with the usual change in albedo of the snow and the air temperatures from winter to spring results in a great increase in radiation melt from winter to spring as will be shown. In the discussions which follow, it is assumed that 200 langleys result in one inch of snowmelt (thermal quality of 98.4 percent). During the winter the thermal quality would usually be greater than this, resulting in even less melt than indicated, while during the spring lesser thermal qualities and greater melts would be the rule. Reference is made to SIPRE Research Paper 8 15/ for a nomograph which expresses total radiation heat supply (both shortwave and longwave) to the snowpack. Parameters of time of year, latitude, sky condition, and albedo of the snow are involved.

5-05.02 Clear-weather melt. - During clear weather, the important variables in radiation melt are: (1) the insolation, (2) the albedo of the snow, and (3) the temperature of the air. Humidity of the air also affects the radiation melt; however, its effect is relatively minor. Figures 1(a) and 1(b) of plate 5-4 are illustrative of daily clear-weather radiation melt amounts for the winter and spring conditions. Back radiation to the snowpack is estimated using the constant ratio, 0.757, for $R_d/\sigma T_a^4$, which is generally applicable for vapor pressures in the vicinity of 6 millibars. In computing longwave radiation emitted by the snowpack, the snowpack was assumed to remain at 32°F. These conditions hold quite well for the spring melt situation, but during the winter lower vapor pressures and snow surface temperatures would be expected. Hence, both the back radiation to the snowpack and the longwave radiation emitted by the snowpack would be less than assumed. Consequently, these assumptions tend to cancel one another, making the net longwave radiation loss from the snowpack during the winter reasonably correct. In figure 1(b), "negative melts" are shown as dashed lines, since the values given in this portion of the figure are dependent upon the snow surface being at a temperature of 32°F, and the indicated loss of heat would indicate this not to be the situation. This section of the figure is strictly applicable only if convection-condensation amounts of heat transfer are sufficient to make up the indicated deficit.

5-05.03 Typical albedoes of the snow during the spring may be taken as 50 percent, and a typical mean daily temperature of 50°F. During the winter, a higher albedo of, say 75 percent, and a mean daily temperature of 30°F are more representative of actual conditions. The points fulfilling these conditions are indicated on the figures. Thus, for clear weather, a typical springtime daily radiation melt of about 1.6 inches is indicated, while during the winter a heat deficit is indicated. (This heat deficit may be partially or wholly made up by convection-condensation.) These conditions of albedo and temperature for winter and spring are assumed in the following paragraphs where the effects of clouds and trees on radiative melt are considered.

5-05.04 Effect of clouds. - The dominant role of clouds in the fluxes of both shortwave and longwave radiation over snow has already been discussed. Since clouds are such a powerful controlling factor in radiative heat exchange, other minor factors such as humidity of the air are often ignored in the estimation of radiative heat exchange on cloudy days. Figures 2(a) and 2(b) of plate 5-4 illustrate the effect of clouds on daily radiation melt during the spring and winter. These figures are simply a combination of shortwave and longwave radiation exchange corresponding to given cloud heights and amounts given by figure 5 of plate 5-1 and by figure 5 of plate 5-3; the amounts of shortwave and longwave radiation with clear skies are as previously mentioned. It will be noted that during the winter the effect of clouds on radiative heat exchange is relatively less than during the spring, due to the lesser radiation melts during the winter time. Also, during the winter, radiation melt tends to increase with increasing cloud cover and lower cloud heights in consequence of the more important role played by longwave radiation during this time of the year.

5-05.05 Effect of forest canopy. - Similar to the effect of clouds, the forest canopy exerts a powerful controlling influence on net allwave radiation exchange between the snowpack and its environment. However, its effect is different from that of clouds, particularly with respect to shortwave radiation. While both the clouds and trees restrict the transmission of insolation, clouds are highly reflective, while the forect canopy absorbs much of the insolation. As a result, the forest canopy tends to be warmed and in turn gives up a portion of the incident energy to the snow. (It radiates, in the longwave portion of the spectrum, directly to the snowpack and also warms the air by convection and radiation which in turn gives up some of its heat to the snowpack.) Clouds, on the other hand, reflect back to space a large portion of the incident radiation, which is thus lost. The role of the forest canopy in radiative heat exchange between the snowpack and its environment is illustrated in figures 3(a) and 3(b) of plate 5-4. These figures represent the typical spring and winter snowmelt conditions in the middle latitudes, as previously specified, under a coniferous forest cover. It shows the variation of the several radiative components and the net allwave radiation with degree of forest cover during clear weather. The curves of these figures are based on the transmission coefficients for insolation given by figure 1 of plate 5-2 and on the net longwave radiation exchange in the forest given by figure 6 of plate 5-3. It may be seen that for the conditions specified, during the spring the maximum radiation melt occurs in the open and the minimum with a canopy density of about 50 percent. During the winter the maximum melt (for the conditions specified) is with 100 percent canopy cover and the minimum with about 20 percent cover. These curves, of course, are merely for average conditions of canopy cover and do not reflect variations in the spacing of trees or take into account clearings in the forest canopy. These effects are discussed in section 5-12 of this chapter.

5-05.06 Effect of slopes. - The effect of slope on allwave radiation exchange over a barren snowfield can be illustrated by an analysis made by Hoeck 22/ for the Alps. This analysis carries on the investigation of the same 25-degree slope gradients previously considered in paragraph 5-03.07 for shortwave radiation alone. It applies to a latitude of 46° 30' N and to an elevation of about 5200 feet msl. Conditions of air temperature and humidity assumed are mean values for the north slopes of the Alps. The effects of variable cloudiness and albedo, as well as the effect of slope, are included in the relationships. These relationships are illustrated in figure 1, of plate 5-6. Figure 1(a) gives the relationship for horizontal surfaces and for east- and west-facing slopes which are the same as a horizontal surface. Figure 1(b) shows the relationship for a south-facing slope of 25-degree gradient and figure 1(c) for a north-facing slope of 25-degree gradient. The two dates used in the foregoing examples for CSSL (plate 5-4) are indicated on these figures for comparison.

5-06. THEORY OF TURBULENT EXCHANGE

5-06.01 General. - Of secondary importance to radiation in the transmission of heat to the snowpack is the process of turbulent exchange in the overlying air. With a downward temperature gradient there is a direct transfer of heat from the air to the snow, and with a downward vapor pressure gradient there is a direct transfer of moisture from the air onto the snow surface, releasing, in addition, its latent heat of vaporization. The reverse processes occur as well; it is the net effect that is of concern. During periods of melting, the temperature of the snow surface remains at 32°F and the corresponding vapor pressure is 6.11 mb. If the air temperature and the vapor pressure immediately above the surface (within the laminar layer--a fraction of an inch in thickness) were known, the flow of heat and moisture to the snow and the resulting melt could be easily computed, using the thermal conduction equation. Such refined measurements, however, are unobtainable with available instruments, and the practical alternative has been to establish the relationship of observed melts to air temperatures and vapor pressures measured at some higher level. These relationships have been determined experimentally for measurements at several sites and for the particular measurement conditions prevailing at the time of the experiment. In order to generalize the equations thus derived to make them hydrologically applicable to other sites and other conditions of measurement, it is of value to consider what is known of the processes of turbulent exchange, since the usual heights of measurement of air temperature and air moisture content are within the region of turbulence, where the vertical distribution of temperature, water vapor and wind speed is governed by the action of eddies.

5-06.02 Basic equation. - The basic equation for turbulent exchange is,

$$Q = A\, dq/dz \qquad (5\text{-}18a)$$

where Q is the flow of some property of the air through a unit horizontal area per unit time, dq/dz is the vertical gradient of this property (q = the property, such as temperature or water vapor; z = height), and A is the exchange or "Austausch" coefficient which will be discussed later. The property q may be any property which is itself unaffected by vertical transport. Since the interest here is in the exchange of heat and moisture to the snow, only the properties of air temperature* and moisture content (often expressed in various ways, as specific humidity, vapor pressure, etc.) and wind speed shall be considered.

