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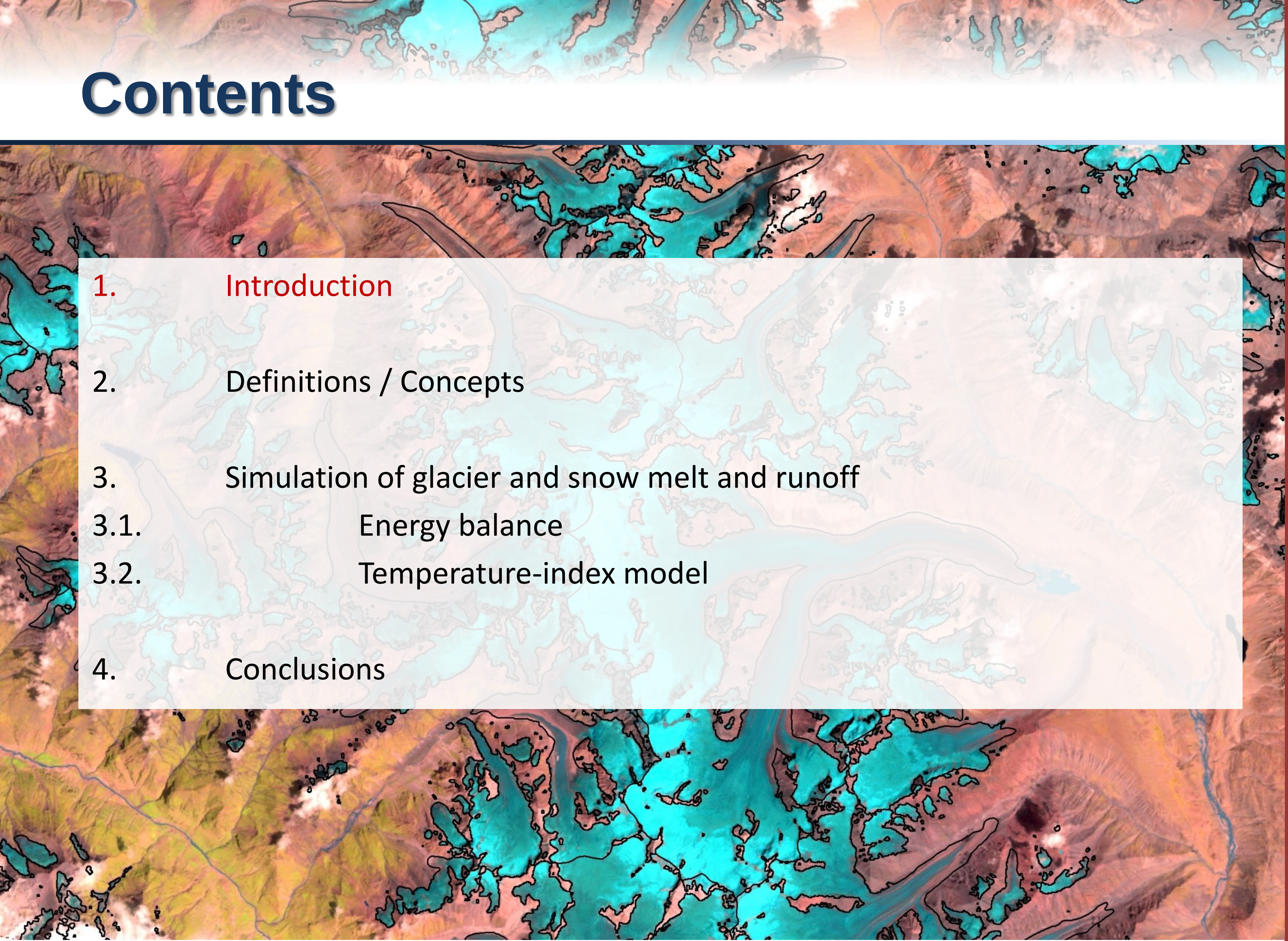
Glacier and snow melt and runoff

Mario Rohrer

IHCAP – Indian Himalayas Climate Change Adaptation Programme
Capacity building programme “Cryosphere” Level-1 (August 18 - September 15, 2014)

Contents



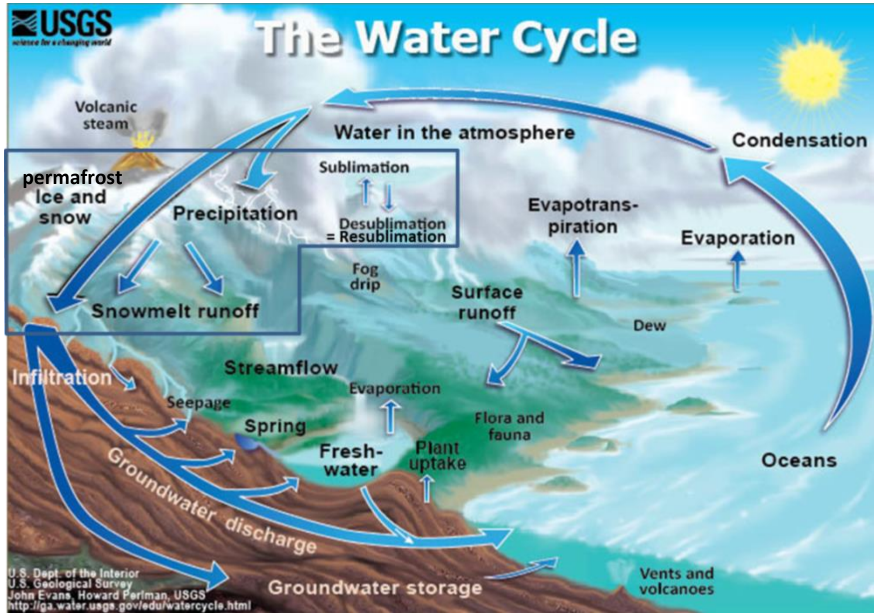
1. Introduction

Why are we interested in snow and glacier melt and runoff?

- Snow and ice are **important parts of the water cycle** in mountainous and high latitudes terrain

- Snow and ice influence the water budget of a catchment by
  storing the water and
  releasing it when melted

- In the Himalayas,
  the runoff generated from snow and ice may represent an
  important part of the total stream flow



U.S. Dept. of the Interior
U.S. Geological Survey
John Evans, Howard Perlman, USGS
http://ga.water.usgs.gov/edu/watercycle.html

The Water Cycle, USGS 2013, http://ga.water.usgs.gov/edu/watercycle.html

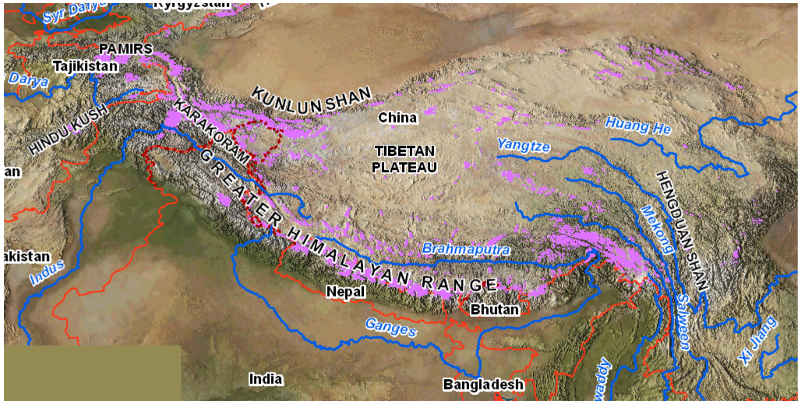
The blue box indicates the part of the hydrologic cycle where water is present as snow or ice.

1. Introduction

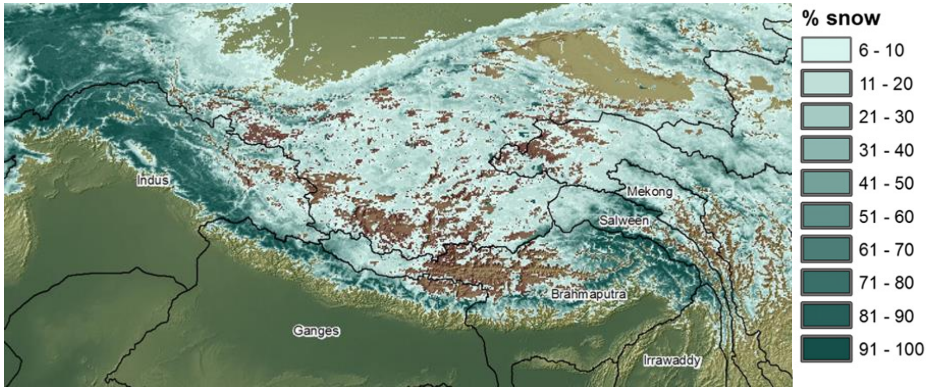
Why are we interested in snow and glacier melt and runoff?

- Runoff from mountain catchments is influenced by the storage of precipitation in forms of glaciers and snow
- The amount of melt per area is often larger for glaciers due to the smaller albedo of ice

As the snow covered area in winter for many Himalayan catchments is much larger than the glacier area (see figure), the contribution of snow may be more important than that of glaciers



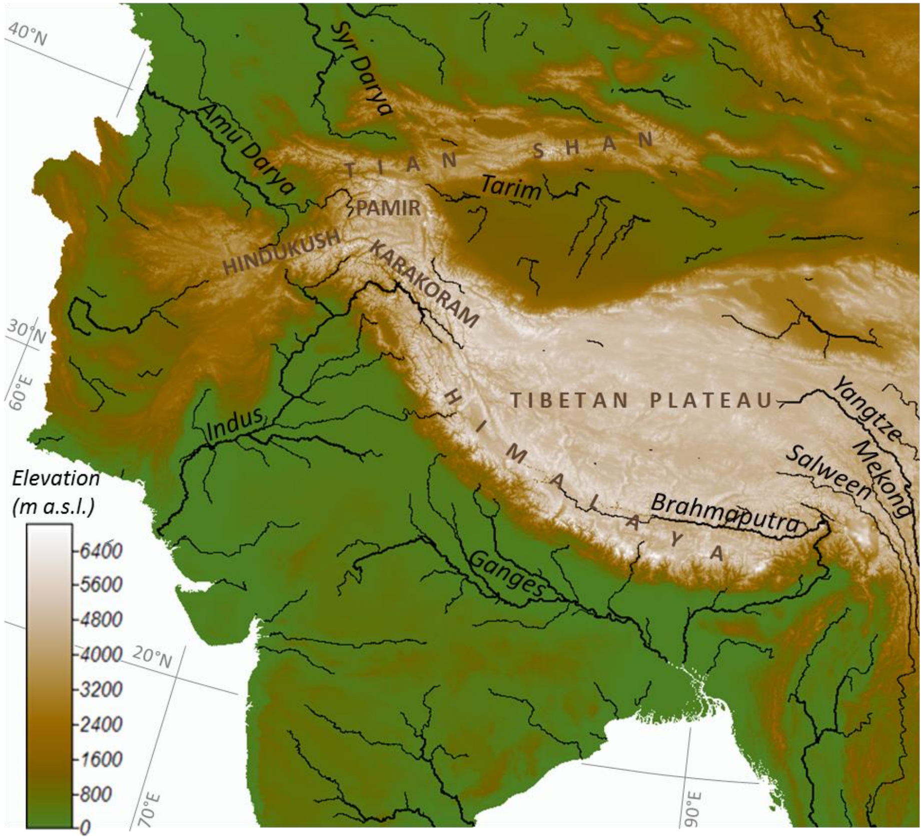
Outlines of High Asian glaciers in pink (USAID)



Seasonal snow cover extent in winter based on MODIS from Mar 2000 to Feb 2008. (Immerzeel et al., 2009)

1. Introduction

The importance of "snow and glacier melt and runoff" in the Himalayas



The map shows some of the **major Asian rivers** originating in mountainous catchments

The **main characteristics** of these watersheds in respect to snow and ice melt and runoff:

- seasonally snow-covered
- glacierized

1. Introduction

Objective of melt modelling and main groups of melt models

Melt modelling is a crucial element in any attempt to

- operational forecast runoff from snow-covered or glacierized areas
- assess changes in the cryosphere associated with climate change

There are two main groups of melt models:

- **energy balance models**
  attempting to quantify melt as a residual in the heat balance equation
- **temperature-index models**
  assuming an empirical relationship between air temperatures and melt rates

-> First: We need some definitions and concepts

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- snow types
- aggregate states of water
- vapor pressure, snow density, snow water equivalent, albedo

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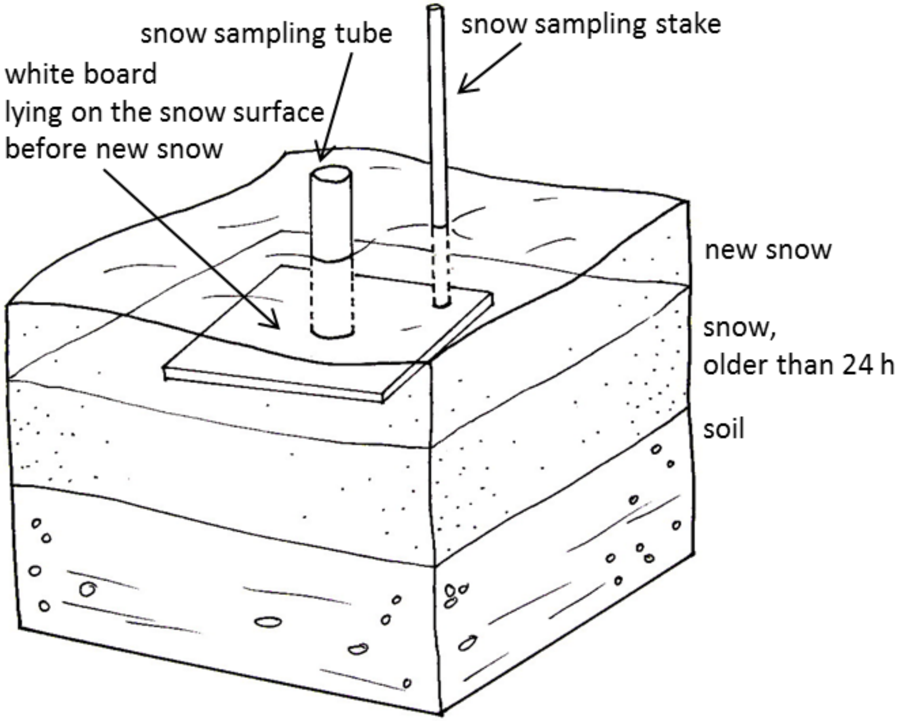
4. Conclusions

2. Definitions/Concepts

new snow, old snow, firn

New snow	Snow deposited within an interval of 24 hours
Old snow	Deposited snow where the transformation is so far advanced that the original form of the ice crystals can no longer be recognized



Sketch of a snowpack with new snow. The white board is used to measure the depth and density of new snow

2. Definitions/Concepts

new snow, old snow, firn

Firn 1) Snow that has survived at least one ablation season but has not been transformed into ice

2) Structurally, the metamorphic stage intermediate between snow and ice, in which the pore space is at least partially interconnected, allowing air and water to circulate


