Snow mass balance, snow melt, and runoff

Air temperature and LWC of snow

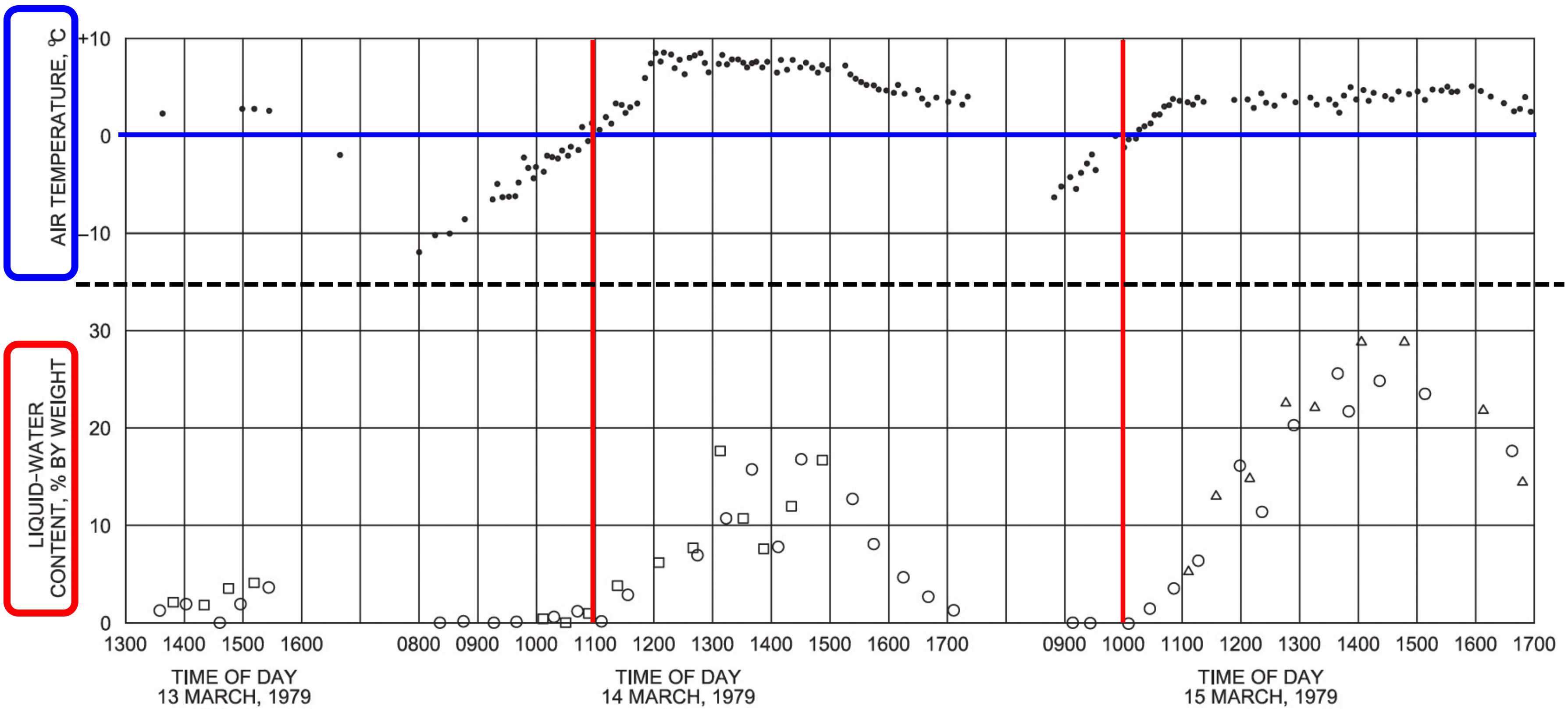
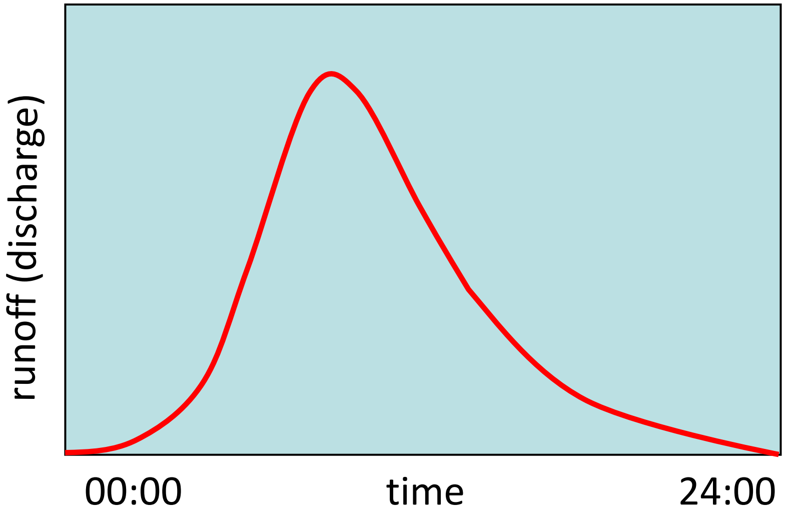


Figure 3.9 Daily variation of snowpack liquid-water content (mass of liquid water per unit mass of snow) determined by three different operators using the freezing calorimetry method at Fraser, Colorado (Jones *et al.*, 1983, copyright 1983 IWA Publishing with permission).

Compare slide #9 in "Snow properties" lecture.

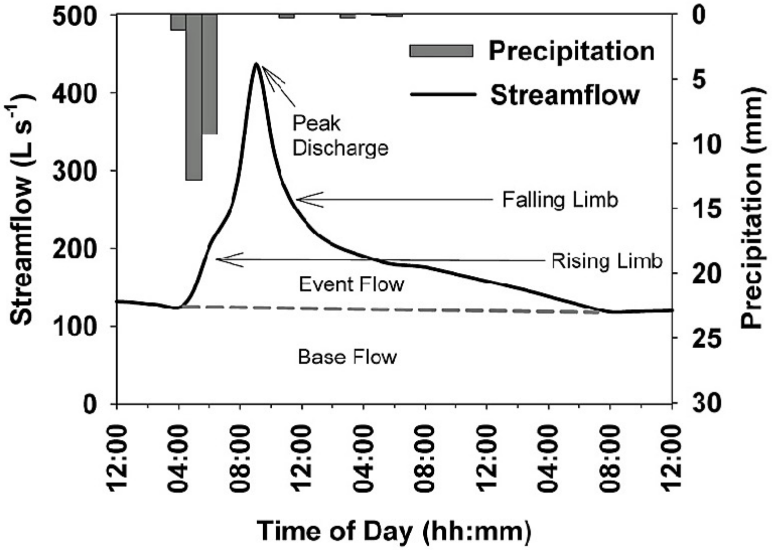
Snowmelt hydrograph

- Rate of water release from the base of the snowpack
- Resembles stream hydrograph
- Strong diurnal cycle
- Lags snowmelt at the surface by hours to days

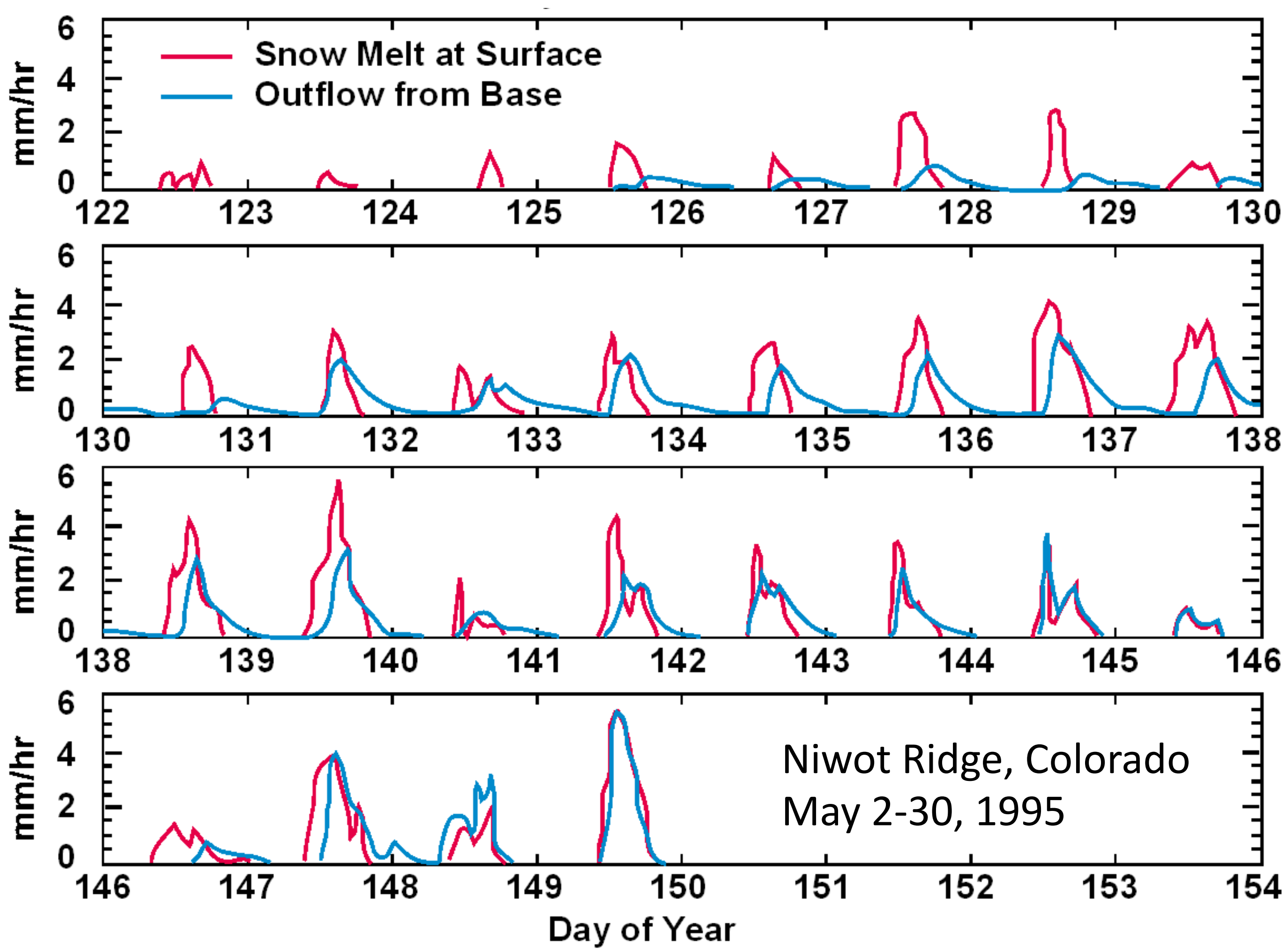


Melt water release from the snowpack during a day

Stream hydrograph



Liquid water transmission



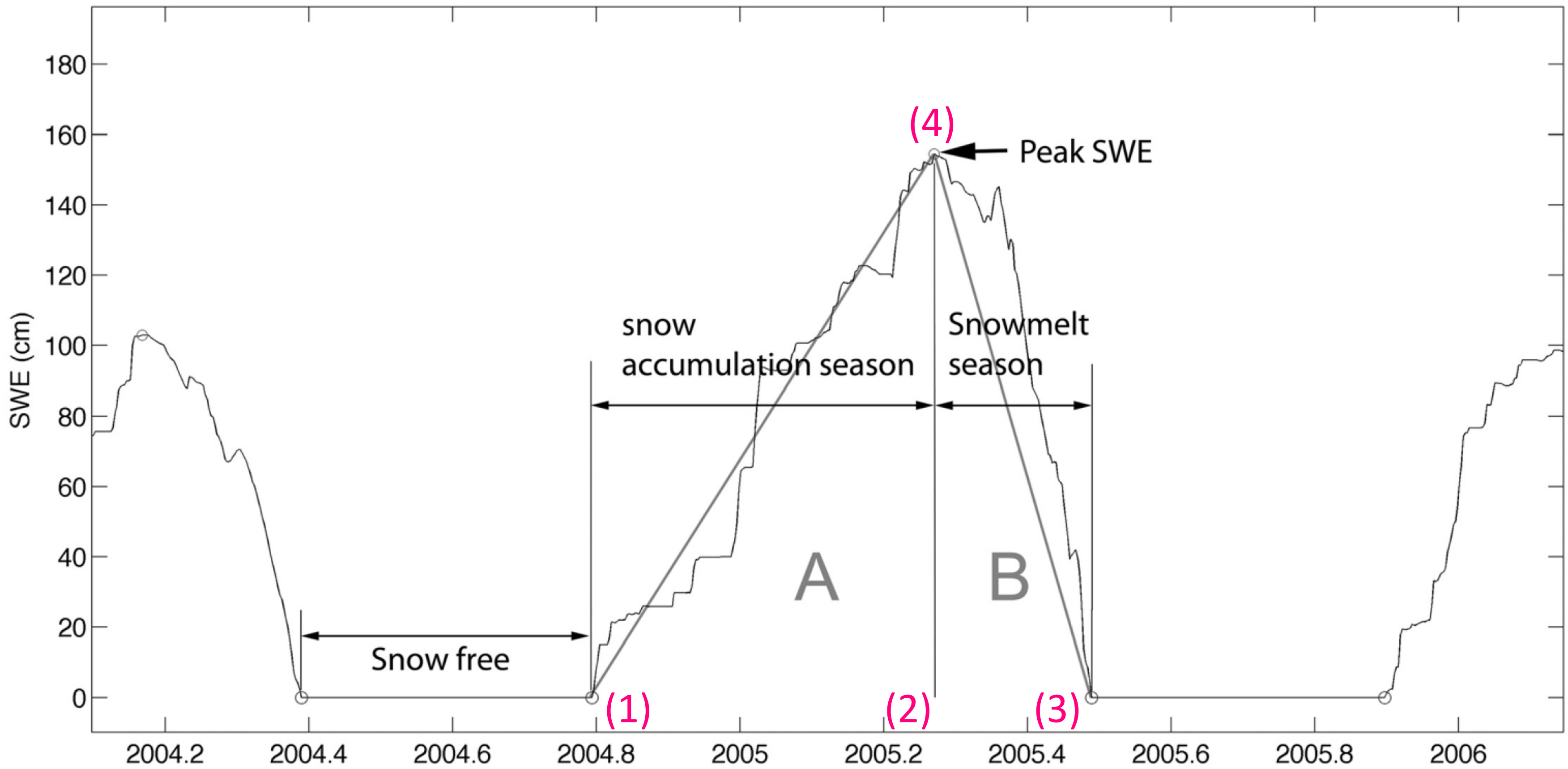
Melt- and rainwater experience a time lag and are attenuated as they move through the snow. Lag and attenuation are functions of depth, density, ice layers, grain size, and refreezing.

Content

Definitions

- Snowpack: accumulated snow on ground at a given time (precipitation and redistribution)
- Input (Accumulation): solid and liquid precipitation, condensation and deposition, and snow transport and redistribution
- Snowmelt: amount of liquid water produced by melting (phase change) during a given time period; water is either retained within the snowpack (LWC) or drained from the snowpack
- Ablation: total mass loss (snow + water) due to sublimation, evaporation, melting, and erosion (snow transport by wind)
- Output: total amount of liquid water (melt + percolating rain) leaving the snowpack at the base during a given time period
- → Ablation ≠ Output !!

Snow Water Equivalent curve



Daily SWE for an accumulation and melt season. (1), (2) and (3) mark the day of initial snow accumulation, peak SWE and snow disappearance, respectively. Metric (4) marks the peak SWE. A and B indicate the accumulation and melt seasons, respectively. [*Trujillo and Molotch*, 2013].

Snowpack mass (water) balance

$$\frac{\Delta(SWE)}{\Delta t} = P - E - O$$

Δ(SWE): $(SWE_{t2} - SWE_{t1})$; [kg m^{-2}] or [mm]

(change in snowpack water equivalent over a time interval)

P: net precipitation rate, including rainfall, snowfall, snow transport by wind (resulting in local accumulation or erosion)

E: net vapor exchange rate between snowpack and atmosphere (sublimation, evaporation, condensation, deposition)

O: outflow rate of liquid water from the snowpack

Energy – water balance link

- Melt (outflow), evaporation/sublimation, and precipitation appear in the energy balance and the water balance equations → Coupling of energy and mass
- Over short time periods, sublimation/evaporation may be negligible (E = 0)
- During dry periods (P = 0), almost all outflow is from melting (when E ≈ 0)
- → then: O = M, where M is the melt rate [kg m^{-2} s^{-1}] or [mm d^{-1}]

$$\frac{\Delta SWE}{\Delta t} = M = O$$

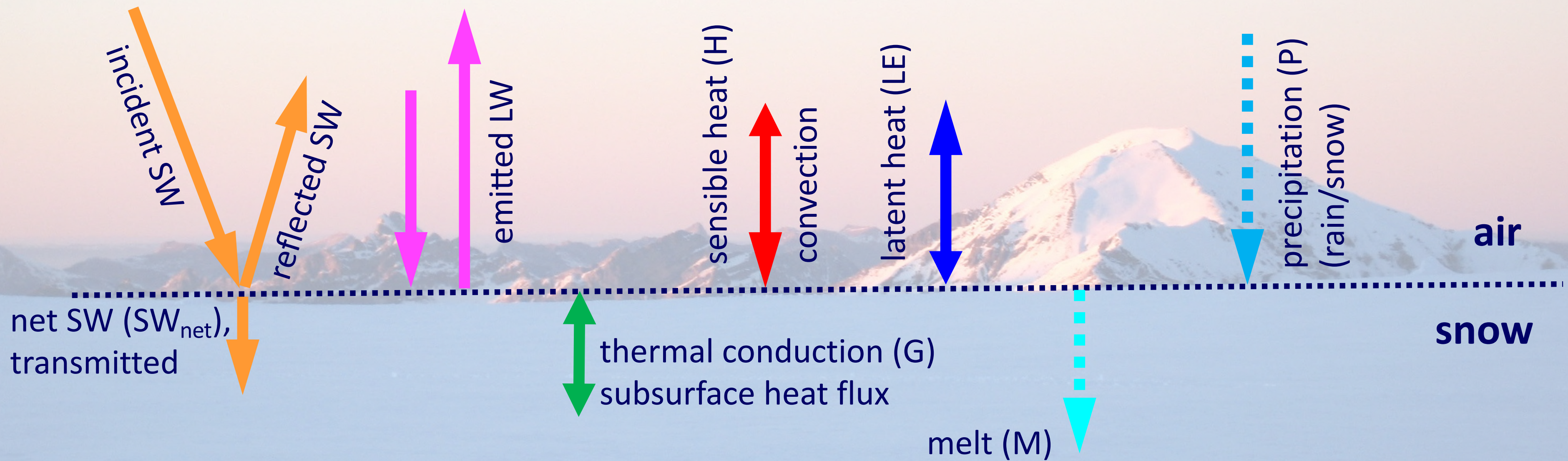
- → By measuring the rate of change in SWE, the melt rate M can be determined
- Melt rates of 20 mm d^{-1} are common, max. 70 mm d^{-1} possible (all expressed in [SWE])

Modeling snow melt – Energy balance (physics-based)

Snow **slab** or **layer:** $dE/dt = (R + G + H + \boxed{LE} + \boxed{M} + \boxed{P}) * A \quad [J\ s^{-1}]$

Net radiation $R = SW⇩ – SW⇧ + LW⇩ – LW⇧$

Transmitted SW : $(1-\alpha)\ SW⇩ = SW_{net}$



Note: individual terms can have a positive or negative sign.

