Chapter 2
Snowmelt Runoff—A Review of Fundamental Processes

2-1. General Introduction

In many regions of the United Sates, snowfall and the resulting seasonal snowcover represent an important source of water. When the snowpacks melt, the snowmelt recharges the groundwater and replenishes surface water storage. Excessive snowmelt runoff can cause flooding, while inadequate snowmelt is often the prelude to later drought.

a. When snow melts, the ice that composes the snow is converted into water. This water is called snowmelt. Since the conversion from ice to water requires the input of energy (or heat), the process of snowmelt is inextricably linked to the flow and storage of energy into and through the snowpack. These linkages between the flow and storage of both water (i.e., ice and liquid water) and energy (or heat) are summarized in Figure 2-1 to facilitate the discussion and to clarify the complicated processes that control snowmelt runoff.

b. The sources of energy that cause snowmelt include both shortwave and long-wave net radiation, convection from the air (sensible energy), vapor condensation (latent energy), and conduction from the ground, as well as the energy contained in rainfall. These energy fluxes are shown in the upper left of Figure 2-1 and are labeled Q_{sn} , Q_{ln} , Q_h , Q_e , Q_g , and Q_p , respectively. These fluxes are usually measured as energy per time per unit area of snow. The energy budget equation that describes the energy available for snowmelt is given in Equation 2-1 below. The total energy available for snowmelt is Q_m.

$$Q_m = Q_{sn} + Q_{ln} + Q_h + Q_e + Q_g + Q_p - \Delta Q_i \qquad (2\text{-}1)$$

ΔQ_i is the rate of change in the internal energy stored in the snow per unit area of snowpack. This term is composed of the energy to melt the ice portion of the snowpack, freeze the liquid water in the snow, and change the temperature of the snow. Thus, during periods of warming, the net flux of heat (ΔQ_i) is into the snow, while during periods of cooling, the net flux (ΔQ_i) is out of the snowpack. Therefore, the amount of energy available to cause snowmelt varies and can be dynamic, depending on the magnitudes of the various energy inputs to the snowpack. Male and Gray (1981) suggest that snowmelt is not homogeneous throughout the snowpack depth and point out that most of the melting occurs at the upper and lower interfaces of the snow (i.e., the interfaces with the atmosphere and the ground).

c. Whenever sufficient energy is available, some snow (ice) will melt and form liquid water (i.e., snowmelt). Since the physical structure of the snowpack is a porous matrix, this snowmelt will be held as liquid water (provided it does not refreeze) in the interstices between the snow grains and will increase snow density and snow water content. The snowpack is commonly called “ripe” when it is isothermal at 0 °C and saturated. Whenever the capacity of the snowpack interstices to hold the liquid water is exceeded, some of the snowmelt will begin to move down-gradient (called direct surface runoff in Figure 2-1) to become a portion of the snowmelt runoff. Additionally, some of the snowmelt may infiltrate into the ground. The amounts that infiltrate depend on inherent soil characteristics, the soil moisture content, as well as whether or not the ground surface is frozen. The infiltrated snowmelt later reemerges as interflow into stream channels, or it percolates into deeper groundwater storage. These snowmelt pathways are delineated in Figure 2-1.

d. Estimates of snowmelt amounts are derived through the use of energy balance equations or by some empirically defined snowmelt index. Determinations of the amounts and the temporal distributions of snowmelt runoff require additional analysis of the storage of the snowmelt in the snowpack and transmission of the snowmelt through the snowpack as well as along the surface of the ground as it courses its way to the stream channel.

e. This chapter will discuss the theoretical basis for snowmelt at a point and from a basin or watershed. Throughout, the overall energy and water mass pathways shown in Figure 2-1 will form the framework for the discussion.