Meteodat GmbH

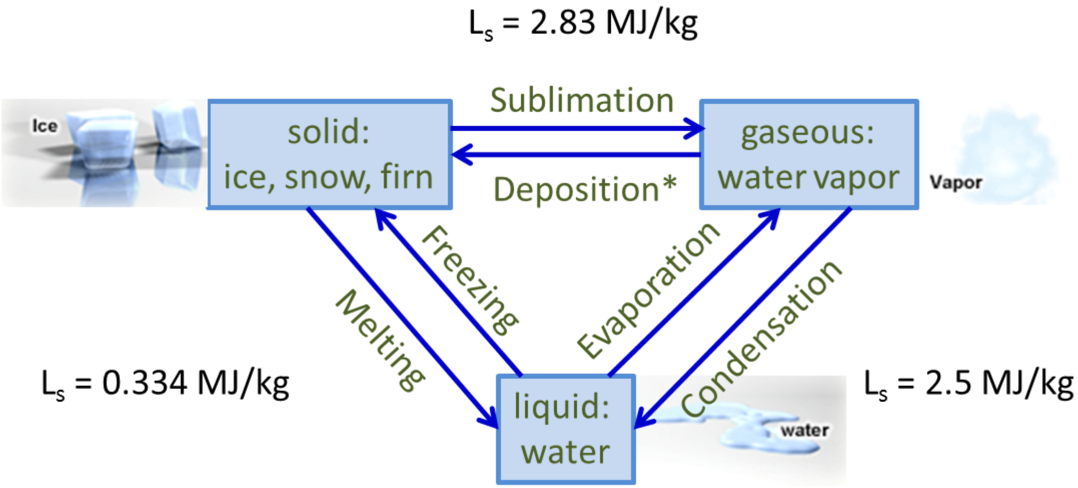
Old snow


SwissEduc

Firn

2. Definitions/Concepts aggregation state of water

Aggregation states and phase changes of water



L_s = 2.83 MJ/kg

solid:
ice, snow, firn

Sublimation

Deposition*

gaseous:
water vapor

Melting

Freezing

Evaporation

Condensation

liquid:
water

L_s = 0.334 MJ/kg

L_s = 2.5 MJ/kg

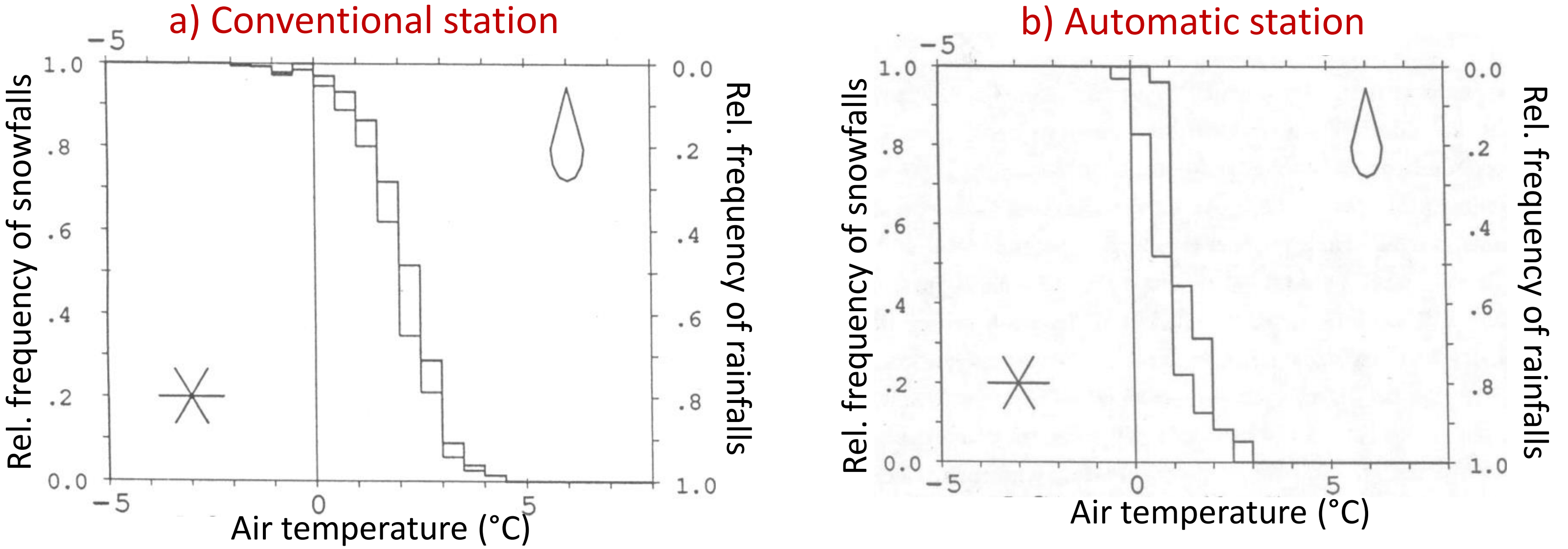
** "deposition"
is also also
called:
"desublimation"
"sublimation"
"resublimation"*

- Latent heat is the 'hidden heat' that is exchanged with the surrounding environment in the event of a phase change of water.

- **Latent heat of sublimation or vaporization is about eight times higher than the latent heat of fusion**

2. Definitions/Concepts aggregational state of precipitation

Aggregation state of precipitation


a) Conventional station

b) Automatic station

Relative frequency of solid, mixed and liquid precipitation in function of air temperature for Davos, Swiss Alps, between 1950 and 1976 at an elevation of 1590 m a.s.l. (Rohrer, 1992)

decisive if precip. falls as snow or rain:

- height of the zero-degree-line in the atmosphere
- altitude of the catchment

$T < 0°C$	snow
$0°C < T < T_{CRIT}$	linear interpolation of the fraction of rain and snow
$T > T_{CRIT}$	rain (e.g. $T_{CRIT} = 2°C$)

2. Definitions/Concepts

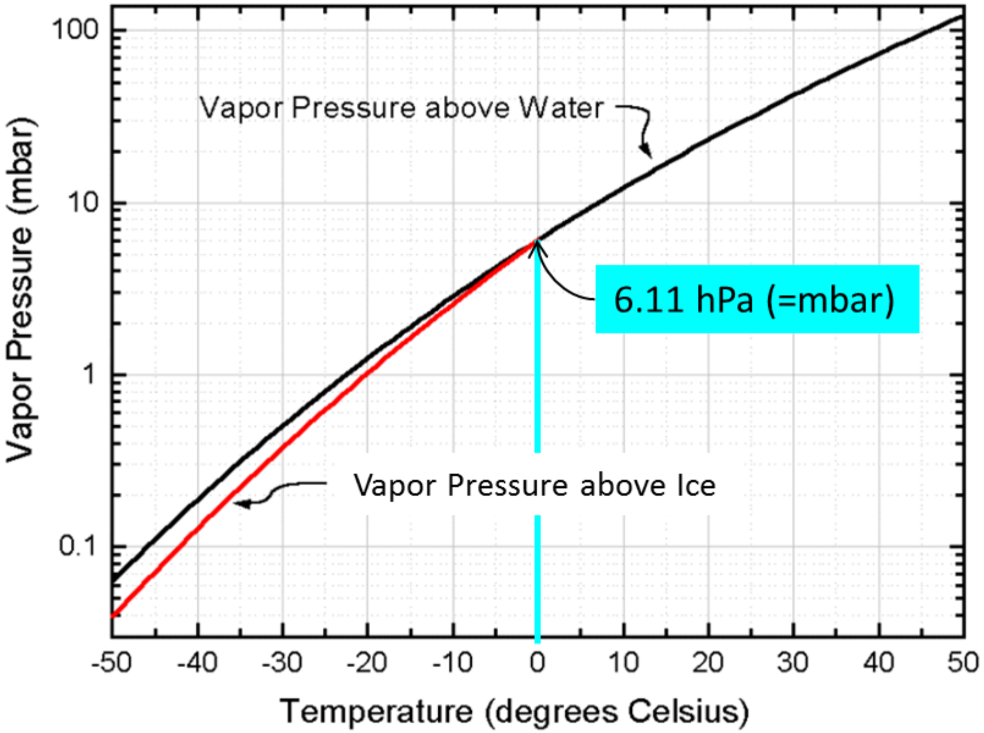
Vapor pressure of water

The vapor pressure is the pressure of the vapor in **equilibrium** with its condensed phase (solid or liquid).

When the rate of water molecules entering the liquid equals the rate leaving the liquid, we have equilibrium. At the equilibrium vapor pressure the air is **saturated** with water vapor.

The vapor pressure of water varies with its temperature

California Institute of Technology



Equilibrium water vapor pressure of ice and water as a function of temperature

2. Definitions/Concepts

vapor pressure

Vapor pressure over snow and ice

The temperature of a snow or ice surface can not exceed 0°C

Thus, the maximum vapor pressure over snow and ice is **6.11 mbar (= hPa)** and in humid climates there are often condensation conditions

Condensation over a melting snow or ice surface leads to:

-> **increase of available melt energy** because of the releasing latent heat

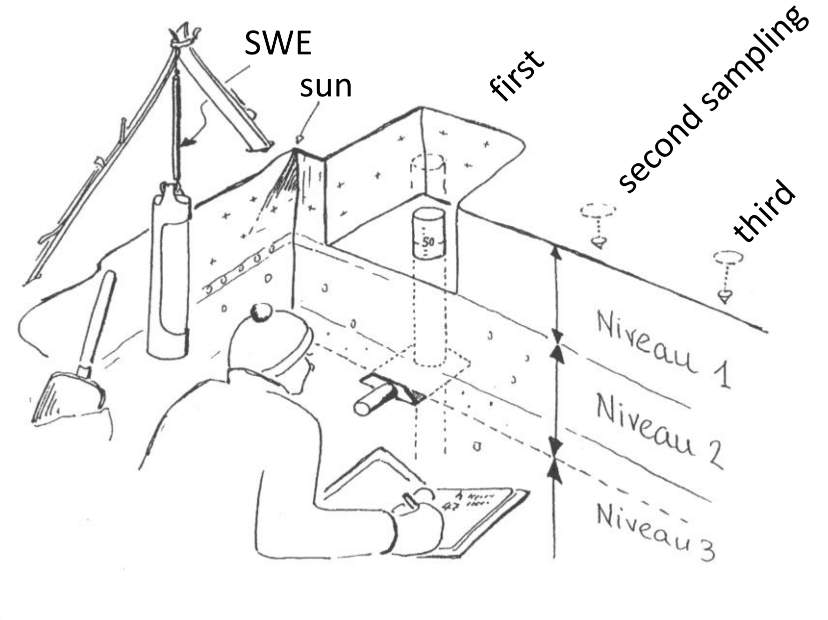
-> increase of moisture of the snow or ice surface
-> often decrease of the albedo
-> increase of the net shortwave radiation
-> **increase of available melt energy**

-> higher air humidity
-> The higher the air humidity, the higher the net longwave radiation
-> **increase of available melt energy**

2. Definitions/Concepts

density of snow and ice

Density of snow and ice



Sampling snow density in a snow pit (G. Kappenberger)

$$\boxed{\rho_s = \frac{m_s}{V_s}}$$

ρ_s	Snow density (kg / m^3)
m_s	Weighted mass of the snow probe (kg)
V_s	Volume of the snow probe (m^3)

S. Schauwecker

Student in a snow pit with a depth of around 2.50 m

S. Schauwecker

Weighting the snow probe with a digital balance

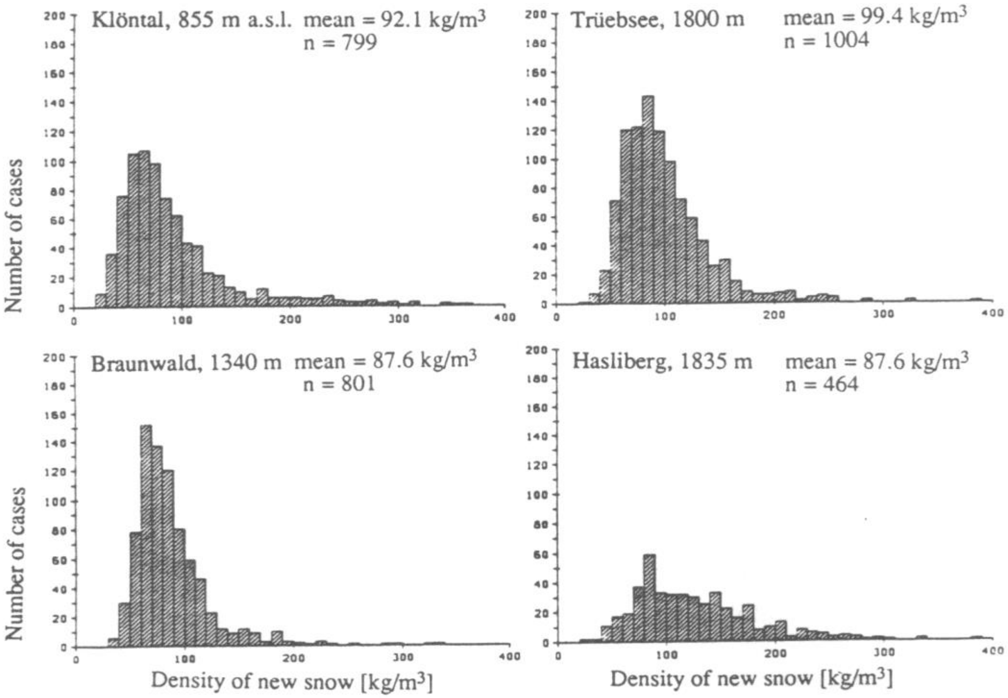
SwissEduc

Weighting the snow probe with a spring balance

2. Definitions/Concepts

density of snow and ice

Density of new snow



*Distribution of **density of new snow** as measured at 4 stations in the Swiss Alps. (Rohrer et al., 1994)*

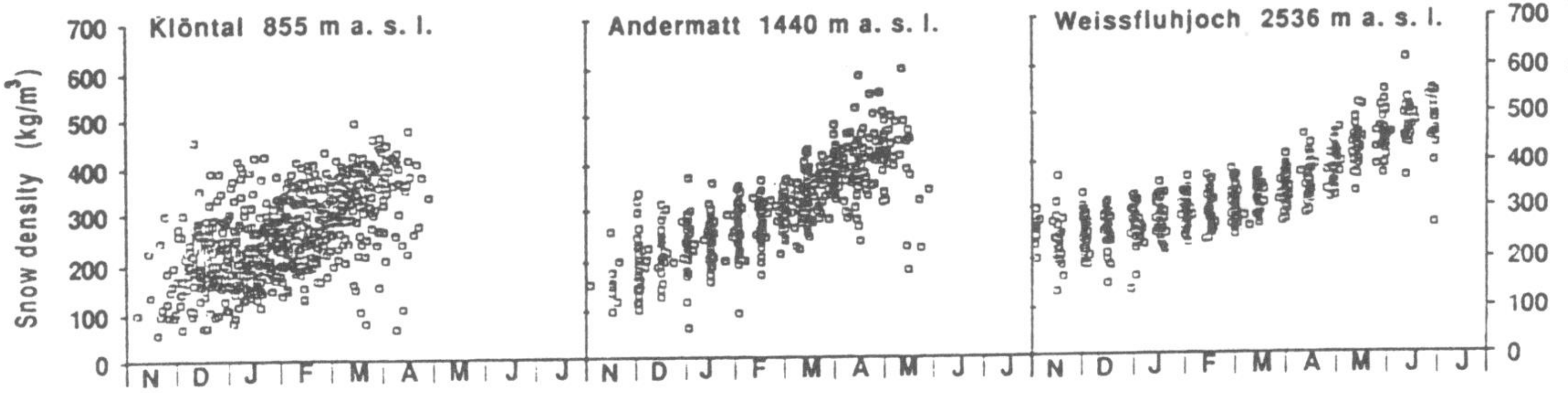
a) Klöntal, 855 m a.s.l.
1967/68 - 1984/85
b) Braunwald, 1340 m a.s.l.
1960/61 - 1984/85
c) Trübsee, 1800 m a.s.l.
1956/57 - 1984/85
d) Hasliberg, 1835 m a.s.l.
1960/61 - 1984/85