A = surface area [m^2]; all fluxes [W m^{-2}]

$\boxed{\ldots}$: terms imply a mass change of the snowpack

Modeling snow melt – Degree day models (parameterized)

- Degree day models = temperature index models – common approach in hydrology
- Simple relations between snow/ice ablation and air temperature
- Generally, they are site specific

$M = c * f(T_a, R, u, \ldots)$; c needs to be calibrated
c = prop. const., T_a = air temperature,
R = (potential) solar radiation, u = wind speed

Basic model:

$$M = DDF \frac{1}{n} \sum_{i=1}^{n} T_i^+$$

$$M = \begin{cases} f_m (T_d - T_0), & T_d > T_0 \\ 0, & T_d \leq T_0 \end{cases}$$

$$M = f_m (T_d - T_0) + aR$$

DDF = degree day factor [mm d^{-1} $^{\circ}C^{-1}$],
T_i^+ = positive air temperatures [°C]
T_0 = freezing temperature (0 °C)
f_m = melt factor [mm d^{-1} $^{\circ}C^{-1}$] ≠ DDF,
a = empirical factor [m^3 J^{-1}]

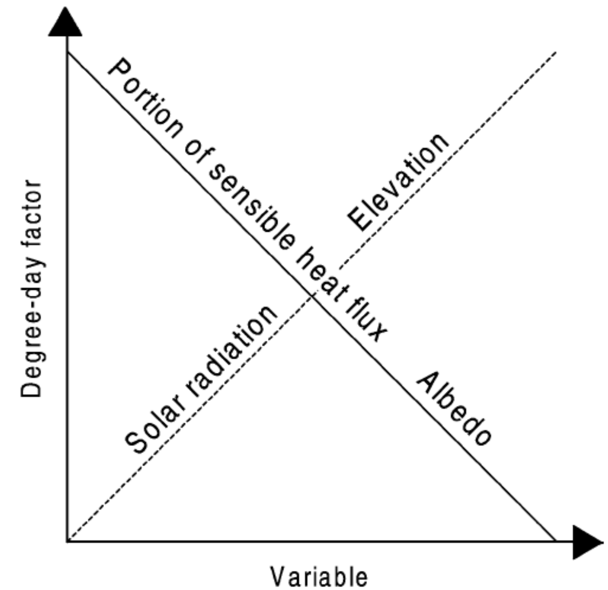


Fig. 1. Schematic plot illustrating qualitatively how degree-day factors are affected as variables increase. Degree-day factors increase as solar radiation and elevation increase and as portion of sensible heat flux and albedo decrease.

Hock 2003, J. Hydrol.

For phys. basis of DD models see Ohmura 2001, J.Appl.Met.

Phases of snowmelt

The process of snowmelt can be split in three phases:

- (1) Warming phase: Absorbed energy raises the average snowpack temperature to a point at which the snowpack is isothermal (no vertical temperature gradient) at 0˚C.

- (2) Ripening phase: Absorbed energy is used to melt snow, but the meltwater is retained in the snowpack in pore spaces by surface tension forces. At the end of this phase, the snowpack cannot retain any more liquid water and is said to be “ripe”.

- (3) Output phase: Further absorption of energy produces liquid water output, which then appears as runoff, infiltration or evaporation.

(1) Warming phase

Energy required to raise the entire snowpack temperature to 0˚C (Q_{cc}) [J m^{-2}]: cold content

$$Q_{cc} = \left|\rho_w \cdot SWE \cdot c_{p,i}(\bar{T}_{snow} - T_{melt})\right|$$

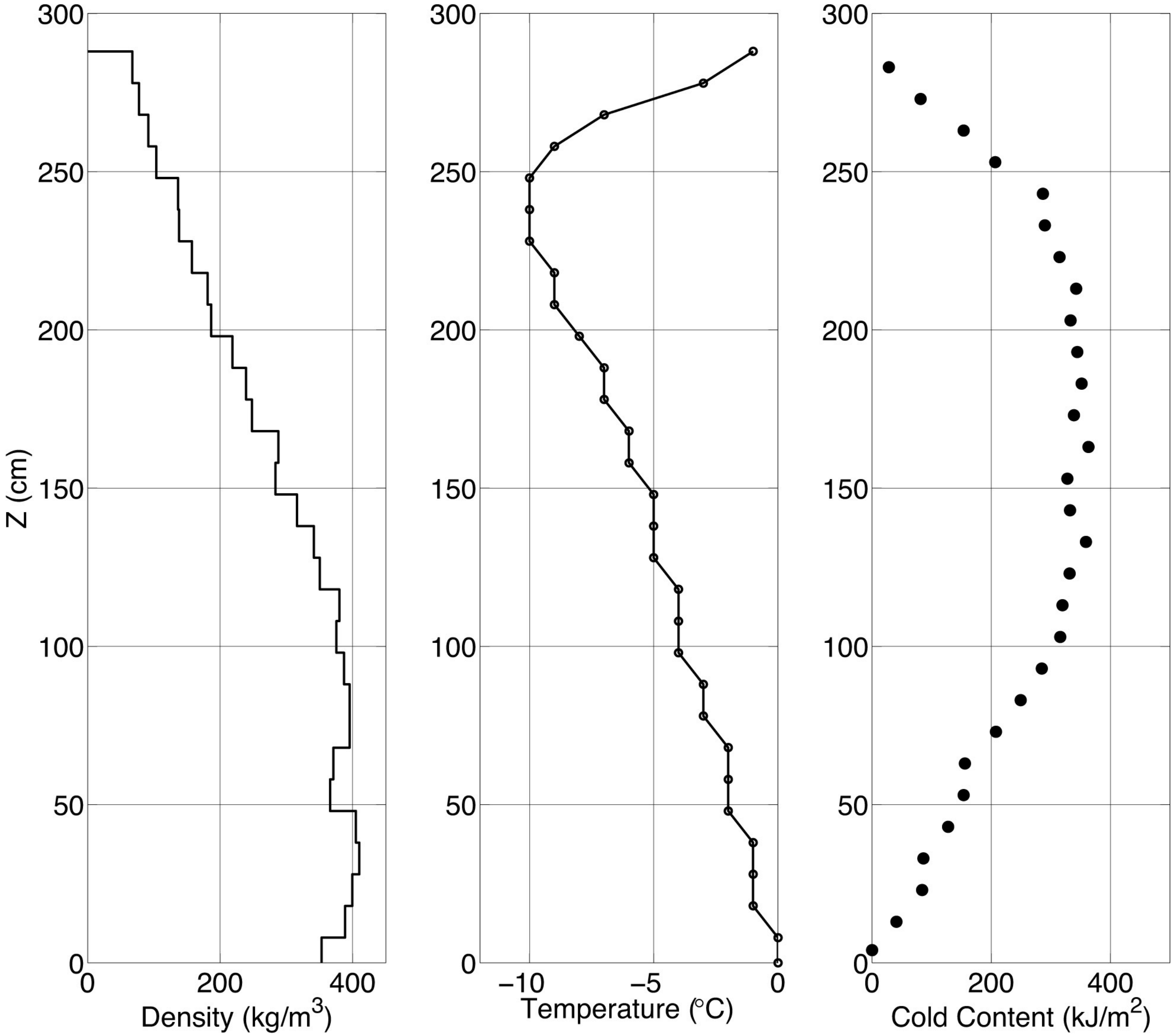
where

$c_{p,i}$ = heat capacity of ice (2100 J kg^{-1} K^{-1} at 0˚C),
T_{snow} = mean temperature of the snowpack (˚C),
T_{melt} = melting point of ice (0˚C),
ρ_w = density of water (approximately 1000 kg m^{-3}),
SWE = snow water equivalent [m] or [mm]

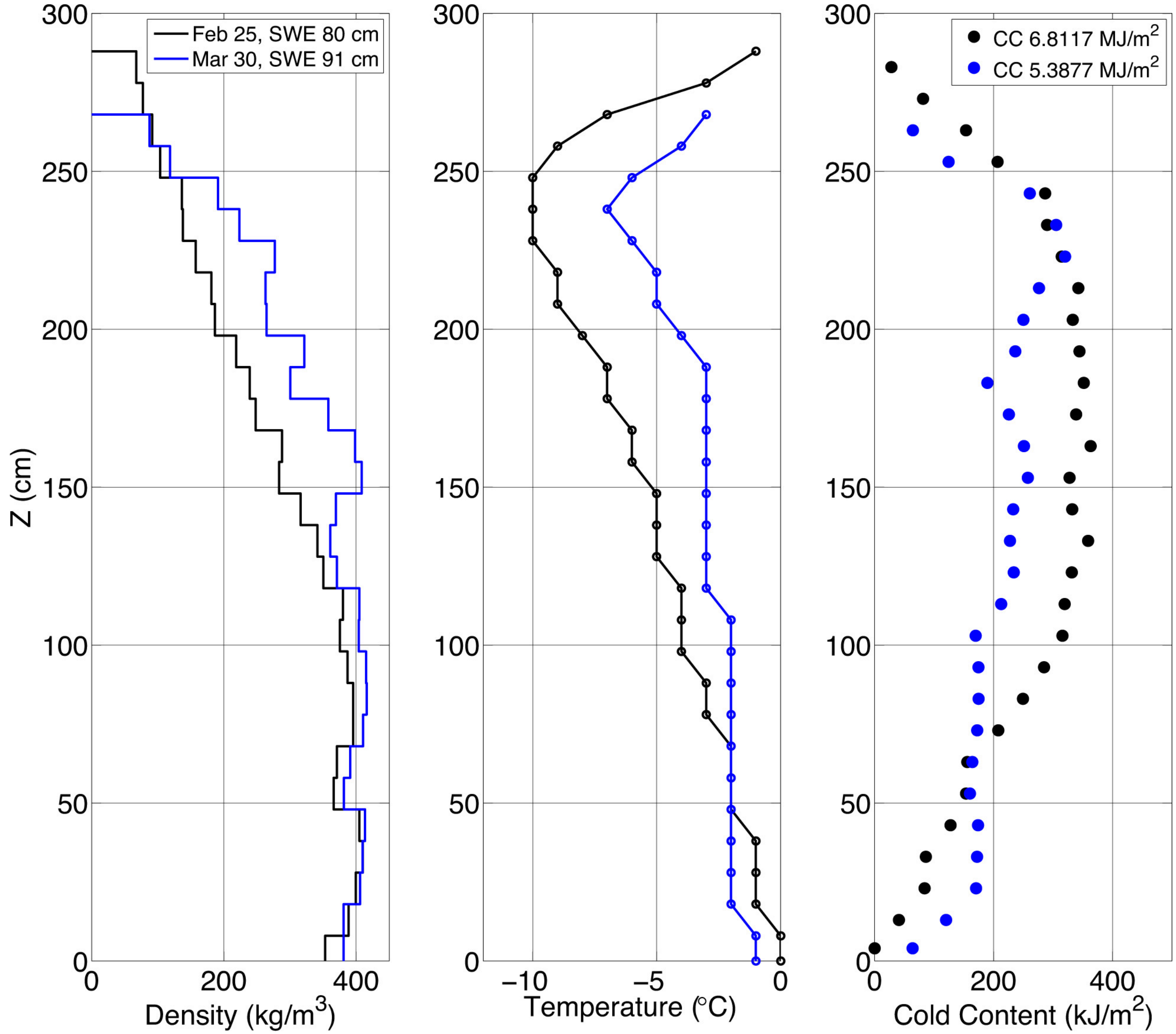
If snowpack water equivalent is unavailable but snow density and depth:
replace ρ_w with the snow density ρ_s, and SWE with snow depth (h_s).

ρ_w SWE = ρ_s h_s [kg m^{-2}]

Snow Pit Example – Feb 25, 2003. Buffalo Pass (CO)



Snow Pit Example – ~1 month difference

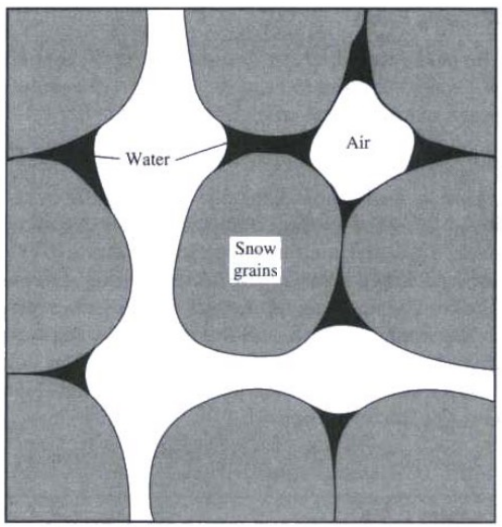


(2) Ripening phase

Available energy (absorbed SW, LW, SH, LH, heat conduction) is used to melt snow, but the meltwater is retained in the snowpack pore space by surface tension and capillary forces.

θ = liquid water content (LWC): ratio of liquid water volume to the total snowpack volume
θ_{ret} = maximum volumetric LWC a ripe snowpack can retain against gravity ("ret": retention)

A typical ripe snowpack has: ρ_s = 500 kg m^{-3} and θ_{ret} = 0.03



Dingman 2002, Figure 5-19

Recall, snowpack density can be written as:

$\rho_s = (1-\phi)\,\rho_i + \theta\,\rho_w$

ρ_i = density of ice: 917 kg m^{-3}

ϕ = porosity, V_a / V_s, i.e.,
the ratio of pore volume to the total snowpack volume

How much water is contained in a ripe snowpack? →

(2) Ripening phase

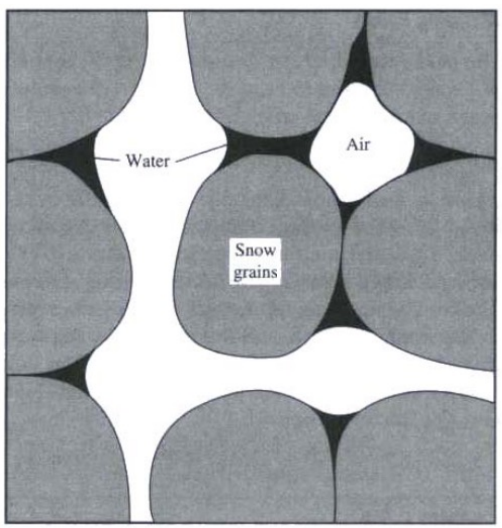
→ How much water is contained in a ripe snowpack? Recall: $\rho_s = (1-\Phi)\,\rho_i + \theta\,\rho_w$

Using $\rho_s = 500$ kg m^{-3} and $\theta_{ret} = 0.03$, → porosity $\Phi = 0.49$

Fraction of pore space filled with water at the end of the ripening phase:

$\theta_{ret} / \Phi = 0.061$ (6.1%) of the available pore volume.

Very little of the pore space in ripe snow is water (just 6% in this example!)



Dingman 2002, Figure 5-19

The net energy Q_{ripe} [J m^{-2}] required to complete the ripening phase is:

$$Q_{ripe} = \theta_{ret}\, h_s\, \rho_w\, L_f$$

L_f = latent heat of fusion (334 kJ kg^{-1})
h_s = snow depth [m]
θ_{ret} = max. vol. LWC of a ripe snowpack
ρ_w = density of water [kg m^{-3}]

(3) Output phase

- Following the ripening phase, further energy input results in more melting and downward melt water percolation immediately resulting in liquid water output.

- The energy Q_{out} [J m^{-2}] required to melt all snow remaining after the ripening phase is:

$$Q_{out} = (SWE - h_{w,ret})\, \rho_w\, L_f$$

where

$h_{w,ret} = \theta_{ret}\, h_s$ [m]

(the liquid-water retaining capacity of the snowpack).

Total energy for complete warming and melting process:

The snowmelt process requires absorption of energy Q of an amount equal to the sum of the energy absorption associated with the 3 individual phases necessary for runoff generation:

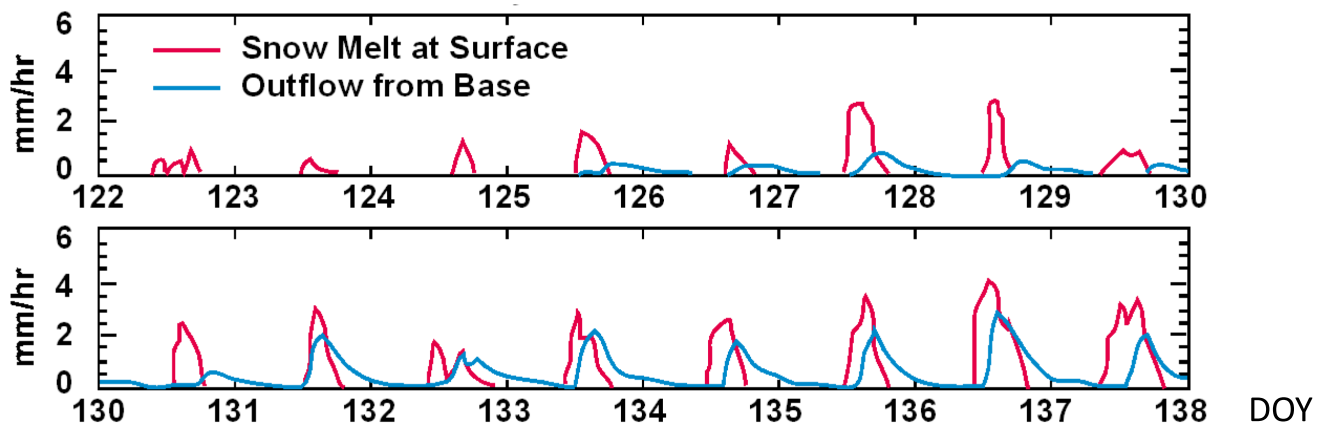
$$Q_{tot} = Q_{cc} + Q_{ripe} + Q_{out}$$

Snowpack storage and lag times

- Outflow results from melt, rainfall, and changes in liquid water holding capacity (e.g., metamorphism and changes in the snow micro-structure).
- Melt water produced near the snowpack surface can refreeze in colder layers below or gets temporarily stored in or on top of internal ice layers and due to capillary forces.
- Transmission time within the snowpack depends on the permeability (condition of pore matrix, connectivity of pores).
- Timing of outflow depends on storage, transmission, and liquid water content (holding capacity) within the snowpack.
- Rainfall on ripe snowpack can move quickly to the soil or produce surface runoff.
- Total delay for liquid water output = Σ of individual time lags.