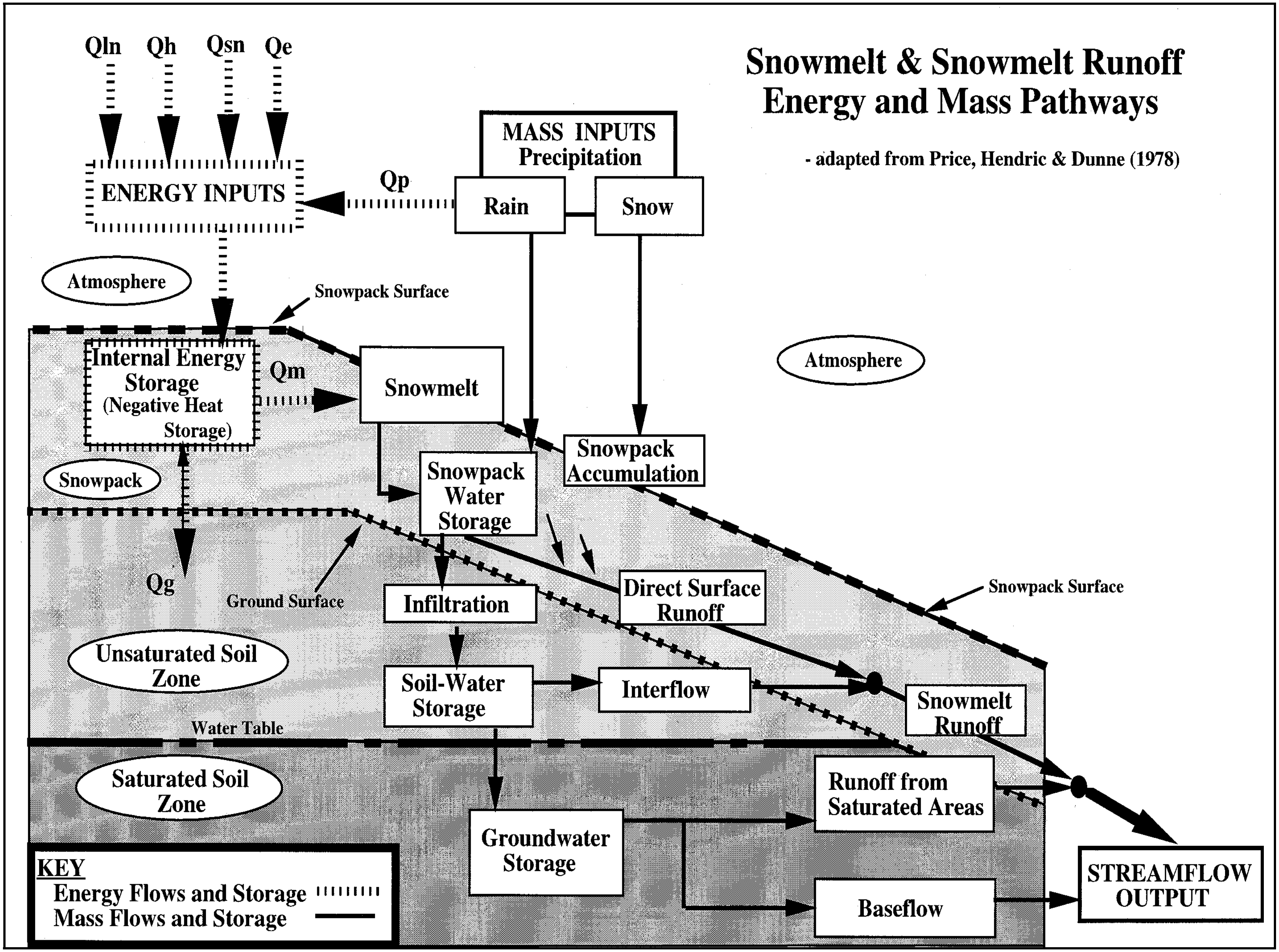

Figure 2-1. Schematic of the snowmelt process. (After Price, Hendrie, and Dunne (1979))

2-2. Energy and Mass Balance of the Snowcover

Evaluating snowmelt theoretically is a problem of heat transfer involving radiation, convection, condensation, and conduction. The relative importance of each of these heat transfer processes is highly variable, depending upon conditions of weather and the local environment. Gray and Prowse (1992) tabulate selected results of the relative contributions of each heat transfer process as a function of site environment. The basic equations and coefficients that describe snowmelt at a point have been derived primarily from various laboratory and field experiments.

a. General. Equation 2-1 summarized the energy sources available to melt snow. The summation of all sources of energy (heat) represents the total amount of energy available for melting the snowpack (Q_m). The amount of snowmelt at a point may be expressed by the general formula given as Equation 2-2

$$M = \frac{Q_m}{334.9\,\rho_w B} \qquad (2\text{-}2)$$

where

M = snowmelt, mm of water equivalent

Q_m = algebraic sum of all heat components, kJ/m2

B = thermal quality of the snow (e.g., ratio of heat required to melt a unit weight of the snow to that of ice at 0 °C)

334.9 = latent heat of fusion of ice, kJ/kg

ρ_w = density of water, kg/m^3

(1) Equation 2-2 may also describe the snowmelt per unit time (for example, mm water equivalent day) when Q_m is expressed in kJ/m^2 per day.

(2) A melting snowpack consists of a mixture of snow (ice) and a small quantity of free (liquid) water trapped in the interstices between the snow grains. The relative proportion of a snowpack that consists of ice determines the thermal quality (B) of the snowpack. A snowpack that contains no free water has a thermal quality of 1.0. However, after melt has begun, there is some free water held within the snow matrix, yielding a thermal quality of less than 1.0. The heat energy required to release 1 g of water is somewhat less than the latent heat of fusion of water (that is the energy required to change state from ice to water; 334.9 kJ/kg or 80 cal/g for pure ice). For a melting snowpack, after free drainage by gravity for several hours, the thermal quality normally averages between 0.95 and 0.97, corresponding to a 3- to 5-percent liquid water in the snow.