- There is no apparent dependency between snow density and elevation
- The mean value for Switzerland is 99.3 kg/m^3
  (based on 48 locations with 22166 measurements)
- The densities range between around 20 kg/m^3 up to more than 300 kg/m^3

2. Definitions/Concepts

density of snow and ice

Density of total snow cover



Seasonal variation of measured snow density of total snow cover at 3 index points in the Swiss Alps (Rohrer et al., 1994)

- *a)* *Klöntal, 855 m a.s.l.*
  1945/46 - 1984/85
- *b)* *Andermatt , 1440 m a.s.l.*
  1946/47 - 1987/88
- *c)* *Weissfluhjoch , 2536 m a.s.l.*
  1947/48 - 1984/85

- There is a steady increase of snow density with progressing season within a band with of about 150 to 220 kg/m^3 at high elevations
- In lower elevations the trend is less pronounced and the band width is much larger

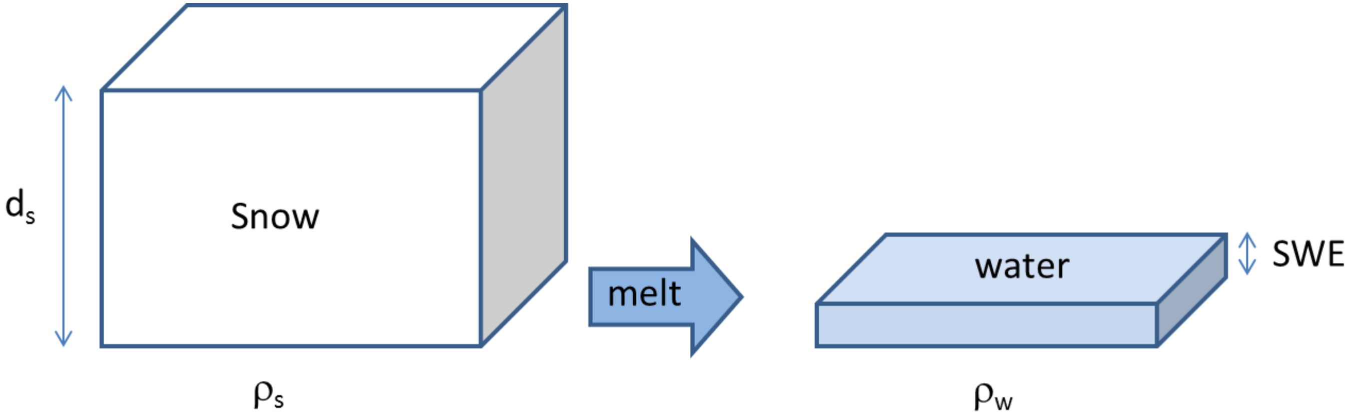
2. Definitions/Concepts

Snow water equivalent

Snowpack **water equivalent** is one of the most important properties of snowpacks needed by snow hydrologists.

The water equivalent of a snowpack represents the **liquid water that would be released upon complete melting of the snowpack**.

Water equivalent is measured directly or computed from measurements of depth and density of the snowpack



2. Definitions/Concepts

snow water equivalent

How to compute SWE?

$$SWE = d_s \frac{\rho_s}{\rho_w}$$

SWE	water equivalent (m)	calculated
d_s	snowpack depth (m)	measured
ρ_s	snowpack density (kg m^{-3})	measured
ρ_w	density of liquid water approx. 1000 kg m^{-3}	given


Meteodat GmbH

measuring snow depth


Meteodat GmbH

measuring snow depth with a GPS and a snow probe

2. Definitions/Concepts

snow water equivalent

Measuring SWE with ground-based methods

Manual method

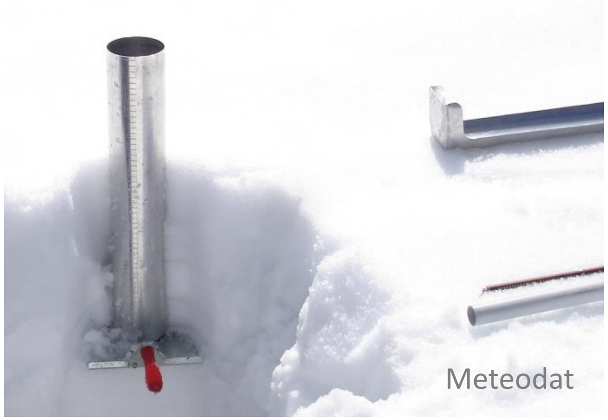
Snow tube

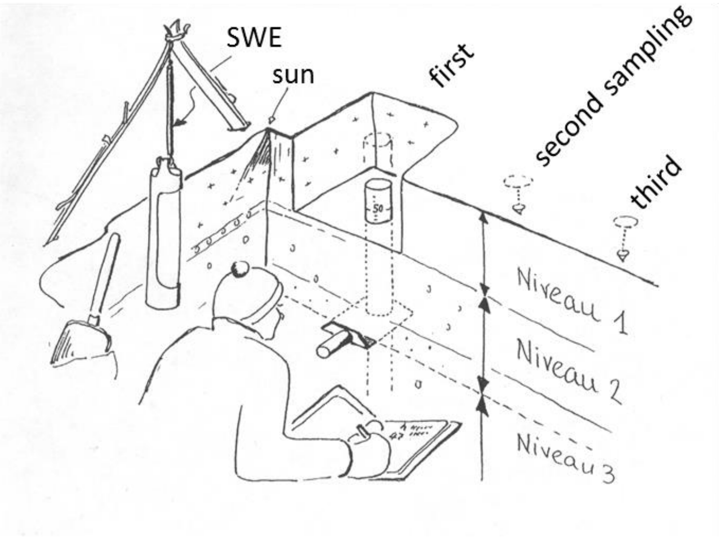
Commonly, the snow tube is a cylindrical snow sampler.

- The ETH-tube is a graduated aluminum tube with a height of 55 cm and a base of 70 cm^2.
- **Disadvantage**: time consuming

Measuring SWE:

1. digging a snow pit to the ground
2. inserting the snow tube vertically
3. measuring height of snow probe
4. determining the net weight of snow probe
5. calculating density (net weight/volume)
6. calculating SWE
7. repeat procedure for all levels and 1-3 columns (see figure)


Meteodat



2. Definitions/Concepts

snow water equivalent

Measuring SWE with ground-based methods

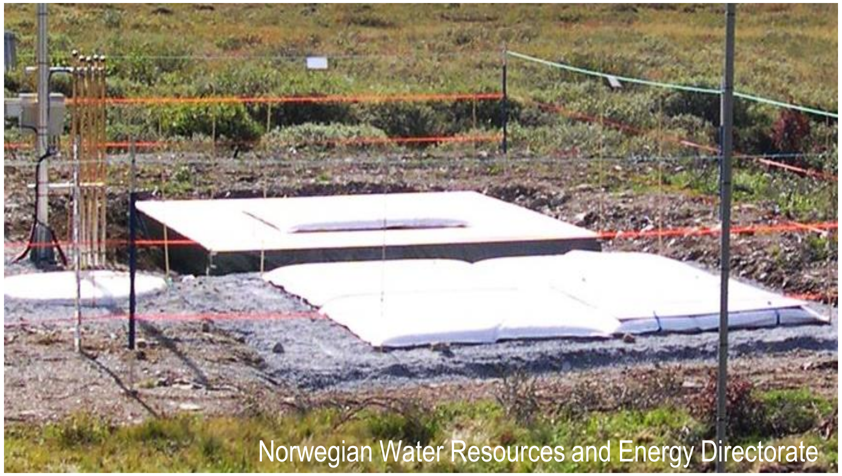
Continuous (automatic) methods

Snow pillow

- most prevalent continuous method of SWE measurement
- measures snow mass by measuring loads on liquid-filled bags (pillow)
- a sensor is measuring the hydrostatic pressure caused by the snow layer
- **disadvantage**: snow bridging e.g. ice layers may support the weight of additional snowpack, which causes underweight conditions

Gamma ray measurement

- radioactive source beneath the snowpack
- Geiger Muller tube suspended above the snowpack
- the gamma radiation passing through the snowpack is absorbed in function to the SWE
- **disadvantage**: radioactive sources are needed


Norwegian Water Resources and Energy Directorate

2. Definitions/Concepts

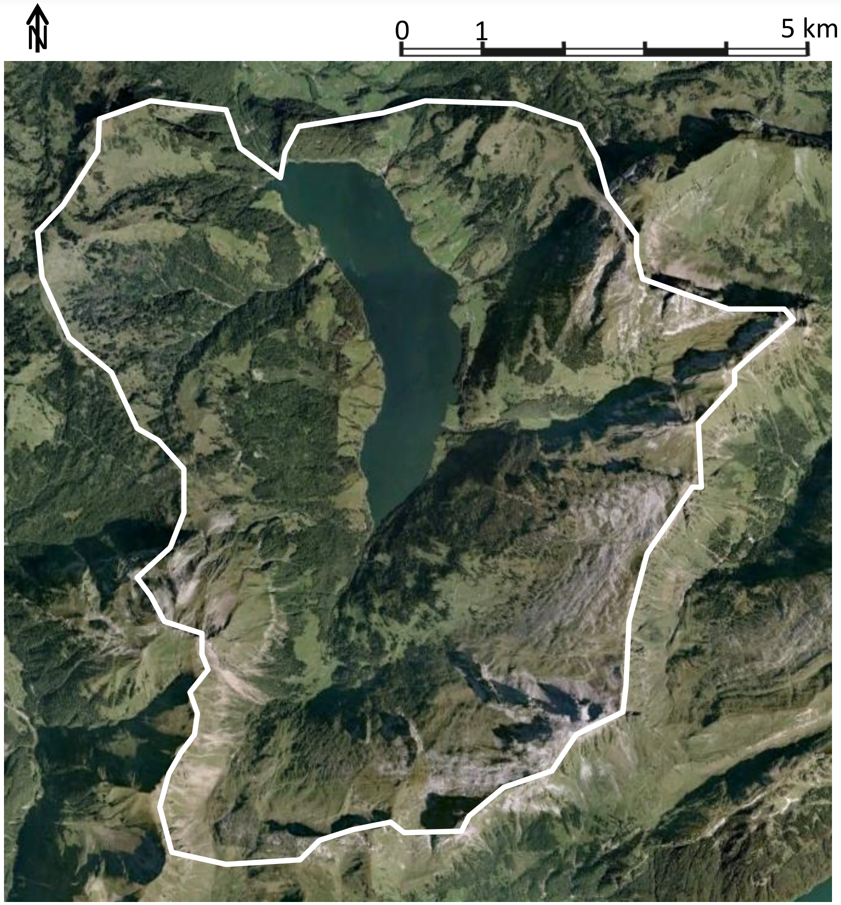
snow water equivalent

Measurements of SWE of a watershed

Spatial variability of snow depth is much higher than the spatial variability of snow density.

→ More snow depth measurements are needed than snow density measurements

→ However, it is important to measure snow depth and density at sites with different elevation and exposition



Wägital watershed in the Swiss Alps (GoogleEarth)

2. Definitions/Concepts

snow water equivalent

Measurements of SWE for a watershed

Example Wägital, Swiss Alps, April 2011

Snow density (n=19)

mean snow density: 0.403 g/cm^3

std of snow density: 0.026 g/cm^3 (4%)

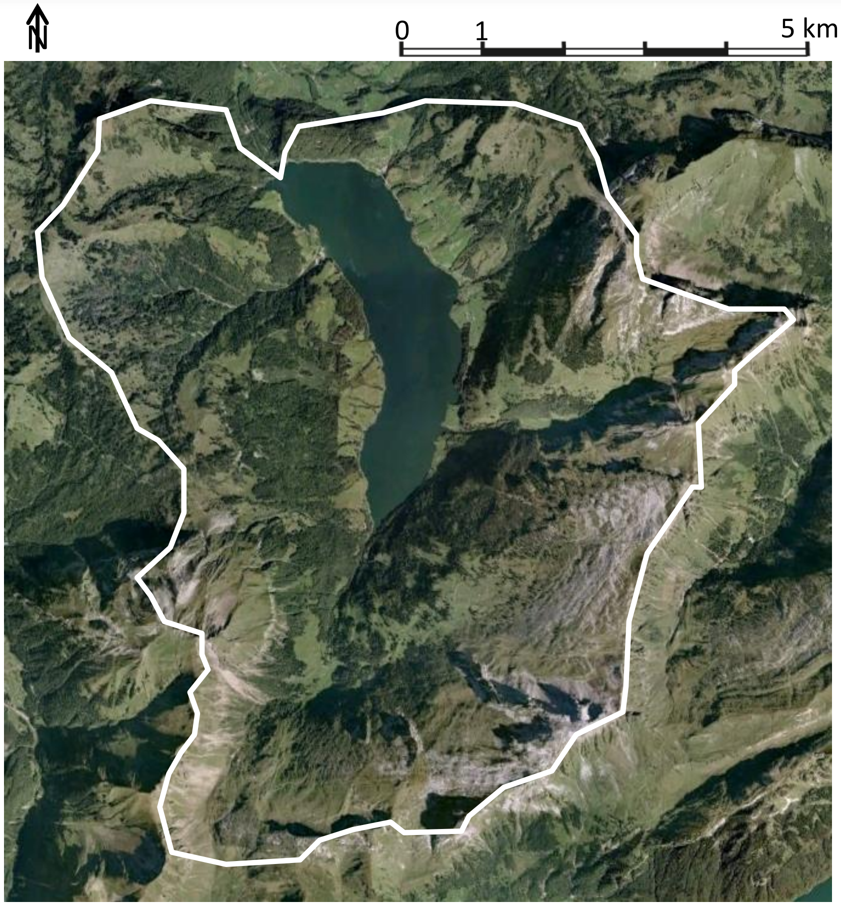
Snow depth (n=19)

mean snow depth: 135 cm

std of snow depth: 69 cm (51%)

Every year in April field measurements are realized at different sample sites:

- snow depth at 30 sites (20-30 samples each)
- snow density at 10 sites (3 samples each)



Wägital watershed in the Swiss Alps (GoogleEarth)

2. Definitions/Concepts

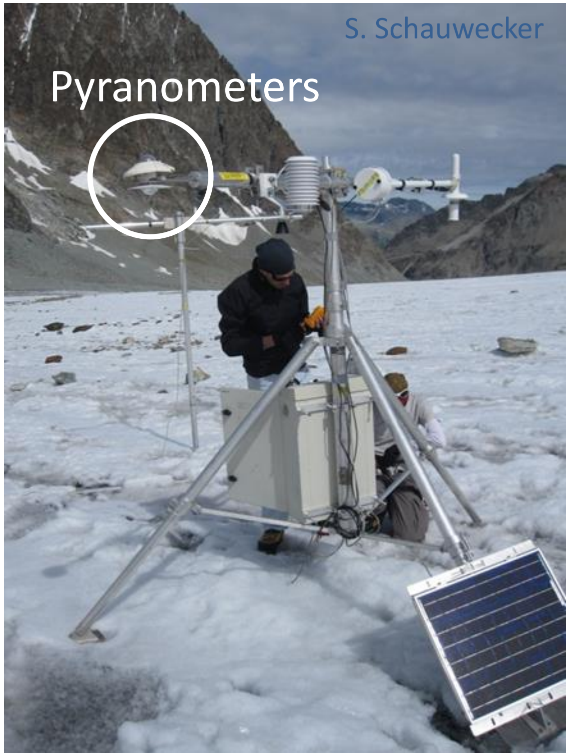
Albedo

Albedo is the ratio of the amount of solar radiation reflected by a surface to the amount incident upon it.