Water movement through snow

Water movement through the snowpack influences the timing and the magnitude of the snowmelt and runoff peak.



Complex series of processes involved:

- energy balance
- snow metamorphism
- snowpack microstructure and stratigraphy
- water movement through saturated/unsaturated porous medium
- refreezing of melt water
- condition of the snowpack prior to introduction of water
- amount of water available at the snow surface

Poorly understood, few experimental studies, depends on the above listed factors.

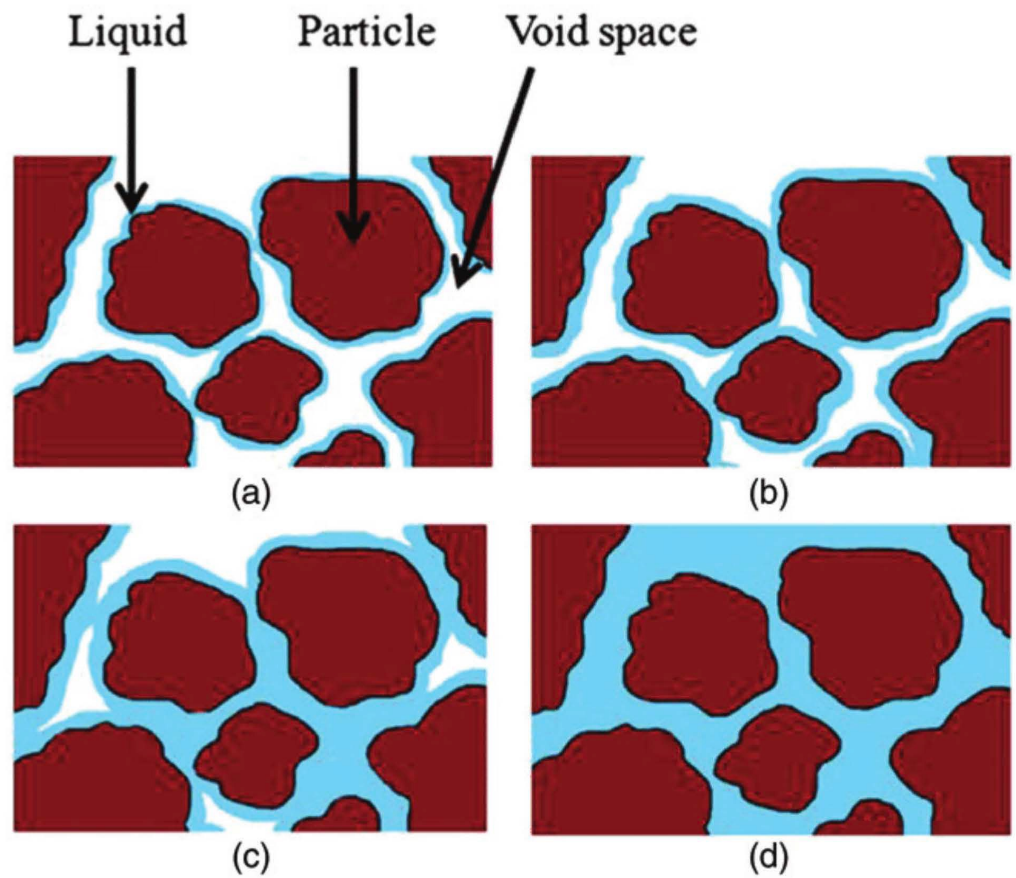
Saturation and water flow regimes in snow

Pendular: < 15%
of pore space is free water (S_w).
Water doesn't drain

Funicular: 15% to 85%
of pore space is free water (S_w).
Water drains by gravity, but
air spaces are continuous

Capillary: > 85%
of pore space is free water (S_w).
Water drains by gravity, and
air spaces are discontinuous

Slurry (saturated): ≈ 100%
of pore space is free water (S_w).
Water drains by gravity,
no more air space



Liquid bridging states
(a) pendular state, (b) funicular state,
(c) capillary state, and (d) slurry state (saturated).
(Kristensen and Schaefer, 1987).

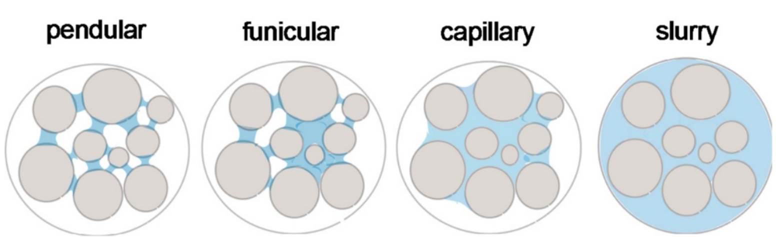
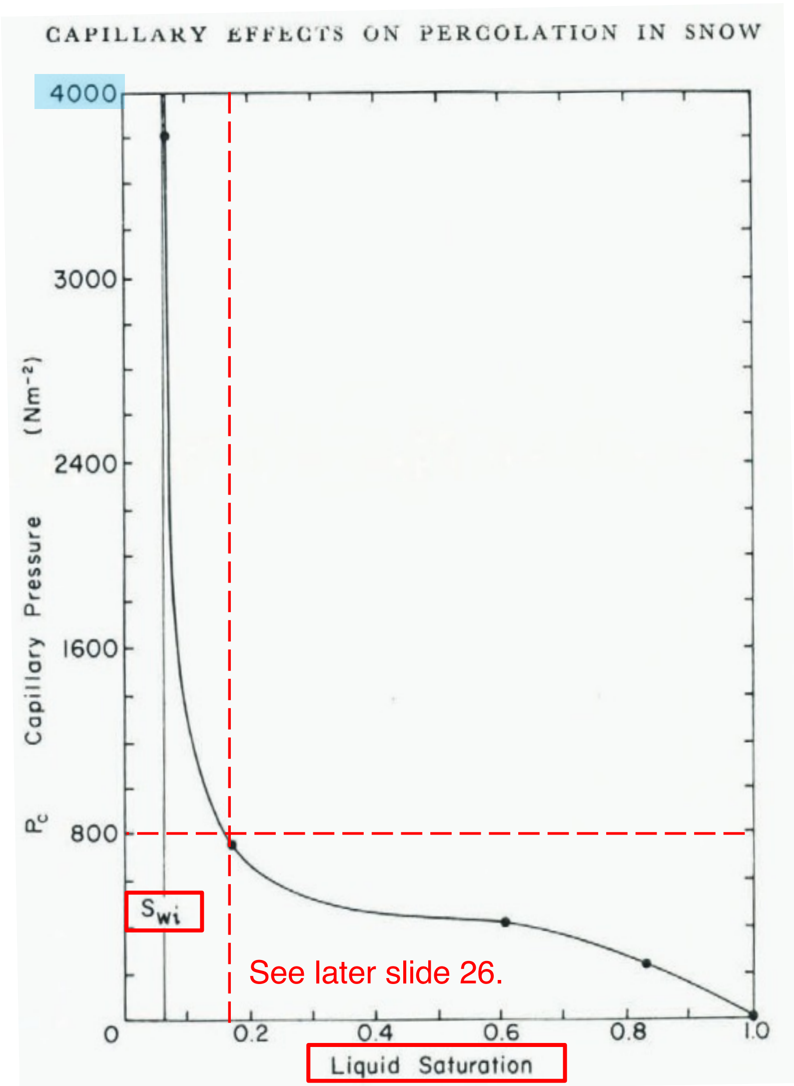


Fig. 8. Different states of saturation of liquid in the granular materials.

Dominant forces

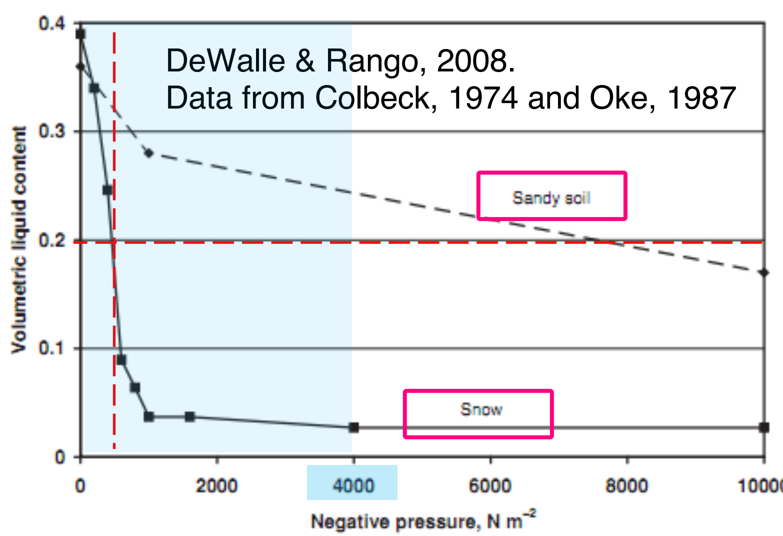
In snow: capillary effects are **small**, gravitational forces are **dominant**.



Capillary rise in snow is only a few cm.

80% of snowpack pore volume is drained with P = –800 N m^{-2}. (This suction force is equivalent to 8 cm of water column only!)

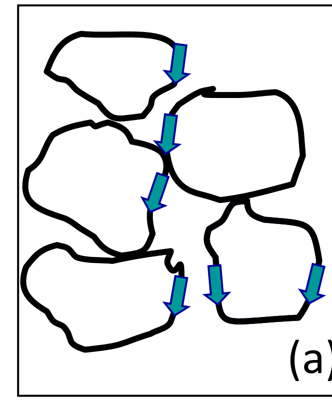
Irreducible liquid water saturation due to capillary retention ("undrainable" water) is S_{wi} = 0.05 – 0.07 [m^3 m^{-3}].
(vol. of water per vol. of pore space)



Vertical flow through homogeneous snow

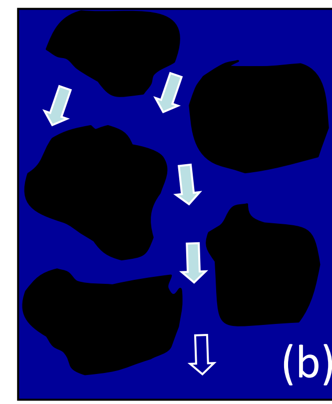
Situation (a):

- At melting temperature, a thin film of water surrounds each snow grain
- Much water can flow through this film (pendular and funicular regime).



Situation (b):

- Once pores are filled, laminar flow can occur
- Very efficient mechanism for draining the snowpack (capillary, slurry).



Flow in porous media: Darcy's law for unsaturated flow (→ Richards Eq.)

$$u_w = -\left(\frac{k_w}{\mu_w}\right)\left(\boxed{\frac{dP_c}{dz}} - \boxed{\rho_w g}\right)$$

pressure gradient term (relatively small)
gravity / acceleration term (relatively large)

u_w = volume flux of water [$m^3\ m^{-2}\ s^{-1}$] = [$m\ s^{-1}$]
k_w = permeability of snow for water [m^2]
μ_w = dynamic viscosity of water [Pa s] = [$kg\ m^{-1} s^{-1}$]
P_c = capillary pressure [Pa]
z = depth below snow surface [m]
ρ_w = density of water [$kg\ m^{-3}$]
g = gravity [$m\ s^{-2}$]

Simplification of Darcy's law

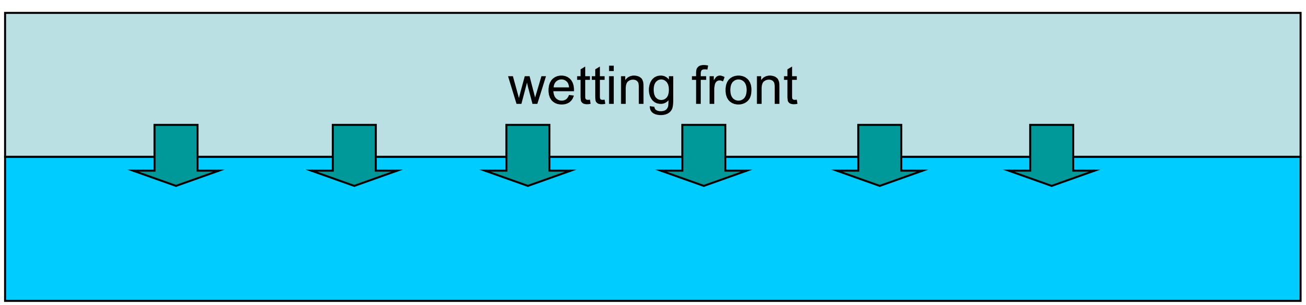
With capillary effects ignored, i.e., no pressure gradient:

$$u_w = -\left(\frac{k_w}{\mu_w}\right)\left(\cancel{\frac{dP_c}{dz}} - \rho_w g\right) \quad \rightarrow \quad u_w = \overbrace{\boxed{\frac{g\,\rho_w}{\mu_w}}}^{\text{known}} k_w = \alpha\, k_w$$

μ_w = 1.79*10^{-3} [Pa s] or [kg m^{-1} s^{-1}] for water at 0°C → α = 5.5 * 10^{6} m^{-1} s^{-1}

- Darcian flow assumes a wetting front.
- Does not account for preferential flow paths or for heat transport due to phase changes

→ Now we need to find k_w (permeability of snow for water [m^2]).



Realistic for snow?

Water permeability in snow

Goal: Find k_w, the permeability of snow for water, which can be expressed as:

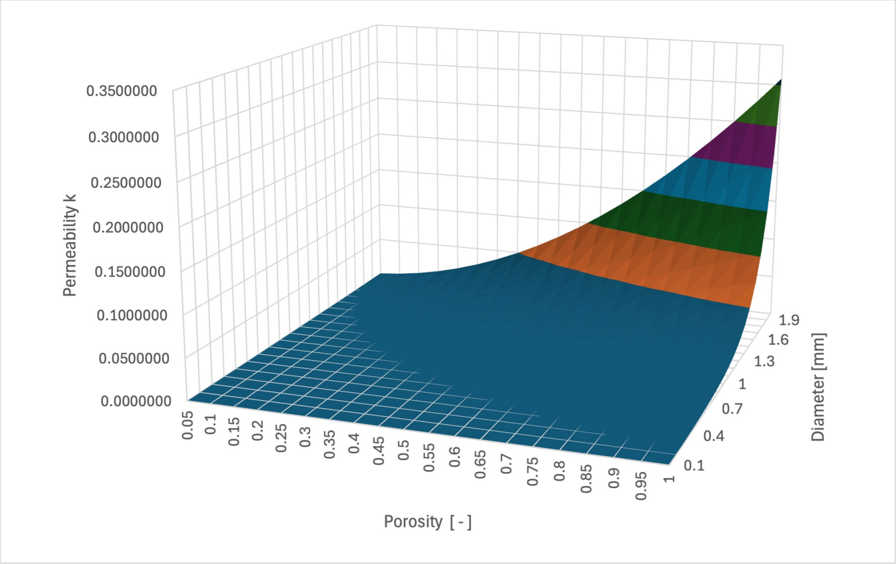
$$k_w = kS^{*\,n}$$

k = intrinsic permeability [m^2], (single phase flow at liquid saturation)
S^* = effective saturation, with $0 < S^* < 1$.
n = empirical const., $2 < n < 4$, with typical value being around 3.

The intrinsic permeability, k, which is related to the mean grain diameter D [m], and the porosity ϕ [m m^{-3}] of the snowpack, is given by:

$$k = 0.077\, D^2\, exp\left(-7.81(1-\phi)\frac{\rho_i}{\rho_w}\right)$$

Empirical, experimental equation.
k = f(D, ϕ)



Permeability and saturation

The permeability is a function of saturation. Recall: $k_w = kS^{*\,n}$

Higher liquid saturation
→ higher k_w
→ higher flow rate

For instance, doubling of S^* increases k_w by a factor of 8 for n=3 ! ($(2x)^3 = 2^3 {*} x^3 = 8 {*} x^3$)

In porous media, the permeability during two-phase (unsaturated) flow is related to the level of liquid saturation, S_w, described by the effective saturation, S^* [-]; S^* is given as:

$$S^* = \frac{(S_w - S_{wi})}{(1 - S_{wi})}$$

S_w = liquid saturation (volume of liquid water per volume of pore space) [$m^3\ m^{-3}$]
S_{wi} = irreducible liquid saturation (irred. vol. of liquid water per vol. of pore space) [$m^3\ m^{-3}$]

Typically, $S^* \approx 0.1 - 0.2$ during normal melting conditions, and $S_{wi} \approx 0.05$.

