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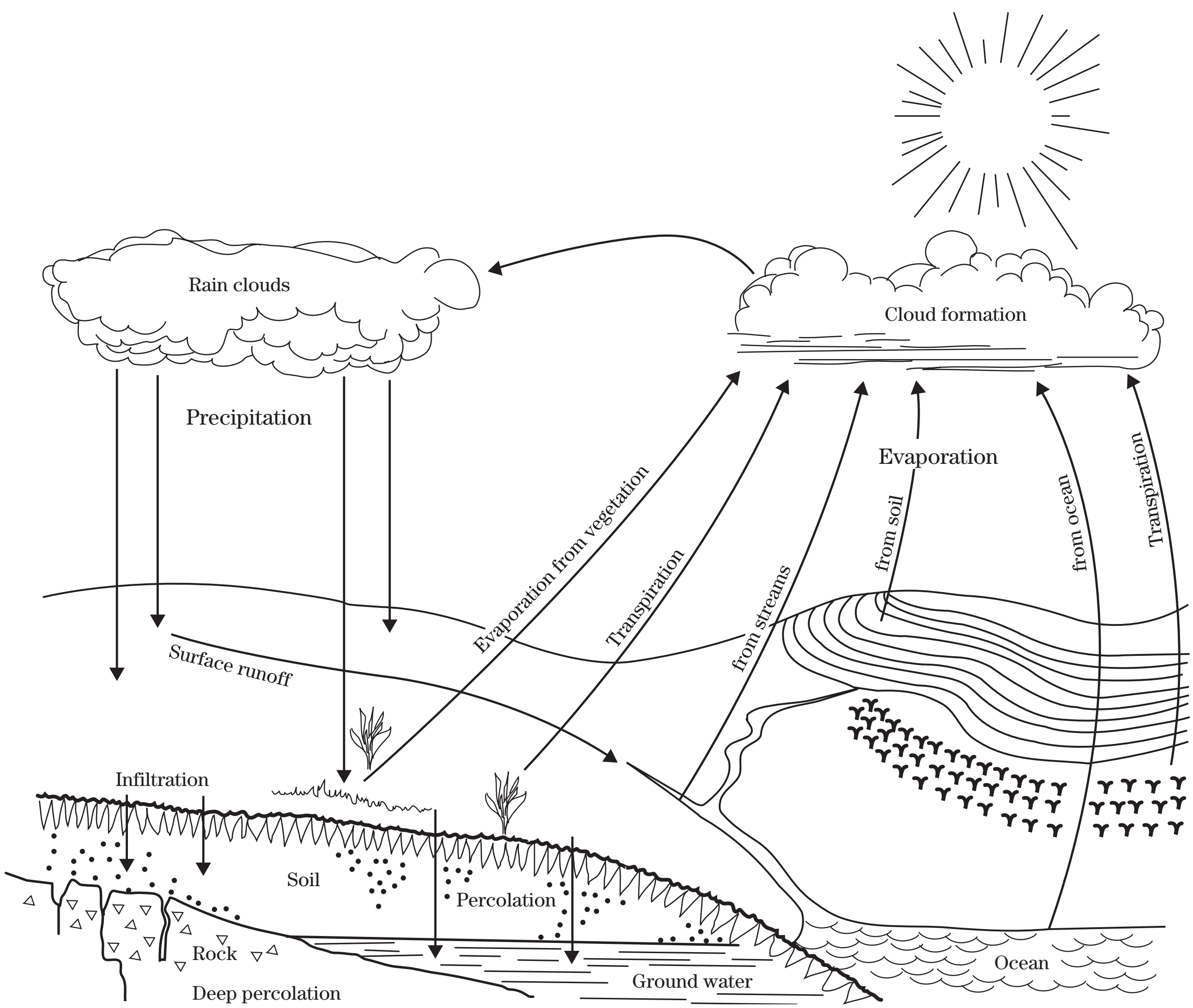
National Engineering Handbook

Chapter 11 Snowmelt



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Chapter 11 Snowmelt

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630.1100 Introduction

This chapter describes the basic physical processes that drive snowmelt and presents methods and guidance for estimating snowmelt runoff volumes and hydrographs for single events. These methods may also be used for short-term forecasts. In addition, a method is presented that may be used to derive flood frequency curves for snowmelt runoff from snow depth and temperature frequency data. Seasonal volume and long-range streamflow forecasting are not described here; the reader is instead referred to other publications for these topics (e.g., USDA SCS 1972b and 1990, Garen 1992) as well as the NRCS National Water and Climate Center:

(http://www.wcc.nrcs.usda.gov/wcc.html)

Snowmelt runoff is a major component of the hydrologic cycle in many regions and is an important consideration for water supply and design flood analysis. In some areas snowmelt event runoff may be more appropriate for the design of water storage facilities and hydraulic structures than rainfall storm runoffs described in National Engineering Handbook (NEH), section 4 (part 630), chapter 10 (USDA SCS 1972a). In addition, the annual peak flow in these areas can arise from either pure snowmelt or rainfall, or a combination of both, leading to a mixed frequency distribution, which is described in NEH, part 630, chapter 18 (USDA NRCS 2000). The modeling methods in this chapter may be used together with the methods described in NEH, section 4 (part 630), chapters 10, 16 (USDA SCS 1972a), and 18 to produce a mixed distribution flood frequency curve.

630.1101 Snowmelt theory

The thermodynamics of snowmelt are well understood and have been thoroughly described in numerous places. Among the early descriptions are those given by Clyde (1931), Light (1941), and Wilson (1941). One of the most thorough studies ever undertaken was that of the U.S. Army Corps of Engineers (COE) (1956), which is still often cited and regarded as a definitive work on the subject of snowmelt dynamics, as well as being a source of equations for practical modeling. This study was the basis of the snowmelt component in the hydrologic model SSARR (U.S. Army COE 1991). The work by Anderson (1968, 1976) has also led to an operational model in use by the National Weather Service (Anderson 1973). More recently, energy balance snowmelt models have been developed to operate on a spatially distributed basis, taking advantage of geographic information systems (GIS) and spatial data sets of elevation, vegetation, soils, and hydrometeorological variables. These include, for example, the models of Marks et al. (1998, 1999) and Tarboton et al. (1995). Descriptions of the snow energy fluxes appear in their papers. Other useful sources of information on snow thermodynamics and melt include Colbeck and Ray (1978), Gray and Male (1981), and American Society of Civil Engineers (1996). Many engineering hydrology textbooks also contain short, but useful, descriptions of snowmelt (e.g., Bedient and Huber 1992, Linsley et al. 1982).