(3) The thermal quality of snow may be far lower for “ripe” snows and in extreme cases where the water cannot drain freely from the snowpack.

b. Radiational energy exchange. Radiational energy is the prime source of energy at the Earth's surface. Some of this energy is classed as solar or shortwave radiation (radiation having wavelengths (λ) between 0.2 and 2.2 μm) and terrestrial or long-wave radiation (wavelengths between 6.8 and 100 μm). The first two terms of Equation 2-1 are sometimes referred to as net radiation Q_n, the sum of net shortwave Q_{sn} and net long-wave Q_{ln} energy fluxes. As the net long-wave exchange is often a loss from the snow surface, Q_n is expressed as

$$Q_n = Q_{sn} - Q_{ln} \qquad (2\text{-}3)$$

(1) Shortwave radiation is the most important source of energy to the snowpack. The net amount of radiant energy that is available to melt snow depends on how much of the radiation is either reflected from or absorbed by the snowpack. The amount of heat transferred to the snowpack by solar radiation varies with latitude, season, time of day, atmospheric conditions, forest cover, and reflectivity of the snow (albedo). The intensity of incident solar radiation just above the Earth's atmosphere and normal to the path of radiation is virtually constant at 1.35 kJ/m^2 per second, the solar constant. In general, less than 50 percent of this incident solar radiation reaches the Earth's surface. As solar radiation passes through the Earth's

atmosphere, it is attenuated through reflection off clouds, scattered by air molecules and particulates in the atmosphere, and absorbed by a number of molecular structures contained in the atmosphere. By far, the greatest change in the portion of solar radiation transmitted through the atmosphere is caused by varying cloud cover, so that direct measures of solar radiation at the ground surface principally show the effect of depletion by clouds. Inasmuch as such measurements are not generally available at a given location, it usually becomes necessary to estimate incoming radiation indirectly from duration of sunshine data, observations of cloud conditions, or diurnal air temperature fluctuations. See Appendix D for further discussion of radiational energy exchange, including several charts showing how radiational flux varies.

(2) Additionally, the local environment has a marked effect upon the amount of solar radiation received on the snow surface. The relative ratio of the daily solar radiation incident upon a snow surface to that on a horizontal surface is a function of the surface slope angle to north or south (or aspect), the latitude, the season, and the amount of diffuse sky radiation relative to direct solar radiation. More complete descriptions of methods for calculating incident radiation and the effects of local environment are given by List (1968), Dozier (1979), Oke (1978), Male and Gray (1981), and Gray and Prowse (1992).

(3) Forest cover can also play an important role in the amount of solar energy that reaches the snow surface. For example, in coniferous forests, the transmission percentage varies with the type, density, and condition of trees. Transmission also varies with the season, because of the change in the shading effect of the trees with the solar altitude. The determination of the amount of sunshine transmitted through the forest is at best an approximation.

(4) The reflectivity of the snow surface plays an important role in the amount of energy available to cause snowmelt. Large portions of the shortwave radiation that reach the snow surface can be reflected. Since snowpack reflectivity varies over a considerable range, it is an important consideration in estimating the amount of solar energy absorbed by the pack. Albedo (A) is defined as the percentage of the incident shortwave radiation that is reflected from the snow surface. Values for albedo range from more than 80 percent for new-fallen snow to as little as 40 percent for melting, late-season, ripe snow.

(5) The amount of energy available for snowmelt from the absorption of shortwave radiation (Q_s) is

$$Q_s = (1 - A) I_i \tag{2-4}$$

where

A = albedo (expressed as a decimal fraction)

I_i = daily incident solar radiation (kJ/m^2 per day)

(6) In the middle latitudes during late spring, the maximum solar radiation for a clear day on a horizontal surface is about 52 MJ/m^2. With a minimum albedo of 40 percent, the resulting possible shortwave radiation melt for an unforested area is on the order of 6.4 cm/day. However, some of the energy absorbed by the snowpack from solar radiation is radiated from the snowpack to the atmosphere as long-wave radiation. Snow is nearly a perfect blackbody, with respect to long-wave (terrestrial) radiation, absorbing all such radiation incident upon it and emitting the maximum possible radiation in accordance with the Stefan-Boltzman law (Equation 2-5).

$$Q_l = \epsilon \sigma T_s^4 \tag{2-5}$$

where

Q_l = total shortwave energy emitted by the snow, kJ/m^2 per second

ϵ = 0.99 for clean snow

σ = Stefan-Boltzman constant, 5.735 × 10^{-11} kJ/m^2 s K^4

T_s^4 = blackbody temperature in Kelvin (K) (temperature of the snow surface)

(7) Consider a melting snowpack having a surface temperature of 0 °C. According to the Stefan-Boltzman law, the snowpack will lose energy at the rate of 0.315 kJ/m^2 per second. Opposed to this is the back-radiation, or long-wave radiation, reflected back from the atmosphere or the forest cover. For clear skies, the heat gain from back-radiation is generally less than the heat loss, so that there is net heat loss from the snowpack by long-wave radiation. With cloudy skies or beneath a forest canopy, however, the back-radiation may be greater or less than the loss from the snowpack, depending principally upon the ambient air temperature.

c. Turbulent transfer.