Low flow rate: ≈ 2 cm hr $^{-1}$, **High flow rate: ≈ 20 cm hr $^{-1}$**

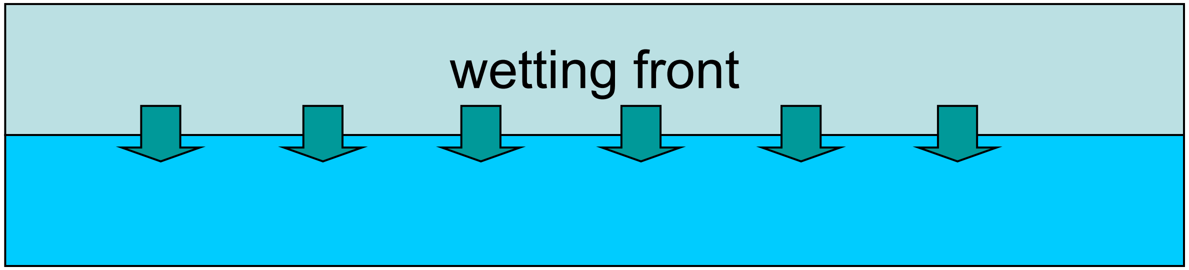
Calculate an example for a given degree of saturation !

Propagation of wetting front

Combining the above theory with the continuity equation allows for predicting of the rate of water movement in snow for a given value of water flux as a function of time at various depths, i.e., the propagation of the wetting front, in contrast to the flow velocity (Colbeck, 1972, 1977):

$$\left.\frac{dz}{dt}\right|_u = \frac{3\alpha^{1/3}k^{1/3}}{\phi(1-S_{wi})}u_w^{2/3}$$

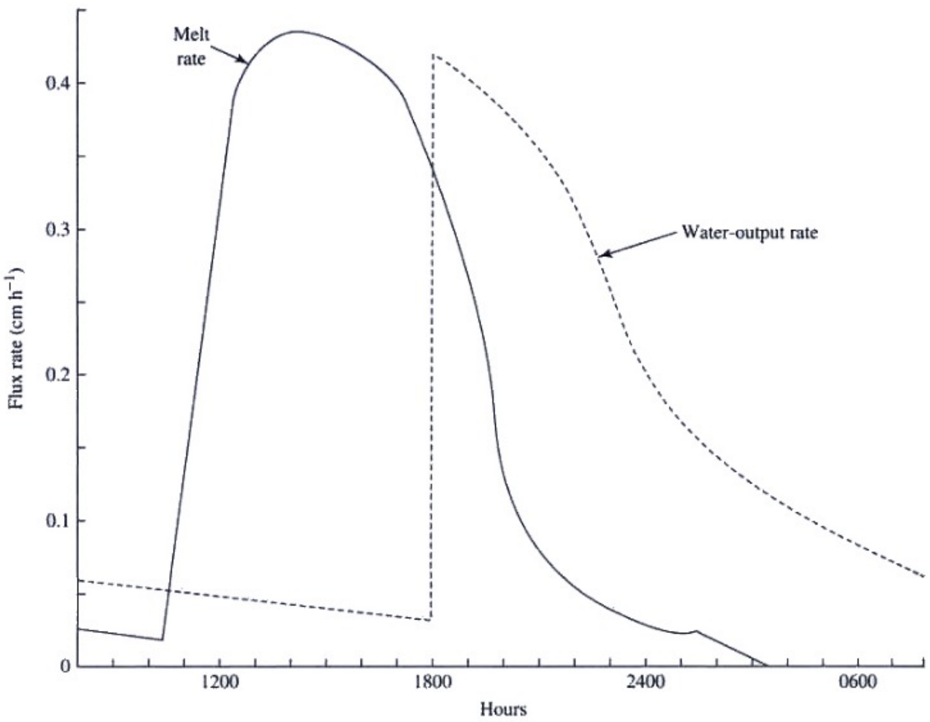
$dz/dt\,\vert_u$	=	speed of wetting front [m s^{-1}]
α	=	$\rho_w g / \mu_w$ (cf. Darcy's Law) [m^{-1} s^{-1}]
k	=	intrinsic permeability [m^2]
u_w	=	flux of water due to melting and/or rainfall
ϕ	=	porosity of snow [-]
S_{wi}	=	irreducible liquid saturation [m^3 m^{-3}]



Recall: this is for homogeneous snow and a uniform water flux.

Melt rate and output rate

- Melt rate peaks in early afternoon.
- Melt starts at the surface, and the melt water percolates through the snowpack.
- Percolation speed increases as melt rate increases.
  → Water from peak melt overtakes melt from earlier.
  → Sharp wave front can develop.
- Lag is depth dependent.



Dingman 2002, Figure 5-26

Liquid water in snow

Application of above theory matches reasonably well with experimental data.

Rapid gravity drainage of liquid water from snow.

85% of accumulated volume was collected within the first 2 hrs of drainage, i.e., preferential flow rather than slow matrix flow.

Experimental evidence for
$2 < n < 4$
in equation for permeability, recall

$$k_w = kS^{*\,n}$$

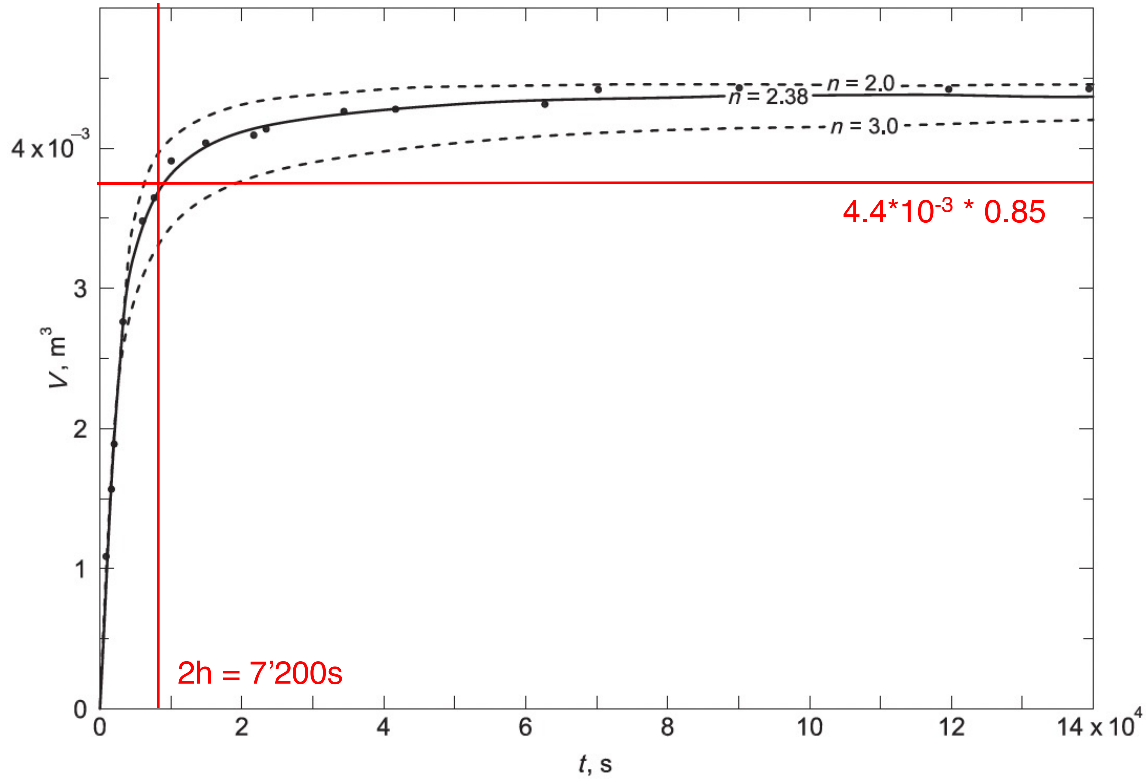
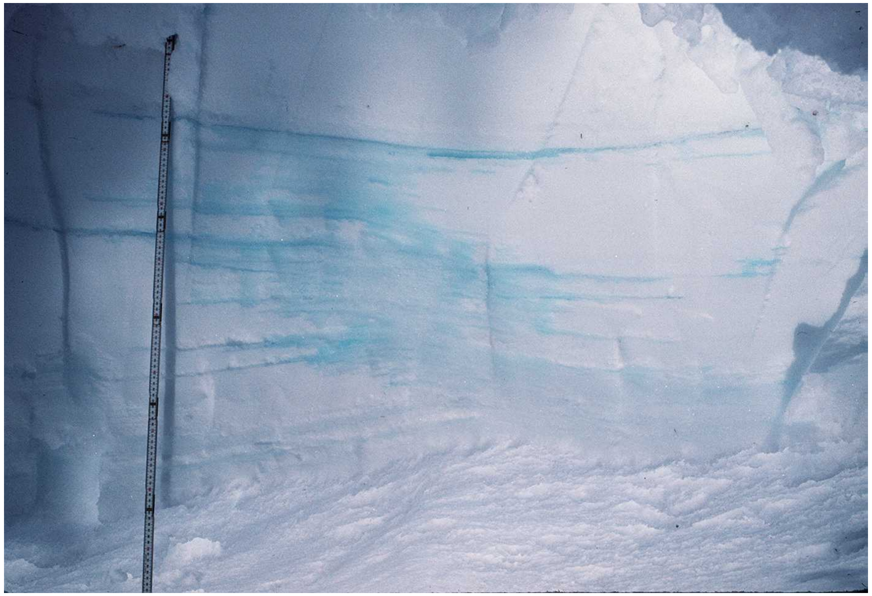


Figure 3.11 Accumulated drainage from melting snow core over time modelled with Darcy's law with the assumption of negligible capillary effects (Denoth *et al.*, 1979, courtesy US Army, Cold Regions Res. Engin. Lab.) showing rapid initial and very slow later drainage of meltwater. Lines show fitted model with varying exponents "*n*" for the snowpack permeability function in Equation (3.11).

But in reality…

- Melt water movement in snow is very complex (often disagrees with theory).
- Snow is spatially heterogeneous (vertically and horizontally).
- Melt processes cause the snow grain size distribution (and other snow properties and conditions) to change rapidly. →→
- Erratic liquid water percolation patterns and behavior !!!



snow pit with dye

snow pit with backlighting

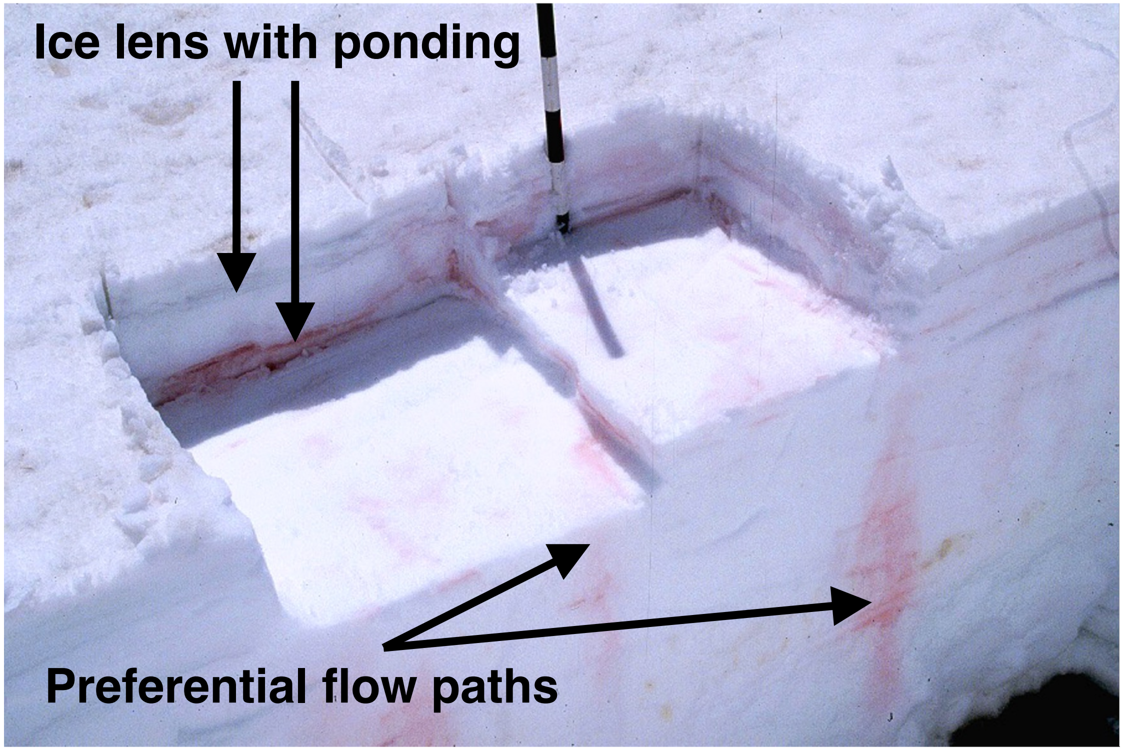
Shimuzu (1970) showed the intrinsic permeability of snow, k_w, varying by a factor of 100

Water movement (percolation) in snow

Preferential flow paths and horizontal shunting

Flow through heterogeneous snow

- Non-linear relationship between liquid water content and water flux in a preferential flow path.
- Flow is much faster than amount of liquid water content would suggest.
- Flow velocities range from 2-60 cm/hr.



Bucket Scheme – A simple model

In its standard version, the SNOWPACK model uses a bucket scheme. It retains water in a layer until the holding capacity (θ_h) has been reached, then water is passed on to the next layer.

$$\theta_h = \begin{cases} 0.0264 + 0.0099\,(1-\theta_i)/\theta_i, & \theta_i \le 0.23 \\ 0.08 - 0.1023(\theta_i - 0.03), & 0.23 < \theta_i \le 0.812 \\ 0, & \theta_i > 0.812 \end{cases}$$

The parameterization of the holding capacity above only considers density, but grain size is often considered as well.

Recent approaches …

… to measure and model liquid water flow in snow:

- Novel measurements of flow velocities in snow
- Richards Equation, Dual Domain Approach, and Ice Lenses

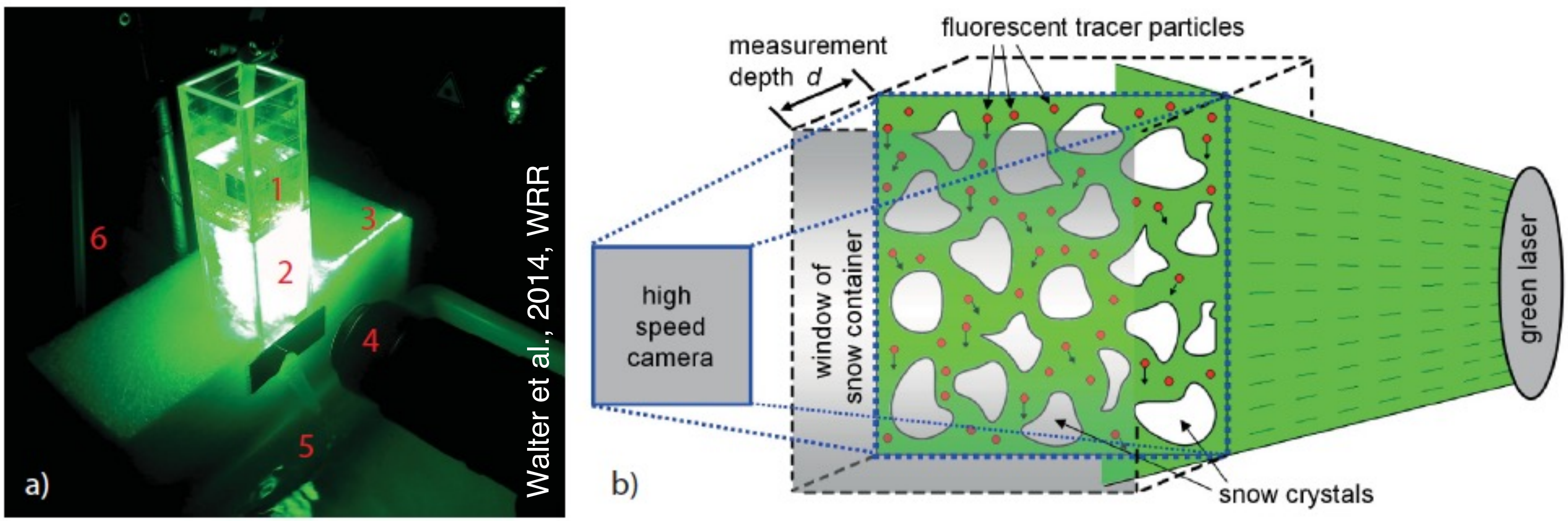


Figure 1. Experimental setup: (a) 1: upper reservoir with seeded water; 2: lower reservoir with wet-snow sample; 3: laser light sheet; 4: lens of high-speed camera with cut-on filter; 5: water basin; and 6: tube for pumping water into upper reservoir. (b) Schematic drawing of the measurement principle which is rotated by 90° clockwise in relation to Figure 1(a).

Reflections and refractions can disturb the images and velocity estimates.

Measurement of water flow in snow

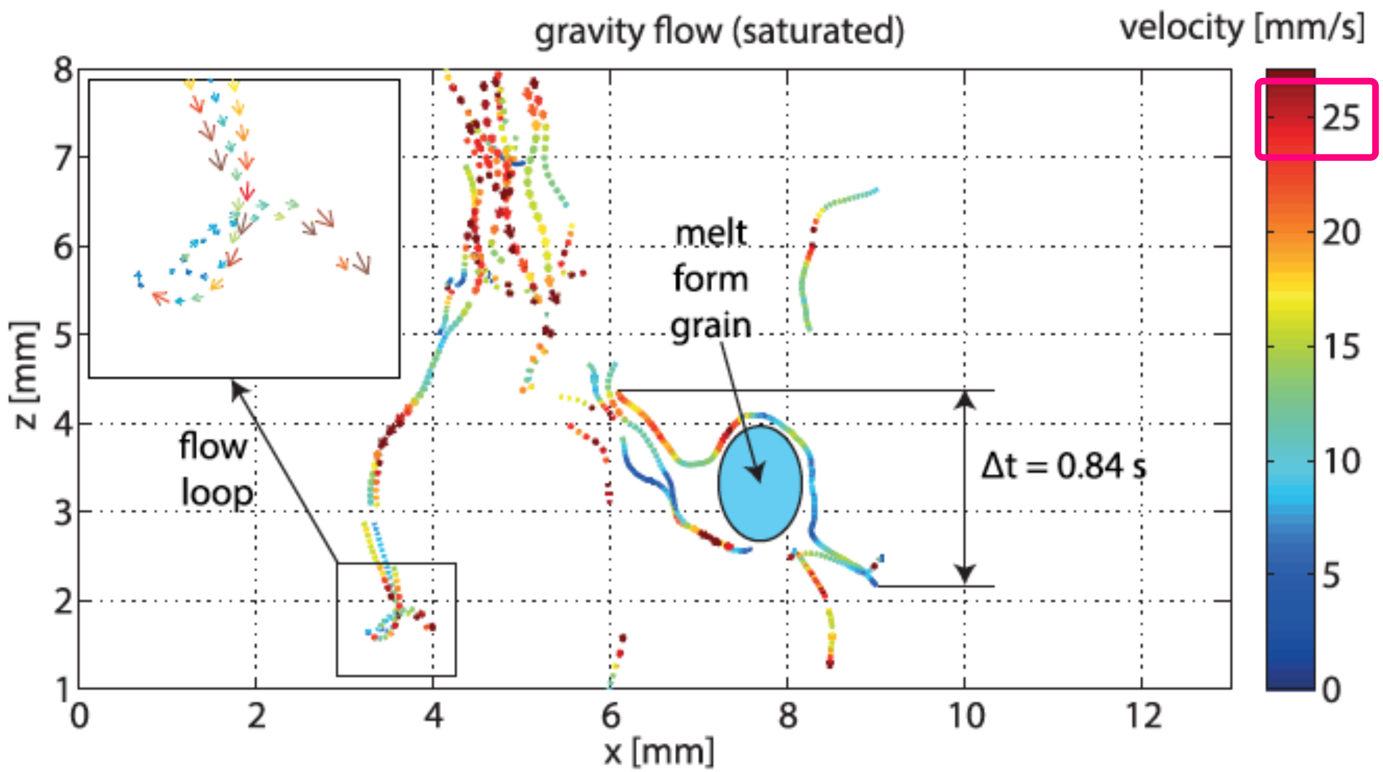


Figure 3. Particle trajectories for the downward saturated gravity-driven flow. Colors of arrows indicate particle velocities. The image shows about 30 individual particle tracks.

