GEOG415  Lecture 8: Snow Hydrology

Snow as water resource

Snowfall on the mountain ranges is an important source of water in rivers.

At Lake Louise, 55% of annual precipitation occurs as snow.

What is the difference between the “fates” of winter and summer precipitation?

A large portion of flow in mountain streams occur in April-June. Is Bow River similar?
To evaluate snowmelt runoff, what do we need know?

(1)

(2)

(3)

Measurement of snowfall

Snowfall measurement is considerably more difficult than rainfall measurement.

Why?

Snow is collected in an electrically heated gauge, and the weight of water is continuously recorded.

Snow gauges require wind shields.

Singh and Singh (2001, Snow and glacier hydrology, Kluwer Academic Publishers. Fig. 4-19.)

Dingman (2002, Physical hydrology. Prentice-Hall. Fig. 5-5.)
Turbulence around the gauge orifice causes the gauge to catch less than actual amount of snowfall.

Gauge catch deficiency = \( \frac{\text{actual} - \text{measured}}{\text{actual}} \)

Extra care should be taken to obtain and interpret snowfall data. See Environment Canada’s web site for guidelines for meteorological measurements.

(www.mb.ec.gc.ca/air/autostations/ab00s07.en.html)
Snowpack measurement

The amount of snow on the ground is expressed as snow water equivalent (SWE), usually in mm.

$$\text{SWE} = \text{snow depth} \times \text{average density of snow}$$

Density of freshly deposited snow is about 0.05-0.10 (mm of water per mm of snow), and it increases as snowpack matures.

To characterize the snow accumulation within a river basin, SWE needs to be surveyed on a number of snow courses using snow tubes.

Why do we need snow courses?

Dunne and Leopold (1978, Fig. 13-1)
What does this diagram indicate?

Snowpack survey results, St. Denis NWA, Saskatchewan.

Why are they so different?
Wind-driven snow transport (blowing snow) is a major process moving snow on the prairie landscape, but it is extremely difficult to quantify the process.

Interception also has major effects on snow accumulation in wooded areas.

**Snowpack evolution**

The density of snow increases by gravitational settling, wind packing, melting, and recrystallization.

(1) Melting of snow near the ground surface causes percolation and refreezing.

(2) Air in the snow pack is saturated with water vapor. Temperature gradient within the snow causes vertical transport of water vapor within the snow pack. How?

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Daffern (1992. Avalanche safety for skiers & climbers. Rocky Mountain Press. Fig. 3-32)
Snowmelt energy balance

Latent heat of fusion \((L_f) = 334 \text{ kJ kg}^{-1}\)

The temperature of snowpack is nearly at 0 °C when snow is melting. Under the constant-temperature condition,

\[
Q_m = Q_s(1 - \alpha) + Q_{lw} + Q_h + Q_e + Q_p
\]

- \(Q_m\): energy available for snowmelt
- \(Q_s\): incoming shortwave radiation
- \(Q_{lw}\): longwave radiation into the snowpack.
- \(Q_h\): sensible heat transfer into the snowpack.
- \(Q_e\): latent heat transfer into the snowpack.
- \(Q_p\): energy input by rain on the snowpack.

The amount of snowmelt \((M)\) is given by

\[
M = \frac{Q_m}{\rho_w L_f}
\]

\(\rho_w\): density of water

The amount of incoming radiation greatly depends on slope aspects and canopy density.

Canopy density =

Dunne and Leopold (1978, Fig. 13-4)
Albedo of snow surface is high and it decreases with time.

Net longwave radiation in absence of forest cover:

$$Q_{lw} = (0.76 \sigma T_a^4 - \sigma T_s^4)(1 - aC)$$

$\sigma$: Stephan-Boltzman constant ($5.67 \times 10^{-8}$ W m$^{-2}$ K$^{-4}$)

$T_a$: air temperature in Kelvin

$T_s$: temperature of snow surface (assumed 273.15 K)

$C$: cloudiness (see DL, p.109)

$a$: empirical constant (see DL, p.109)

Net longwave radiation when forest cover is present:

$$Q_{lwf} = F(\sigma T_f^4 - \sigma T_s^4) + (1 - F)(0.76 \sigma T_a^4 - \sigma T_s^4)(1 - aC)$$

$F$: forest canopy density

$T_f$: temperature of the trees (commonly assumed equal to air temperature)
In this example, $Q_s$ and $\alpha$ are fixed. At a high air temperature (10 °C), longwave radiation emitted by the forest canopy has major effects on net radiation input ($Q_n$).