(1) Energy is also exchanged between the snowpack and atmosphere through the processes of convection and condensation. Depending on the climatological and local weather conditions, the relative importance of these processes differs widely. For example, during clear weather in the spring, energy exchange by the process of turbulent exchange from the atmosphere is of secondary importance compared with radiation for snowmelt. However, during a winter rain on snow, turbulent exchange is the dominant heat exchange process. Turbulent exchange involves the transfer of sensible heat from warm air advected over the snowfield (convection), and also the latent heat of condensation of water vapor from the atmosphere condensed on the snow surfaces. Computation of the transfer of sensible and latent heat from the atmosphere is complex from a theoretical standpoint, and exchange coefficients are derived empirically from controlled experiments.

(2) The principal variables affecting convective (sensible) heat exchange are the temperature gradient of the atmosphere measured above the snow surface and the corresponding wind speed. Similarly, the primary variables affecting condensation (latent) heat exchange are the vapor pressure of the atmosphere and the snow surface and the corresponding wind speed. Equations 2-6 and 2-7 describe sensible and latent heat transfer, respectively (Gray and Prowse 1992).

$$Q_h = D_h u_z (T_a - T_s) \qquad (2\text{-}6)$$

$$Q_e = E_e u_z (e_a - e_s) \qquad (2\text{-}7)$$

where

D_h = bulk transfer coefficient for sensible heat transfer, kJ/m^3 °C

u_z = wind speed at a chosen height above the snow surface, m/s

T_a = temperature at the air surface, °C

T_s = temperature at the snow surface, °C

D_e = bulk transfer coefficient for latent heat transfer, kJ/m^3 Pa

e_a = vapor pressure of the air surface, Pa

e_s = vapor pressure of the snow surface, Pa

d. Heat conduction from the ground. Heat entering the snow from the ground (Q_g) by solid conduction is a very small component to the overall energy budget, especially compared with the radiational and turbulent exchange at the air/snow interface. This ground heat component can be neglected over short periods of time (less than 1 week). Although the daily melt caused by ground heat is small, it can amount to a significant quantity of water over an entire snow season. Most lumped, conceptual models use constant daily values in the range of 0-5 J/m^2 per second. Ground heat flow can also be estimated using soil temperature gradients measured near the surface in an equation for steady-state, one-dimensional heat flow by conduction:

$$Q_g = k \frac{dT_s}{dz} \qquad (2\text{-}8)$$

where

k = thermal conductivity of the soil

$\frac{dT_g}{dz}$ = temperature gradient from soil to snow *e. Heat convected by rain.* The heat convected from the snow by rainfall is

$$Q_p = C_p \rho_w P_r (T_r - T_s) / 1000 \tag{2-9}$$

where

C_p = specific heat of rain, kJ/kg °C

ρ_w = density of water, kg/m^3

P_r = rain quantity, mm/unit time

T_r = temperature of the rain, °C

T_s = snow temperature, °C

The temperature of the rain is assumed to be equal to the air temperature or, if available, the wet-bulb temperature. The specific heat C_p is equal to 4.20 kJ/(kg °C) for rainfall and 2.09 kJ/(kg °C) for snowfall.

f. Internal energy. By definition, if the cold content or heat deficit of the snowpack is positive, the snowpack's temperature is below freezing. The internal energy Q_i can be changed and the heat deficit reduced by the heat released when melt or rainwater freezes within the snow cover. This phenomenon is prominent during diurnal temperature cycles with refreezing at night because of radiational cooling. Melt and rainwater will continue to freeze within the snow cover until the total heat deficit reaches zero. When the total heat deficit reaches zero, the snow cover will become isothermal at 0 °C. This internal energy is calculated by the following expression (Gray and Prowse 1992):

$$Q_i = d_s (\rho_i C_{pi} + \rho_l C_{pl} + \rho_v C_{pv}) T_m \tag{2-10}$$

where

d_s = depth of snow

ρ = density, ρ_i = 922 kg/m^3, ρ_l = 1000 kg/m^3

C_p = specific heat, $C_{pi} \cong$ 2.1 kJ/kg °C; $C_{pl} \cong$ 4.2 kJ/kg °C

T_m = mean snow temperature, °C

The subscripts *i*, *l*, and *v* refer to the ice, liquid, and vapor phases, respectively. The contribution of the vapor phase is assumed negligible.

2-3. Snowpack Meltwater Production and Movement

As was pointed out earlier (see Figure 2-1), before snowmelt becomes runoff from a watershed, a number of processes occur. These processes involve a change in character of the snow crystals, changes in snowpack temperature and density, and the movement of meltwater through the pack. The changes in the internal energy of the snowpack are relatively small and are usually neglected in deep packs, where other energy components dominate. For shallow snowcovers, however, these phenomena become more important.

a. Character of the snowpack. The formation of the snowpack begins with the deposit of new-fallen snow of relatively low density (i.e., specific gravity). With time, the snowpack changes; the delicate crystals of snow become coarse grains, and the density of the pack increases. The metamorphosis from a loose, dry, and subfreezing snowpack of low density to a coarse, granular, and moist snowpack of high density is sometimes spoken of as “ripening” of the snowpack. A ripe snowpack is said to be “primed” to produce runoff when it becomes isothermal at 0 °C and its liquid-water-holding capacity has been reached. At this point, the only storage effect of the snowpack is that of “transitory” storage, resulting in a temporary delay of liquid water in transit through the pack. Although ripe snow is usually the relatively dense, coarse-grained snowpack characteristic of the spring, there is no restriction on the time of year that the snowpack may yield liquid water to the underlying ground surface. Midwinter rainfall or snowmelt may satisfy the “cold content” and liquid-water-holding capacity of the snowpack. After those deficiencies have been met, any further input of liquid water at that time will pass through the snowpack as drainage by gravitational force. Figure 2-2 shows the features of a deep snowpack during a winter-spring season. Changes in depth, density, and snow temperature can be seen as the season progresses. Note the midwinter rainstorm in December where the snowpack became isothermal in only a few hours.