5-06.03 Derivation of practical equation. - To put the basic equation in a form usable for hydrologists, it is necessary to know the variation of air temperature, humidity and wind speed with height in the zone of turbulent mixing. The matter of vertical gradients under a variety of conditions of stability of the air has been extensively investigated (Sverdrup 34/, Johns Hopkins Publications in Climatology 36/) and the profiles of these properties during conditions of atmospheric stability, such as prevail over snowfields, have been shown to follow a

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*Strictly speaking, potential temperatures should be used, as the equation is applicable only to properties of the air which do not change with vertical motion. If, however, temperatures are measured within ten feet of the snowpack, as they usually are, the error in using measured air temperatures is negligible for practical purposes, amounting to less than 0.05°F for the movement of the air between the level of measurement and the snow surface.

power law distribution. (A logarithmic profile has been found to more adequately represent the distribution under neutral or unstable conditions, but it is disregarded in this presentation as being uncharacteristic of conditions over snowfields.) According to a power law, the ratio of values of some property of the air, q, at two heights above the snow surface is equal to some power of the ratio of the heights, z, themselves. It is expressed by the relationship

$$q_2 / q_1 = (z_2 / z_1)^{1/n} \tag{5-19a}$$

If z_1 is unity, then

$$q = q_1 z^{1/n} \tag{5-19b}$$

where q is the value of the property at level z and q_1 the value at unity level. (The values of the property q in the above equations represent the differences between the values measured at their respective heights and the values of the properties at the snow surface. For wind speed and for temperatures measured in degrees C, the snow surface value is ordinarily zero, and the measured values of these elements may be used directly in the equations. For temperatures measured in degrees F and for vapor pressures, the snow surface values must be subtracted from the measured values to make the relationship valid.) Differentiating equation 5-19b,

$$dq/ dz = (q_1 / n) z^{(1-n)/n} \tag{5-20}$$

and substituting in equation 5-18a,

$$Q = A (q_1/n) z^{(1-n)/n} \tag{5-18b}$$

which expresses the eddy exchange of the property at any level z from measurements of the property made at unity level. The exchange coefficient, A, also applies to the level z. The value of the exchange coefficient varies with height as is discussed in the following paragraph.

5-06.04 Under equilibrium conditions, the gradients of temperature and moisture assumed by the air above the snow are such that the eddy transfer of heat and moisture are constant with height up to normal heights of measurement. Consequently, in equation 5-18a, the exchange coefficient must vary inversely with the gradient, dq/dz. That is,

$$A / A_1 = (dq/dz)_1 / (dq/dz) \tag{5-21a}$$

and, from equation 5-20,

$$A = A_1 z^{(n-1)/n} \tag{5-21b}$$

Where A_1 is the exchange coefficient for the 1-foot level. Substituting this value of A in equation 5-18b,

$$Q = A_1 \, q_1 \, / \, n \qquad (5\text{-}18c)$$

Sverdrup 34/ has shown that the exchange coefficient at a given level is directly proportional to the wind speed at that level. Thus,

$$A_1 = k \, v_1 \qquad (5\text{-}22)$$

where v is the wind speed and the subscripts indicate unity level; k is a proportionality constant. Substituting this value of A_1 in equation 5-18c,

$$Q = (k/n) \, q_1 \, v_1 \qquad (5\text{-}18d)$$

Using the power law (equation 5-19b) to express the variation of the property q and the wind speed v with height, equation 5-18d may be given as,

$$Q = (k/n) \quad (z_a z_b)^{-1/n} \, q_a v_b \qquad (5\text{-}18e)$$

where the subscripts a and b are used to identify the levels of measurement of the property and the wind speed respectively, since the two may be different.

5-06.05 Condensation melt. - Considering now the specific case of the exchange of moisture by eddy diffusion. Since the exchange coefficient expresses the mass (of air) exchanged per unit time and area, it is necessary to know how much of the property being considered --here water vapor--is contained in unit mass of air. Thus specific humidity (mass of water vapor per unit mass of air) must be used to express the humidity gradient. Since the specific humidity is approximately given by the expression, $(0.622/p)e$, where p is the atmospheric pressure and e is the vapor pressure of the air, values of vapor pressure may be substituted for specific humidity. Thus the moisture transfer, Q_e, is given by the equation,

$$Q_e = (k/n) \, (z_a z_b)^{-1/n} (0.622/p) \, e_a v_b \qquad (5\text{-}23a)$$

Equation 5-23a thus gives the moisture transfer and hence the amount of condensate given up by the air to the snow surface. (In the above equation, e_a represents the difference in vapor pressure between the air at level z_a and the vapor pressure of the snow surface. If the vapor pressure of the air is greater than that of the snow surface, e_a is positive and condensation results. If the vapor pressure of the air is less than that of the snow surface, e_a is negative and evaporation occurs. In what follows, where condensation ~~is~~ discussed, evaporation is considered simply as negative condensation.) In addition to the condensate itself, for every gram of water condensed on the snow surface, approximately 600 calories of heat energy are released (latent heat of vaporization). This is sufficient to melt 7.5 grams of snow having a thermal quality of 100 percent (ratio latent heat of vaporization to latent heat of fusion of water equals 600/80). Adding this melt to the condensate gives a total of 8.5 grams of melt-plus-condensate. Equation 5-23a thus gives,

$$M_e = 8.5\ (k/n)\ (z_a z_b)^{-1/n} (0.622/p)\ e_a v_b \qquad (5\text{-}23b)$$

where $\underline{M_e}$ represents the total melt plus condensate.

5-06.06 Convection melt. - In the case of heat transfer by eddy exchange, $\underline{H_c}$, the measured property of the air used in equation 5-18e becomes air temperature; the specific heat of the air, $\underline{c_p}$, must also be included in the equation to properly convert the temperature measurements into heat units. That is,

$$H_c = (k/n)\ (z_a z_b)^{-1/n}\ c_p\ T_a\ v_b \qquad (5\text{-}24a)$$

In cgs units, the equivalent snowmelt is equal to the net heat exchange divided by 80 (for pure ice at 32°F).

$$M_c = (1/80)\ (k/n)\ (z_a z_b)^{-1/n}\ c_p\ T_a\ v_b \qquad (5\text{-}24b)$$

where $\underline{M_c}$ is the resultant melt in grams.

5-06.07 Elevation effect. - The coefficient $\underline{k}$ in the foregoing equations is a complex function itself. Among other things, its value is dependent upon the density of the air. Since density of the air varies with elevation, so does the value of $\underline{k}$, and hence it is not a constant for all locations. In order to make this coefficient independent of air density, and hence elevation, another term giving the variation of density with elevation, may be included in the equation, the sea-level density of the air being implicitly included in the value of the coefficient. Thus the term, p/p_o, where $\underline{p}$ is the atmospheric pressure at the elevation of the site and p_o is sea-level pressure, may be included in the equation to represent the variation in density of the air with elevation. Including this term in equations 5-23b and 5-24b,

$$M_e = 8.5\ (k'/n)(z_a z_b)^{-1/n} (0.622/p_o)\ e_a v_b \qquad (5\text{-}23c)$$

$$M_c = 1/80\ (k'/n)\ (z_a z_b)^{-1/n} (p/p_o)\ c_p\ T_a\ v_b \qquad (5\text{-}24c)$$

where $\underline{k'}$ is now a constant applicable to sea level pressure ($p = p_o$).