(a) The energy balance

If all the heat fluxes toward the snowpack are considered positive and those away considered negative, the sum of these fluxes is equal to the change in heat content of the snowpack (ΔH) for a given time period. That is,

$$\Delta H = H_{rs} + H_{rt} + H_s + H_l + H_g + H_p \qquad [11\text{–}1]$$

where:
H_{rs} = net solar radiation
H_{rt} = net thermal radiation
H_s = sensible heat transfer from air
H_l = latent heat of vaporization from condensation or evaporation/sublimation
H_g = conducted heat from underlying ground
H_p = advected heat from precipitation

The solar radiation (H_{rs}) is the net of incoming minus reflected solar radiation. The reflection is because of the albedo of the surface, which varies with the age of the snow (decreases with age), the sun angle (lower in midday than in the morning and evening), and the contamination of the snow by dirt and debris (which reduces the albedo). The albedo is higher in the visible parts of the spectrum (0.28–0.7μ) than it is for the near infrared (0.7–2.8μ). For freshly fallen, clean snow, the visible albedo is very high (about 0.95–0.98), whereas the infrared albedo is somewhat lower (about 0.7–0.8). The thermal radiation (H_{rt}) is primarily the net of incoming radiation from the atmosphere, clouds, and surrounding vegetation minus the outgoing blackbody radiation from the snowpack itself.

Sensible heat transfer occurs when the air temperature is different from the snowpack temperature. If the air is colder, H_s is negative (heat leaves the snowpack), and if the air is warmer, H_s is positive (heat enters the snowpack). Latent heat is the energy released during a phase change of water from vapor to liquid to solid when condensation onto the snowpack occurs, or conversely, it is the energy extracted from the snowpack when evaporation or sublimation from the snowpack occurs. Condensation or evaporation/sublimation depends on the humidity of the air and the water vapor pressure gradient between the air and the snow surface. If the humidity is high, such that the vapor pressure of the air is greater than that at the snow surface (i.e., at the temperature of the snow), the vapor pressure gradient is towards the snow, and condensation will occur, in which case H_1 is positive. If the air is dry, evaporation and/or sublimation will occur, and H_1 will be negative. Sensible and latent heat transfers are enhanced under windy conditions.

Conduction of heat between the snowpack and the underlying soil occurs if there is a temperature difference between the two, H_g being positive if the snow is colder than the soil, and H_g being negative if the snow is warmer than the soil. Advected heat from precipitation H_p is positive if the temperature of the precipitation is warmer than the snow and negative if it is colder.

When the snowpack is in thermal equilibrium, $\Delta H=0$; a negative energy balance will cool the snowpack, while a positive energy balance will warm it. The snow cannot be warmer than zero degrees Celsius, and melt cannot occur in significant amounts until the entire snowpack has reached this temperature. Once the entire snowpack is isothermal at zero degrees Celsius, positive values of ΔH will result in melt:

$$M = \frac{\Delta H}{80B} \qquad [11\text{–}2]$$

where:
M = melt (cm)
ΔH = heat flux (cal/cm^2)
B = thermal quality of snowpack

The value 80 (cal/cm^3) is the latent heat of fusion. The thermal quality of the snowpack is the fraction of its water content that is in the solid phase. For a melting snowpack, B generally is in the range of 0.95 to 0.97, corresponding to 3 to 5 percent liquid water (U.S. Army COE 1956).

Table 11–1 summarizes each of the terms in the energy balance equation and their relative importance.

Table 11–1 Relative importance of energy balance terms

Term	% ΔH	Comments
H_{rs}, H_{rt}	60 – 90%	Controlled by terrain, season, cloud cover, shading, air temperature, humidity. H_{rs} and H_{rt} are generally of about the same magnitude, but different sign. H_{rt} is usually negative and dominates in winter. H_{rs} is positive and dominates in spring. During a crossover period in early spring before the onset of melt, H_{rs} and H_{rt} cancel each other, and the net is near zero.
H_s, H_l	5 – 40%	Controlled by temperature and humidity gradients and wind speed. H_s and H_l are usually of opposite sign, so they tend to cancel. That is, it is usually either warm (H_s +) and dry (H_l -) or cold (H_s -) and humid (H_l +). Sometimes H_s and H_l are of the same sign, but the magnitude is small (e.g., cold and dry). Occasionally both are positive and large (i.e., warm and humid), usually during high winds, such as during rain-on-snow events.
H_g	2 – 5%	Usually small because the temperature of the ground is generally about the same as the temperature of the snow. During melt, both ground and snow are at 0 °C, so $H_g = 0$.
H_p	0 – 1%	Heat content of precipitation is relatively small compared to latent heat required to melt snow, unless precipitation volume is very large and precipitation temperature is significantly greater than 0 °C.

(b) Energy sources and the behavior of snowmelt

To be able to understand and describe the behavior of snowmelt in a given situation, such as its amount and timing during a significant melt event, it is necessary to know which energy sources are dominant. The specific combination of temperature, precipitation, humidity, wind, and cloudiness during an event determines the streamflow response. It is possible, for example, for two events to have similar air temperatures and perhaps even precipitation amounts, yet have different responses because of the effects of the other hydrometeorological factors on the energy fluxes.

These considerations are particularly important during rain-on-snow events. In this situation, sensible and latent heat (H_s and H_l) can become substantial, if not dominant, sources of energy for snowmelt. This was clearly illustrated by Marks et al. (1998) for the February 1996 flood on the Willamette River in Oregon. It was the combination of warm temperatures plus high humidity and wind that supplied much of the energy for snowmelt, particularly in open areas. Immediately after the event, however, the energy balance returned to a more normal situation, dominated by net radiation.

This illustrates why temperature alone cannot always adequately represent the energy dynamics involved in a snowmelt runoff event. It is therefore important to know for any given event whether it was generated by clear weather snowmelt or by rain-on-snow. If it is known that the flows of interest are rarely affected by rain, then the energy balance is simpler, and temperature-based methods are likely to be adequate. If, on the other hand, it is known that the largest flows are caused by rain-on-snow, then it becomes more complex to model and predict, in that knowledge of not only temperature, but also of several other hydrometeorological variables is necessary to describe the snowmelt and runoff behavior of the event.

630.1102 Data sources

Data for evaluating snowmelt can come from hydrometeorological stations or remote sensing. Station data are available primarily from the National Weather Service (NWS) at the National Operational Hydrologic Remote Sensing Center (http://www.nohrsc.nws.gov) and the NRCS at the National Water and Climate Center (http://www.wcc.nrcs.usda.gov/wcc.html), with smaller networks run by various other agencies, such as the Forest Service. In addition to snow water equivalent, temperature, and precipitation, many sites monitor snow depth, and a few are equipped with humidity, wind, and solar radiation sensors. Much of these data are available in near real-time and are therefore available for both short- and long-term forecasting as well as for historical analysis. The NWS also has data from remote sensing, such as snow covered area maps from satellite images, and snow water equivalent from flight lines obtained by sensing gamma radiation from low-flying aircraft.

630.1103 Modeling snowmelt

Two basic approaches are used to model snowmelt for daily or shorter time steps. The most thorough method is to measure or estimate each term in the energy balance equation and to simulate the energy fluxes within the snowpack. This method is data intensive and sometimes cannot be done because of inadequate data or if this level of detail is unwarranted for the purpose at hand. The alternative is a melt index approach, the most common of which is the degree-day method, in which air temperature is used to index all of the energy fluxes. While the index approach has limitations, it is nevertheless commonly used because of its simplicity.