Walter et al., 2014, WRR

Measurement of water flow in snow

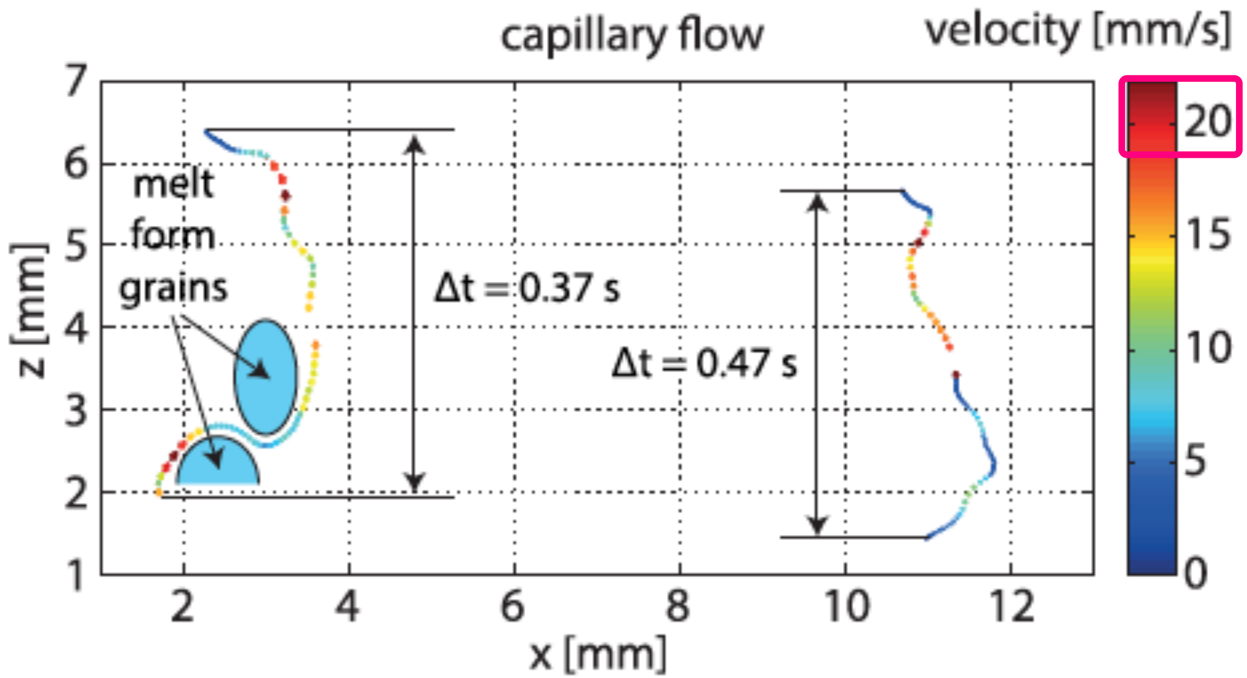


Figure 4. Particle trajectories for the upward flow driven by capillary forces. Colors of arrows indicate particle velocities.

Velocities are similar to saturated downward flow (cf. Fig.3, previous slide, i.e., quite fast).

Walter et al., 2014, WRR

Measurement of water flow in snow

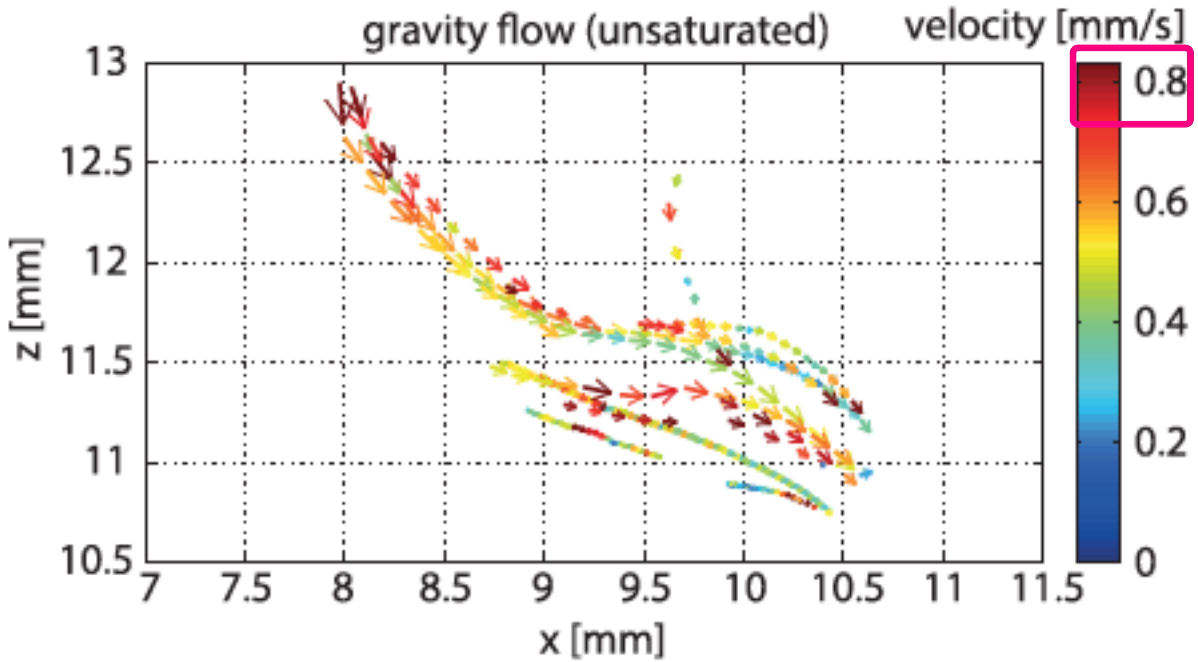
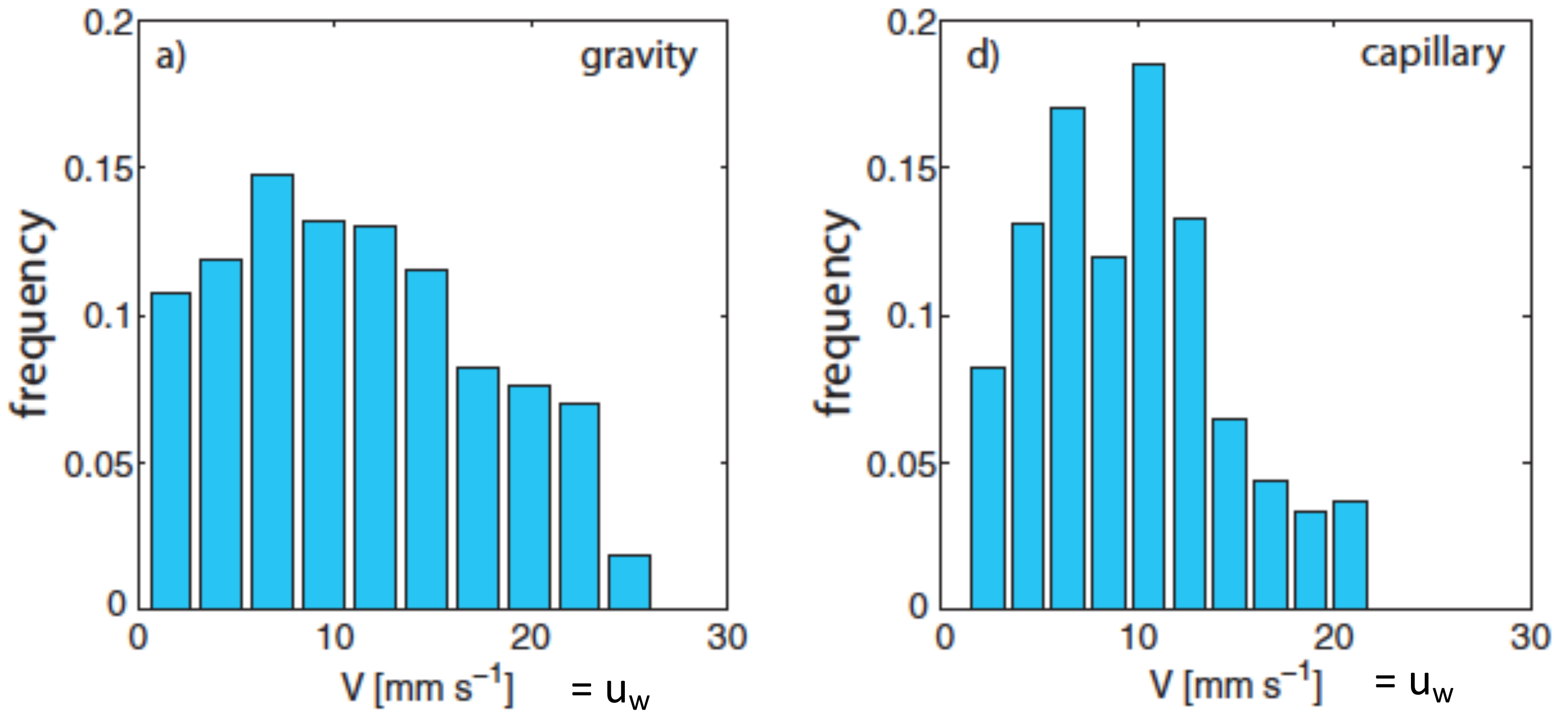


Figure 5. Particle trajectories for the downward unsaturated gravity-driven flow. Colors of arrows indicate particle velocities. The figure represents a zoom into a region of the recorded images where particles were successfully tracked.

Velocities are much smaller in downward UNSATURATED (probably gravity) flow than in both downward saturated and upward unsaturated flow (cf. previous slides).

Walter et al., 2014, WRR

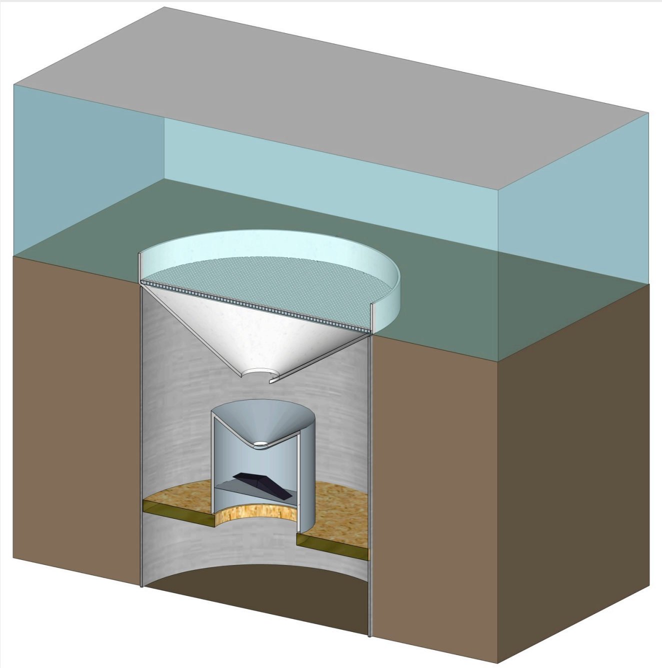
Distribution of flow velocities



Walter et al., 2014, WRR

Measuring liquid water output

Snow lysimeter: → Snowpack liquid water output
Soil moisture probes: → Soil moisture and temperature
Stream runoff: → Discharge gauging stations



Formation of ice lenses

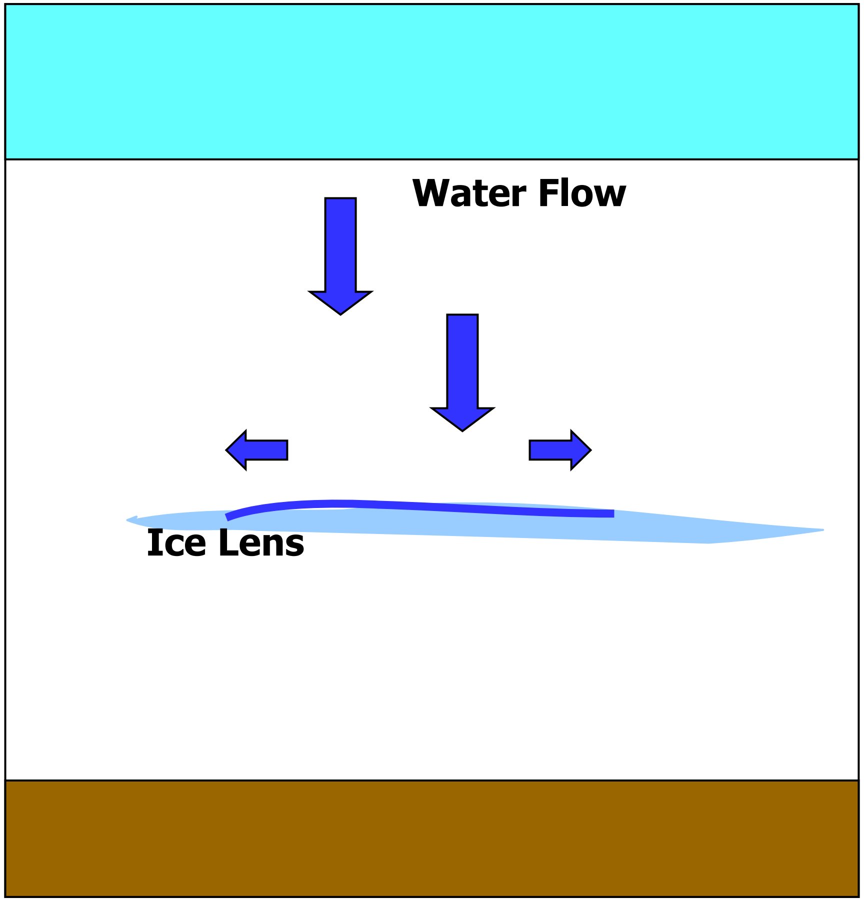
Ice Layers:

Develop from refreezing of internal melt

Relatively impermeable to subsequent flow and water vapor

Forces ponding of water and lateral flow

Flow fingers = preferential flow paths and ice lenses can redirect flow over several m^2 [*Kattleman*, 1989]



Dual Domain Richards Approach

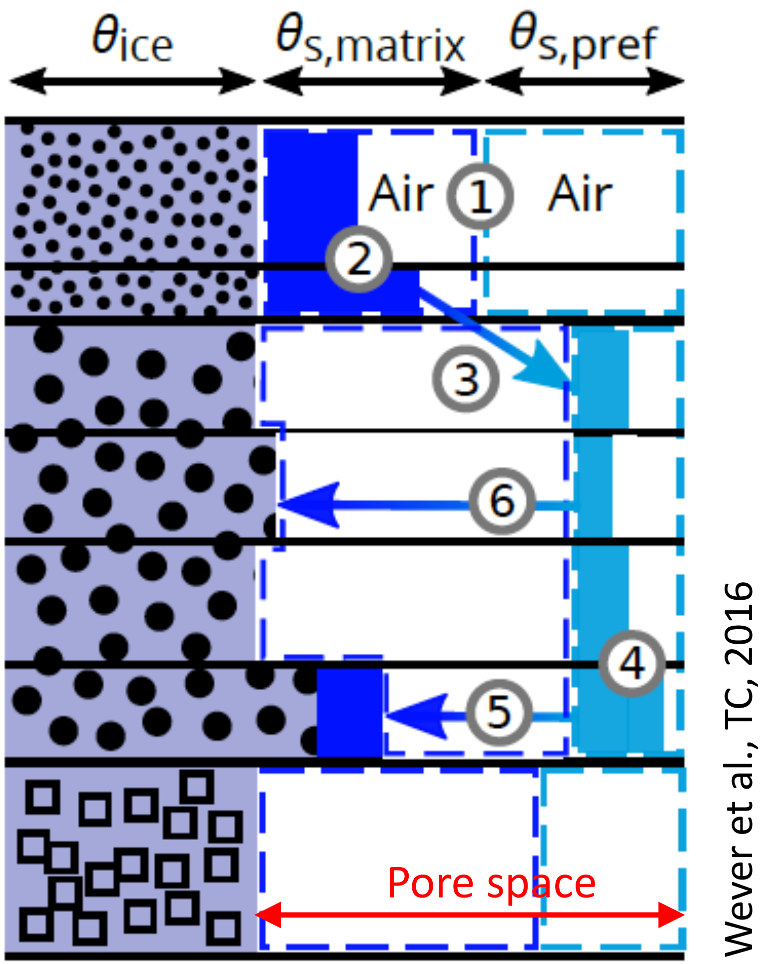


Figure 1. Schematic overview of the dual domain implementation for the SNOWPACK model, in which the pore space that can be occupied by liquid water is separated into a part for matrix flow ($\theta_{s,matrix}$) and a part representing preferential flow ($\theta_{s,pref}$). The numbers refer to processes described in the text.