5-06.08 Combined equation. - Since many of the terms of equations 5-23c and 5-24c are common to both, the two equations may conveniently be combined into a single expression of convection-condensation melt, $\underline{M_{ce}}$. Thus

$$M_{ce} = \frac{k'}{n} (z_a z_b)^{-1/n} \left(\frac{1}{80} \frac{p}{p_o} c_p T_a + 8.5 \frac{0.622}{p_o} e_a \right) v_b \qquad (5\text{-}25)$$

This is the theoretical expression for the snowmelt resulting from eddy transfer of heat and moisture to the snowpack. It is based on the concept that the exchange coefficients for both the exchange of heat and moisture are the same. In the sections which follow, the values of the coefficients for convection and condensation melt will be arrived at separately by experimental means.

5-07. CONDENSATION AND EVAPORATION

5-07.01 General. - Numerous investigations have been made of actual amounts of water vapor condensed upon or evaporated from exposed water surfaces. Most studies have been primarily interested in water losses due to evaporation. 2/ 35/ In snow hydrology, in addition to the amounts of water evaporated or condensed, the latent heat of vaporization involved in the change of state from gas to liquid or vice versa is also of concern. In the case of water surfaces, such heat may be absorbed or given up without immediate hydrologic effect; on snow surfaces this heat is of considerable hydrologic significance, for it is capable of producing a more than sevenfold increase (or decrease in the case of evaporation) in the amount of water available for runoff over that actually condensed (or evaporated). In what follows, the discussion shall be concerned with condensation, it being understood that evaporation is merely the negative case. The experimental methods are similar in either case. The condensation or evaporation amounts are determined volumetrically or by weighings. Simultaneously, measurements of vapor pressure and wind speeds are made. The equation which is used to relate vapor pressure and wind speed to the amounts of condensate is of the form,

$$q_e = k_e (e_a - e_s) v_b \qquad (5\text{-}26)$$

where q_e is amount of condensate, e_a and e_s are the vapor pressure of the air and snow surface respectively, v_b is the wind speed, and k_e is a coefficient relating the two.

5-07.02 Condensation over snow. - Detailed experiments of condensation and evaporation over snow have been made by several investigations. Among these are studies made at CSSL (see Res. Note 25) and by deQuervain at the Weissfluhjoch Institute in the Swiss Alps. 7/ The method used has been to place pans filled with snow into the snowpack so that the surface of the snow in the pans is flush with the surrounding snow surface, and to note the moisture gain or loss in the pans by periodic weighings. It is to be remembered, however, that where snow is involved, for every unit of water vapor condensed, additional heat of vaporation is released, capable of melting 7.5 times this amount of snow. Thus to represent the condensate plus its accompanying melt, assuming this latent heat is completely effective, the constant k_e above must be multiplied by 8.5 (1 + 7.5), and the condensation melt equation for any particular site becomes

$$M_e = 8.5\, k_e \; (e_a - e_s)\, v_b \tag{5-27a}$$

5-07.3 Generalization of equation. - Since the vapor pressure of the air and the wind speed vary with height above the snow surface, the value of the coefficient k_e is dependent upon the height of measurement of these meteorological elements. By assuming the power law variation of vapor pressure and wind speed with height, discussed previously (par. 5-06.03), it is possible to generalize equation 5-27a so that the coefficient k_e is constant regardless of heights of measurement and is applicable to any site over open snow. The value of the exponent n in the power law (equations 5-19) has been the subject of considerable investigation. A good summary is found in Sutton's Micrometeorology 33/ where a spread of values is reported on from a variety of conditions of stability of air. Over snow, however, an inversion generally exists and values of n are fairly constant. Power law exponents derived by Walsh* as a best fit to wind speed gradients at CSSL range from 4.8 to 7.1, with an average value of 5.8. Sverdrup found by observations on Isachsen's Plateau 34/, that an n of 5.6 applied to the vertical distributions of wind speed, air temperature and vapor pressure. Recent experimentation by the Snow Investigations tends to confirm a value of $n = 6$ as an adequate representation of the vertical distribution of these meteorological parameters above a melting snow field in an open, unforested site. Thus

$$e_a - e_s = (e_1 - e_s)\, z_a^{1/6}$$

and

$$v_b = v_1 z_b^{1/6}$$

where the subscript 1 refers to the various properties measured at a standard reference level of one foot (or other unit). Substituting these relationships in equation 5-27a,

$$M_e = 8.5k_e\, (z_a z_b)^{1/6}\, (e_1 - e_s)\, v_1 \tag{5-27b}$$

Letting $k_e' = 8.5\, k_e\, (z_a z_b)^{1/6}$, where k_e' is now a constant (the value of the coefficient when wind speed and vapor pressure are measured one foot above the snow surface),

$$M_e = k_e' \; (e_1 - e_s)\, v_1 \tag{5-27c}$$

Moreover, since

$$e_1 - e_s = (e_a - e_s)\, z_a^{-1/6}$$

* See Miscellaneous Report 6.

and

$$v_1 = v_b z_b^{-1/6}$$

vapor pressure and wind speed measured at any levels (z_a and z_b) can be reduced to their equivalent values at the one-foot level, and

$$M_e = k_e' \ (z_a z_b)^{-1/6} \ (e_a - e_s) \ v_b \qquad (5\text{-}27d)$$

where k_e' is a constant relating condensation melt (plus condensate) to the condensation parameter for values of wind speed and vapor pressure measured at the one-foot level.

5-07.04 Evaluation of constant. - The value of k_e in equation 5-26 has been evaluated by relating measurements of moisture exchange, q_e, to simultaneous products of vapor-pressure gradient and the first power of the wind speed, $(e_a - e_s) \ v_b$, by a linear least-squares regression. From this, the coefficient k_e' of equation (5-27d) was then evaluated by substituting actual heights of measurements of wind speed and vapor pressure. Values of the constant, k_e' so determined (for daily melts in inches, for vapor pressure measured in millibars and wind speed in miles per hour, heights of measurement in feet) are as follows:

Central Sierra Snow Laboratory (Res. Note 25)	0.0540
Weissfluhjoch Institute (deQuervain)	0.0770

Figure 2, plate 5-5 illustrates this relationship, giving daily melts in terms of mean daily wind speed and vapor pressure at one-foot level, using the CSSL coefficient. (Melts for other heights of measurement may be determined from figure 2 by use of figure 5, plate 5-5.)

5-08. CONVECTIVE HEAT TRANSFER FROM THE AIR

5-08.01 Unlike radiative heat transfer and the transfer of moisture between the air and snow surface, convective heat transfer from the air to the snow surface cannot be measured and has to be evaluated indirectly. This has been done by considering convection melt as a residual in the general snowmelt equation. Since the total melt from experimental areas is measurable, and the melts from the processes of radiation and condensation can be computed as previously explained, the difference between the total melt and these two computed melts is considered due to convective heat transfer. By relating quantities so determined to the meteorological parameters of air temperature and wind speed by the equation,

$$M_c = k_c \ (T_a - T_s) \ v_b \qquad (5\text{-}28a)$$

where $\underline{M_c}$ is the convective melt quantity, $\underline{T_a}$ and $\underline{T_s}$ are the temperatures of the air and snow surface, and v_b the wind speed, the convective melt coefficient, k_c, can be determined. This was done at CSSL using data from a special snow lysimeter constructed expressly for this purpose (see Res. Note 25). The convective melt coefficient obtained reflects the conditions peculiar to that site and the particular instrumentation involved. By a process similar to that described in paragraph 5-07.03, the foregoing equation can be generalized for application to any heights of measurement and to any site over open snow. Thus

$$M_c = k_c' \ (p/p_o)(z_a z_b)^{-1/6} \ (T_a - T_s) \ v_b \qquad (5\text{-}28b)$$

where $k_c' \ (p/p_o)(z_a z_b)^{-1/6}$ equals $\underline{k_c}$ in equation 5-28a. The term $\underline{p/p_o}$ (where $\underline{p}$ is atmospheric pressure at the observation level, and $\underline{p_o}$ is sea level pressure) is introduced to correct for the variation of air density with elevation, since the coefficient $\underline{k_c}$ is directly proportional to the air density. Values of $\underline{p/p_o}$ as a function of elevation are obtainable from figure 6 of plate 5-5. Values of $z^{-1/6}$ for various heights of measurement are obtainable from figure 5 of plate 5-5. The value of the convection coefficient, k_c', of equation 5-28b was determined for an open site from the CSSL experiments to be 0.00629 for daily (24 hour) melt rates expressed in inches, temperatures measured in degrees F, wind speeds in miles per hour, and heights of measurement in feet. Figure 1, plate 5-5 shows this relationship for measurements of air temperature and wind speed at one-foot level. (Melts for other heights of measurement may be determined from figure 1 by use of figure 5, plate 5-5.)