(a) Energy balance approach

Because of the large amount of data and the complexity of the processes involved, the energy balance approach is best implemented with computer models. Using fast computers with large disk storage capacities, along with geographic information systems (GIS) and spatial data layers of elevation, soils, vegetation, and hydrometeorological inputs, such models are now feasible for operational use. For example, the model of Marks et al. (1998, 1999) is documented and has been applied to several watersheds in the Western United States. This model can be integrated into a complete hydrologic simulation model (Schumann and Garen 1998, Garen et al. 2001, Garen et al. 2002). This type of modeling, however, requires considerable effort in data preparation, hence is warranted only when a very detailed and accurate simulation is needed.

Equations used by the U.S. Army Corps of Engineers in the HEC-1 model (U.S. Army COE 1998) are index equations that include the most important parameters for the rainy and non-rainy periods. Instead of modeling the energy balance, regression analysis was used to determine the coefficients for the significant measured parameters, such as temperature, wind, and radiation. The resulting equation for non-rainy periods in partially forested areas is:

$$M = C\left[.002I_i\left(1-a\right)+\left(.0011v+.0145\right)\left(T_a-T_F\right)+.0039v\left(T_d-T_F\right)\right] \qquad [11\text{–}3]$$

and for melt during rainy periods:

$$M = C[.09 + (.029 + .00504v + .007P)(T_a - T_F)] \qquad [11\text{–}4]$$

where:
- M = melt (inches/day)
- I_i = incident solar radiation on a horizontal surface (langleys/day)
- a = albedo of the snow
- v = wind speed (miles/hour) 50 feet above the snow surface
- T_a = air temperature (°F)
- T_F = freezing temperature (°F, allowed to vary from 32 °F for spatial and temporal fluctuations)
- T_d = dewpoint temperature (°F)
- P = rainfall (inches/day)
- C = coefficient to account for variations

(b) Degree-day method

The degree-day method is a temperature index approach that equates the total daily melt to a coefficient times the temperature difference between the mean daily temperature and a base temperature (generally 32 °F or 0 °C).

$$M = C_M\left(T_a - T_b\right) \qquad [11\text{–}5]$$

where:
- M = snowmelt in in/d (mm/d)
- C_M = the degree-day coefficient in in/degree-day F (mm/degree-day C)
- T_a = mean daily air temperature °F (°C)
- T_b = base temperature °F (°C)

The coefficient C_M varies seasonally and by location. Typical values are from 0.035 to 0.13 inches per degree-day Fahrenheit (1.6 to 6.0 mm/degree-day C). A value of 0.060 inches per degree-day Fahrenheit (2.74 mm/degree-day C) is often used when other information is lacking. C_M has also been related to snow density and wind speed (Martinec 1960) and to accumulated degree-days and elevation (Rosa 1956). These variations reflect the different energy dynamics and changing snowpack conditions over time and space. The fact that it varies like this demonstrates that this single index (temperature) cannot represent all of the relevant processes, so to compensate, the degree-day coefficient must change with the changing conditions. During rain-on-snow, the degree-day method must be used with caution as it most likely is not valid. It is most applicable to clear weather melt in forested watersheds.

630.1104 Snowmelt runoff

(a) Regional analysis

Several methods can be used to do a regional analysis of snowmelt runoff. For seasonal volumes the reader should refer to NEH, section 22 (USDA SCS 1972b). In some areas it may not be possible to separate the snowmelt runoff events from the rain or rain-on-snow events. In these cases the normal procedure would be to regionalize the runoff peaks or volumes without regard to cause. Methods for statistical regionalization are described in NEH, part 630, chapter 18 (USDA NRCS 2000).

Where the major flood events are from rainfall during the snowmelt season, snowmelt is commonly treated as baseflow or quick return flow and the events are modeled as rainfall runoff using methods described in NEH, section 4 (part 630), chapters 10 and 16 (USDA SCS 1972a). Rain-on-snow events may also be modeled (Marks et al. 1998, U.S. Army COE 1998, Martinec et al. 1994, U.S. Bureau of Reclamation 1966), but the curve number method of chapter 10 is not an appropriate method of determining the losses. Snowmelt baseflow must not be ignored when modeling for dam design with such models as TR-20 (USDA SCS 1992), WinTR-20 (USDA NRCS 2004 draft), or SITES (USDA NRCS 2001).

In some areas, such as the prairies of eastern Montana, snowmelt events can be separated quite easily from rainfall events strictly by season. Snowmelt typically occurs during February, March, and April when precipitation amounts are generally quite small. The dates of individual events are noted and compared with precipitation and temperature records to verify the cause. Snow-on-ground records can also be accessed and checked. The record for a crest-stage gage in northern Montana is shown in figure 11–1 as an example of a primarily snowmelt runoff stream. All of the runoff events except the one in July of 1970 were the result of snowmelt.

Frequency analysis can be done for the peak flow and runoff volumes from the separate causes. Figures 11–2 and 11–3 show results of a regional frequency analysis for runoff volume in the eastern Montana region (Van Mullem 1994). These figures may be used to estimate hydrographs directly, similar to the methods in NEH, section 4 (part 630), chapter 21 (USDA SCS 1972a), or to calibrate a snowmelt runoff model.

Figure 11–1 Crest-stage record for a snowmelt runoff stream in Montana (from USGS open file report 78-219, 1978)

06138800 Spring Coulee near Havre, MT

Location—Lat 48°25'20", long 109°51'49", in NW1/4 sec. 25, T.31 N., R.14 E., Hill County, Hydrologic Unit 10050005, at culverts on county road, 13 miles (20.9 km) southwest of Havre.
Drainage area—17.8 mi^2 (46.1 km^2).
Records available—May 1959 to September 1973.
Gage—Crest-stage gage installed May 26, 1959. Altitude of gage is 2,670 ft (814 m), from topographic map.
Annual maximum data—

Water year	Date	Gage height (ft)	Discharge (ft^3/s)
1959	eMar. 11, 1959	5.60	257
1960	Mar. 17, 1960	---	a10
1961	---	---	(c)
1962	Mar. 18, 1962	1.32	48
1963	Feb. 3, 1963	1.35	46
1964	---	---	(c)
1965	Apr. 6, 1965	2.01	77
1966	Mar. 9, 1966	6.24	345
1967	Mar. 22, 1967	3.41	146
1968	---	---	(c)
1969	Mar. 26, 1969	5.56	255
1970	July 13, 1970	.73	22
1971	Feb. 12 1971	5.75	263
1972	Mar. 13, 1972	(f)	a2
1973	Feb. 28, 1973	(f)	a3

a About.
c No evidence of flow during year.
e Prior to installation of gage.
f Peak discharge did not reach bottom of gage.