(1) Changes that take place within the snowpack are caused by several physical processes:

(a) Heat exchange at the snow surface.

(b) Percolation of meltwater or rain through the snowpack.

(c) Internal pressure attributable to the weight of the snow.

(d) Wind.

(e) Temperature and vapor pressure variation within the snowpack.

(f) Heat exchange at the ground surface.

As each new layer of snow is deposited, its upper surface is weathered by radiation, rain, and wind. The undersurface of the new layer may also be affected by ground heat. As a result, the snowpack is stratified, showing distinct layers and ice planes or lenses that separate individual snowstorm deposits. The interior of the pack is subjected to the action of percolating water and diffusing water vapor.

(2) During the melt season, on clear nights, a relatively shallow surface layer of the snowpack generally cools considerably below 0 °C, owing to the loss of heat to the sky by long-wave radiation; the liquid water may freeze in this layer to as much as 25.4 cm (10 in.) deep, but below this surface layer the liquid water remains unfrozen.

b. Drainage of snowmelt through the snowpack. Snowmelt moves through the snowpack vertically and horizontally. However, after the liquid-water conditioning of the snowpack has taken place, the movement of water through the pack is mostly straight downward to the ground/snow interface. Ice layers within the snowpack, however, tend to deflect the path intermittently, thereby resulting in an irregularly stepped pattern (see Figure 2-2). Analysis of meltwater movement through snow is more complicated than infiltration in a more static medium such as soil. The snowpack medium changes continuously as snow grains change in shape and size. In addition, as the snow melts and refreezes, impermeable ice layers form. Colbeck (1978) and Yosida (1973) have shown that as meltwater drains through the snowpack, there exists a wetting front that is isothermal at 0 °C and a lower layer in the snowpack below 0 °C. These wetting fronts may not be a uniform wave. Vertical flow fingers form around inhomogeneities in the snowpack (Marsh and Woo 1984). Because of these inhomogeneities in the snowpack, it is typically beyond the focus of operational snowmelt models to determine representation values of permeability and the effective porosity of the snow. The net storage effect on water draining through the snowpack is a time delay to runoff, on the order of 3 to 4 hr of storage time for moderately deep packs. In general, the time delay caused by transitory storage in the snowpack may be ignored when considering snowmelt or rainfall runoff from project basins whose areas exceed 518 or 777 sq km (200 or 300 square miles). This applies only to mountainous regions where slopes are adequate to ensure free horizontal drainage. Where horizontal drainage is inadequate (as in the Great Plains, in contrast to the mountainous region of the western United States), the delay to runoff caused by the snowpack may be much larger than for the vertical transit of water through the pack alone. Anderson (1973) has developed empirical relationships that represent drainage of snowmelt in the snowpack by using a time lag and attenuation.

2-4. Meltwater Infiltration

The ground conditions (both the soil mantle and underlying groundwater aquifers) are important in evaluating snowmelt runoff.

a. Unsaturated zone.

(1) The soil mantle functions as a reservoir, storing water, when available, to be used during periods when potential evapotranspiration exceeds current supply. In snow hydrology, there will be essentially no direct runoff until the soil storage is filled to its field capacity,