5-09. SUMMARY OF CONVECTION-CONDENSATION MELTS

5-09.01 Comparison with other investigations. - In present use by snow hydrologists in the computation of condensation or convection melts are the results of experiments and the melt equations presented by other investigators, notably Sverdrup 34/ and deQuervain 7/. Since the form of their equations is similar to those given herein, the coefficients for convection and condensation melt may be compared as follows:

Investigator	Condensation coefficient (k_e')	Convection coefficient (k_c')	Ratio: k_c'/k_e'
Snow Investigations (1954)	0.0540	0.00629	0.12
Sverdrup (1936)	0.0674	0.0215	0.32
deQuervain (1951)	0.0770		

Light, 28/ who used a theoretical equation advanced by Sverdrup, arrived at coefficients essentially the same as those of Sverdrup's given above.

All coefficients above are for wind speeds, air temperatures, and vapor pressures measured at the one-foot level; they are for temperatures measured in degrees F, vapor pressures measured in millibars, and wind speeds in miles per hour; they are for melts expressed in inches per day (24 hours).

5-09.02 As may be seen, the condensation coefficients of the other investigators given above are of the same order of magnitude as those of Snow Investigations. The convection coefficients, however, are more than three times as great as those determined by the Snow Investigations. Sverdrup's coefficients were derived from experiments over snow on Isachsen's Plateau, in which ablation measurements furnished total melt data, from which was subtracted computed radiation melt, leaving a residual melt assigned to combined condensation and convection.* Sverdrup, being unable, in the absence of measured condensation amounts, to separate condensation-convection melt except on a theoretical basis, assumed that the exchange coefficients of moisture and heat are the same (as described in section 5-06). This assumption is becoming less satisfactory with accumulating experimental evidence. 3/ 18/ A further check on the condensation constant comes from deQuervain's investigations at Weissfluhjoch Institute by experimental methods similar to those of Snow Investigations. The ratios, k_c'/k_e', given above afford a means of comparing the relative magnitudes of the convection and condensation melts as found by the several investigations. The value of this ratio, as given by Bowen 5/, ranged from 0.32 to 0.37, the smaller value of the ratio being associated with stable atmospheric conditions (such as are found over snow fields). Thus Bowen's ratio agrees with that of Sverdrup. In summary, theory accords reasonably with experiment in considerations of moisture exchange. For convective heat exchange there is notable discrepancy. The smaller exchange coefficient indicated by Snow Investigations is recommended in view of the more detailed measurements of meteorological parameters (including radiation) available at present.

5-09.03 Combined equation. - Many of the terms of equations 5-27d and 5-28b are common to both, and the two equations may conveniently be combined into one equation of convection-condensation melt. Using the coefficients derived at the CSSL, this equation is as follows:

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*Light's equation is theoretical, based on one advanced by Sverdrup which assumes a logarithmic variation of air properties with height, in contrast to the power law variation cited previously and also used in Sverdrup's experimental work. Light's equation, in application, employs a basin factor of less than unity, by which the theoretical equation must be multiplied to reproduce measured basinwide melt amounts. (Radiation melts are not considered). It would be reasonable to expect this factor to be greater than unity, since it must include the not inconsiderable melt from radiation. Its empirical value of approximately 0.65 would seem to indicate that the coefficients of the theoretical equation are too large.

$$M_{ce} = (z_a z_b)^{-1/6} \left[0.00629\ (T_a-T_s)(p/p_o) + 0.0540\ (e_a-e_s)\right] v_b \quad (5\text{-}29a)$$

or

$$M_{ce} = 0.00629\ (z_a z_b)^{-1/6} \left[(T_a-T_s)(p/p_o) + 8.59\ (e_a-e_s)\right] v_b \quad (5\text{-}29b)$$

The above equation is for daily (24-Hour) melt in inches where the temperatures are in degrees F, vapor pressures in mb, and wind speed in mph and height in feet. They are for melt from ripe snow packs and apply strictly only to unforested sites. This combined relationship is given graphically in figures 3 and 4 of plate 5-5. Figure 4 expresses the total melt in terms of temperature and relative humidity.

5-09.04 Discussion. - The foregoing coefficients for convection and condensation melt were derived from measurements made at a point in the open. To apply these results to other areas, the effects of terrain on air temperature, vapor pressure and wind speed must be considered. In forested areas wind speeds are less than in the open; hence if convection-condensation melt in the forest is to be determined, based on measurements of wind speed made in an open area, coefficients smaller than those given herein would certainly be expected. Moreover, the power law variation of wind speed with height applies to open areas only. In the forest no such simple relationship of wind speed with height applies. Since each forested area presents a different problem due to differences in type and spacing of trees, topography, etc., no general relationships can be given. Areal variations in air temperature and vapor pressure are less than those of wind speed but are also of some consequence. In the computation of basinwide melts due to convection and condensation, the variation of both air temperature and vapor pressure with elevation is an important consideration. It is to be pointed out that, unlike solar radiation, the temperature and the vapor pressure of the air are not independent of the resultant melt they produce. For a given airmass, an increase in wind speed tends to produce more convection-condensation melt but at the same time tends to reduce the temperature and vapor pressure of the air near the snow surface. Thus there is a tendency for the value of the measured gradients to vary inversely with the wind speed. In general, the relationships given in this section should not be extrapolated to values outside the range of those indicated in the several figures which illustrate them.

5-10. CONDUCTION OF HEAT FROM THE GROUND

5-10.01 General. - In the preceding section the major heat fluxes to the snowpack have been considered. Ordinarily these heat fluxes--radiation and convection-condensation--are all that need be considered in the determination of daily melt quantities. Yet there is still another heat flux which, although negligible in daily computations of melt, becomes significant when the melt season as a whole is considered. This flux is the conduction of heat upward to the snowpack from the underlying ground. This source of heat has special hydrologic significance since it can cause melting during the winter and early spring when melt at the snow surface is non-existent. Thus the melt due to this cause is capable of priming the underlying soil in advance of the actual melt season, and may also help to ripen the snowpack, readying it for melt.