Figure 11–2 Spring season snowmelt, 25-year, 7-day runoff volume (inches)

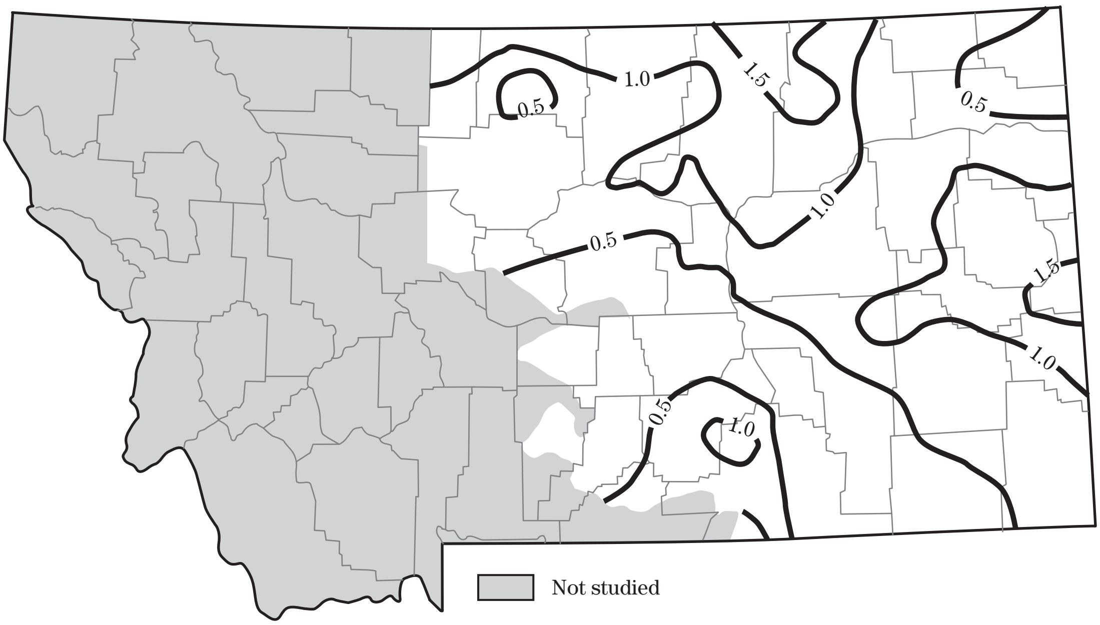
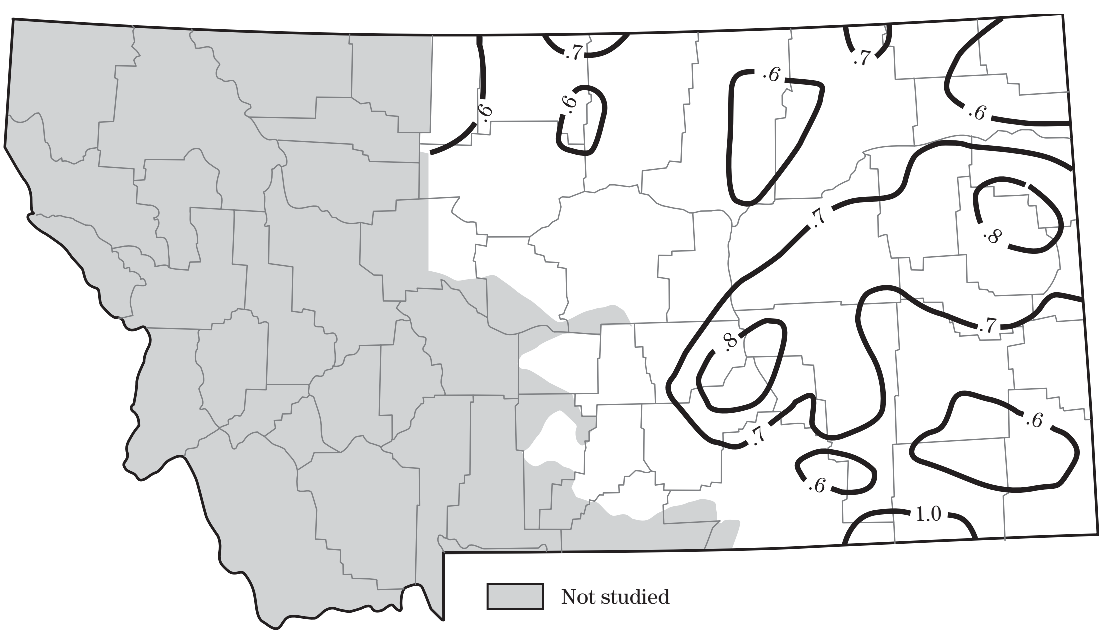


Figure 11–3 Spring season snowmelt, 3-day, 7-day runoff ratio



(b) Spatial variability of snow cover

Snow cover information is an important element in all hydrologic problems that involve snowmelt. The areal extent of the snow cover determines the area contributing to runoff at any given time during the melt period. For mountain basins the areal extent of the seasonal snow cover decreases gradually during the snowmelt season, which may last for several months. The depletion pattern varies with the terrain. Elevation is the dominant variable for snow cover depletion because of the higher accumulation of snow with elevation (U.S. Army COE 1956). Within an elevation zone, aspect, slope, and forest cover all are important variables. For mountain areas, similar patterns of depletion occur from year to year and can be related to the snow water equivalent (SWE) at a site, accumulated ablation, accumulated degree-days, or to runoff from the watershed (Martinec et al. 1994, U.S. Army COE 1991, Anderson 1973).

Prairie snowpack is not uniform either and varies because of aspect and wind as well as by cover type. South facing slopes have less snow, and north facing slopes have more. Windswept areas and ridges may be nearly bare, while drifts in draws and coulees may be deep (Cooley 1988). Depletion patterns in prairie areas are more variable, so it is more difficult to develop depletion curves from historical data. Steppuhn and Dyck (1973) showed that with sampling stratification (i.e., sampling by cover and landscape type) fewer measurements of snow depth and density were needed to determine the SWE over a watershed accurately. Emerson (1988) applied this sampling technique to a watershed in North Dakota. The resulting SWE map is shown in figure 11–4.

Another way to represent spatial variability of snow cover is with a depth distribution pattern (Donald et al. 1995). The depth distribution pattern shows the percentage of the watershed in each range of SWE values. Figure 11–5 shows the depth distribution pattern for the watershed in figure 11–4. The depth-distribution data, when accumulated, results in a cumulative distribution function. Figure 11–6 is the resulting cumulative distribution from figure 11–5.