The reflectance is averaged in the approximately spectral range 0.15 – 2 µm.

$$\boxed{\propto = \frac{S_{out}}{S_{in}}}$$

S_{out}	reflected shortwave radiation
S_{in}	incoming shortwave radiation
α	albedo



Automatic weather station with two pyranometers (one looking upwards, one looking downwards) measuring incoming and reflected shortwave radiation at one point on Haut Glacier d'Arolla, Swiss Alps.

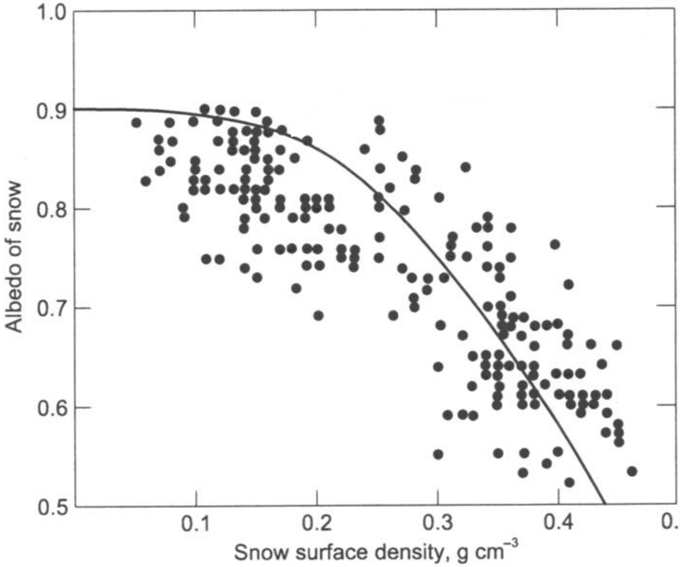
2. Definitions/Concepts

Albedo of snow

Albedo is controlled by the **properties of the surface**, for example grain size, density, light-absorbing impurities or liquid water content.

When snow is fresh and crystals are pristine, the albedo may be as high as about 90%.
As snow ages, the crystal structure often becomes more rounded, due to wet or dry **metamorphism** or from becoming **windblown**. Also **particle accumulation** on the surface and **sun angle** are important to consider.

Anderson (1976) developed an expression for albedo in function of snow surface density

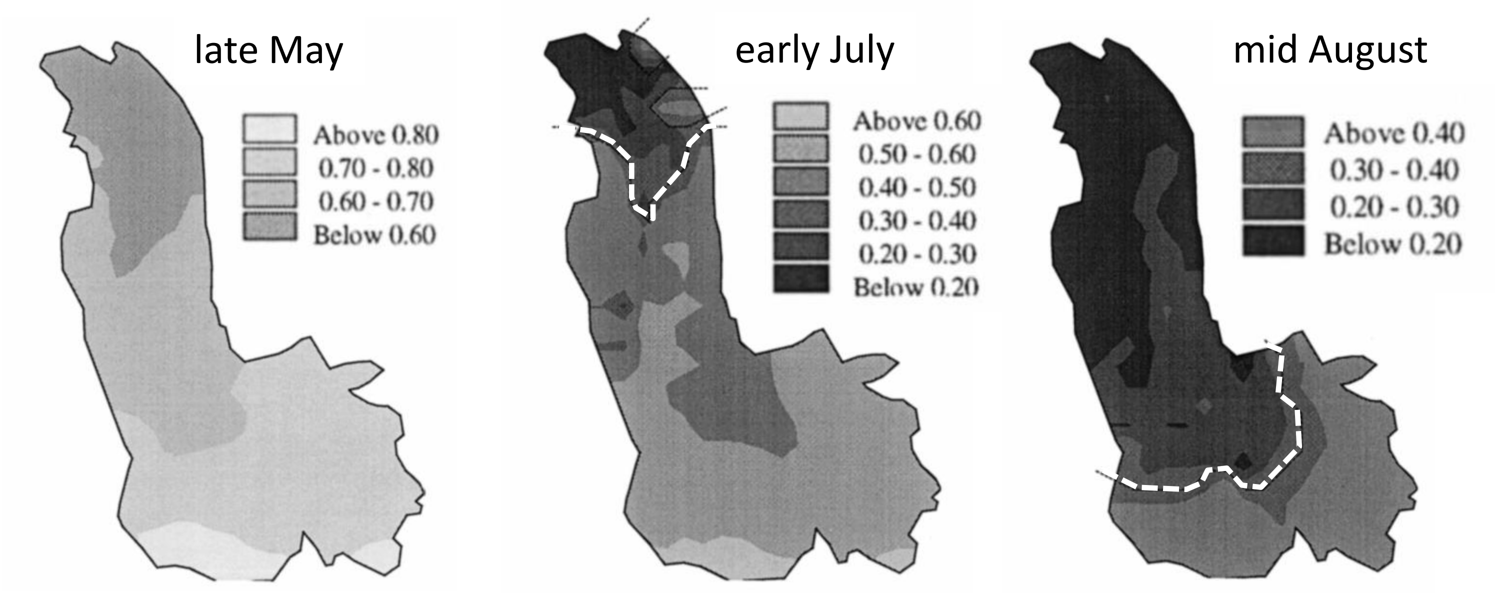


Snowpack albedo as a function of snow surface density (DeWalle and Rango, 2008, modified from Anderson, 1976).

2. Definitions/Concepts

albedo

Albedo of a glacier


late May

early July

mid August

Maps of albedo variation across Haut Glacier d'Arolla (Swiss Alps) in 1993. The line marks the approx. position of the transient snowline below which the glacier is not snow covered. (Brock et al., 2000)

- Albedo α of a glacier is not constant in space and time
- Albedo α of ice is generally lower than α of snow
- The decrease of α resulted mainly from the transition from a complete glacier-wide snow cover to a mixture of surface types

2. Definitions/Concepts

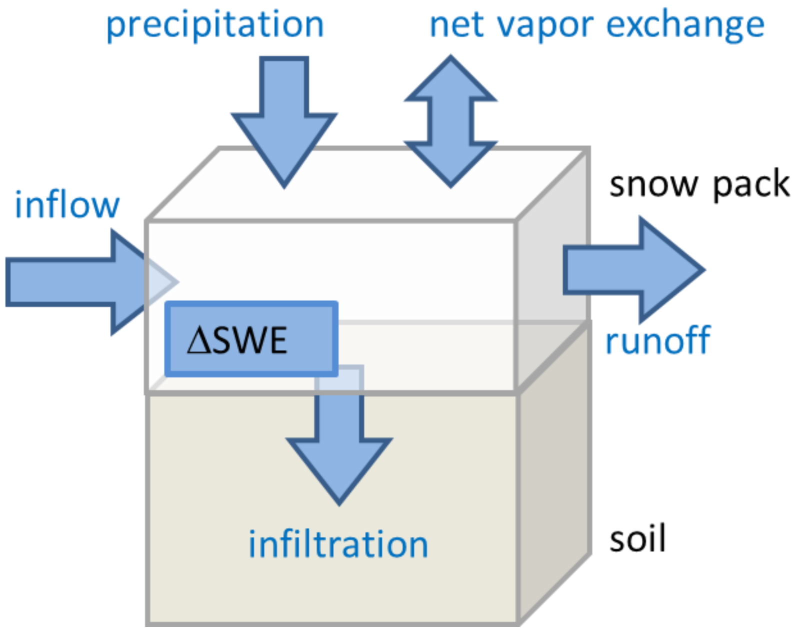
snowpack water balance

Conservation of mass

$$\frac{dSWE}{dt} = \sum inputs - \sum outputs$$
$$= +P \mp E - F + I - R$$

Unit: $\frac{m^3}{m^2 s}$ or $\frac{mm}{d}$ (volume per area and time)

SWE	water equivalent of the snowpack (mm)
P	net precipitation inputs from rainfall, snowfall (mm/d)
E	net vapor exchange between snowpack and environment by sublimation, evaporation, and condensation (mm/d)
F	infiltration to the ground (mm/d)
I	inflow of liquid water to the snowpack (mm/d)
R	runoff of liquid water from the snowpack (mm/d)



Scheme of a snow pack and the main processes leading to a change in the snow water equivalent (SWE)

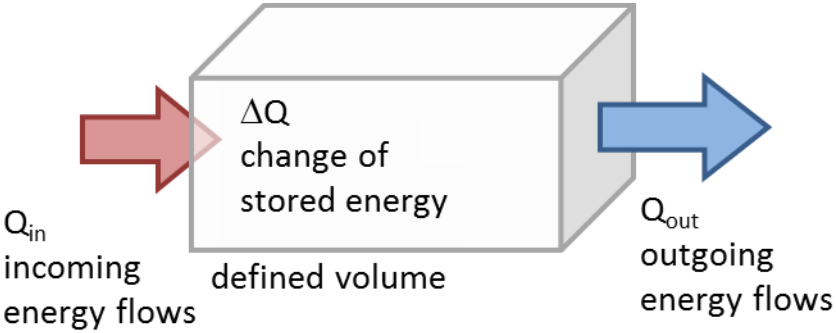
2. Definitions/Concepts

energy balance

Conservation of energy

General speaking, an **energy balance** is the relation describing the change in the amount of energy stored within a defined volume owing to flows of energy across the boundary of the volume.

- The **surface energy balance** is that of an interface or degenerate volume, where the thickness is approached to zero.
- If we consider a volume of snow or ice, a **change in the amount of stored energy** will result in a **change in the temperature or the phase**, or both, of the material in the volume.



$$\boxed{\frac{dQ}{dt} = \sum Q_{in} - \sum Q_{out}}$$

Unit: $\frac{J}{m^2 s}$ or $\frac{W}{m^2}$ (energy per area and time)

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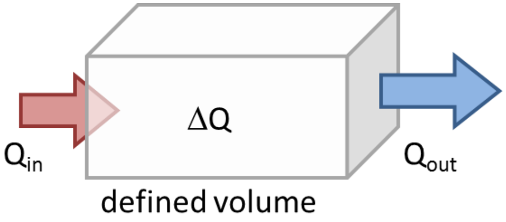
3. Simulation of glacier and snow melt and runoff

*Mainly **two types of models** are used to simulate melt and runoff*

It is difficult to build and maintain networks of stations to **record directly the melt rates**. Therefore, mainly two types of melt models have been developed:

1. **Energy balance equation** for the snow and ice surface to determine the melt rates as a residual of the energy balance. Detailed calculation of all components of the surface energy balance are required, which is a difficult task.

-> Melt is a function of all the components of the energy balance



2. A widely used empirical method is the **temperature index model**, where the air temperature delivers all the information about the energy balance.
   These approaches don't require detailed field measurements of the components of the energy balance.

-> Melt is a function of air temperature

3.1. Energy balance model

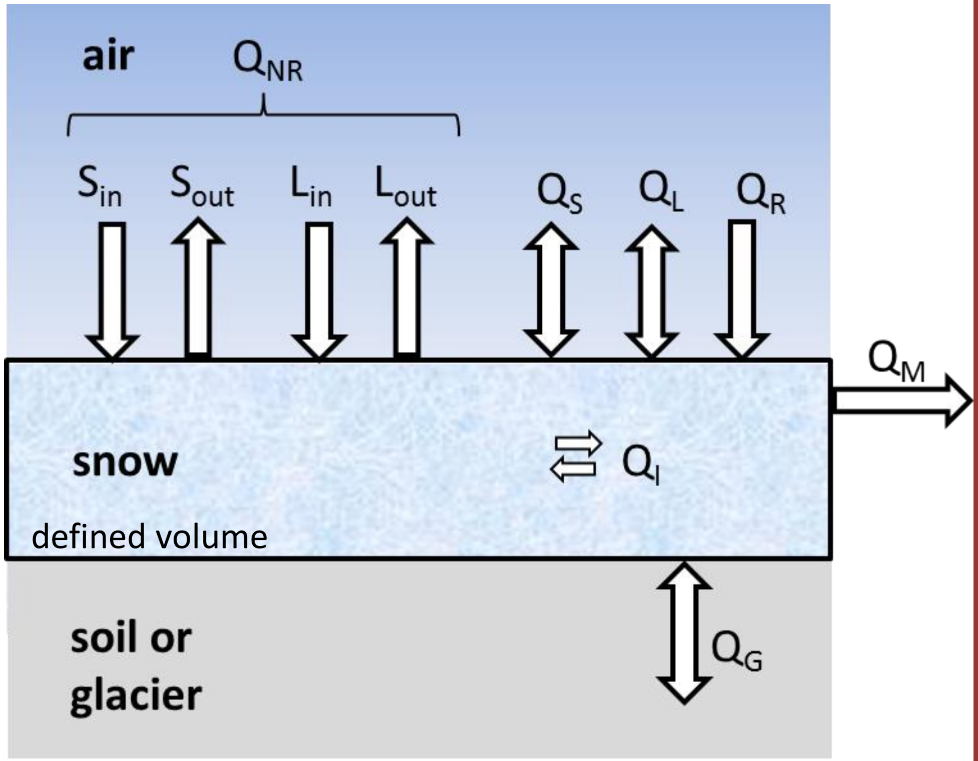
equation

The idea is to determine the melt rates as a residual of the energy balance

Ice and snow melt can be calculated using an energy balance model for the snowpack or the glacier surface.

In this model each of the relevant energy fluxes at the snow or glacier surface is computed from **physically based** calculations using **in-situ measurements** of necessary variables.