5-10.02 Ground-temperature gradients. - The flux of heat upward from the underlying ground to the snowpack during the winter and spring months results from thermal energy that is stored in the ground during the summer and early fall when no snow cover exists. During the summer months, the ground surface is heated, primarily by solar radiation and as a result the thermal gradient is directed into the ground. In consequence of this thermal gradient, heat is conducted downward into the ground, the amount being dependent upon the thermal gradient and the conductivity of the ground itself. Thus the ground may be warmed to a considerable depth. During the winter, with snow on the ground, this process is reversed. The ground surface is cooled to 32°F (or below) and the thermal gradient is directed upward. Figure 5 of plate 5-6 shows the annual variation in soil temperature at several depths for an area having a shallow winter accumulation of snow. This figure was prepared from soil temperature data secured in Minnesota by Algren. 1/ In figure 4, plate 5-6, these data are plotted to show vertical temperature gradients through the soil profile. Mean monthly values are given for the months September through April. Since snow is a good insulator,* the ground is shielded from the sub-freezing air temperatures of winter in the case of deep snowpacks. Before the deep pack is established, however, the ground may become frozen to some depth. Ground that may have been frozen before the deposition of the permanent snowpack will generally be thawed by the conduction of heat from greater depths once the protective snow cover is deposited. It will be noted in figure 4 that during brief transition periods in the spring and the fall the situation exists in which there is a dual temperature gradient in the ground. During the fall, the ground surface may be suddenly cooled by early snow fall and (or) cold air temperatures, resulting in an upward gradient near the surface, while deeper in the ground the gradient is still directed downward. Heat flows in both directions from the depth of maximum temperature, further warming deeper layers and conducting heat to the surface of the snowpack. In the spring, the gradients are directed in the opposite directions. The ground surface becomes warmed as the snow cover disappears and heat flows downward; at the same time heat is being conducted upward from greater depths to some intermediate area of minimum temperature. Typical ground temperature gradients determined from data secured at CSSL (see Tech. Bull. 16) for the months January through April are given in the following table:

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* See chapter 8.

Mean Ground Temperature Gradients, CSSL
(in ^{o}F per foot)

Month	Three feet to surface	One foot to surface
January	1.6	2.1
February	1.4	1.9
March	1.3	1.6
April	0.9	1.2

These data are for silty clay loam such as is found in the meadows at CSSL where the data were obtained. They are an average of five years of record. Data for earlier and later months are not given because of the transition periods previously discussed.

5-10.03 Thermal conductivity of the ground. - The quantity of heat transported by the conduction process across unit area of a plane parallel to the ground surface during unit time is given by the equation,

$$H_g = k\ dT/dz \tag{5-30}$$

where dT/dz is the temperature gradient in a direction perpendicular to the plane, and k is a proportionality factor known as the thermal conductivity. For soils, the thermal conductivity varies with the composition, the density, and also with the moisture content; generally speaking the thermal conductivity of a soil varies directly with its density and its moisture content. The heat flux by conduction is the result of both the temperature gradient and the thermal conductivity; however, these two terms are not mutually exclusive. The better the thermal conductivity of the soil, the less the temperature gradient, other things being equal, since the better conduction tends to equalize temperature differences. On the other hand, it is possible to maintain quite steep temperature gradients in a substance having a low thermal conductivity. Some typical values of thermal conductivities are given in the table below for a silty clay loam.

Thermal Conductivity for Unfrozen Fairbanks Silty Clay Loam

(after Kersten 25/)

Mositure Content	Density in lbs per cubic ft.		
	80	90	100
2.5%	0.0005		
18.0%	0.0019	0.0025	0.0031
25.0%	0.0022	0.0028	0.0036*
30.0%	0.0025	0.0029*	
40.0%	0.0027*		

*Saturated soil.

Reference is made to the report by Kersten 25/ for an extensive tabulation of thermal conductivities of many different soils.

5-10.04 Observed melt quantities. - Data on the melt occurring at the bottom of the snowpack as a result of the conduction of heat from the ground are difficult to obtain and are not ordinarily available. Special measurements were made, however, during one year at CSSL from which this melt could be computed (see Tech. Bull. 16). Measurements of the compaction and melt of the bottom layers of the snowpack were determined by means of a slide wire settling meter. Densities of these same layers were determined from deep pit measurements. From these data the change in water equivalent of the bottom layer through the season could be computed. These data are given in figure 5 of plate 5-6, along with the computed water equivalents. The monthly ground melt amounts from this figure are as follows:

Month	Inches
January	0.11
February	0.37
March	0.69
April	0.77
May	0.96

The total seasonal loss from 1 January to 9 June amounted to 3.26 inches. It will be noted that the melt rate accelerated as the season progressed. This may be partially explained by the increase in the thermal conductivity of the soil as it became progressively more moist; however, thermal conductivities computed from these melt data and available temperature gradient data exhibit too wide a variation (by comparison with Kersten's data). It is probable that, early in the season (January and February) some of the heat conducted to the bottom of the pack is being consumed in ripening the pack, without producing melt and runoff. This is to say, some of the heat is being used to bring the bottom layer to 32°F and to saturate it to its maximum free water holding capacity. These data, while not universally applicable, are felt to be generally indicative of the magnitudes and variations in the ground melt. It may thus be seen, considerable water is available to prime the soil in advance of the main melt season. For areas having greater temperature gradients and/or thermal conductivities, even greater melts from ground heat would be expected. Rough approximations of this amount may be made for other areas from the foregoing data on thermal conductivities and temperature gradients.

5-11. HEAT CONTENT OF RAIN WATER

5-11.01 Derivation of equation. - When rain falls on the snowpack it is cooled to the temperature of the snow, the quantity of heat involved being given up to the snow. For snowpacks isothermal at

32^oF, this release of heat results in snowmelt, while for colder packs this heat tends to raise the snow temperature to 32^oF. The amount of heat given up to the snow by the rainwater is directly proportional to the quantity of rainwater and to its temperature excess (above that of the snowpack). Considering a melting snowpack, for every degree centigrade the rainwater is in excess of the snow temperature (zero ^oC), and for every gram of rainwater, one calorie of heat is available. Thus, each centimeter depth of rainfall releases one langley (calorie per square centimeter) for each degree centigrade above freezing. That is,

$$H_p = (T_r - T_s) P_r \qquad (5\text{-}31a)$$

where H_p is the heat released by the rainfall in langleys, T_r and T_s are temperatures of the rainwater and snowpack, respectively, in degrees C, and P_r is the depth of rainfall in centimeters. English units may be substituted for the temperature and rainfall measurements in the above equation. Thus,

$$H_p = 5/9\ (T_r - T_s)\ 2.54P_r$$

$$= 1.41\ (T_r - T_s)\ P_r \qquad (5\text{-}31b)$$

where T_r and T_s are now in degrees F, and P_r is in inches. H_p is still in langleys. For snow having a thermal quality of 100 percent, the resultant melt, M_r, in inches, is given by,

$$M_r = \frac{1.41}{203.2}\ (T_r - 32)\ P_r$$

$$= 0.00695\ (T_r - 32)\ P_r \qquad (5\text{-}32)$$

This equation is shown graphically in figure 2 of plate 5-6. It may be seen from this figure that the melt from rainfall is relatively minor for normal rainfall temperatures when compared with the quantity of rainfall itself. For example, one inch of rainfall at a temperature of 46^oF produces only 0.1 inch of melt.

5-11.02 Latent heat of fusion. - When rain falls on a subfreezing snowpack, an additional quantity of heat is also given to the pack. The rainwater that is frozen within the pack releases its latent heat of fusion (80 cal/g) to the snow. Thus for each inch of rainfall, 203.2 langleys are given up to the pack. In view of this large source of heat and the small specific heat of snow, it may be seen that subfreezing snowpacks cannot prevail during rainstorms of any considerable magnitude. For example, a snowpack six-feet deep having a mean density of 40 percent and a mean temperature of -10^oC could be brought to zero degrees C by 1.8 inches of rainfall at zero degrees C. (This topic is discussed further in chapter 8.)

5-11.03 Rain temperature. - The surface wet-bulb temperature is generally considered to be a suitable estimate of the temperature of the rain. Becuase of the usual near-saturated conditions that exist during rainstorms, dewpoint, or for that matter air temperature, may be used as rain temperature with very little error. Some hydrologists make the refinement of considering the mean temperature of, say, the kilometer of air above the snow surface. This usually lower temperature assumes the raindrops to be more nearly represented by the temperatures of the region through which they pass; that is, there is a lag in the change in the temperature of the falling rain as it attempts to assume that of its environment.