Figure 11–4 Snow water equivalent determined by ground survey in the West Branch Antelope Creek watershed on February 27, 1979 (Emerson 1988)

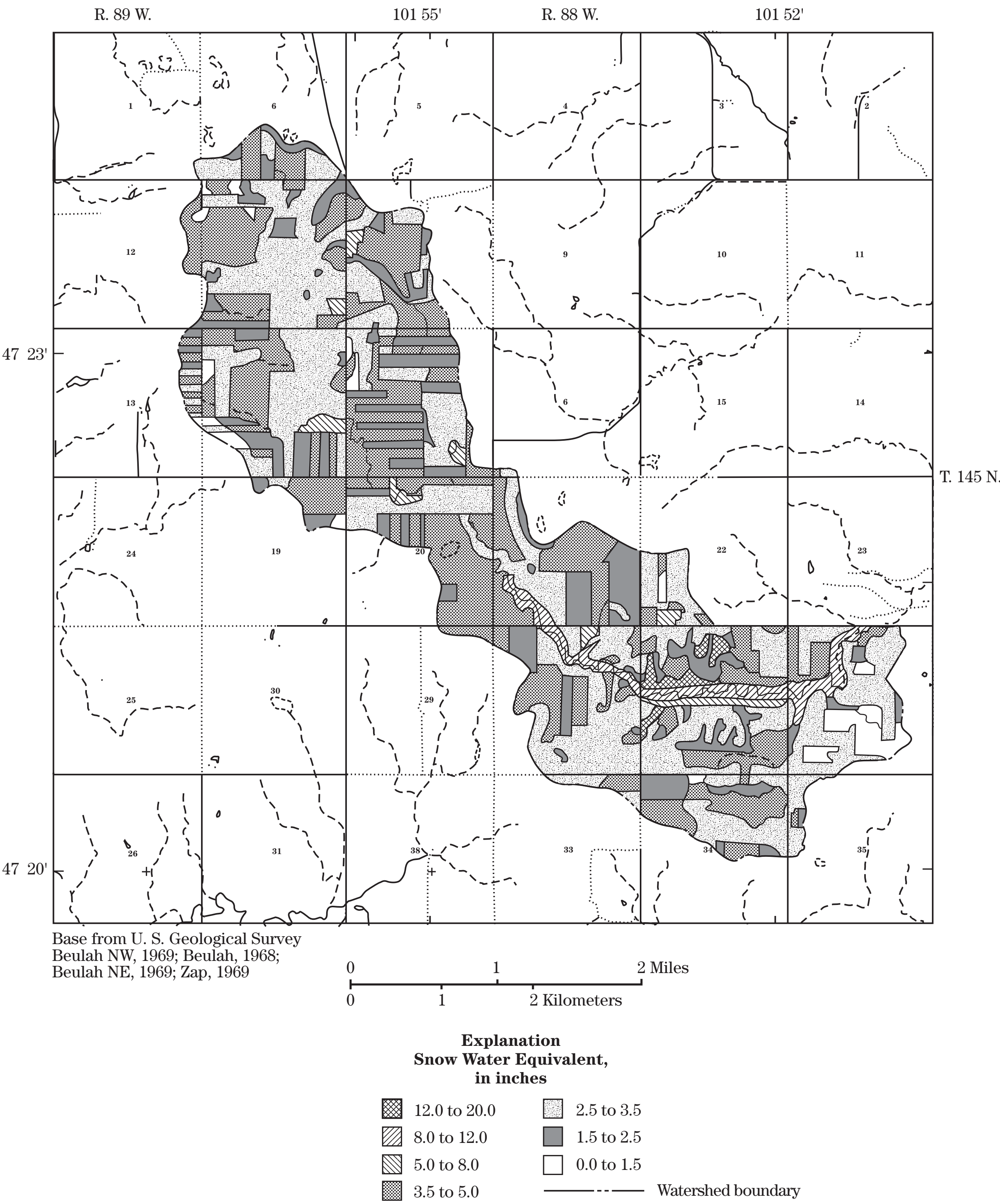


Figure 11–5 Snow water equivalent distribution for the data shown in figure 11–4 (mean SWE is 3.6 inches)

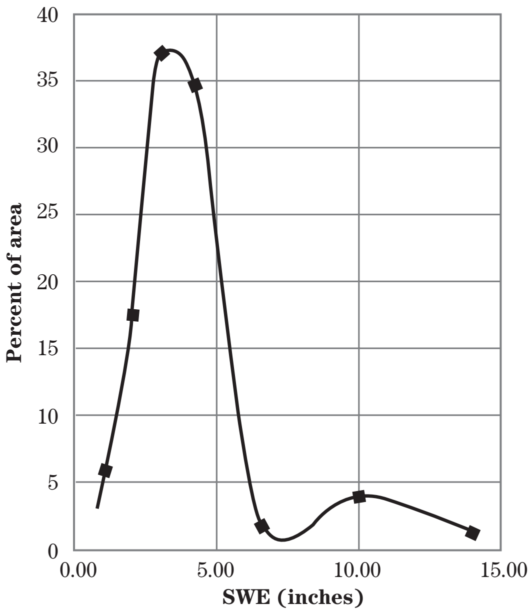
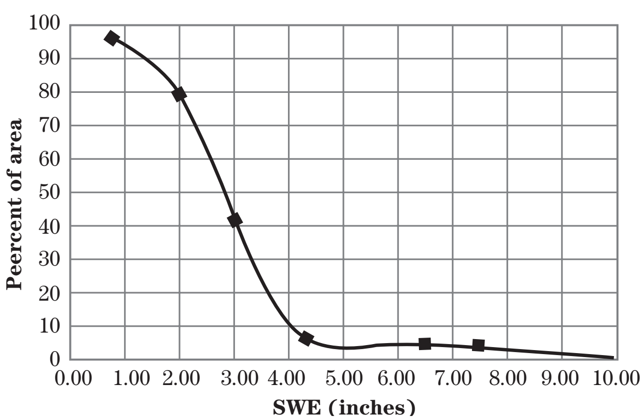


Figure 11–6 Cumulative SWE distribution derived from figure 11–5



(c) Temperature and precipitation during the melt period

Current conditions and forecasted weather conditions can be used for short-term snowmelt runoff forecasts (up to 7 days). For longer or more distant periods, average conditions are often assumed. Another technique is to use a historical period of about 30 years to obtain a wide range of possible outcomes that may then be statistically analyzed.

To simplify computations for frequency event modeling, regionalization of temperatures during the melt period may be done by making Temperature-Duration-Probability (TDP) studies. TDP is the frequency analysis of maximum temperatures for several durations. The maximum daily mean temperatures, in degree-days, during the usual melt period are found for each of several durations for each year in the period of record. A frequency analysis is then made for each duration.