The **energy available for melt Q_M** and thus, melt rate is obtained from the energy balance of fluxes at the snow / glacier surface, attempting to quantify melt as residual in the heat balance equation



Schematic of the energy balance for a snowpack

3.1. Energy balance model

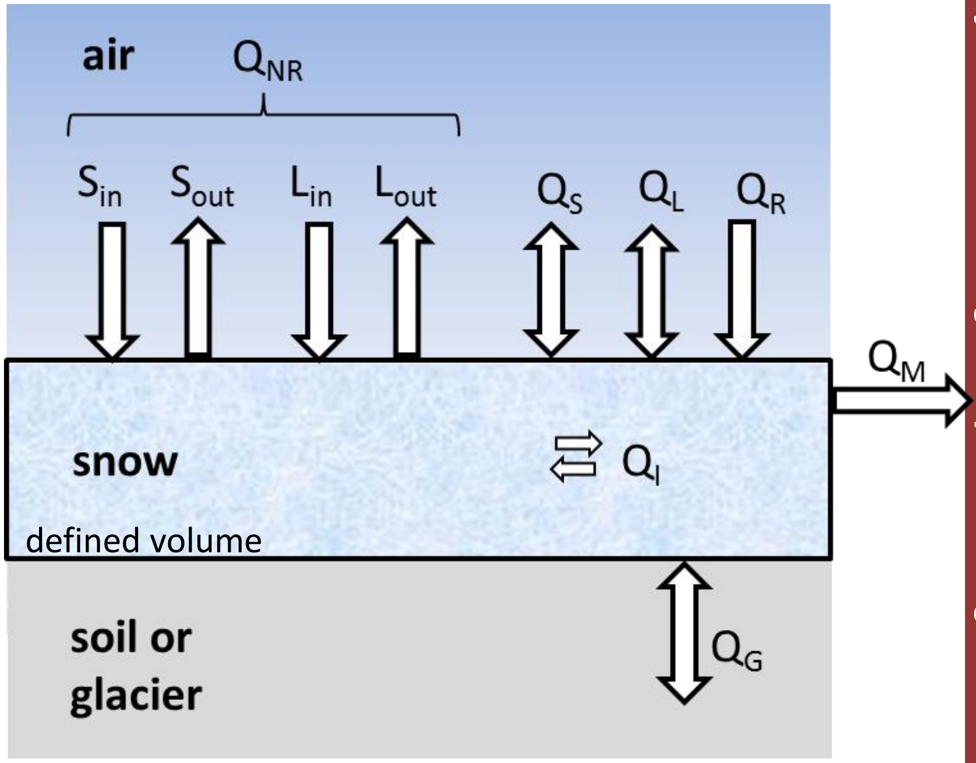
Equation

$$\frac{dU_I}{dt} = \sum Q_{in} - \sum Q_{out}$$

$$Q_I = Q_{NR} + Q_S + Q_L + Q_R + Q_G + Q_M$$

Take care with positive/negative signs!
Which fluxes are positive / negative?

U_I	snowpack internal sensible and latent heat storage
Q_{NR}	net radiant energy exchange
Q_S	sensible heat exchange with the atmosphere
Q_L	latent heat exchange of vaporization and sublimation with the atmosphere
Q_R	heat provided by rain
Q_G	heat from conduction in the ground
Q_I	change in snowpack internal sensible and latent heat storage
Q_M	Loss of latent heat of fusion due to meltwater leaving the snowpack



Schematic of the energy balance for a snowpack

3.1. Energy balance model

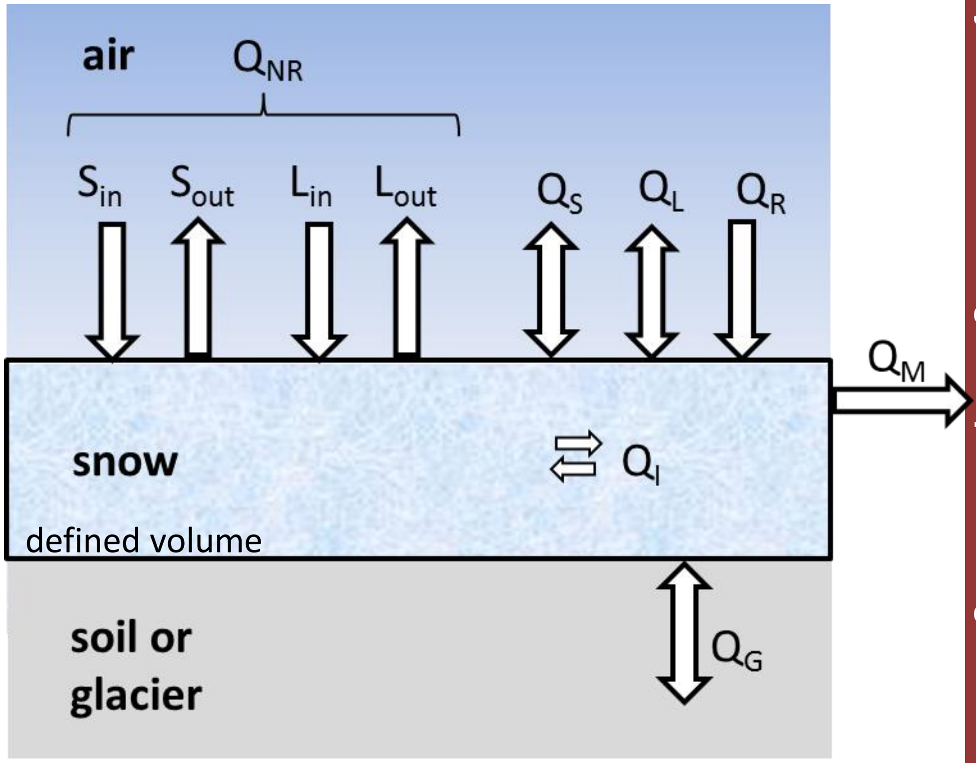
equation

General definitions

$$\frac{dU_I}{dt} = \sum Q_{in} - \sum Q_{out}$$

$$Q_I = Q_{NR} + Q_S + Q_L + Q_R + Q_G + Q_M$$

- An energy flux is a rate of transfer of energy through a surface.
- The SI unit of the energy fluxes is Watts per area (Wm^{-2}), where 1 Watt is 1 Joule per second.
- Energy fluxes directed towards the surface are defined positive.



Schematic of the energy balance for a snowpack

3.1. Energy balance model

equation

We are interested in the energy available for melt

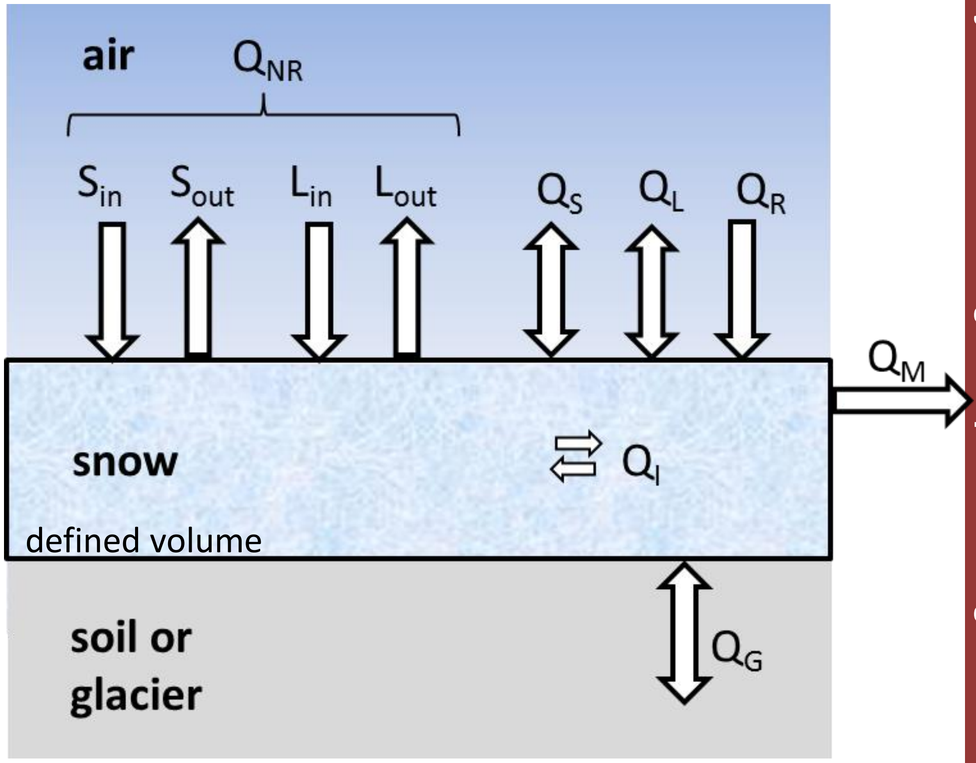
$$Q_I = Q_{NR} + Q_S + Q_L + Q_P + Q_G + Q_M$$

Melt energy can be calculated as a residual if all fluxes are known (computed, measured or neglected)

The **amount of melt** is then calculated based on the melt energy:

$$M = \frac{Q_M}{\rho_w L_f}$$

ρ_w	density of water (kg m^{-3})
L_f	latent heat of melting (334 kJ kg^{-1} at 0°C)
Q_M	heat used for melt (Wm^{-2})
M	Melt of ice or snow (m s^{-1})



Schematic of the energy balance for a snowpack

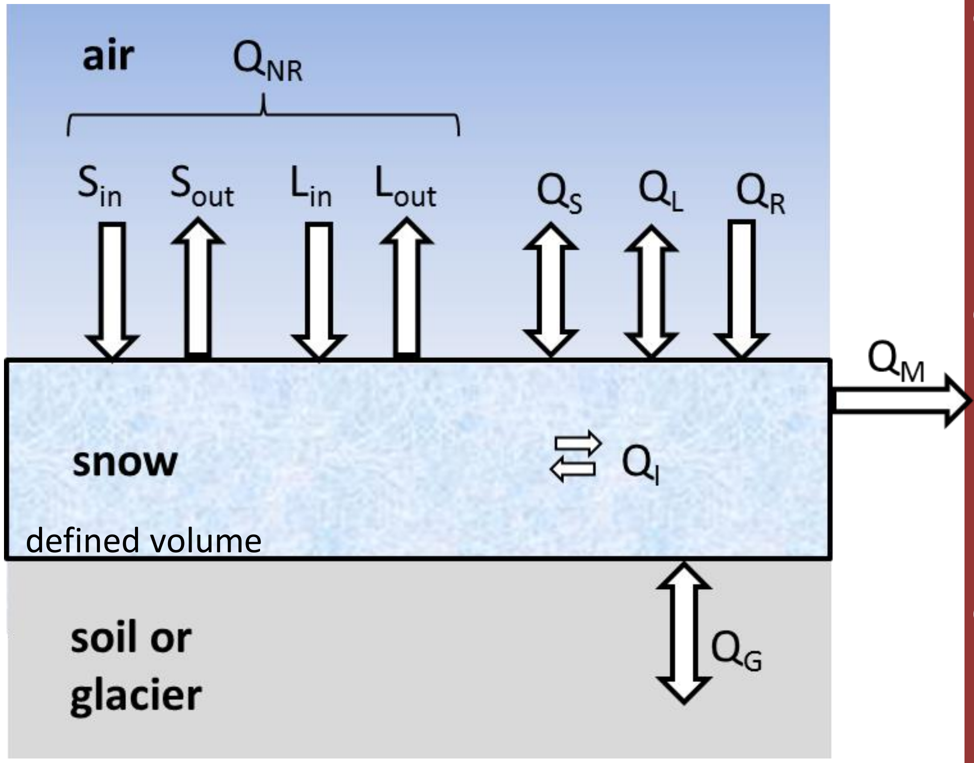
3.1. Energy balance model

equation

Disadvantages of using the energy balance to model melt

$$Q_I = Q_{NR} + Q_S + Q_L + Q_P + Q_G + Q_M$$

- The measurement of all components is challenging
- A parameterization of turbulent fluxes (Q_s and Q_L) is necessary
- Regionalisation is a demanding task



Schematic of the energy balance for a snowpack

3.1. Energy balance model

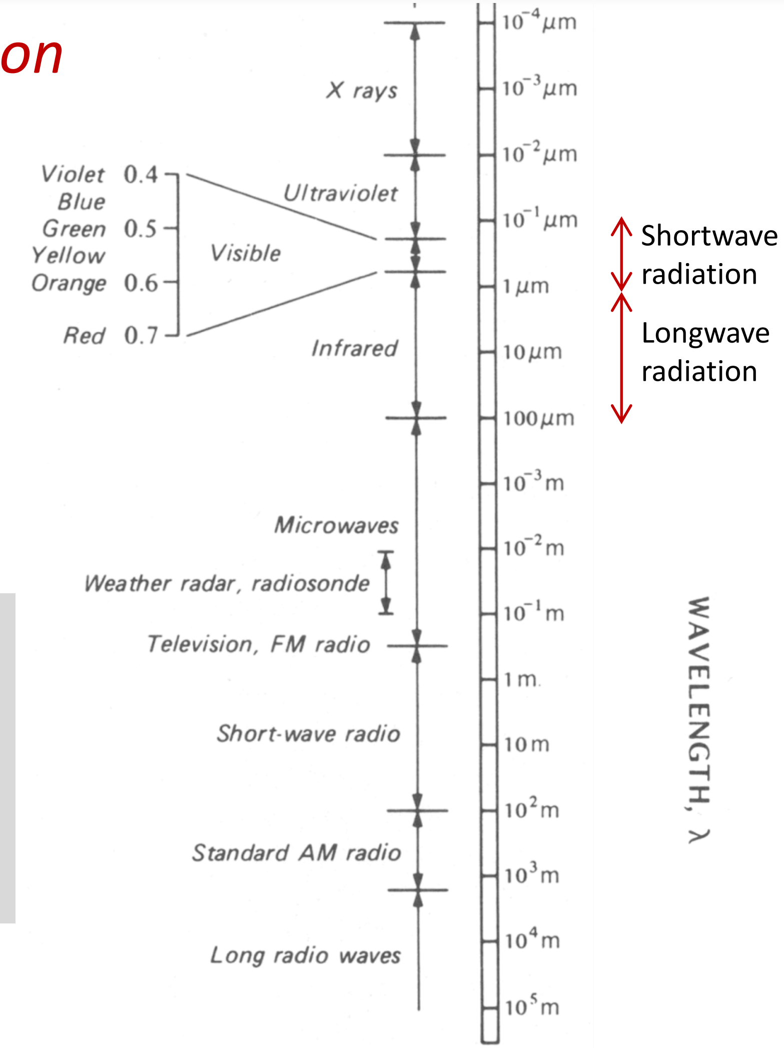
Net radiation

Net radiation Q_{NR} is the sum of net shortwave S and longwave radiation L:

$$\begin{aligned} Q_{NR} &= S_{net} + L_{net} \\ &= (S_{in} - S_{out}) + (L_{in} - L_{out}) \end{aligned}$$

S_{net}	net incoming shortwave radiation
L_{net}	net incoming longwave radiation
S_{in}	incoming shortwave radiation
S_{out}	reflected shortwave radiation
L_{in}	incoming longwave radiation
L_{out}	outgoing longwave radiation

Shortwave radiation:	Wavelenth of 0.15 - 2 µm
Longwave radiation:	Wavelenth of 2 - 100 µm



The electromagnetic spectrum (Oke, 1987)