5-12. INTERRELATIONSHIPS BETWEEN COMPONENT MELTS

5-12.01 Examples. - By way of summary of this chapter, the figures of plates 5-7 and 5-8 are presented. These figures give the average component heat fluxes to the snowpack for the months November through June and for three days during the spring melt season for conditions as they exist at the CSSL basin. The meteorological conditions used to compute the fluxes given in plate 5-7 are the mean values for the five years 1946-1947 through 1950-1951. All heat fluxes are expressed in inches of melt (200 ly equal one inch melt), and are for basinwide conditions; the basin is assumed to have 100 percent snow cover for the entire period. Negative melts represent heat losses from the pack, the same correspondence between heat and melt quantities mentioned above being used. The relationships between observed meteorological conditions and resultant snowmelt previously presented and discussed in this chapter are used to calculate the component melts. It will be noted in figure 5 of plate 5-7 that for the months November through February, the total heat flux to the snowpack is on the average, negative. During March the total flux becomes positive; however, the amount of heat is small. During April there is a sudden acceleration in the heat supply to the snowpack which continues through June. By considering these heat fluxes, it can thus be seen why the snowpack accumulates through March and ablates thereafter. During March and April, some of the net heat supply to the pack is consumed in ripening the pack; by the latter part of April, however, melt is generally as indicated by the figures. Individual figures on these plates are discussed in the following paragraphs.

5-12.02 Figure 1 of plate 5-7 gives a general picture of the variation of insolation with the season at this mid-latitude station. It shows how the insolation received at the outer limits of the earth's atmosphere is, on the average, depleted by the atmosphere and by clouds. Furthermore, it shows the radiation actually absorbed by the snowpack, forest cover and snowpack albedo being considered. It is interesting to compare this absorbed radiation with that originally available at the outer limit of the earth's atmosphere, noting how little of this available energy is actually directly absorbed by the snow. This is especially true during the winter when frequent new snowfalls maintain a high albedo.

5-12.03 In figure 2 of plate 5-7, the absorbed radiation of figure 1 is converted to equivalent snowmelt assuming that 200 ly produce one inch of melt. This conversion is made so that the melt quantities due to radiation can be combined with other melt components which are expressed directly in inches of melt. Actually, this is somewhat misleading since the amounts given are not the actual melt that would be realized from the snowpack but are, rather, the heat fluxes to the snow. The actual resultant melt is dependent upon the thermal quality of the snow as has been discussed. Losses of heat from the snowpack are shown as negative melt quantities. Also shown in figure 2 is the average longwave radiation loss for this area, expressed in inches of melt. It includes the effects of cloud and forest cover, of temperature and humidity of the air, and of the changes in the temperature of the snow surface itself. Shortwave and longwave radiation melt are combined into a single curve showing the variation in radiation melt through the period of interest.

5-12.04 Figure 3 of plate 5-7 shows the variation of convection and condensation melts with time and also a combined convection-condensation melt curve. These values were computed from mean observed air temperatures and vapor pressures for the years of record at CSSL. A constant wind speed was assumed throughout, since no regular seasonal trend of wind speed was discernible from the available data.

5-12.05 Figure 4 of plate 5-7 shows melts resulting from conduction of heat from the ground and from the heat content of rainwater falling on the snowpack. The former was computed from mean ground temperature gradient data for CSSL and estimated thermal conductivities for the ground as given in Technical Bulletin 16. While the temperature gradient tends to decrease throughout the period under consideration, there is an increase in the conductivity of the ground during the spring melt season as a result of the increase in moisture content. This results in the rise of ground melt during the period March-May. Rain melt was computed using the rainfall amounts from the water balance for this area (see chap. 4). The quantities of heat given (in inches of melt) are largely the result of rain falling on the sub-freezing snowpack and releasing its latent heat of fusion. The additional increment of heat which results from the rainwater being cooled to 32°F is practically negligible.

5-12.06 Figures 1, 2, and 3 of plate 5-8 give a detailed, hour-by-hour picture of the principal heat fluxes to the snowpack for three representative days during the spring melt season. A clear day, a partly-cloudy day, and an overcast day are illustrated. The data presented in those figures are from some special lysimeter studies of snowmelt made at CSSL during the 1954 spring snowmelt season (see Res. Note 25). They are for melt at an open, unforested site and for days without precipitation. Ground melt is not included, as the lysimeter introduced an artificial effect with respect to this melt component.

5-12.07 Discussion. - In the preceding section of this chapter, each of the sources of thermal energy involved in the melting of the snowpack has been examined separately. Yet these heat fluxes are not independent of one another; rather they are quite interdependent, the degree of relationship being influenced by the terrain involved. For this reason it is well that these heat fluxes also be examined collectively and with consideration of the terrain. Moreover, the discussion has thus far been concerned with heat exchange and melt as a specified point only. Areal considerations are also involved in a complete understanding of heat exchange as it affects snowmelt. In this section, these broader aspects of snowmelt will be examined.

5-12.08 Since all the thermal energy involved in melting the snowpack has its ultimate source in the solar radiation reaching the earth, the other processes serve merely as intermediate means of heat transfer. Yet it is the measurement of some of the manifestations of solar energy, such as air temperature and vapor pressure, that are most commonly used as indexes of snowmelt. The earth's atmosphere is warmed but slightly by solar radiation passing through it, as was pointed out in the section on solar radiation. Over snowfields this absorption is increased, since a portion of the reflected beam is also absorbed. Nevertheless, the degree of heating of the air by solar radiation is relatively small. Of greater consequence is the heating of air which results from air passing over lands and objects which are heated by solar radiation. These give their heat to the air by the processes of convection and longwave radiation, the air being much less transparent to the latter than it is to solar radiation. Similarly, water surfaces and ground surfaces containing water may be heated by solar radiation and by the transfer of heat from the air by the process of convection and longwave radiation and give off energy in the process of evaporation, the latent heat of vaporization being consumed in the process. Thus water vapor is added to the air and may subsequently condense upon the snow surface, releasing its latent heat.

5-12.09 In view of the foregoing, it may be seen that over barren snowfields of great areal extent, and in the absence of advection of energy, the air can neither be heated appreciably above 32°F nor can its vapor pressure exceed 6.1 millibars (saturated vapor pressure at 32°F). Since the upper limit of snow surface temperature is 32°F and its vapor pressure thus restricted to 6.1 millibars in this situation, it cannot warm the air above this temperature either by convection of heat or longwave radiation, nor can it add moisture to the air in excess of its own vapor pressure. Thus any appreciable convection and condensation melts require bare ground or water surfaces and/or forest cover which can serve as exchange mechanisms to convert radiant energy into sensible heat and moisture. (Some heat may result from subsidense; however, this is ignored in the discussion which follows.)

5-12.10 Advection of thermal energy. - Considering the snowmelt within a given drainage basin, heat and moisture may be advected either into or out of the basin depending upon the forest cover and areal extent of the snow cover in the basin relative to its environment. If the basin under consideration is but a relatively small part of a much larger homogeneous area, then very little advection would ordinarily be expected. The air leaving the basin should have about the same temperature and moisture content as the air entering. Only in the case of exceptionally warm and moist or cold and dry air masses passing over the area would advective effects be of any consequence. Even then the basin under consideration would add or subtract only its small incremental share to the modification taking place in the airmass in passing over the larger homogeneous area. Advection plays a more important role in snowmelt where the basin under consideration is adjacent to and leeward of an area having markedly different characteristics. Thus considerable heat and moisture may be advected into a basin situated to the lee of a non-snow-covered area or of an open water area. This situation is exemplified by the snowfields of the mountain ranges along the Pacific Coast. Here air coming off the ocean first passes over the valleys to the west before reaching the snowfields. Another situation favorable to the advection of heat and moisture into a given drainage basin occurs where the area to the windward is more heavily forested than is the drainage basin itself. Thus, in the windward area the air may be warmed and moisture added to a greater extent than within the basin, the net result being the advection of moisture and heat into the area. A more or less barren drainage area to the lee of a heavily forested area would result in considerable advection during clear weather.