Figure 11–7 shows TDP curves when degree-days accumulated over the entire duration at a particular frequency are plotted against days. The lines can be represented with a power function:

$$T_D = aD^b \qquad [11\text{–}6]$$

where:

T_D = accumulated degree-days for a duration of D days
a = value of 1-day maximum temperature
D = duration as number of days
b = slope of the line

Figure 11–7 TDP curves for 50-year return period at three stations in eastern Montana

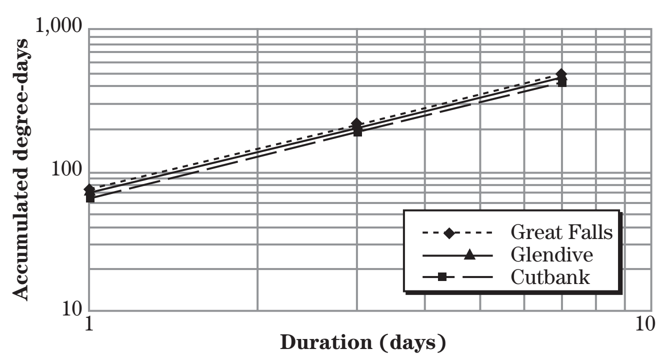


Figure 11–7 shows that TDP curves in a region are quite uniform. This enables them to be easily regionalized (Van Mullem 1998).

The sequence of daily temperatures within a given duration can be determined by estimating an average temporal distribution of total accumulated degree days.

The diurnal variation in temperature may be estimated by finding the average variation for the time of the year and the location and then applying that to the mean values. A model that uses the diurnal variation and has a time step less than a day results in a more accurate representation of runoff and better prediction of peak discharge than a daily model. Example 11–4 in 630.1105 shows the application of the diurnal variation in a hydrograph model.

If precipitation occurs during a runoff period, it must first be determined whether the precipitation is rain or snow. Snow is added to the remaining SWE while rainfall on the snowpack either fills available void space within the pack and remains there (as a liquid or it may freeze), or it may percolate through the pack and be available for infiltration and runoff. Rainfall on the snowpack can result in a heat exchange that contributes to snowmelt; however, the melt from rainfall is relatively small compared to the quantity of the rainfall itself.

The importance of rainfall during the snowmelt period is a regional factor. It is important in the Pacific Northwest, but it may be ignored for the short melt period on the northern Great Plains.

(d) Infiltration and losses

Snowmelt as determined from the degree-day equation is generally assumed to be the total ablation of the snowpack, and evaporation and condensation are ignored for short-term runoff modeling. The difference between the melt volume and the runoff volume is considered a loss and is assumed to be infiltration into the soil and groundwater storage. These losses are not expected to return to the stream during the event, but may contribute to baseflow.

Infiltration losses under a snowpack are difficult to base on the soil and cover characteristics because of varying frozen ground conditions (Guymon 1978). Instead, the infiltration or loss parameter may be selected based on calibrating the model so that runoff volumes computed from a known volume of snowmelt agree with measured volumes of runoff from a watershed.

$$\text{Runoff} = C_R M \qquad [11–7]$$

The runoff coefficient (C_R) is the ratio of runoff to snowmelt (M). It takes care of all the losses between the snowmelt and the outflow from the watershed. The coefficient varies widely from watershed to watershed from as little as 0.1 to more than 0.9. The ratio may be related to soil and cover types and to total precipitation (Farnes 1971). It also varies seasonally, generally decreasing as evapotranspiration losses increase as the melt progresses (Martinec et al. 1994).

Infiltration equations generally express the infiltration rate either as a function of time or of cumulative infiltration amount. Those equations that use time (e.g., Horton's equation) are not very suitable for modeling. Any of several equations that relate infiltration rate to cumulative infiltration amount may be used with snowmelt. These equations include the uniform loss rate, exponentially declining loss rate, and the Green-Ampt equation. The various infiltration methods are described thoroughly in many standard hydrology textbooks (e.g., Bedient and Huber 1992). Note that the curve number equation described in NEH, section 4 (part 630), chapter 10 (USDA SCS 1972a), is used as an infiltration model in NEH, section 4 (part 630), chapter 16 (USDA SCS 1972a), but is not recommended to determine losses from snowmelt.

Because the moisture and frost conditions of the soil are not known, the simpler methods are probably adequate. For infiltration loss estimates, the HEC-1 model (U.S. Army COE 1998) uses either the constant rate or the exponentially declining loss rate methods. The SRM model (Martinec et. al 1994, also at http://hydrolab.arsusda.gov/cgi-bin/srmhome) uses the runoff coefficient method. These methods are illustrated in examples 11–1 to 11–3.

Example 11–1 Runoff coefficient method

Given: The daily mean temperatures, a beginning SWE of 2.46 inches, a melt rate coefficient of 0.06 inch per degree-day, and a runoff coefficient of 0.5.

Find: The estimated daily runoff in watershed inches.

Solution:

Date	Average watershed temperature (°F)	Degree -days 1/	Total available SWE 2/ (in)	Estimated melt 3/ (in)	Estimated runoff 4/ (in)
April 5	32	0	2.46	0	0
April 6	35	3	2.46	.18	.09
April 7	34	2	2.28	.12	.06
April 8	36	4	2.16	.24	.12
April 9	48	16	1.92	.96	.48
April 10	43	11	0.96	.66	.33
April 11	42	10	.30	.30 5/	.15
April 12	40	8	0	0	0
Totals				2.46	1.23

1/ Degree-days = T – 32.
2/ Available SWE = previous SWE – preceding days melt.
3/ Using C_M = .06 in the equation $M = C_M(T - 32)$.
4/ Using C_R = 0.50 in the equation Runoff = $C_R M$.
5/ Melt is limited to the available SWE.

Example 11–2 Constant loss rate

Given: The melt rates from example 11–1 and a constant loss rate of 0.23 inches per day.

Find: The estimated runoff in watershed inches.

Solution:

Date	Snowmelt (in)	Infiltration (in)	Runoff (in)
April 5	0	0	0
April 6	.18	.18	0
April 7	.12	.12	0
April 8	.24	.23	.01
April 9	.96	.23	.73
April 10	.66	.23	.43
April 11	.30	.23	.07
April 12	0	0	0
Totals	2.46	1.22	1.24

Example 11–3 Exponentially declining loss rate

Given: The maximum loss rate declines according to the equation

$$i_t = \frac{i_o}{r^{c\sum I}} \qquad \Sigma I > 0$$

where:

i_o = initial maximum loss rate
i_t = maximum loss rate at time t
r = rate of decline
c = exponent parameter
$\sum I = \sum_{j=1}^{t} i_j$ = accumulated loss up to time t

Find: The runoff in inches for i_o = 0.25 inch per day, r = 4, and c = 0.1.

Solution:

Day	Snowmelt (in)	Accumulated loss (in)	Maximum loss rate (in/d)	Actual loss (in)	Runoff (in)
April 5	0	0	0.250	0	0
April 6	0.18	0	0.250	0.18	0
April 7	0.12	0.18	0.244	0.12	0
April 8	0.24	0.30	0.240	0.24	0
April 9	0.96	0.54	0.232	0.23	0.73
April 10	0.66	0.77	0.225	0.23	0.43
April 11	0.30	1.00	0.218	0.22	0.08
April 12	0	1.22	0.211	0	0
Total	2.46			1.22	1.24

Note that for this small amount of melt the loss rate did not decline much.