Accounting for Preferential Flow!

If pressure head exceeds water entry pressure head of layer below water moves from matrix to preferential flow (2).

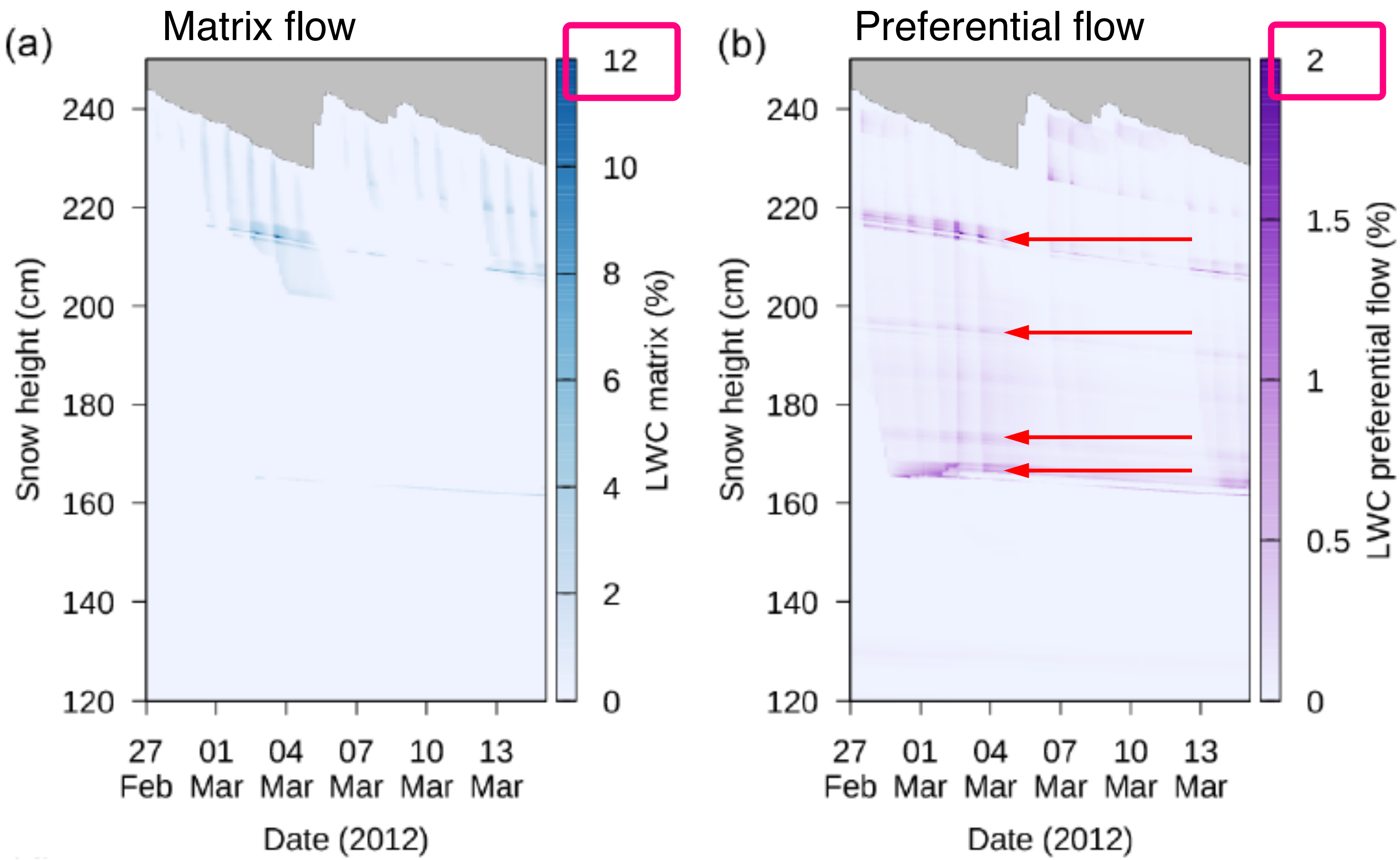
If the saturation in the preferential flow path exceeds a threshold, water moves back to the matrix domain (5).

Only the matrix part is allowed to undergo phase changes, and ice layers form when water moves back from preferential flow to matrix flow and refreezes.

Preferential flow always remains in the liquid phase. Refreezing of preferential flow water is mimicked by moving water from preferential flow to the matrix flow domain (6).

Richards Equation:
Describes movement of water in unsaturated porous media.

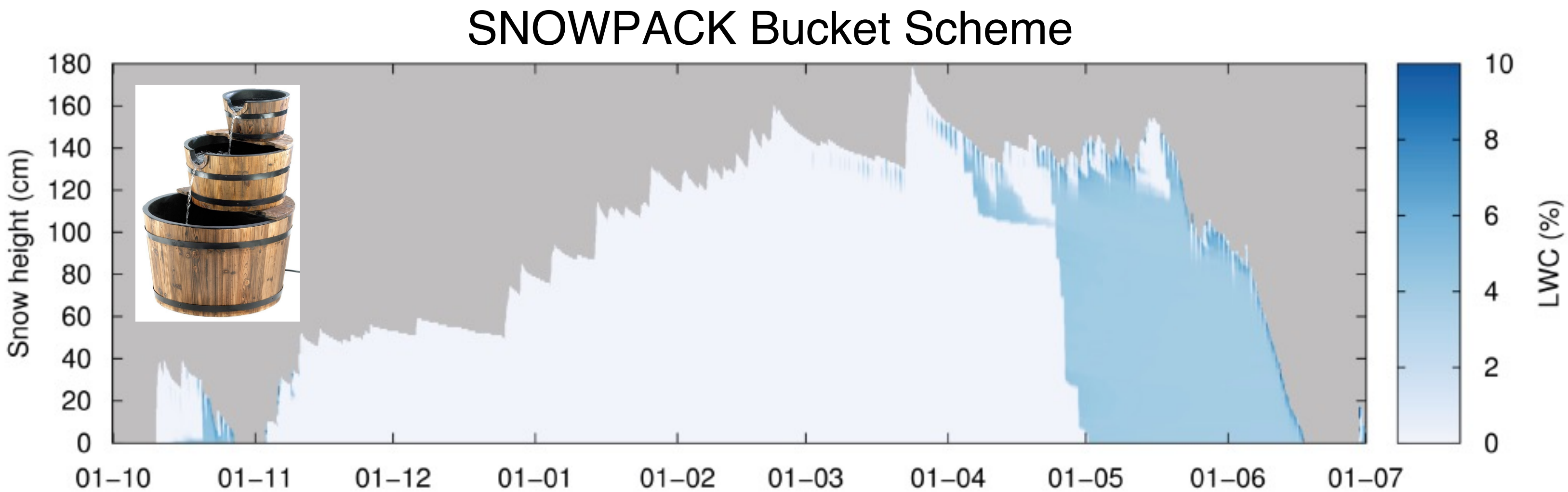
Matrix flow vs. preferential flow

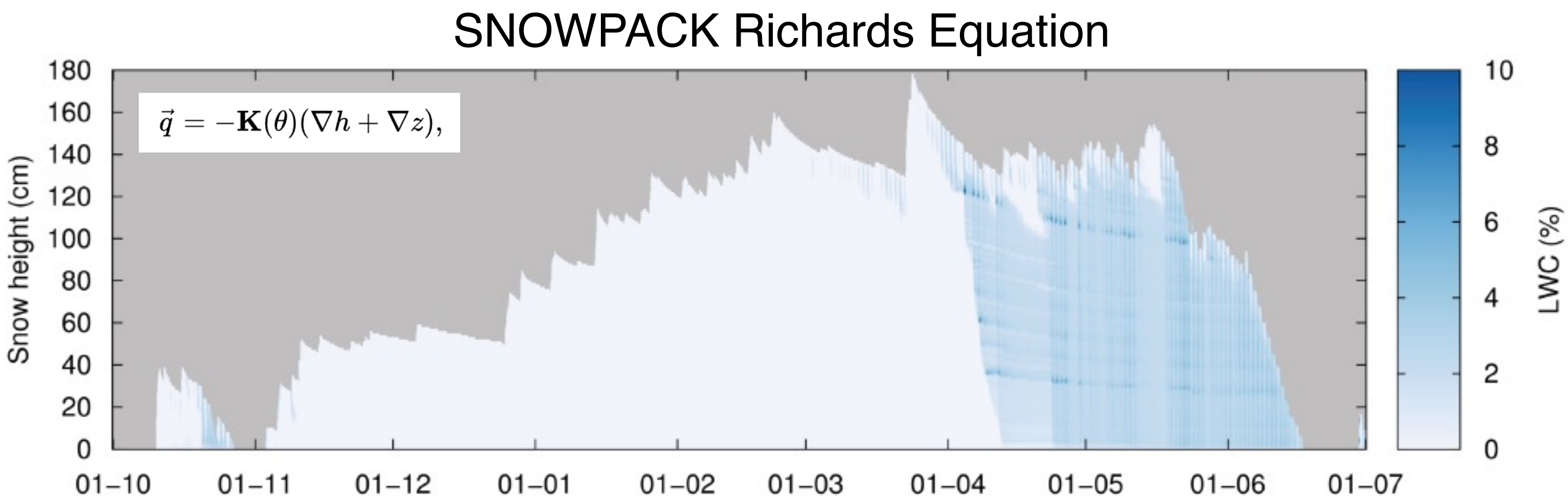


Liquid water as preferential flow is faster than in matrix flow.
Preferential snow drains the snowpack faster; matrix flow retains more water.
Preferential flow reaches deeper sub-freezing layers. → Ponding and ice layers.

Wever et al., 2016

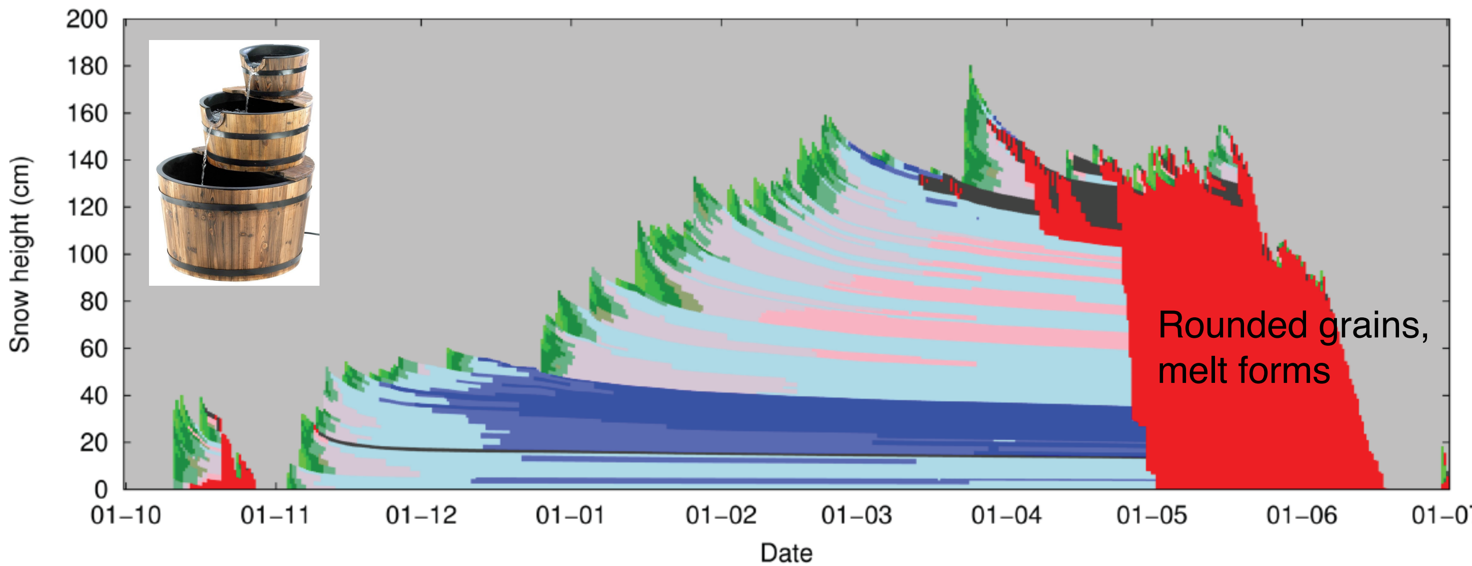
Internal differences in LWC



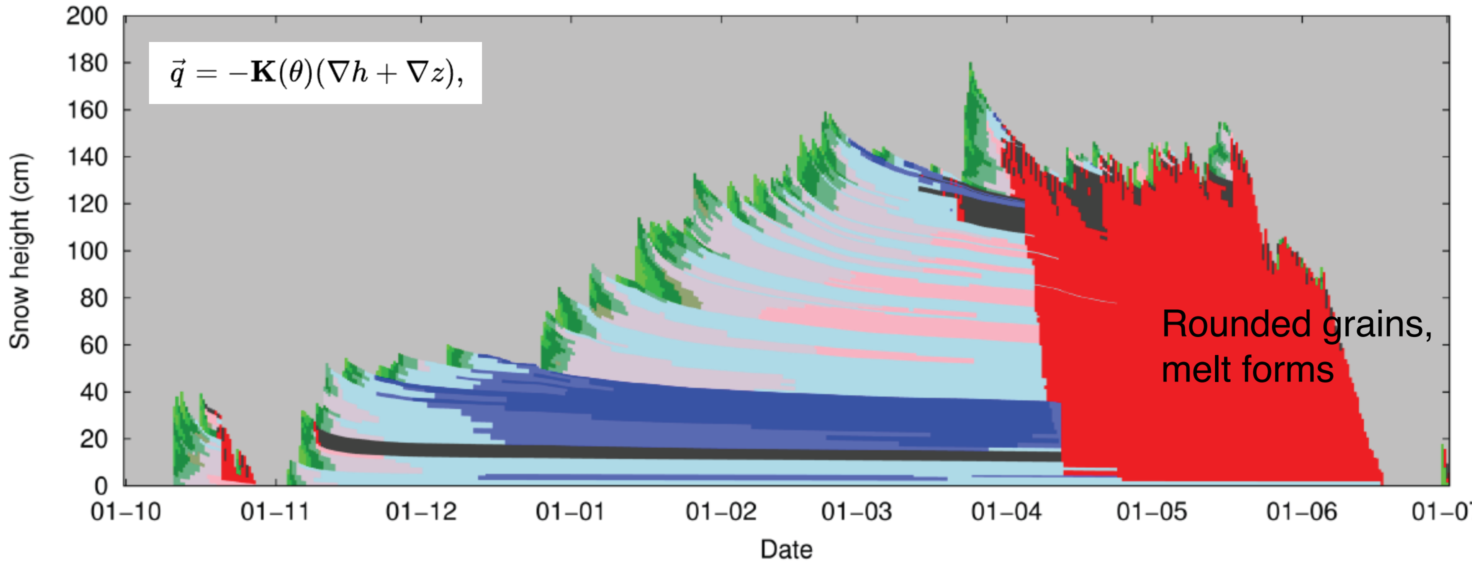


- Faster wetting of entire snowpack and thus earlier runoff formation (closer to observations).
- Ponding of water at grain size (type) transitions, i.e., layer interfaces.

Internal differences in grain type

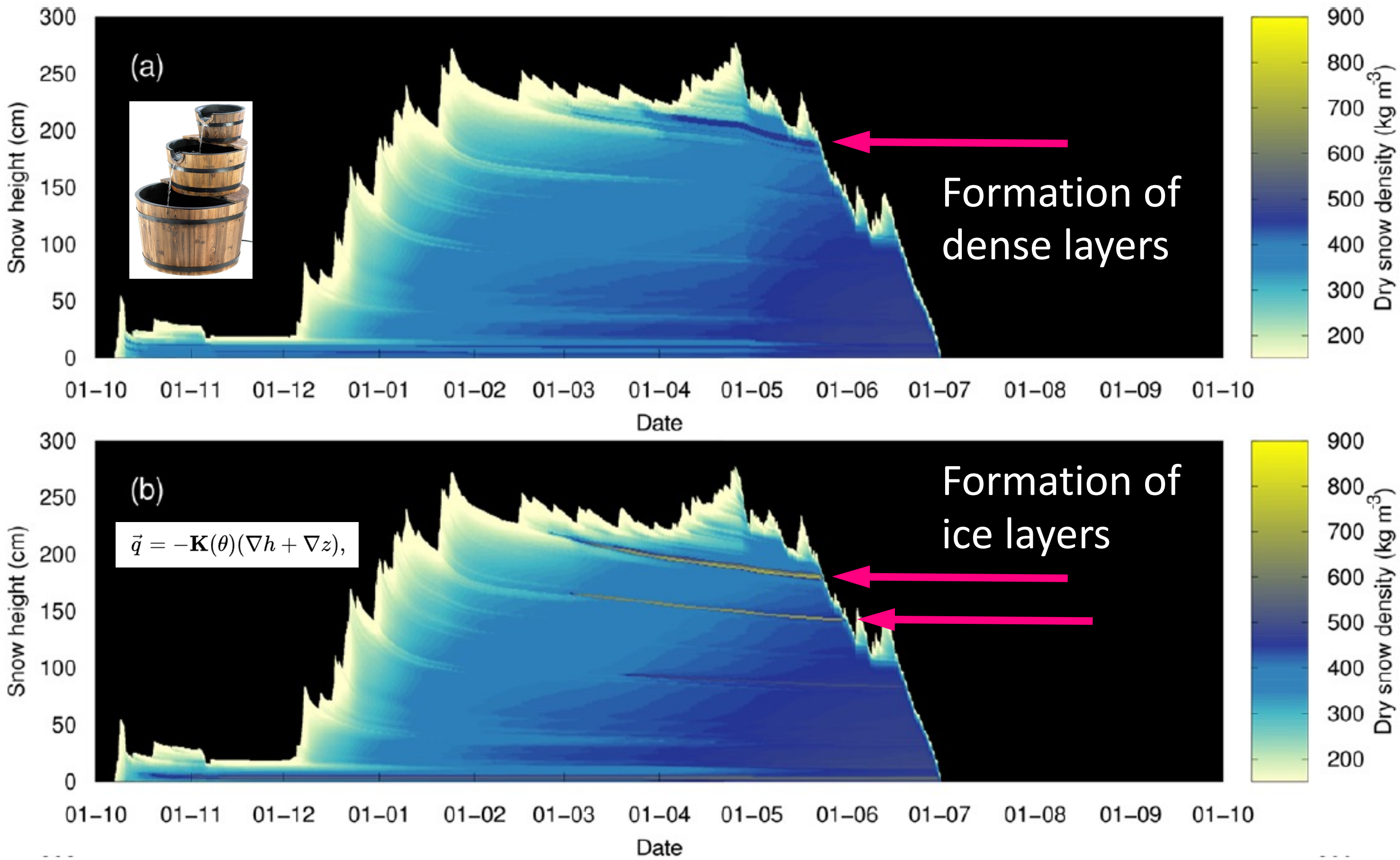


(a)