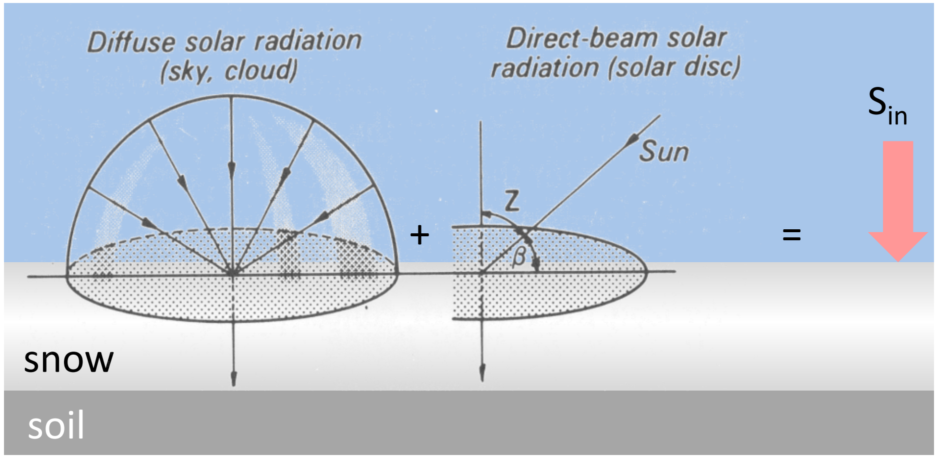
3.1. Energy balance model

shortwave radiation

Incoming shortwave radiation

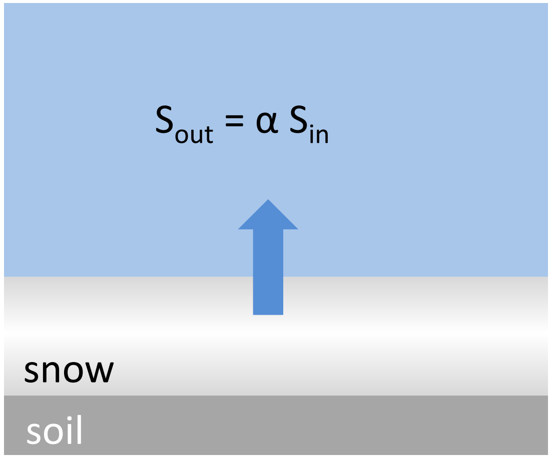
- Incoming solar radiation is the radiant energy flux arriving to the earth surface (the surface not necessarily being horizontal)
- Shortwave radiation is split in **direct and diffuse radiation** (see figure)
- It depends mainly on the time (hour, season), latitude, topography and cloud cover

modified after Monteith (1973)



Reflected shortwave radiation

S_{in}	incoming shortwave radiation
S_{out}	reflected shortwave radiation
α	albedo



$S_{out} = \alpha S_{in}$

3.1. Energy balance model

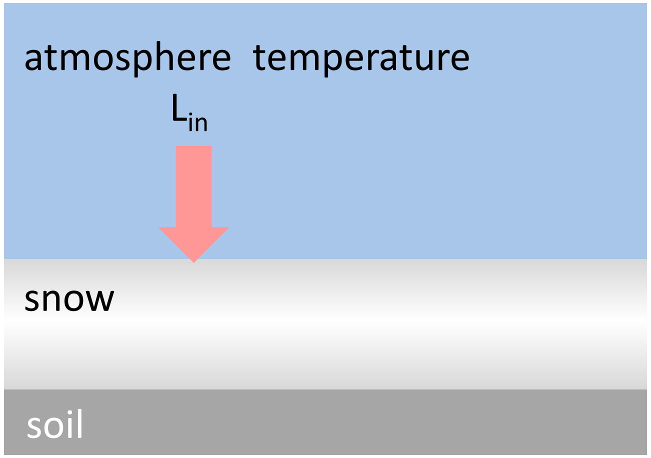
longwave radiation

Longwave radiation is calculated using **Stefan-Boltzmann's law**

Incoming longwave radiation

Emitted from atmospheric components (vapor, aerosols, clouds) as a function of their temperature

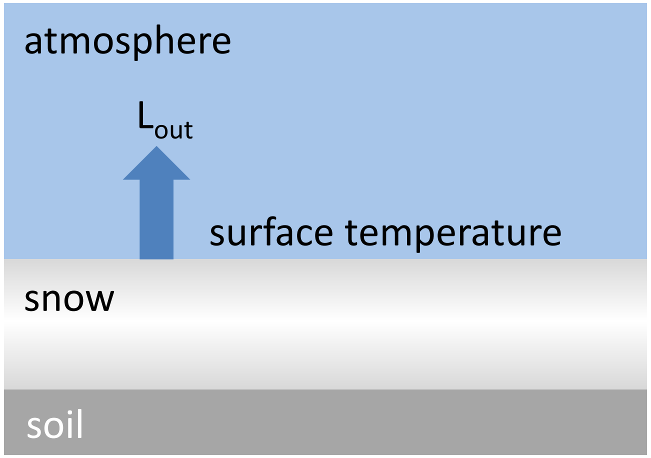
$$\boxed{L_{in} = \epsilon_a \sigma {T_a}^4}$$



Outgoing longwave radiation

Emitted from the glacier or snow surface as a function of its temperature and the properties of the surface

$$\boxed{L_{out} = \varepsilon_e \sigma {T_e}^4}$$

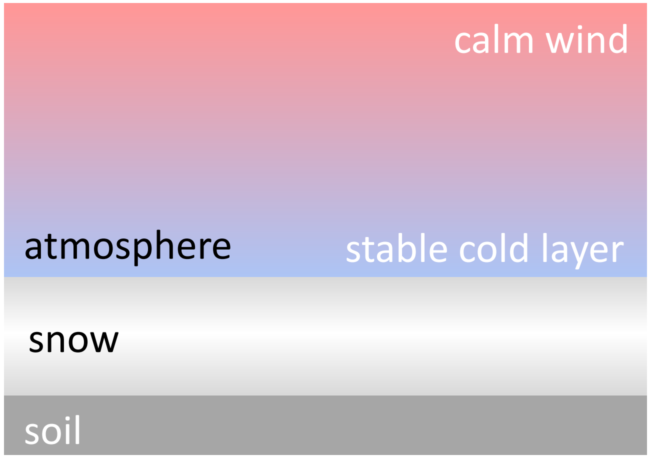


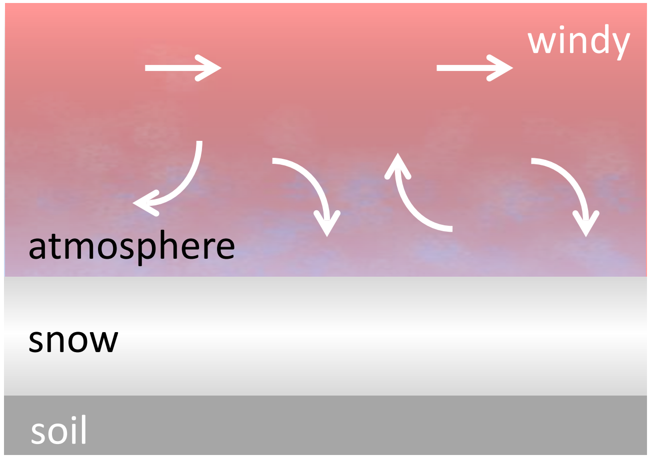
$T_{a,e}$	temperatures of the lower atmosphere (a) and of the snow or ice surface (e) (K)
$\varepsilon_{a,e}$	emissivity of the atmosphere (a) on the snow or ice surface (e) (-)
σ	Stefan-Boltzmann constant

3.1. Energy balance model

turbulent heat fluxes

Sensible and latent heat fluxes are both turbulent heat fluxes





- in general, the atmospheric boundary layer over snow and ice is stably stratified because the underlying surface is colder than the air
- without wind, the stable layer does not break up and turbulent fluxes are minimal

- winds create turbulence
- cooled air near the surface mixes with warmer air above
- warm air and water vapor can be brought to surface

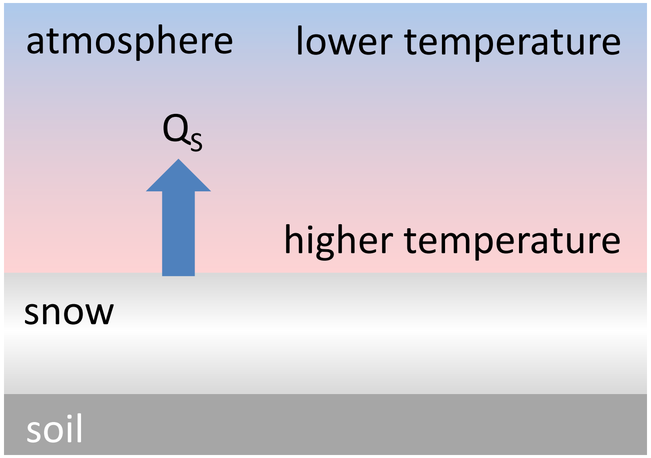
3.1. Energy balance model

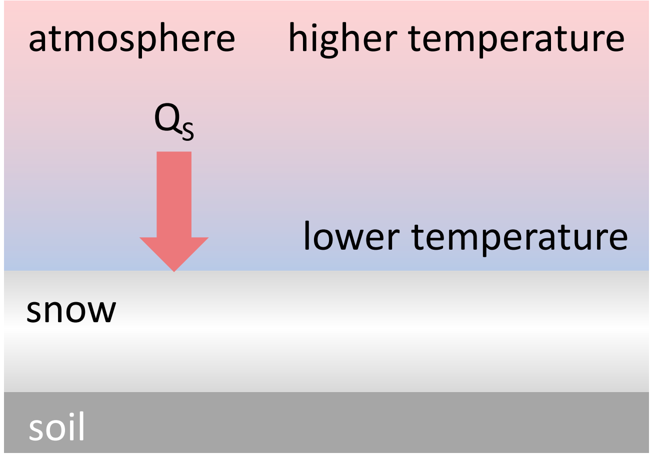
sensible heat flux

Sensible heat flux

Heat energy transferred between the surface and air mass when there is a **difference in temperature** between them.

It depends on: temperature difference between atmosphere and snowpack surface, wind speed, surface roughness and the stability of the air





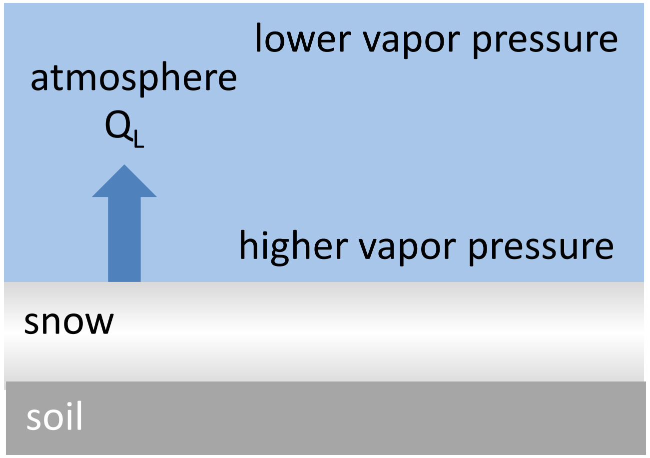
- wind strong enough to induce turbulent fluxes is required (see previous slide)

3.1. Energy balance model

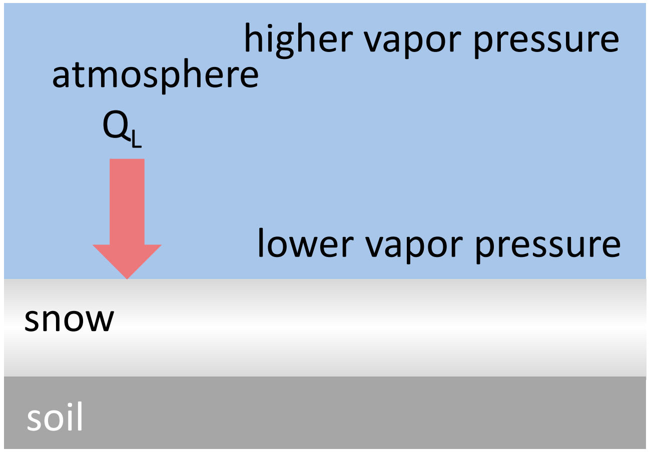
latent heat flux

Latent heat flux

The exchange of heat between the surface and air mass due to the change of phase of the water contained in the two media (transfer of vapor from one to the other) when there is a **difference in water vapor.**



- moisture from snow is diffused to the atmosphere above (sublimation)
- latent heat is lost from the snow
- snow stays cool even if air temperatures are rather warm



- deposition of moisture from atmosphere to snow surface
- latent heat is gained by the snow
- surface will warm
- warming may start the melting process
- wind strong enough to induce turbulent fluxes is required

3.1. Energy balance model heat provided by rain

Heat provided by rain

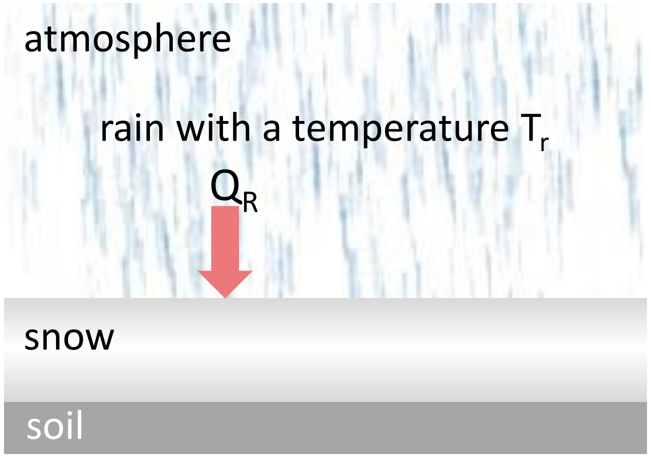
Rainfall can influence the energy budget by

- **sensible heat additions** due to the heat added by a volume of relatively warm rain

additionally, if the snowpack is below 0°C:

- **release of the latent heat** if rainfall freezes on a sub-zero snowpack

For small rain intensities, the sensible heat additions due to heat added by rain drops can often be neglected.


atmosphere

rain with a temperature T_r

Q_R

snow

soil

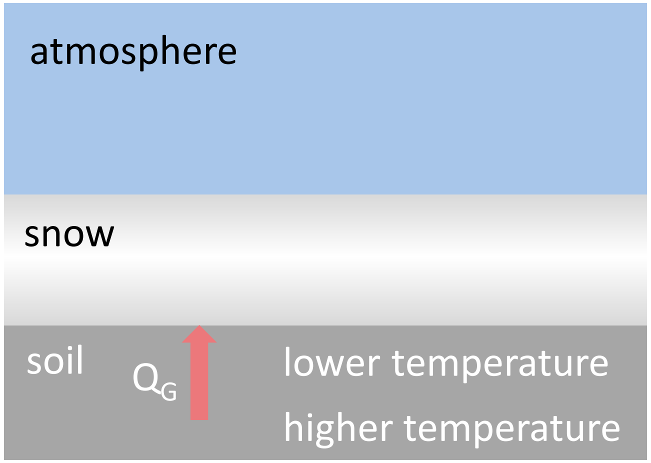
3.1. Energy balance model

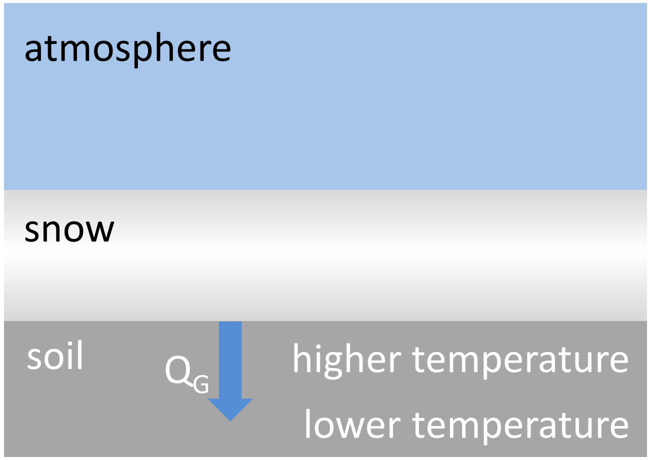
ground heat conduction

Ground heat conduction

- The conduction depends on the **thermal conductivity** of the soil and the **temperature gradient** within the ground
- Ground heat conduction represents generally a very minor energy source for melt.
- Reasons: Soil is generally a poor conductor of heat and temperatures in the soil are often low