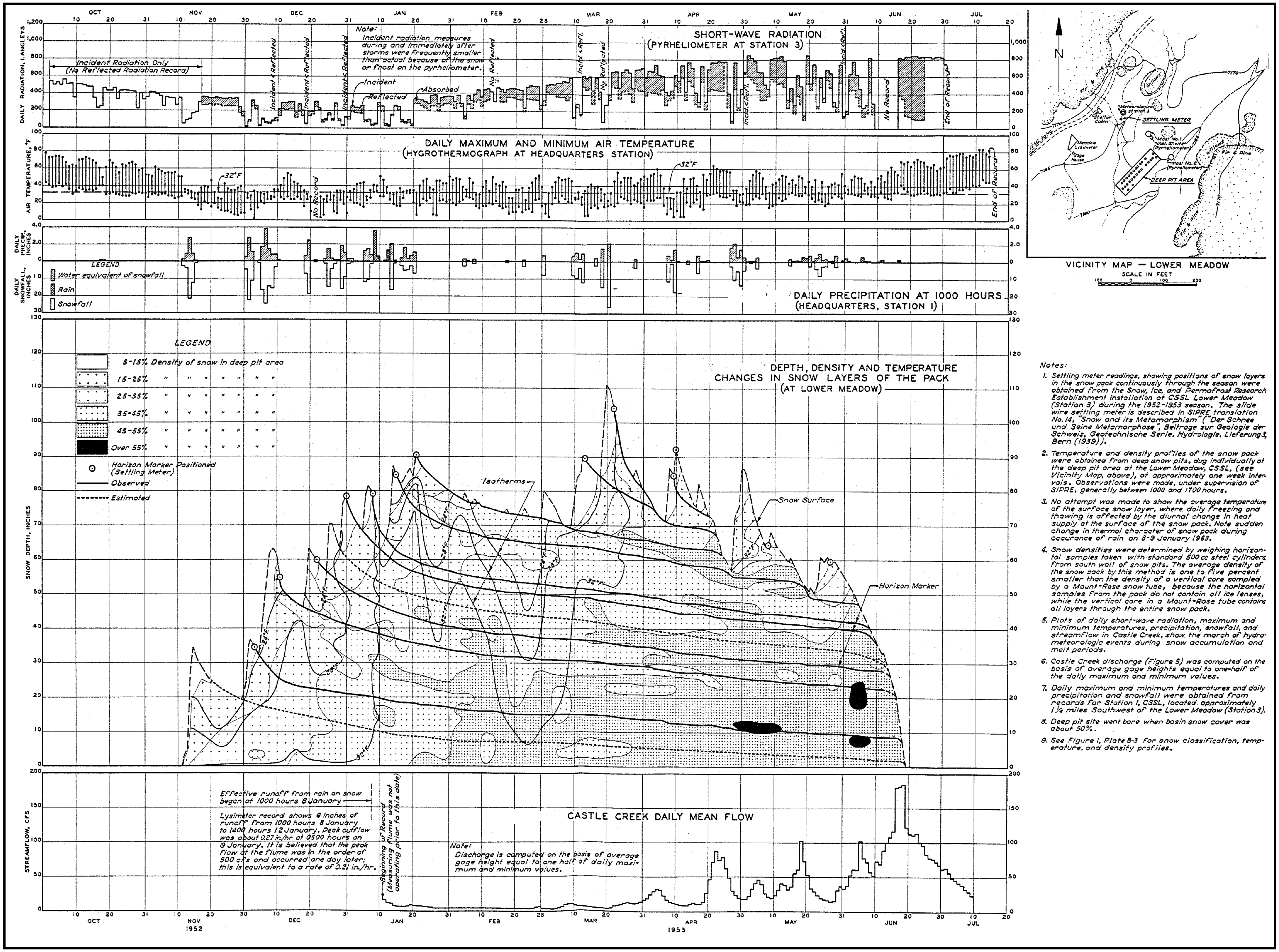

Figure 2-2. Snowpack characteristics (Plate 8-2, *Snow Hydrology*)

which is the total amount of water that can be held against gravity. The theoretical maximum capacity of the soil to withhold water permanently is determined by the difference between the “permanent wilting point” and the “field capacity” of the soil. For typical mountain soils, the maximum soil storage capacity ranges from approximately 10.2 to 20.3 cm (4 to 8 in.) of water in the zone from which stored water may be removed by transpiration or evaporation. After the field capacity of the soil has been reached, excess water may pass through the soil under gravitational force and appear later as subsurface or base flow components of streamflow. The time delay of transitory storage in the soil is integrated in the total basin storage effect.

(2) Direct measurements of soil moisture in project basins are generally lacking. While attempts are being made in some areas to obtain electrical resistance-type measurements of soil moisture beneath the snowpack, certain limitations currently restrict their use to qualitative interpretation. The principal difficulties are problems with calibration, lack of “buffer effect,” inconsistency of results, disintegration of the sensing unit, and unrepresentativeness of individual samples. Accordingly, basin soil moisture conditions are generally estimated from indirect relationships involving earlier precipitation, duration of rainless days, groundwater levels, stream discharges, time of year, or other factors associated with soil moisture variation. For areas of deep snow accumulation, as in the mountains of the western United States, the soil moisture deficit is satisfied early in the snowmelt period, and in many areas it may often be satisfied in the fall from rainfall or snowmelt. In the latter case, the soil beneath the snowpack remains at or above the field capacity throughout the winter, and any loss by evapotranspiration will usually be supplied by winter snowmelt or rainfall. For years in which the soil moisture capacity is not filled by fall or winter rainfall or snowmelt, it is necessary to estimate the condition of the soil from preceding hydrometeorological events.

b. Frozen soil. Cases are known where losses are reduced significantly because of frozen ground, thus increasing runoff. The ground is generally unfrozen beneath deep mountain snowpacks because of the flow of heat from the ground, together with the insulating effect of the snowpack. Frozen ground will occur during winter or early spring, in areas where snowpacks are shallow, and where prolonged periods of subfreezing air temperatures prevail. Such conditions are characteristic of the northern Great Plains regions of the United States. Gray and Prowse (1992) state that the infiltrability of frozen ground is the single most important factor affecting the apportioning of snow water between direct runoff and soil waters in most northern regions.

(1) In general, the effect of frozen ground is to inhibit infiltration. In cases where the soil pores are small, liquid water entering the ground will refreeze within the surface layer and will retard further infiltration. Accordingly, the concept of satisfying soil moisture deficits for unfrozen soil would not apply and, in addition, the basin time delay for water in transit would be considerably reduced.