Formation of ice layers in SNOWPACK

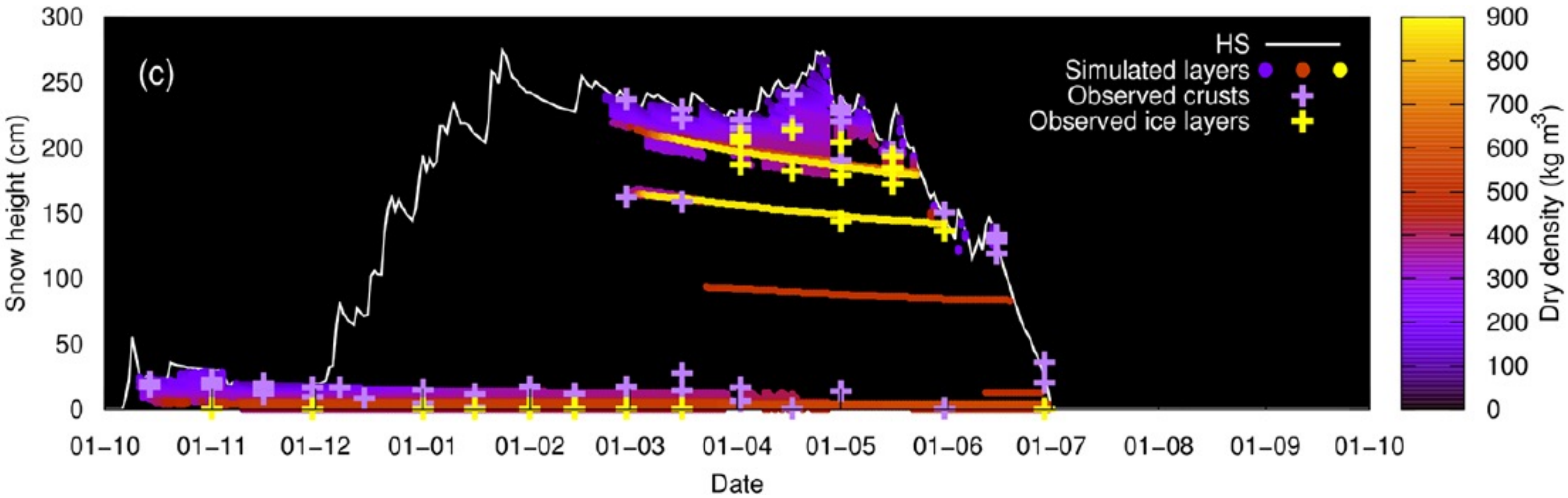
Preferential flow is necessary to have sufficient water transport to sub-freezing layers, where water can pond and then form an ice layer. The comparison shows the formation of ice layers in panel (b) as opposed to just denser snow layers in (a).



Wever et al., 2016

Agreement with observations

Simulated ice layers and high-density crusts compare well with observations.



Wever et al., 2016

Example of snowmelt simulation

Snowmelt dynamics observed and simulated with the Snowpack and StreamFlow models in sub-catchments of the Dischma Valley, Davos, GR.
BK: Bucket Scheme, RE: Richards Equation

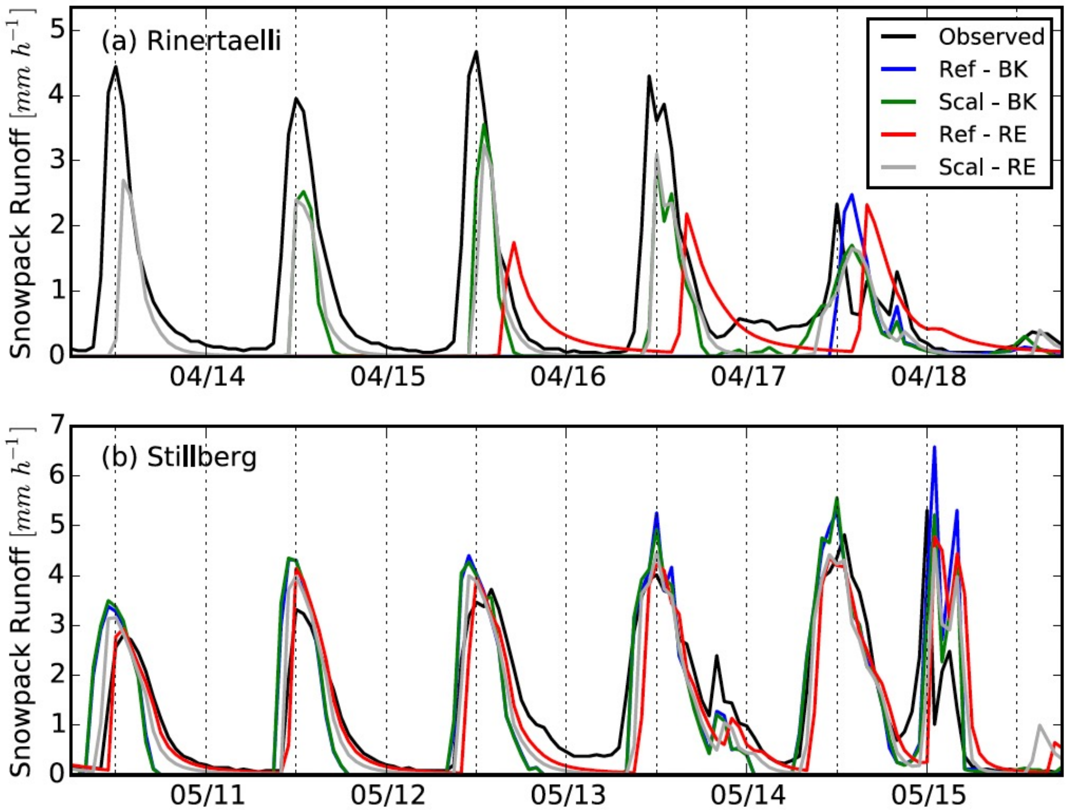


Figure 5. Comparison of observed and simulated liquid water output at the base of the snowpack in (a) Rinertaelli and (b) Stillberg for the four model configurations.

Brauchli et al. 2017, WRR

Modulating effect of snow on runoff

An existing snow cover may:

- Initially retard runoff formation
- Lead to overall increased runoff
- Change timing of peak runoff

Rain On Snow (ROS) events:

- Drastically change snowpack properties
- Can lead to surface or internal crusts
- Have had disastrous consequences

Def.: runoff excess = runoff – precipitation

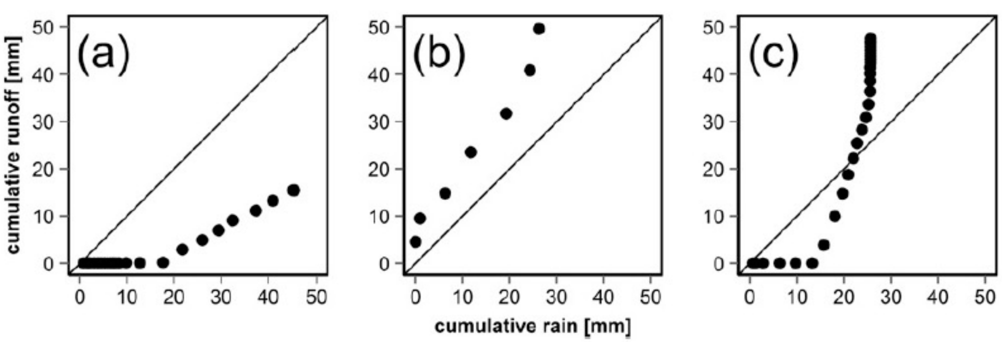


FIG. 4. Temporal trajectory of the ratio between cumulative snowpack runoff and cumulative rain input at hourly time steps for three ROS event patterns.

Würzer et al., JHM, 2016

Rain on snow events

Oroville Dam nearly failed in 2017 after a ROS event.
188’000 evacuees,
USD 1 Billion damage!

The role of EB terms for ROS

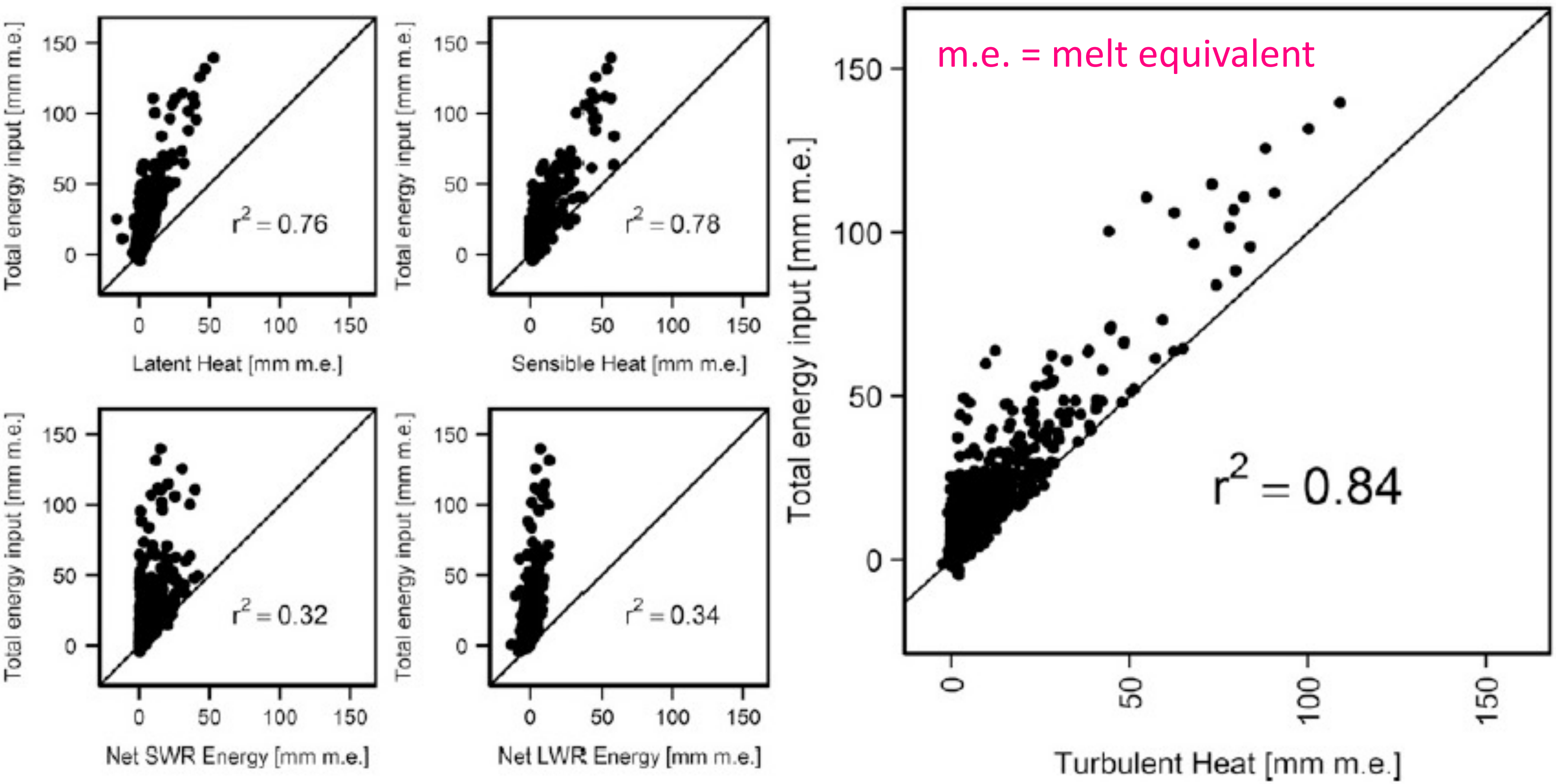


FIG. 5. Contribution of single energy balance terms to the total energy input to the snowpack during all ROS events analyzed: latent heat, sensible heat, net shortwave radiation, net longwave radiation, and turbulent (sensible plus latent) heat flux (mm m.e.). Ground heat flux and heat advected by rain are not shown.

Würzer et al., JHM, 2016

Runoff excess, time lag, rain intensity

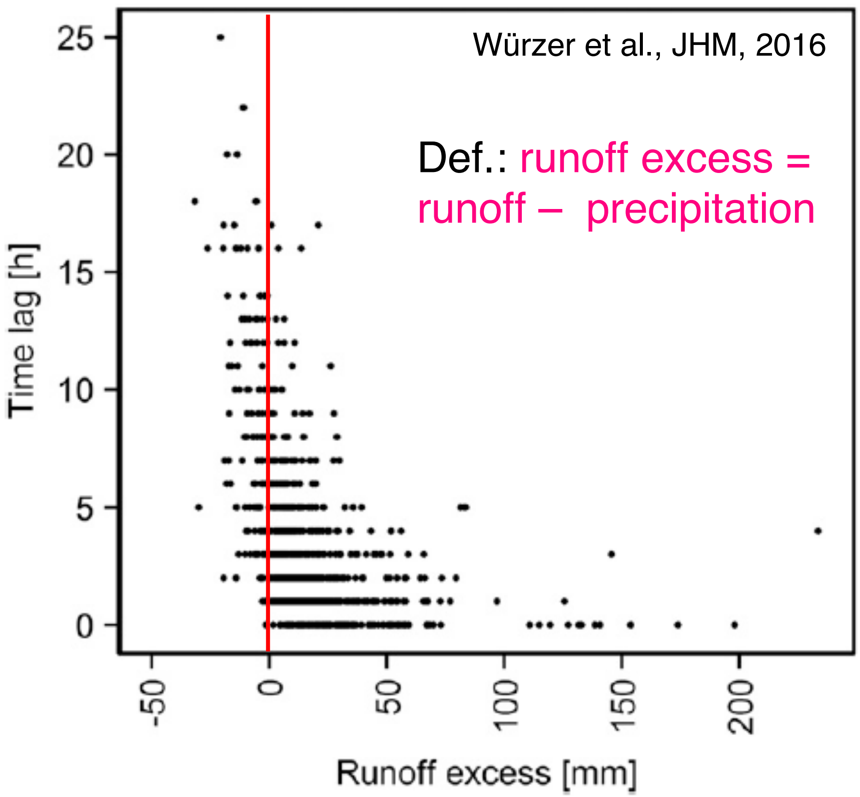


FIG. 6. Time lag between the start of rainfall and the start of snowpack runoff as it relates to runoff excess. Events without runoff are excluded from the figure (no time lag).

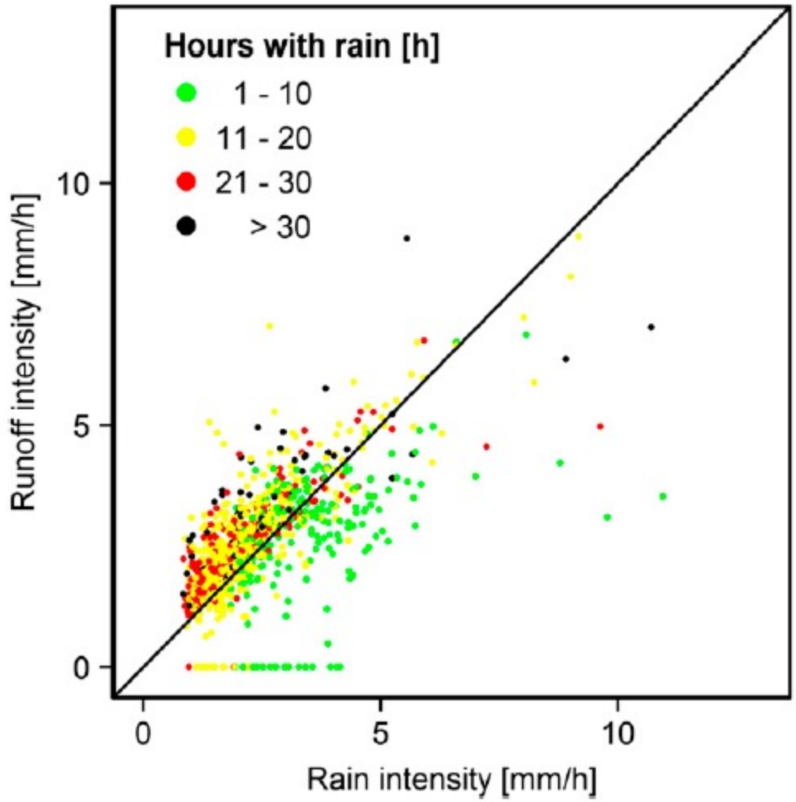


FIG. 9. Runoff intensities (averaged over the duration of runoff) vs rainfall intensities (averaged over the duration of rainfall) for rainfall events of different length.

Events with large time lag typically do NOT produce large runoff excess.

Events with short time lag typically produce LARGE runoff excesses.

For intense and short rain events, the snow cover attenuates discharge intensities.

For moderate and longer rain events, the snow cover amplifies discharge intensities.