630.1105 Runoff hydrographs from snowmelt

(a) Unit hydrograph method

Snowmelt runoff hydrographs may be developed by using the unit hydrograph method as described in NEH, section 4 (part 630), chapter 16 (USDA SCS 1972a). To obtain realistic hydrographs with this method, the computational interval needs to be about 0.133 the time of concentration (T_c) (see chapter 16). For small watersheds this requires intervals much less than the 24-hour daily melt values. Because the melt rate varies over the day, this variation must be used to obtain realistic snowmelt hydrographs from small watersheds. The variation in melt rate is approximated by the variation in temperature. For illustration, a sine curve will be used here to describe the variation of temperature within a day. This function is often used (e.g., Anderson 1973, US Army COE 1998) although it does not represent nighttime temperatures very realistically (it causes the temperature to increase before sunrise); other reasonable functions could also be used. Example 11–4 illustrates the application of the diurnal variation in a hydrograph model.

Using the sine curve, the temperature at any time may be determined from:

$$T = T_a + A\{\text{Sin}[15°(t + C)]\} \qquad [11–8]$$

where:
- T = temperature at time t
- T_a = mean temperature for the day
- A = amplitude, $(T_{max} - T_{min})/2$
- t = hour of the day
- C = time shift in hours

Figure 11–8 is an example of the hourly temperature where T_{max} is 75 °F, T_{min} is 45 °F, and the time shift is 16 hours. This places the minimum temperature at 0200 and the maximum temperature at 1400. In general, the time shift is computed as 30 – maxhr, where maxhr is the desired hour (24-hour clock) of the maximum temperature. The melt for any period of Δt hours is

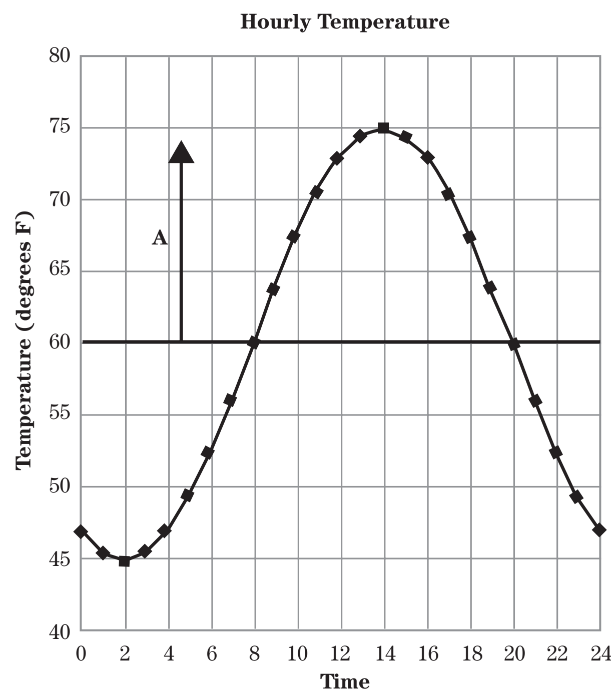
$$M = \frac{\Delta t}{24} C_M (T - T_b) \qquad [11–9]$$

where:
- C_M = daily melt coefficient
- T_b = base temperature

The hourly runoff values from example 11–4 can then be entered into a hydrograph model, such as TR–20 (USDA SCS 1992), WinTR-20 (USDA NRCS 2004 draft), or SITES (USDA NRCS 2001) (the runoff is entered as a rain table with the CN=100), and the snowmelt hydrograph is produced. Although melt and runoff can be computed for shorter time increments, the hourly values are satisfactory since the TR-20 model interpolates for the shorter computational interval.

Figure 11–9 shows snowmelt hydrographs from a 10-square-mile watershed with T_c of 3.35 hours. Both hydrographs have the same volume of runoff. One is computed from the runoff distribution in example 11–4; the other has a uniform rate of melt over the 24 hours.

Figure 11–8 Hourly temperatures (T_{max} = 75 °F and T_{min} = 45 °F)



Example 11–4 Application of diurnal variation in a hydrograph model

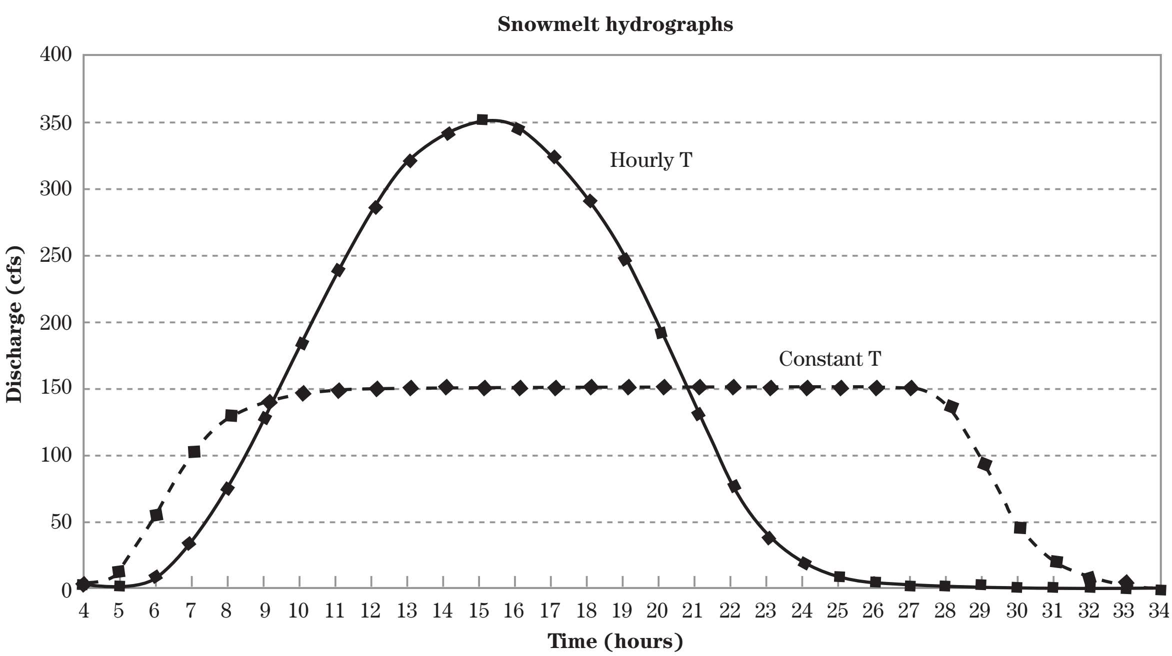
Given: The hourly temperatures shown in figure 11–8.

Find: The hourly melt for $C_M = 0.06$ and the hourly runoff for a constant loss rate of 0.05 in/ hr.