5-12.11 On the other hand, heat and moisture may be advected from a snow-covered basin. Even air initially warmer than 32°F and having a vapor pressure in excess of 6.1 millibars, may, in passing over the basin, be warmed above 32°F by convection and longwave radiation from the trees and barren areas, and made more moist by evaporation from snow-free areas and transpiration from trees. Of course air initially having a temperature and dewpoint less than 32°F can be modified to at least those temperatures by even a barren snowfield.

5-12.12 Local effects. - Considering now snowmelt in a basin in which no advection of heat energy takes place: the condition may be approached in actual situations where weak pressure gradients exist or where a high pressure area stagnates over the basin being considered, and where the basin is but a small part of a much larger area having similar conditions of snow and forest cover. Since the air leaving the basin has, under these conditions, the same heat content as the air entering, all thermal energy, other than solar radiation, used in melting the snowpack must be generated within the basin. This situation is thus referred to as the "local climate" or "radiation climate." Since the air temperature remains constant, the heat added to the air by convection and longwave radiation from trees and snow-free ground surfaces is just balanced by the heat given up by the air to the snow by the same processes. Thus, in this situation, the total energy represented by the snowmelt and the water lost to the atmosphere by evapotranspiration is equal to the net amount of solar radiation absorbed within the area. (See supplement to Research Note 19.) One would thus expect the melt under these conditions of a local climate to vary directly with the density of the forest, the over-all albedo decreasing with increasing forest cover. Heavily forested areas, however, tend to result in advection of heat and moisture from the area, whereas more barren areas favor advection into the area. This will be considered further in the following paragraph.

5-12.13 Forest effects. - Is the melt rate greater in forested or in barren areas? This is a question that has long been debated by those concerned with snowmelt. The answer is simply this: Sometimes it is greater in the forest and sometimes it is greater in the open, depending upon the size of the clearing and other factors. With the background of this chapter it is possible to qualify this statement. But first the question of deposition of snow should be dealt with. Since relative melt rates in the forest and in the open are often judged by the disappearance of snow, the usually greater deposition in the open tends to bias the answer in favor of the forested areas where there is generally less snow to begin with. However, generalizations must be made with as much caution in the case of deposition as with melt. While small forest clearings usually collect the maximum snowpack, larger open areas may be scoured of snow which is then deposited in greater depths near and under the surrounding trees (see chap. 3 for a discussion of deposition effects). Melt rates are generally considerably greater in large clearings than they are in forested areas. These areas not only have approximately the same air temperatures and vapor pressures as do the forested areas, due to large-scale mixing of the air, (even though the heat and moisture are given to the air primarily as a result of the interception of solar radiation by the trees), but the higher wind speeds encountered over the snow in the open result in greater melts due to convection and condensation. Since the melt component due to absorbed solar radiation is considerably greater in the open than it is in forested areas, the greater convection-condensation melt component in the open, coupled with this greater radiation melt, results in greater melt in the open. (Of course evaporation might prevail, in which case heat losses in the open due to this cause would exceed those in the forest. The heat transfer due to evaporation is, however, relatively small.) Longwave radiation loss in the open is, of course, greater than in the forest; however, since a part of the solar energy absorbed by the forest canopy is re-radiated to space and part is used up in warming the air and transpiring moisture which in turn are conveyed to the snow surface, the greater loss of longwave radiation in the open is usually more than made up for by the greater shortwave radiation gain.

5-12.14 It has been observed that the last snow patches to disappear in the spring are usually found in small clearings in the forest; that is, clearings having a diameter about the same as the heights of the surrounding trees. The reason for this is two-fold: (1) the greater deposition of snow in these sheltered areas and, (2) the lesser melt rates in these areas. The first of these reasons is dealt with in chapter 3. The second is the result of the shading of the sites from direct solar radiation by the surrounding trees, while, at the same time, longwave loss is affected to a lesser degree. Then too, convection-condensation melts are less than in larger open areas by virtue of the lesser wind speeds in these areas. In summary, then, it may be said that melt rates are greatest in large open areas and least in small forest clearings, the melt rate in the forest being intermediate between these two. It is to be emphasized, however, that the greater melt rates of large open areas is still contingent upon the existence of surrounding forests or bare ground for a supply of sensible heat and moisture. Large, non-forested, snow covered areas such as are found in parts of Canada would not have as high melt rates as would a large clearing in the otherwise forested area.

5-12.15 From the foregoing the true relationship between air temperatures, vapor pressures, and snowmelt can be deduced. It may be seen that upper air temperatures and humidities should <u>not</u> be good indexes of snowmelt since they do not adequately reflect the conversion of radiant energy into sensible heat and moisture. The air may be warmed and its humidity increased in passing over trees and bare ground warmed by radiation, and it may in turn convey this thermal energy to the snowpack without any change indicated in the upper air. Only surface temperatures and humidities adequately reflect this heat transfer. Moreover, since more of the radiant energy is manifest in the air temperature and humidity in forested areas than in barren areas, temperature and humidity indexes should be expected to increase in accuracy with the degree of forest cover. These effects are considered in detail in the following chapter which deals with the practical computation of snowmelt by means of indexes.

5-13. SUMMARY

5-13.01 The relationships presented in this chapter are summarized below in outline form. The general snowmelt equation is given, followed by the equations which give the component melts. References are made to figures which illustrate these relationships where applicable. Symbols used are defined in the text.

<u>GENERAL SNOWMELT EQUATION</u> (M)

$$M = H_m/203.2 \qquad (5\text{-}2a)$$

(See fig. 1, pl. 1)

where,

$$H_m = H_{rs} + H_{rl} + H_c + H_e + H_g + H_p$$

SHORTWAVE RADIATION MELT (M_{rs})

Insolation

(See figs. 3 & 4, pl. 1)

Effect of Clouds

$$I/I_c = 1 - (1-k)\ N \qquad (5\text{-}6b)$$

where,

$k = 0.18 + 0.024z$

(see Fig. 5, pl. 5-1)

Effect of forest cover

(See fig. 1, pl. 5-2)

Effect of slope

(See Fig. 6, pl. 5-1)

Albedo

(See figs. 2, 3, & 4, pl. 5-2)

Absorption

$$I_d = I_a e^{-kd} \qquad (5\text{-}8)$$

(See fig. 5, pl. 5-2)

LONGWAVE RADIATION MELT (M_{rl})

Radiation from clear skies (R_d)

$$R_d = \sigma T_a^4\ (a + b\sqrt{e_a}) \qquad (5\text{-}9)$$

(See figs. 2 & 3, pl. 5-3)

Radiation from snowpack (R_u)

(See fig. 2, pl. 5-3 -- Black body curve)

Net radiation exchange -- clear skies (R_c)

(See Fig. 4, pl. 5-3 -- over melting snow)

Effect of Clouds

$$R = \sigma (T_c^4 - T_s^4) \quad \text{(overcast)} \qquad (5\text{-}12)$$

(See fig. 4, pl. 5-3: black body curve)

$$R = R_c\ (1 - kN) \qquad (5\text{-}13)$$

where,

$k = 1 - 0.24\ z$
(See fig. 5, pl. 5-3)

Effect of forest cover

$R_f = \sigma\,(T_a{}^4 - T_s{}^4)$ (Solid canopy) (5-16)

$R = FR_f + (1\text{-}F)\ R_c$ (partial canopy) (5-17)
(See fig. 6, pl. 5-3)

CONVECTION MELT (M_c)

$M_c = k_c\ (T_a - T_s)\ v_b$ (5-28a)

$M_c = k_c{}'\ (p/p_o)(z_a z_b)^{-1/6}\ (T_a - T_s)\ v_b$ (5-28b)
(See fig. 1, pl. 5-5)

CONDENSATION MELT (M_e)

$M_e = 8.5\ k_e\ (e_a - e_s)\ v_b$ (5-27a)

$= k_e'\ (z_a z_b)^{-1/6}\ (e_a - e_s)\ v_b$ (5-27d)
(See fig. 2, pl. 5-5)

GROUND MELT (M_g)

$H_g = k\ dT/dz$ (5-30)
(See figs 4 & 5, pl. 5-6)

RAIN MELT (M_p)

$M_r = 0.00695\ (T_r - 32)P_r$ (5-32)
(See fig. 2, pl. 5-6)

5-14. REFERENCES

1/ ALGREN, A.B., "Ground temperatures as affected by weather conditions," Heating, Piping & Air Conditioning, June 1949, pp. 111-116.