Melt water flow paths (water routing)

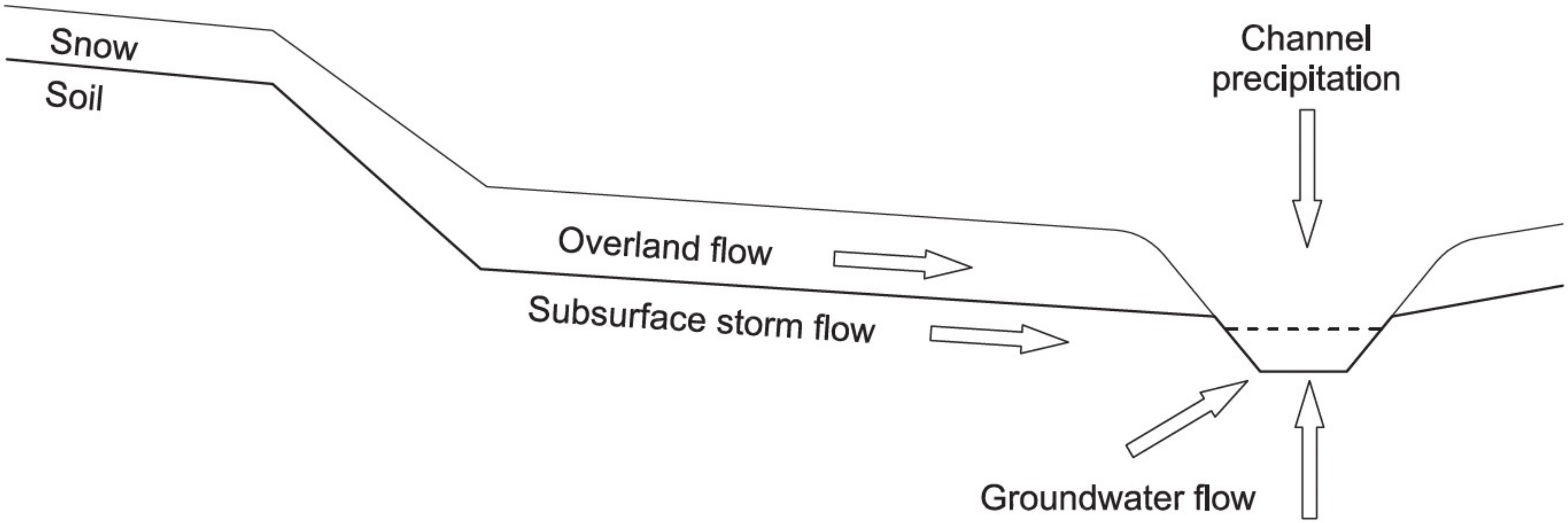


Figure 9.1 Diagram showing four major hydrologic flow paths possible for snow-fall and meltwater.

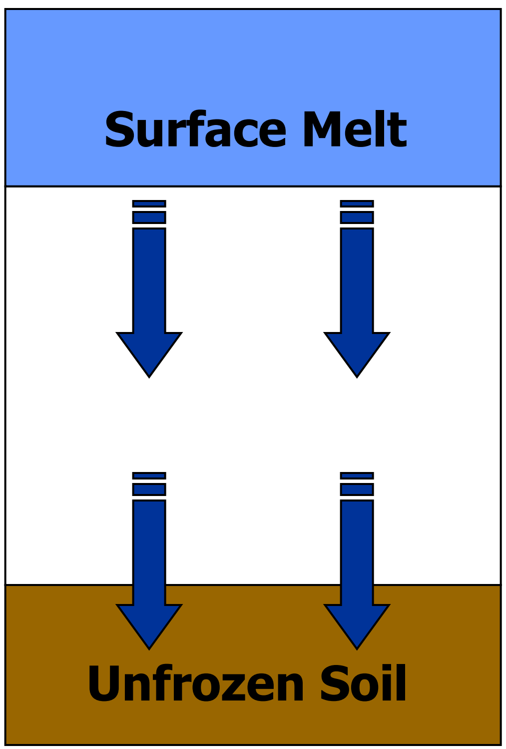
Infiltration rate vs. output rate. Depends on the state of soil.

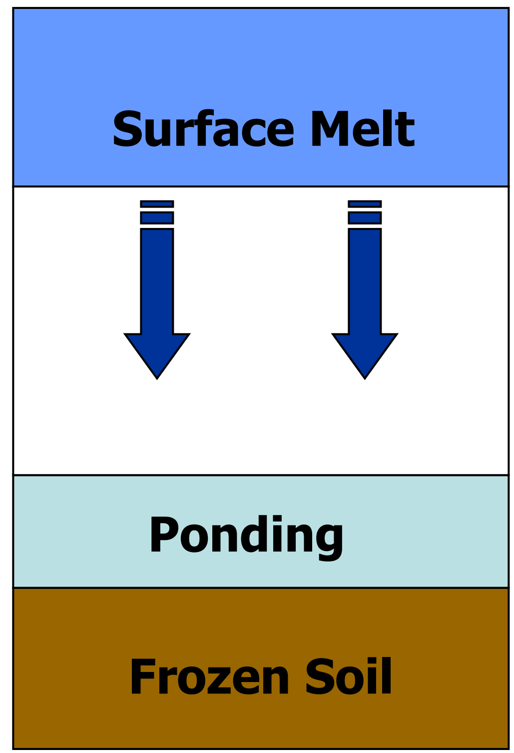
Fate of snowmelt

Depends on slope, snowpack, and soil conditions.

Snowmelt encountering thawed, permeable soil, at a rate less than the infiltration rate, will enter the soil. In this case it behaves much like rainfall would.

Snowmelt encountering frozen soil at the base of the snowpack, or other impediments to infiltration, may pond at the snow/soil interface.

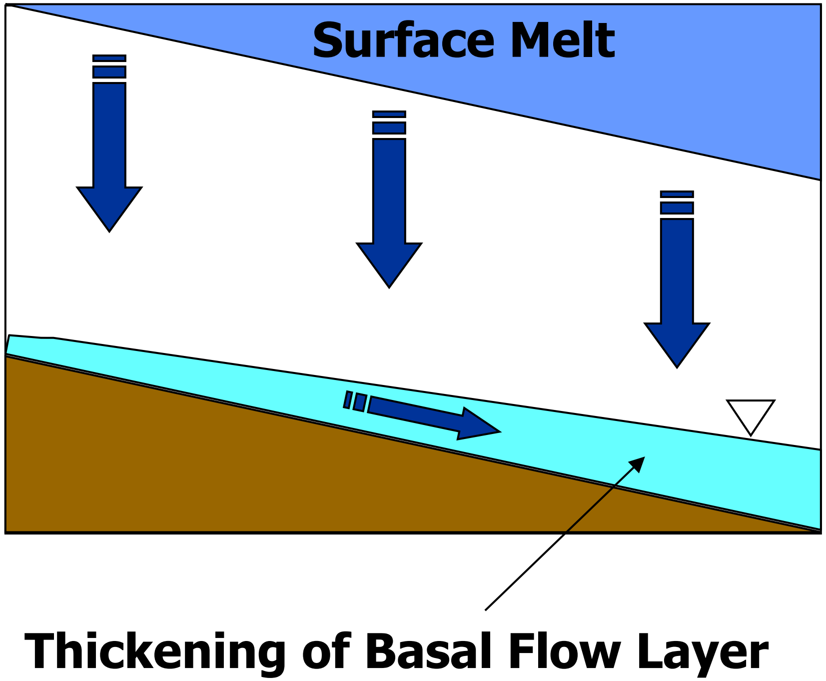




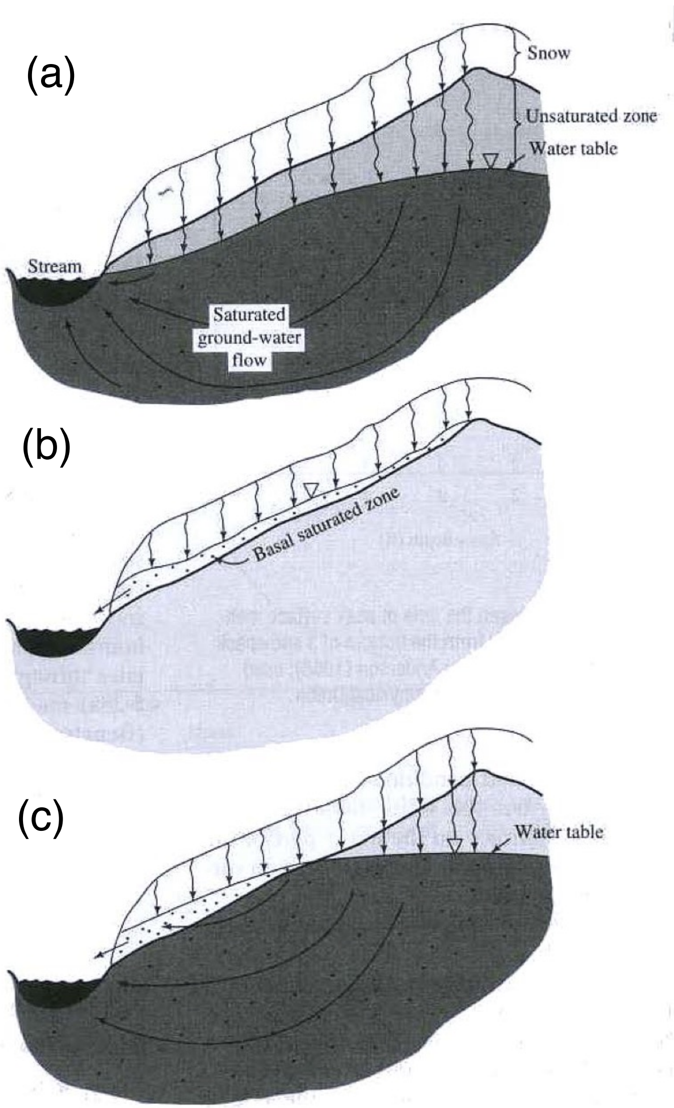
Fate of snowmelt

Sub-nival flow on a slope

- Lateral flow of basal ponded water may develop, depending on slope. If snow is still present, lateral flow is still through a porous medium.
- Liquid water at the base of a snowpack causes rapid destruction of small snow grains, leaving larger grains, allowing for more rapid flow (positive feedback).



Modes of runoff generation



The water table is deep, and percolating meltwater infiltrates to raise the water table inducing increased groundwater flow to the stream.

Saturation excess, or the water table is at the soil surface, or the soil surface is impermeable (e.g., frozen ground), then percolating meltwater forms a basal saturated zone.

The lower portion of the water table rises above the ground surface into the snowpack. Water in the upper part of the slope moves as in (a), and the water in the lower part of the slope behaves as in (b).

Dingman 2002, Figure 5-28

Basal ice development

On shallow slopes, ponded meltwater may refreeze (at the base of the snowpack), forming ice layers that may impede further meltwater infiltration into the soil.

Figure 9.4 Rapid delivery of meltwater as overland flow on frozen ground in a pasture swale that has turned the snow into a slush layer. See also color plate.

Frozen ground and stream flow

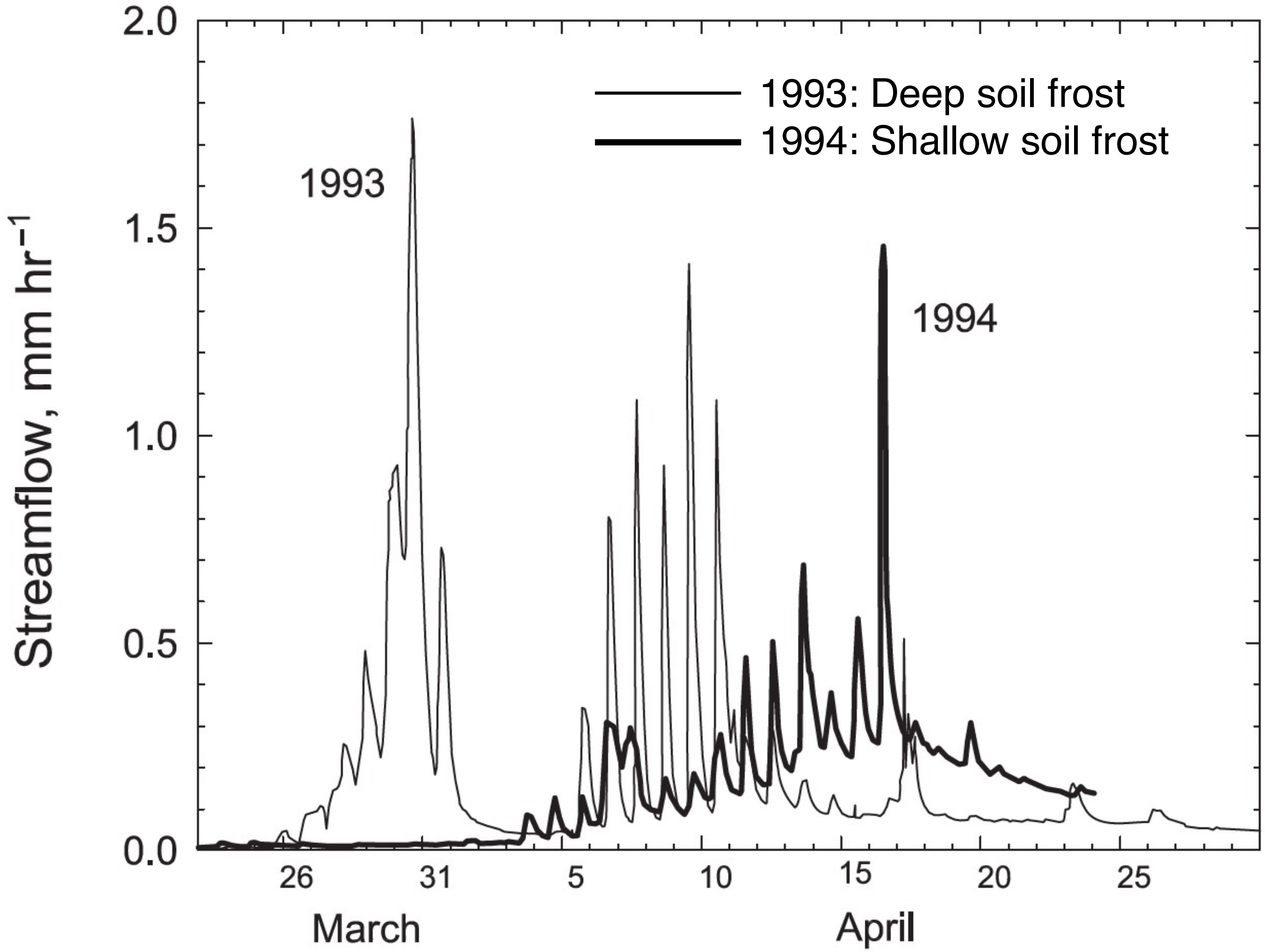


Figure 9.3 Streamflow from snowmelt in 1993 with deep soil frost and 1994 with shallow soil frost on a Vermont agricultural watershed (Shanley and Chalmers, 1999 © 1999 John Wiley & Sons Limited, reproduced with permission).

Stream flow hydrograph for alpine catchment

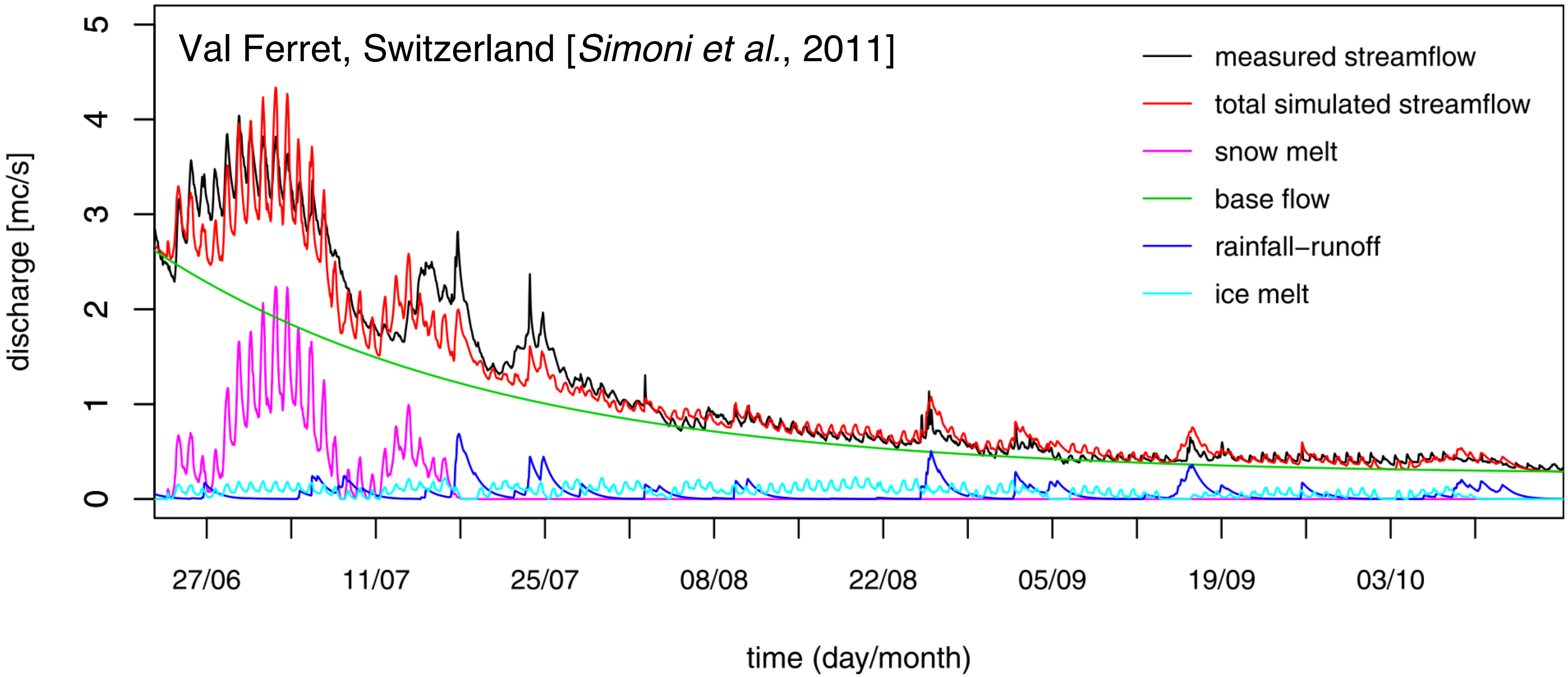


Figure 12. Simulated discharge (red line) obtained with the lumped model using spatially averaged air temperatures compared to streamflow measurements (black line). The four modeled components contributing to the streamflow are displayed with different colors: the snowmelt runoff (pink line), the ice melt runoff (light blue line; see (3)), the rainfall runoff (dark blue line; see (5)), and the base flow (green line; see (6)).

Snowmelt in the Alps…

Next:


Accumulation,
snow transport,
spatio-temporal
variability of snow
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