(a) Air temperature = 0°C
(b) Air temperature = 10°C

Dunne and Leopold (1978, Fig. 13-6)
Sensible and latent heat transfer are affected by temperature gradient \((T_a - T_s)\), vapor pressure deficit \((e_a - e_s)\), and wind speed \((u)\). Many formula to calculate sensible and latent heat have been published (see DL, p.476). The following have been developed by Canadian researchers.


\[
Q_h = -0.92 + 0.076 u_{10} + 0.19 T_{max}
\]

\(Q_h\): sensible heat \((\text{MJ m}^{-2} \text{ d}^{-1})\)
\(u_{10}\): wind speed at 10 m \((\text{m s}^{-1})\)
\(T_{max}\): daily maximum air temperature \((\degree \text{C})\)

\[
Q_e = 0.080(0.18 + 0.098 u_{10})(e_2 - e_s)
\]

\(Q_e\): latent heat \((\text{MJ m}^{-2} \text{ d}^{-1})\)
\(e_s\): mean daily vapor pressure at the snow surface
(assumed equal to 6.1 mb for melting snow)
\(e_2\): mean daily vapor pressure \((\text{mb})\) at 2 m.

Heat input by rain on the snowpack can be evaluated from

\[
Q_p = PT_a c_w r_w
\]

\(P\): depth of rainfall \((\text{m})\)
\(c_w\): specific heat of water \((= 4.19 \text{ kJ kg}^{-1} \degree \text{C}^{-1})\)

However, the amount of snowmelt resulting from \(Q_p\) is usually much smaller than the amount of rain.
In this example, $Q_n$, $Q_h$, and $Q_e$ are shown as equivalent amounts of water. How are they calculated?

The total melt is primarily controlled by radiation in the first half of the period when temperature was low. In the second half, sensible heat dominates the energy input.

[Graph showing melt as a function of days with categories for net radiation, sensible heat, latent heat, and total melt]

Dunne and Leopold (1978, Fig. 13-9)

Implication?
Temperature index method

Air temperature has a major influence on snowmelt. → use air temperature as an index

Seen this type of approach somewhere else?

\[ M = a + kT_a \]

\( k \): degree-day factor

Both \( a \) and \( k \) needs to be calibrated. Why?

\[ M = 1.0 + 0.06T_a \]

Dunne and Leopold (1978, Fig. 13-11)
Snowmelt runoff

A snowpack needs to become “ripe” before it starts generating snowmelt runoff.

- field capacity of snow?
- temperature profile?
- effects of rain on snow?

Snowmelt runoff can reach stream channels via various paths. In Case (b), the frozen soil becomes essentially impermeable (concrete frost) and severely limit the infiltration of snowmelt water.

Dunne and Leopold (1978, Fig. 13-14)
Infiltration capacity of frozen soil

After extensive study of snowmelt infiltration, Canadian researchers have developed an empirical equation to predict the amount of infiltration in frozen soil in the prairies.


The amount of infiltration is dependent on the premelt soil water content (= ice content). Ice crystals block the flow of water, and significantly reduces infiltration rate. Therefore, drier soil has higher infiltration capacity.

\[ INF = 5(1 - \theta_p) \text{SWE}^{0.584} \]

- INF = total infiltration (mm)
- \( \theta_p \) = (water content) / (porosity) of the 0-30 cm soil layer before snowmelt
- SWE = snow water equivalent (mm) on the ground

The equation predicts the infiltration capacity of cultivated soil reasonably well, but it does not work in soils with root holes or cracks (macro pores).