(2) The factors that affect the role of frozen ground in snow hydrology include frost types and hydraulic properties, changes in the routing of water through a watershed because of frozen ground, and the features of a streamflow hydrograph during a storm on frozen ground. Several structurally and hydrologically different types of frost may form when the soil freezes. The type of soil frost primarily depends on the moisture content at the onset of freezing and, to a lesser extent, the type of soil that freezes. Dingman (1975) found four types of frost commonly mentioned in the literature, those being concrete, granular, honeycomb, and stalactite. Only concrete and granular frosts occur in sufficient quantity or remain long enough to be of any hydrological significance. Concrete frost is most common in bare or sparsely covered soil and is the predominant frost type in soils frozen deeper than a few inches. Granular frost is most common in soils with higher organic material content and shallow freezing.

(3) The most important hydrological features of frozen ground are the change in the soil's permeability and, to a lesser extent, the volume of water bound in the frozen soil. Concrete frost is generally impermeable, although it may be interrupted by discontinuities in the soil surface. Frozen soil can also retain significant amounts of water in the soil column, particularly during the spring thaw when the subsurface frost layer holds meltwater above it (Alexeev et al. 1972).

(4) Frozen ground interferes with the normal path and time of travel of water through the watershed to the stream channel. The rate at which water infiltrates into the soil depends greatly on the conditions at the surface, as infiltration capacity is determined by soil type, soil moisture condition, and soil frost conditions. When the rate of water delivered to the soil exceeds its infiltration capacity, the excess water at the surface becomes overland flow. This surface water may also be stored in surface depressions until it can infiltrate, flow overland, or evaporate.

(5) The extent to which frost interrupts the normal routing processes of a watershed depends upon the extent of frozen ground. Dunne and Black (1971) found the areal extent of concrete frost in pastureland to be important to the timing of runoff. Discontinuities in concrete frost are common in forested areas, allowing more infiltration during frost conditions (Trimble, Sartz, and Pierce 1958). A general progression of frost occurrence by land-use type has been identified by several investigators (Storey 1955, Pierce 1956, Dingman 1975). The susceptibility of a land area to freezing is inversely related to the amount of ground cover and proportional to the degree of compaction of the soil. The general sequence of land-use types in accordance with their degree of frost susceptibility is bare cultivated ground, grassland, pasture, softwood stands, and hardwood stands. The proportions of these land-use types in a watershed influence the extent of frost, and, thus, the change in how the watershed responds.

(6) Water can run off frozen ground during rain on bare ground, rain on snow, and during snowmelts. In each of these events, the frozen ground effect depends on its extent at the event's beginning and the persistence of the frost throughout.

(7) The principal effects of frozen ground on the outflow hydrograph of the watershed are faster response with higher peak flow and greater volume in the total hydrograph. Simulations compared with hydrographs of actual storms over frozen ground show a distinct quickening of response and an increase in the peak outflow during the storm (Anderson 1978, Stokely 1980, Peaco 1981). The water's inability to enter into the soil reduces the amount of groundwater storage and increases the total volume of the hydrograph (Haupt 1967). Dingman (1975) suggests that a bimodal hydrograph could result if there is substantial frost melting during the storm. Upon the melting of the frost, overland flow could be reduced and infiltration increased, causing a dip in the hydrograph until the interflow and baseflow response appeared. In operational snowmelt runoff models, the effect of soil frost is accounted for in a soil moisture routine by controlling soil parameters and transfer coefficients. The state factor for frozen ground is usually a frost index. Examples of the use of a frost index are given by Anderson and Neuman (1984) and Molnau and Bissell (1983).

c. Saturated zone. Delay of runoff by ground and channel storage is a basic hydrological phenomenon. Direct evaluation of groundwater storage through the use of well records is impractical in mountainous areas because of the wide variability of conditions in a drainage basin. Streamflow-recession analysis is a way to indirectly evaluate basin storage. Volumes of water "generated" in a given period can be determined by use of standard recession analysis techniques.

2-5. Glacier Effects on Runoff

The presence of glaciers in a watershed or larger basin significantly affects runoff volume, frequency, and variability (Lawson 1993). Partial glacierization of a basin by as little as a few percent of cover can cause moderate to extreme variations in peak runoff magnitude and frequency over days, years, and decades. Runoff is not directly related to precipitation within a glacierized basin, so it is, at present, difficult to predict because of a lack of glaciohydrologic data and a limited, rather rudimentary knowledge of the glaciohydrologic processes controlling runoff.

a. Runoff from glacierized areas of a basin is generally greater than that from nonglacierized areas with similar precipitation, often by 3 to 10 times. The majority of runoff from ice-covered areas comes during the melt season, generally from mid-May to mid-September. At progressively higher latitudes or in higher elevation basins, the time of flow is reduced. Because glaciers act as natural storage reservoirs that retain a large proportion of the winter precipitation in their accumulation areas, they generally moderate annual streamflow. During warm, dry, and sunny summers, water released from storage by melting of ice compensates for reduced runoff from precipitation. During cooler and wetter periods, the proportion of runoff from precipitation increases and supplements reduced meltwater runoff.