Solution:

Hour	Temperature	Melt	Runoff	Hour	Temperature	Melt	Runoff
0	47.01	0.038	0.000	13	74.49	0.106	0.056
1	45.51	0.034	0.000	14	75.00	0.108	0.058
2	45.00	0.033	0.000	15	74.49	0.106	0.056
3	45.51	0.034	0.000	16	72.99	0.102	0.052
4	47.01	0.038	0.000	17	70.61	0.097	0.047
5	49.39	0.043	0.000	18	67.50	0.089	0.039
6	52.50	0.051	0.001	19	63.88	0.080	0.030
7	56.12	0.060	0.010	20	60.00	0.070	0.020
8	60.00	0.070	0.020	21	56.12	0.060	0.010
9	63.88	0.080	0.030	22	52.50	0.051	0.001
10	67.50	0.089	0.039	23	49.39	0.043	0.000
11	70.61	0.097	0.047	24	47.01	0.038	0.000
12	72.99	0.102	0.052				
Totals						1.719	0.568

Figure 11–9 Snowmelt hydrographs for example 11–4 comparing hourly rate with a constant daily rate for 10 mi^2 watershed



Note: This plot was calculated using HEC–1 with the runoff data from example 11–4 above.

(b) Recession curve method

This method is especially applicable to deep snowpack areas where the melt period extends for more than a week or two. It also fits the concept of subsurface flow, which dominates in forested watersheds, rather than overland flow.

Runoff during the recession of a hydrograph when there is no snowmelt or rainfall may be represented by the equation

$$Q_n = Q_0 k^n \qquad [11\text{–}10]$$

where:
Q_n = discharge after n days
Q_0 = initial discharge
k = recession constant

This equation is in standard texts. The runoff on any day during the melt period may then be represented as

$$Q_{n+1} = C_R M_{n+1}\left(1 - k_{n+1}\right) + Q_n k_{n+1} \qquad [11\text{–}11]$$

where:
C_R = runoff coefficient
M_{n+1} = snowmelt for day n+1

This relationship has been modified by Martinec, et al. (1994) to consider the fact that the coefficient k increases with decreasing discharge so that

$$k_{n+1} = xQ_n^{-y} \qquad [11\text{–}12]$$

where:
x and y = constants determined for a given basin by analysis of the recession curves

The recession curve method is used by the SRM model to predict daily flow values. Peak flow for any day may be estimated from the daily flow by considering the normal daily fluctuation.

(c) Water movement through snow

Snowmelt occurs in the upper layer of snow and generally percolates very slowly to the ground surface. The percolation rate is highly variable with a typical range from 3 inches per hour to 3 feet per hour (.08–.9 m/h). The rate is dependent on the internal structure of the snow, the condition of the snowcover, and the amount of water available at the surface. For deep mountain snowpacks the additional time required for the meltwater to reach the stream channel can be significant and will vary during the snowmelt season.

The percolation rate for wet snow is typically 1 to 3 feet per hour (Wankiewicz 1978). For the shallow depths of 1 to 2 feet on the prairie, the time required for percolation is about 1/2 to 1 hour. This additional time needs to be added to the time of concentration or lag time of the watershed for hydrograph modeling.

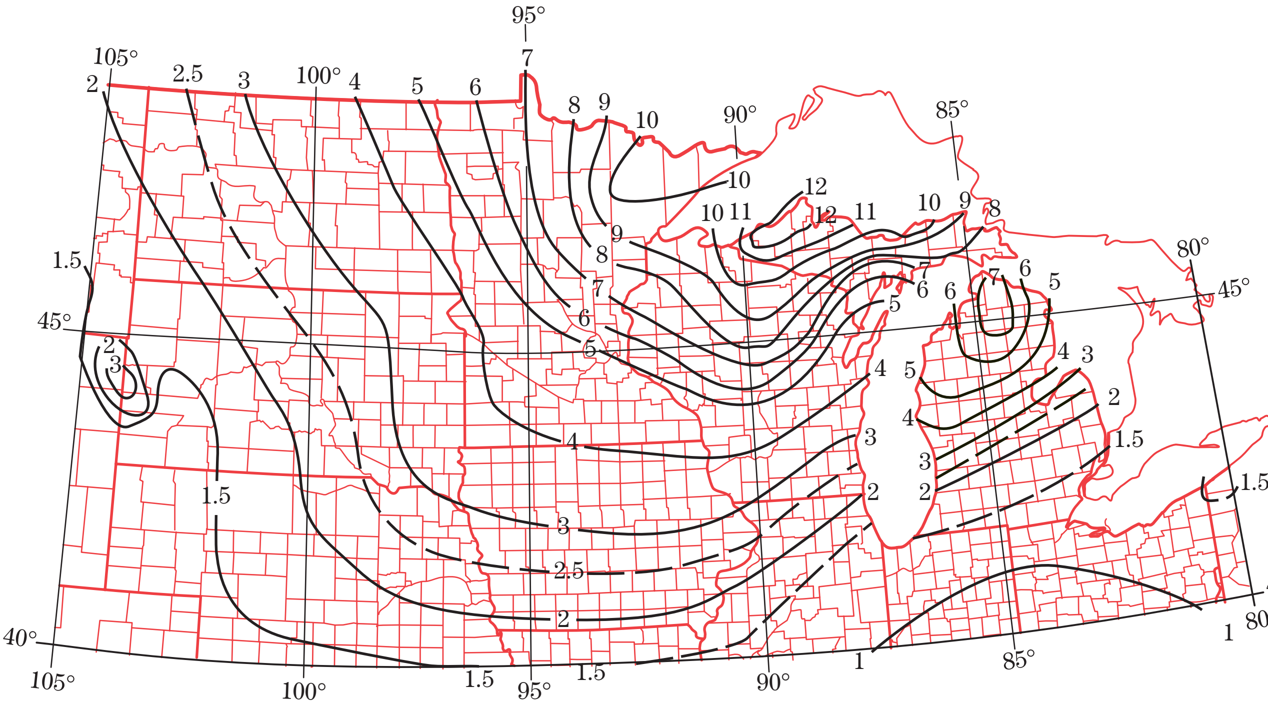
After reaching the ground surface, the travel time is again highly variable and dependent on the snow and surface conditions. Although additional delays of significant time may occur at a site, overall the flow of meltwater after reaching the bottom of the snowpack is almost as fast as the flow of rainwater on the soil without snowcover. This is attributed to the creation of channels at the snow-soil interface (Obled and Harder 1978). Therefore, no additional time needs to be added to the time of concentration for overland flow under snow.

(d) Snowmelt runoff by frequency

If the amount of SWE by frequency and the temperature during the melt period for a watershed is known, the runoff by frequency may be determined. Figure 11–10 is an example of a snow water equivalent by frequency map. The map is taken from the U.S. Department of Commerce Weather Bureau Technical Paper 50 (USDOC 1964), which includes maps for both the March 1 to 15 and March 16 to 31 periods with probabilities from 50 percent to 1 percent. Snow depth frequency maps (Van Mullem 1992) may be converted to SWE by assuming an average snow density.

Van Mullem (1998) gives a procedure applied in eastern Montana to compute runoff for selected frequencies.

Figure 11–10 Maximum March 16–31 snow water equivalent (inches) expected to be equaled or exceeded once in 25 years



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