-> The ground heat can often be neglected





Permafrost situation

3.1. Energy balance model

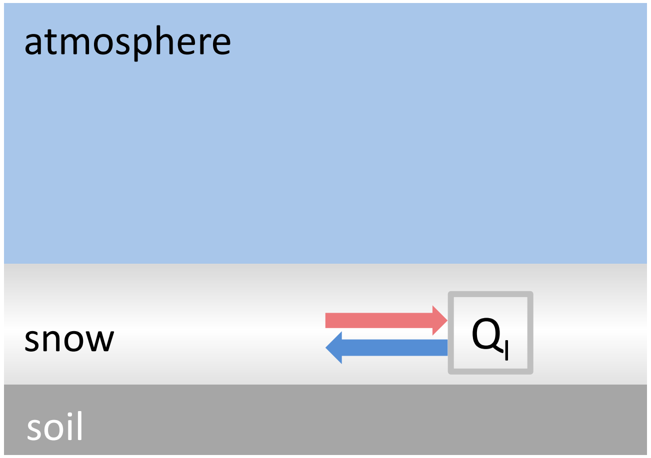
Internal energy

Internal sensible heat changes are mainly represented by

- **changes in snowpack temperature and mass over time** integrated over the depth of the snowpack

Once the snowpack is isothermal at 0°C and melt begins, the importance of this energy component diminishes
-> for a melting surface this component is often neglected

"cold content" is the energy that needs to be added to yield melt of a snowpack



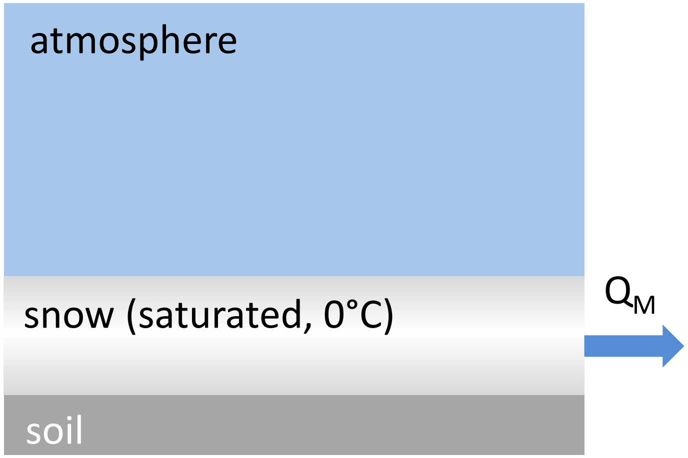
3.1. Energy balance model loss of latent heat due to melt

Loss of latent heat due to melt

Melt energy Q_M is the loss of latent heat of fusion when liquid water drains from the snowpack.

The conditions that meltwater drains from the snowpack:

- energy is must be first used to warm the snowpack to an **isothermal 0°C condition**
- snowpack has a certain **liquid-water holding capacity**
  the liquid water content is often assumed to be 10%
  (e.g. 5 mm water within a snowpack of 50 mm SWE of a horizontal snowpack)



Contents

3.2. Temperature-index model

introduction

The main questions in the following slides:

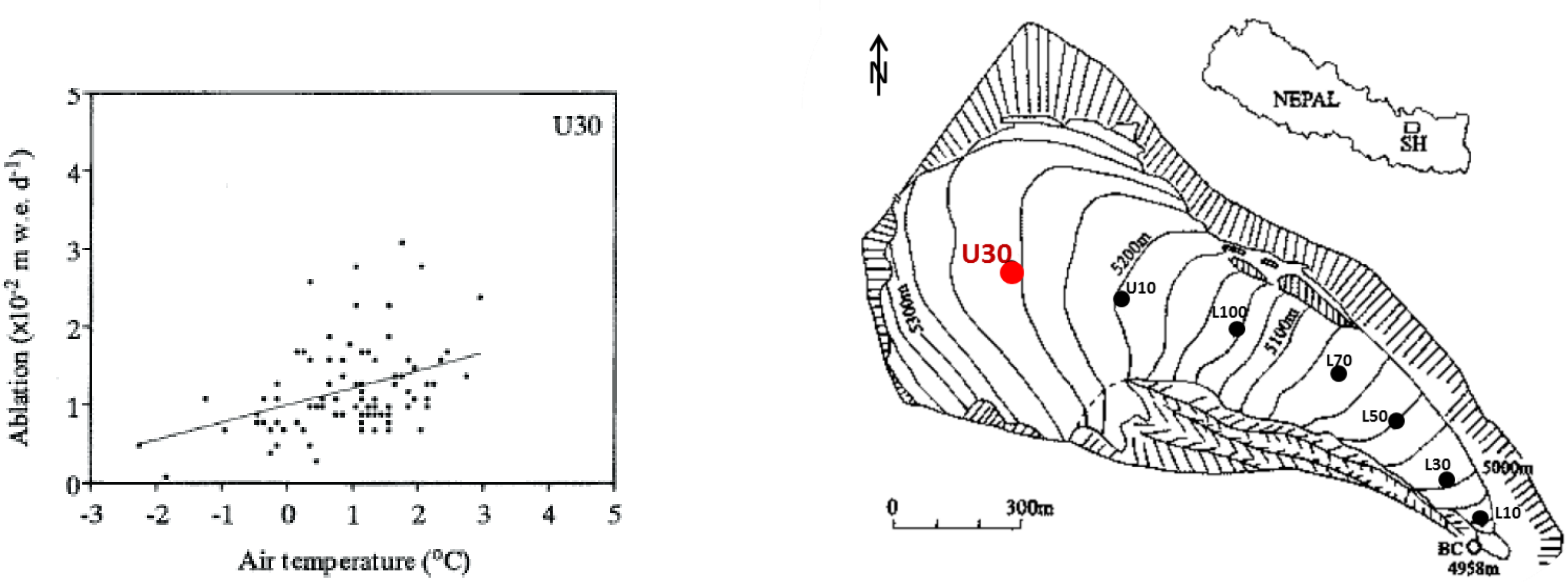
- What is a temperature-index model?
- How to compute melt / runoff using a DDF
- Why is the temperature-index model working despite of its simplicity?
- Advantages / limitations
- Examples of DDFs for clean ice, debris and dust covered ice surfaces
- Improved temperature-index models
- Applications

3.2. Temperature-index model

definition

What is a temperature-index model?

Temperature-index melt models - also called degree-day models – are based on the assumption of an **empirical relationship between melt and air temperature** based on a strong and frequently observed correlation between these quantities.



Daily ablation rate versus daily mean air temperature on Glacier AX010, Nepalese Himalayas from June to August 1978. The thin line represents the regression line (Kayastha, 2000)

3.2. Temperature-index model

computation

IHCAP – Indian Himalayas Climate Change Adaptation Programme: Level-1 course

Computing melt using a DDF

The main assumption of the degree-day approach is that there is no melt below a **threshold temperature**, while above the threshold, melt is correlated to air temperature assuming a **linear relationship** between **ablation and positive temperature sums**.

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

M	amount of snow (or ice) melt (mm d^{-1})
DDF	Degree-day factor (mm d^{-1}°C^{-1})
T_d	average daily air temperature (°C)
T_0	base temperature at which snow melt occurs (°C), usually taken as 0°C

3.2. Temperature-index model

computation

Computing melt using a DDF

The amount of melt expected in several time steps (assuming that T_0 is 0°C) is expressed by:

$$\boxed{\sum_{i=1}^{n} M = DDF \sum_{i=1}^{n} T^{+}\,\Delta t}$$

M	amount of snow (or ice) melt (mm)
DDF	Degree-day factor (mm $d^{-1}°C^{-1}$)
T^{+}	positive air temperatures of each time interval ($°C^{-1}d^{-1}$)
n, Δt	amount n of time intervals Δt

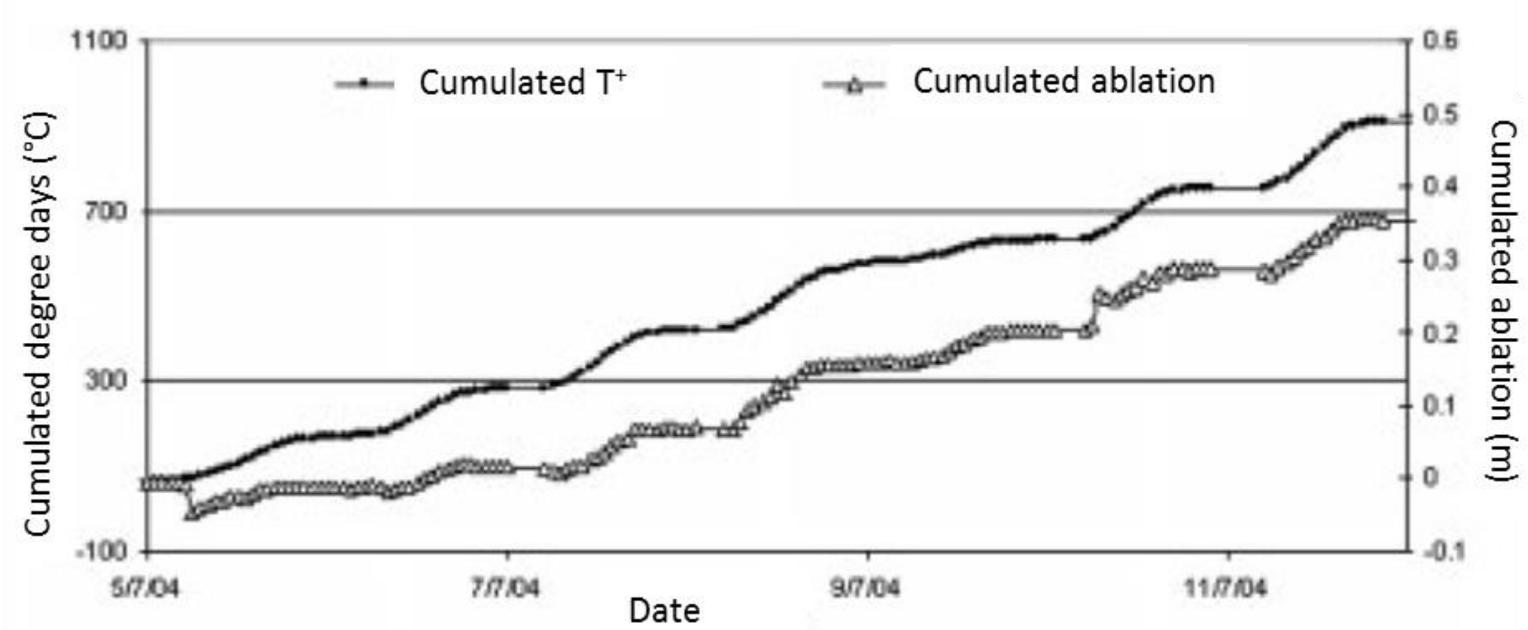
Commonly, a daily **time interval** is used for temperature integration, although any other time interval (hourly or monthly) can also be used for determining degree-day factors.

The threshold temperature is based on the concept that significant melt occurs only above a certain measured air temperature.

3.2. Temperature-index model why is it working?

Temperature-index model is an extremely strong parameterisation of the energy balance. Why is this method working?

First of all: Air temperature and melt show a strong and frequent correlation



Comparison between the cumulated hourly positive degrees and the cumulated ablation data measured by an automatic ablatometer on Baltoro Glacier (at an elevation of 4178 m a.s.l.). The debris covering the ice was 4 cm. (Mihalcea et al., 2006).

3.2. Temperature-index model why is it working?

What is air temperature? Which are the mechanisms causing temperature variations that would lead to variations in melt using the temperature-index model?

Air temperature is a physical measure of the thermal condition of air which are generated by various components
(advection, convection, mixing, radiative processes, turnover of latent heat in melting, condensation and evaporation)

(a)

Variations of air temperature are caused by two mechanisms:

a) **Variations in the heat balance conditions**
as determined by the heat balance equation of the underlying ground surface

b) **Variations in the advection**
of air masses of different thermal conditions

(b)

*Observed temperature by a sensor mounted 2.00 m above ground level, **ventilated** and shielded from incoming and reflected solar radiation (a) on a Glacier in the Swiss Alps and (b) in a watershed in the Chilean Andes*

3.2. Temperature-index model why is it working?

For which components of the energy balance temperature is decisive and under which conditions?

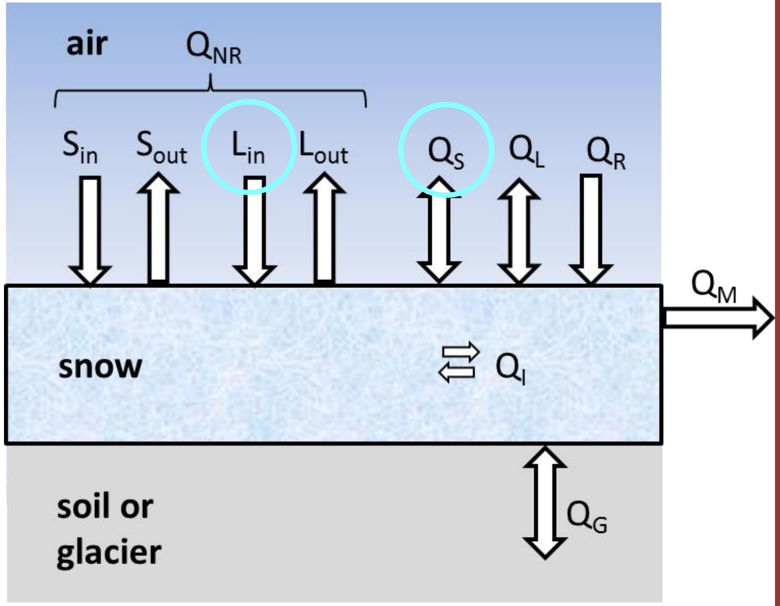
Air temperature is a decisive measure of

- sensible heat flux

$$Q_S = f(T)$$

- to some extent of the incoming longwave radiation

$$L_{in} \cong f(T)$$



Taking a constant degree-day factor (DDF) is based on the assumption of an in-time **constant relative contribution to melt** of every component of the heat balance.

-> changes in weather types, climatic conditions or physical conditions of the underlying land surface are **affecting the relative contribution** of the heat balance components and cause therefore variation in the DDF.

3.2. Temperature-index model why is it working?

How are the correlation of air temperature to melt and the DDF for air temperature measurements of different sites?

DDF as obtained for the snow-free part of Aletsch-Glacier during the melt period from 31 July to 27 August 1965, using the air temperature from various climatological stations with different site characteristics. (Lang, 1990)

Air Temperature Observation Site			Correlation Coefficient	DDF Melt Factor (cm °C^{-1} day^{-1})
Location	Altitude (m a.s.l.)	Distance (km)		
Mountain	3576	15	0.64	3.70
Slope	2220	1	0.75	4.18
Glacier	2200	0	0.77	7.25
Valley	549	60	0.82	5.11

Interesting:

- The air temperature at the **valley station**, although at a maximum distance, provides the best information on the variation of daily melt rates, as indicated by the correlation coefficients
- considerable **differences in the DDF**

why?