(1) Year-to-year variations in runoff also vary with percentage of glacier cover within the basin. Calculations of the coefficient of variation (CV) for annual runoff from partly glacierized basins suggest that there is minimum variability when ice covers between 35 and 45 percent of the basin area. The CV then is less than those of nearby nonglacierized basins, which tend to vary (and have a similar CV) as the precipitation totals vary. In addition, the CVs for monthly variations in runoff are lowest at the height of the melt season, but highest early in the season as the glacier's drainage system develops. In contrast, diurnal fluctuations in glacially fed rivers are greater than those in nonglacial rivers. They reflect primarily total energy input, which determines melt rates of the snowcover and glacier ice and, secondarily, the nature of the drainage system within the snowpack and glacier as it develops through the melt season.

(2) In addition, the timing of the peak diurnal discharge, as well as its magnitude, varies with the time of the melt season, occurring progressively earlier in the day with a larger magnitude and daily range later in the melt season. Peak seasonal flows are typically delayed compared with adjacent nonglacierized basins. For example, in the northwestern United States, discharge peaks in July or August, whereas it peaks in May in nonglacierized basins. This response reflects reduced albedos as the snow cover melts and maximum melt rates later in the year reflecting the minimal cloud cover and low precipitation of the region.

b. The effect of rainfall on runoff may differ from nonglacierized basins as well, reflecting the state of the drainage system within the snowpack and glacier. Early in the melt season, when drainage is incompletely developed, peak runoff from rainfall may lag significantly, whereas late in the season, when it is well developed, water movement through the glacier-covered area may be rapid and the response in runoff almost immediate. Flooding commonly follows heavy rainfalls after extended periods of high ice-melt rates, particularly when the drainage system is well-developed. Sudden, sometimes catastrophic, flooding also results from the unexpected release of water stored within or under the glacier or from drainage of ice-dammed lakes. Finally, snowfalls (rather than rain) at any time in the ablation season interrupt ice melt because of the reduced surface albedo, causing a decrease in runoff over several days or more.

c. The processes of snow metamorphosis and snowmelt upon the glacier follow those described elsewhere in this manual. A significant difference in defining snowmelt runoff, however, exists because the snowpack lies on glacier ice, which may be impermeable owing to seasonal freezing, and will therefore require warming or melting of internal passageways before runoff can actually take place. This process is not well-documented and its nature, the factors controlling it, and its rate cannot currently be defined. It is clear that runoff is significantly delayed because meltwater is stored within the snowpack above the ice. Only by surface drainage does the snowmelt slowly reach streams draining the ice-covered areas. Meltwater produced by ablation of glacier ice similarly is delayed from reaching basin streams by the early season absence of a well-connected drainage system inside and below the glacier. Therefore, while hydrometeorological analyses can be used to predict melting rates for the ice surface as it is gradually exposed by snowmelt, accurately predicting daily, monthly, or seasonal discharges remains elusive.

d. Drainage within (englacially) and below (subglacially) the glacier is inherently complex. In a general sense, water flows englacially either within the ice grains, eventually forming capillaries or small tubes that intersect larger ice-walled conduits within the glacier, or through larger drainage features, including crevasses, fractures, and moulins, that intersect or feed conduits at depth within the ice. The ice-walled conduits progressively join larger conduits at depth, forming an upward branching network. Once at the bed, water moves in conduits and cavities incised into the bed or into the overlying ice. These conduits are tributaries to larger tunnels discharging at the ice margin. Some water also flows as a thin film at the ice/bed interface or as groundwater in subglacial sediments. However, the configuration, distribution, and variability of this drainage system are incompletely described and are generally speculative.

e. Conceptual models specific to glacierized basins attempt to predict runoff by separating the procedure into two steps: one to calculate meltwater production and the other to calculate drainage, both from the glacierized and nonglacierized portions of the basin. Meltwater production is simulated by considering the physical processes and their effect on melt rate. Drainage from the glacierized part of the basin, however, is poorly simulated by existing models, mainly because of the lack of empirical data and lack of a sufficient understanding of the processes controlling flow rates and volume, and water storage. In some models, a linear reservoir with a retention time built in is used to simulate glacier drainage. In others, the glacier is considered an extremely thick snowpack and treated as such. Operational physical models that may or may not treat glacier drainage and storage separately include those of Anderson (1973) as modified by Nibler (cited in Fountain and Tangborn 1985), Baker et al. (1982), Lang (1980), Lang and Dayer (1985), Tangborn (1984, 1986), and those in Power (1985). In addition, none of the models is strictly physically based, but incorporate statistical treatments where process relationships are unknown. These conceptual models illustrate the present approaches to glacierized basin runoff predictions. In general, none of the operational models accurately predict the frequency and magnitude of peak, seasonal, and annual flows, and each of the models is basin-specific. Only Lang and Dayer (1985) apply their model to hourly and daily predictions of runoff; overall, their model best predicts seasonal runoff for an operational scheme.