2/ ANDERSON, Ernest R., "Energy-budget studies," Geological Survey Circular 229 (Water Loss Investigations: Volume -- Lake Hefner Studies Technical Report), 1952, pp. 71-119.

3/ ANDERSON, E.R., L.J. Anderson, and J.J. Marciano, "A review of evaporation theory and development of instrumentation," (Interim Report: Lake Mead Water Loss Investigations), Report 159, Navy Electronics Laboratory, San Diego, Calif., February 1950.

4/ ANGSTROM, A., "On the radiation and temperature of snow and the convection of the air at its surface," Arkiv. f. Math. 13, Nr. 21, 1919.

5/ BOWEN, I.S., "The ratio of heat losses by conduction and by evaporation from any water surface," Phys. Rev., Vol. 27, June 1926, pp. 779-787.

6/ BRUNT, David, Physical and Dynamical Meteorology, Cambridge Univ. Press, 1952.

7/ de QUERVAIN, M., "Evaporation from the snowpack," (translated from the German in Snow Investigations Research Note 8, October 1952).

8/ DUNKLE, R.V. and others, "Non-selective radiometers for hemispherical irradiation and net radiation interchange measurements," Univ. of Calif., Div. of Engrg. Research, Report No. 9, (Report Code NR-015-202), October 1949.

9/ ELSASSER, Walter M. "Heat transfer by infrared radiation in the atmosphere," Harvard Meteor. Studies, No. 6, 1942.

10/ FALCKENBERG, G., "The infrared absorption constants of some meteorologically important substances," (Absorptionskonstantan einiger meteorologisch wichliger, Korper fur infrarote wellen; test in German). (Meteoro. Zeitschr.) Vol. 45, September 1948, pp 334-337.

11/ FRITZ, Sigmund, "Solar radiant energy and its modification by the earth and its atmosphere," Compendium of Meteorology, American Meteor, Society, Boston, Mass., 1951, pp. 13-33.

12/ FRITZ, Sigmund and Torrence H. MacDonald, "Average solar radiation in the United States," Heating and Ventilating, Vol. 46, July 1949, pp. 61-64

13/ GEIGER, Rudolf, The Climate near the Ground, (A translation by Milroy N. Stewart and others, of the second German edition of Das Klima der Bodennahen Luftschicht with revisions and enlargements by the author.) Harvard Univ. Press, Cambridge, Mass. 1950.

14/ GERDEL, R. W., "Penetration of radiation into the snowpack," Trans. Amer. Geophys. Union, Vol. 29, No. 3, June 1948, pp. 366-374.

15/ GERDEL, R. W., M. Diamond, and K. J. Walsh, "Nomographs for computation of radiation heat supply," SIPRE Research Paper 8, Corps of Engineers, February 1954.

16/ GIER, J.T. and R.V. Dunkle, "Total hemispherical radiometers," AIEE Transactions, Vol. 70, 1951.

17/ GOSS, John R., and Frederick A. Brooks, "New constants for the empirical expression for downcoming atmospheric radiation under cloudless sky," (Paper presented at 134th Nat'l Meeting, AMS, Berkeley, California, December 1954.)

18/ HALSTEAD, Maurice H., "The relationship between wind structure and turbulence near the ground," Publications in Climatology, Vol. 4, No. 3 (The Johns Hopkins Univ., Lab. of Climatology), Seabrook, New Jersey, 1951.

19/ HAMON, Russell, W., Leonard L. Weiss, and Walter T. Wilson, "Insolation as an empirical function of daily sunshine duration," Mon. Wea. Rev., Vol. 82, No. 6, June 1954, pp. 141-146.

20/ HAND, I.F., "Pyrheliometers and phrheliometric measurements," U.S. Wea. Bureau, Washington, D. C., 1946.

21/ HAURWITZ, B., "Insolation in relation to cloud type," Jour. of Meteor., Vol. 5, No. 3, June 1948, pp. 110-113.

22/ HOECK, Erwin, "The influence of radiation and temperature on the melting process of the snow cover," (Der einfluss der strahlung und der temperatur auf den schmelzprozess der schneedeck; text in German), Bietr. Geol. Schweiz - Geotech. Serie-Hydrologie, 1952.

23/ HUBLEY, Richard C., "Measurements of diurnal variations in snow albedo on Lemon Creek Glacier, Alaska," Jour. of Glaciology, Vol. 2, No. 18, October 1955, pp. 560-563.

24/ JOHNSON, Francis, S., "The solar constant," Jour. of Meteor., Vol. 11, No. 6, December 1954, pp. 431-439.

25/ KERSTEN, Miles, S., "Laboratory research for the determination of the thermal properties of soil (final report)," Engrg. Exper. Station, Univ. of Minn., June 1949.

26/ KIMBALL, H.H., "Variations in total and luminous solar radiation with geographic position in the United States," Mon. Wea. Rev., Vol. 47, No. 11, November 1919, pp. 769, 793.

27/ KLEIN, Wm. H., "Calculation of solar radiation and the solar heat load on man," Jour. of Meteor., Vol. 5, No. 4, August 1948, pp. 119-129.

28/ LIGHT, Phillip, "Analysis of high rates of snow-melting," Trans. Amer. Geophys. Union, part I, 1941, pp. 195-205.

29/ MacDONALD, T.H., and Norman B. Foster, "Pyrheliometer calibration program of the U.S. Weather Bureau," Mon. Wea. Rev., Vol. 82, No. 8, August 1954, pp. 219-227.

30/ MANTIS, Homer T. (ed.), "Review of the properties of snow and ice," (SIPRE Report 4), Univ. of Minn., Inst. of Technology, Engrg. Exper. Station, March 1951, (Chaps. 5-7), pp. 52-85.

31/ MILLER, David H., "Snow cover and climate in the Sierra Nevada, California," Univ. of Calif. Publications in Geography, Vol. 11, Univ. of Calif. Press, Berkeley, 1955.

32/ NEWMANN, U., "Insolation in relation to cloud amount," Mon. Wea. Rev., Vol. 82, No. 11, November 1954, pp. 317-319.

33/ SUTTON, O.G., Micrometeorology, McGraw-Hill Book Co., Inc., 1953.

34/ SVERDRUP, H.U., "The eddy conductivity of the air over a smooth snow field," Geofysiske Publikasjoner, Vol. 11, No. 7, 1936, pp. 1-69

35/ SVERDRUP, H.U., Oceanography for Meteorologists, Prentice-Hall, Inc., New York, 1942.

36/ The Johns Hopkins University, Laboratory of Climatology, "Micrometeorology of the surface layer of the atmosphere; the flux of momentum, heat, and water vapor, final report," Publications in Climatology, Vol. 7, No. 2, Seabrook, New Jersey, 1954.

37/ THRELKELD, J.L. and R.C. Jordan, "Solar radiation during cloudless days," Heating, Piping and Air Conditioning, February 1955, pp. 117-122.

38/ WILSON, Walter T., "An outline of the thermodynamics of snow melt", Trans. Amer. Geophys. Union, Part I, 1941, pp. 182-195.