3.2. Temperature-index model why is it working?

Why is the correlation of air temp. and melt higher for the valley station?

Important: the air temperature measurement gives information about the thermal conditions of the air at **a particular point** and influenced by **two components:**

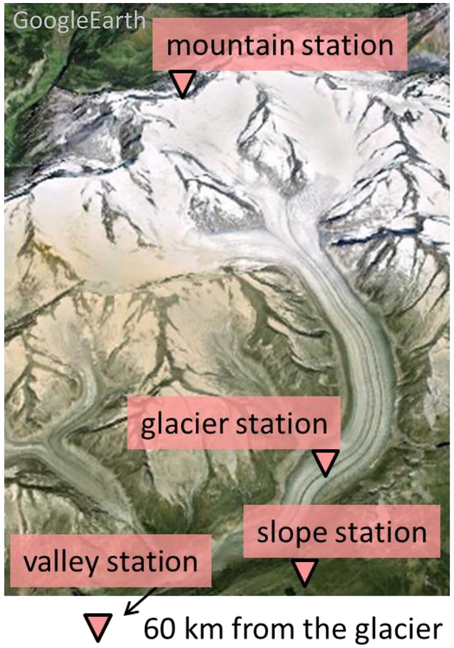
a) **heat balance at the ground surface close to the observational site**
b) **advective component**

The melting processes on the glacier are generally dominated by the radiative energy fluxes. The valley station and glacier station have a similar climatology. However, air temperature at the valley station is much more influenced by net radiation compared to the well exposed mountain station where advection and «free atmosphere» conditions are dominant.

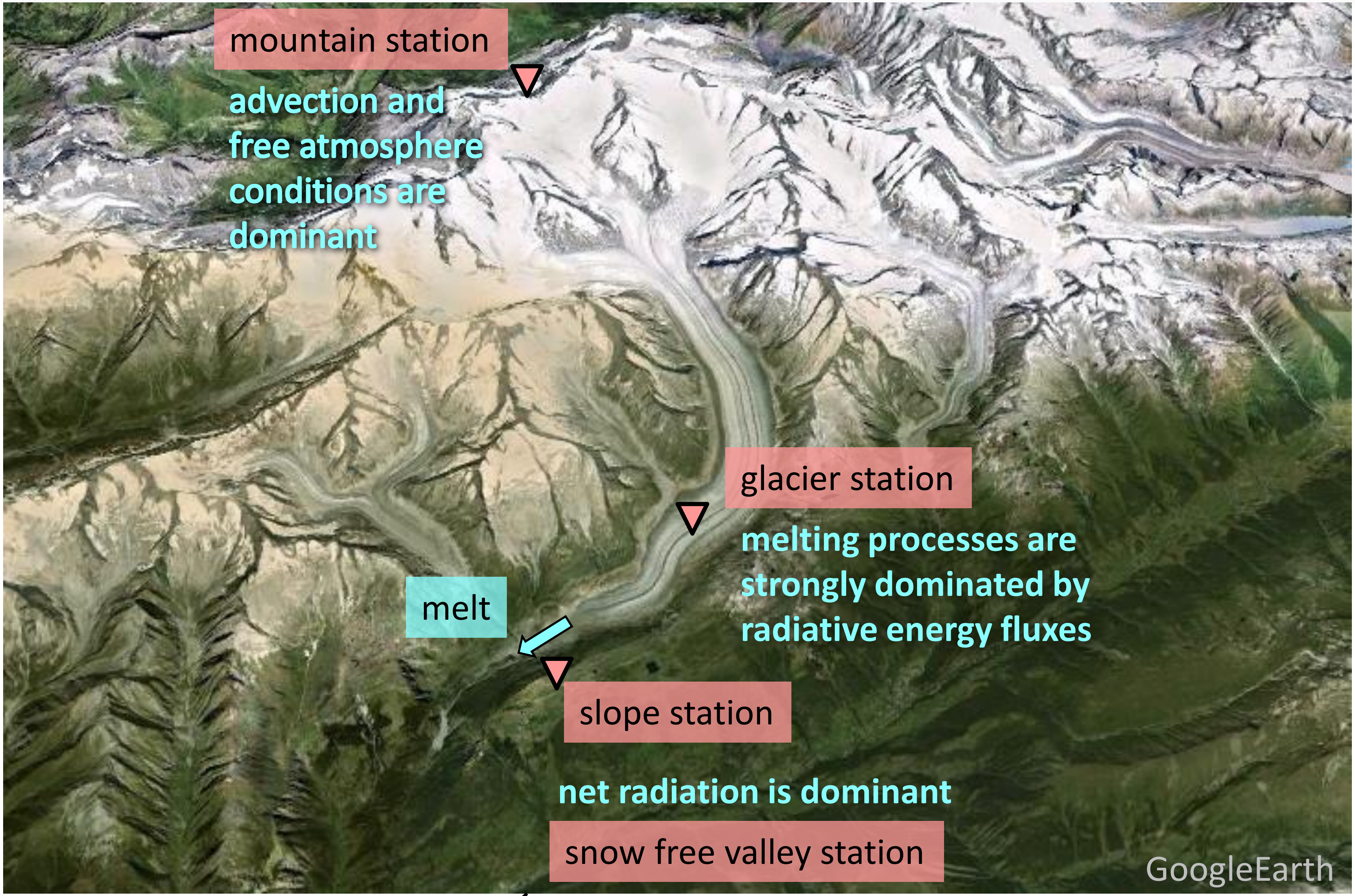
This may lead to the fact that:

i. Highest information content
   T observed at the valley station, 60 km away from the glacier

ii. Lowest information content
   T observed at the mountain station



3.2. Temperature-index model why is it working?



60 km from the glacier

3.2. Temperature-index model main advantages

Which are the main advantages of the temperature-index model?

Temperature-index methods were applied first by Finsterwalder and Schunk (1887) and are still widely used by hydrologists around the world

The main advantages:

- wide availability of air temperature data
- relatively easy interpolation, extrapolation and forecasting possibilities of air temperature
- generally good model performance despite their simplicity
- computational simplicity
- robust method

3.2. Temperature-index model

limitations

Which are the main limitations of the temperature-index model?

The temperature-index model is subject to two main limitations:

- The **DDF depends on station data** (both, discharge and temperature data needed). With every change in position of the station, the DDF has to be **re-calibrated**. A calibration is also needed if the DDF is applied to another region.

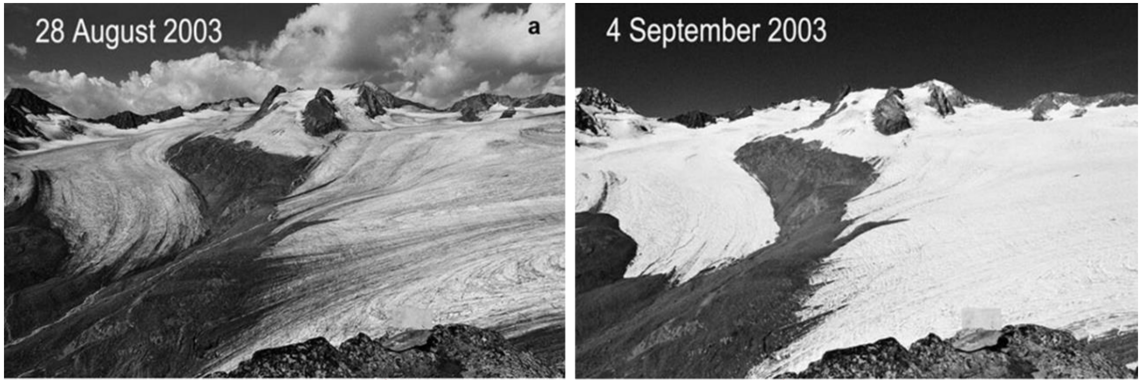
- The dependency of melt / runoff on temperature is **not necessarily constant in the future**. For projections of e.g. water availability in the future, the assumption of a temporal constant DDF may lead to large uncertainties.

3.2. Temperature-index model

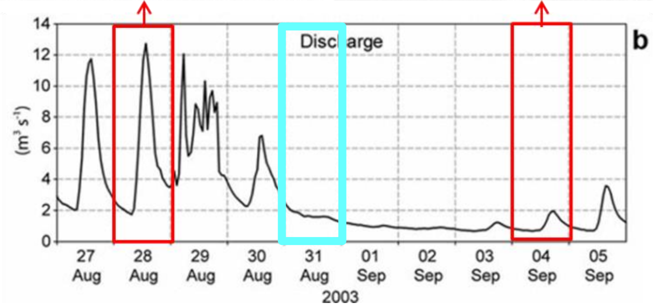
example: ice

Which is the difference in DDF for snow and ice?

Degree-day factors for snow tend to be considerably **lower than those for ice**, due to generally lower albedo (and thus higher net shortwave radiation) of ice.
The example of Vernagtbach glacier shows that **melt decreases after the snowfall** event.



snowfall event of
14 mm SWE
on 31 August



(a) Photographs of the western part of the Vernagtach basin (28 August and 4 September 2003).
(b) Hydrograph of the Vernagtbach stream for the period 27 Aug-6 Sep 2003. (Escher-Vetter & Siebers, 2007)

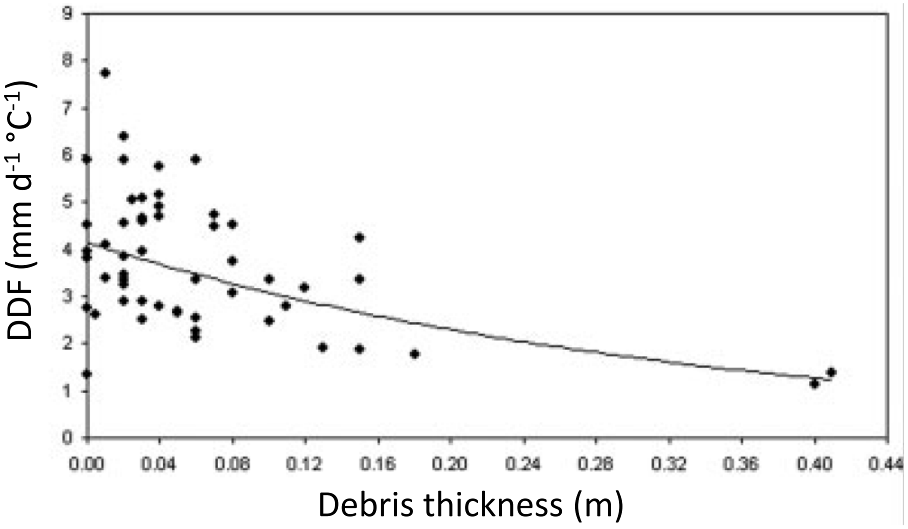
3.2. Temperature-index model example: debris

How does debris influence melt?

Debris has a strong influence on the surface energy balance and thus melting of the underlying ice.

This barrier to heat transfer causes the ablation rate to decrease with increasing debris thickness once a **critical thickness** (of some few cm) is exceeded.

Photograph of the debris covered Baltoro glacier (Concordia area) taken in summer 2004 by C. Mayer (Mihalcea et al., 2006)



Variation of calculated DDF vs. debris thickness on Baltoro glacier for 56 installed ablation stakes in June-July 2004 (Mihalcea et al., 2006)

3.2. Temperature-index model

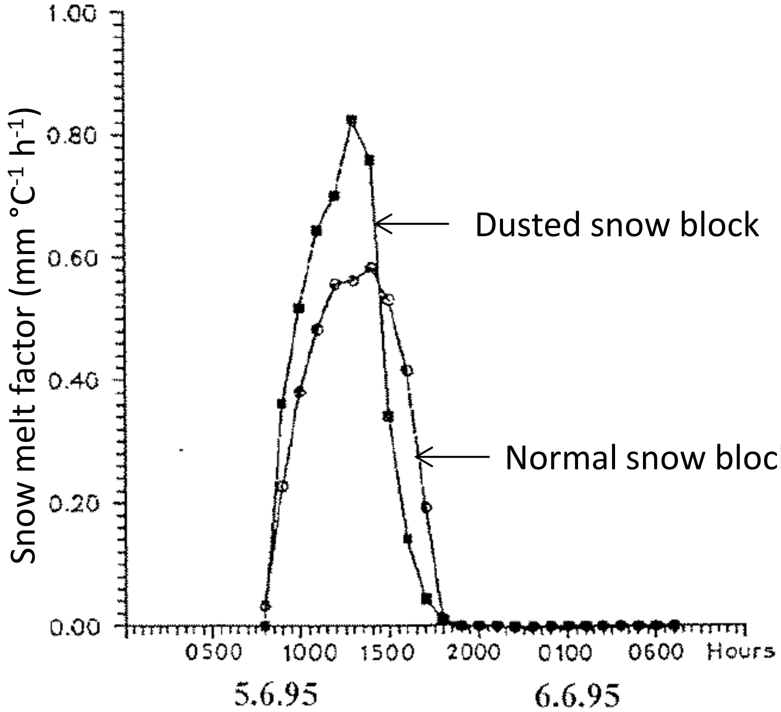
example: dust

How does dust influence melt?

The focus of the study of Singh & Kumar (1996): Influence of a 2 mm dust layer on DDF on a glacier in Garhwal Himalayas

They could show that **runoff from snow with uniformly spread black dust is higher** than from snow with a clean surface.

Dusting or blackening of the snow surface by dark material **reduces the albedo**.
The **increased absorption of shortwave solar radiation** leads to accelerated melt rate.



Diurnal variation in snowmelt factor for normal and dusted snow blocks (observed on 5.6.95/6.6.95) (Singh and Kumar, 1996)

3.2. Temperature-index model values for DDF

Some values for the DDF for snow on Himalayan glaciers

Glacier	• ablation measurements • temperature measurements	DDF_{snow} ($mm\ d^{-1}°C^{-1}$)	
Ürümqi No.1	• unknown • on the ablation zone	3.1	Liu et al. (1996)
Quiongtailan (Tien Shan)	• ablation measured using stakes • on the glacier	3.4	Zhang et al. (2006)
Baishuihe No.1 (Hengduan mountains)	• ablation measured using stakes • on the glacier	5.9	Zhang et al. (2006)
Dokriani (Garhwal Himalayas)	• runoff from snow blocks over a glacier • over snow covered glacier	5.7 - 5.9	Singh and Kumar (1996)
Glacier AX010 (Shorong Himal)	• ablation is calculated using a mass-balance model • on the ablation zone	7.3 - 11.6	Kayastha et al. (2000)

Why are the DDF values for snow different among the sites?

3.2. Temperature-index model

improved temperature-index models

Which modifications of temperature-index models have been proposed?

Aim of modifications: capturing more accurately seasonal and diurnal variations in degree-day factors
The main strategies to improve the temperature-index models are different for snow cover and glaciers and need to be discussed separately:

snow cover

In order to capture well so called "advective melt situations", it was proposed to complement the DDF with turbulent heat fluxes, playing an important role in melt when temperature and water vapor content are high.

glacier

The most common addition to temperature-index-type models has been the **incorporation of measured short-wave radiation or net radiation**.
Further index variables like vapor pressure, sunshine duration, wind speed etc. can be included in addition to air temperature and radiation.

3.2. Temperature-index model

improved temperature-index models

Hock (1999) proposed the extension of the temperature-index approach under consideration of the daily potential direct radiation variations.

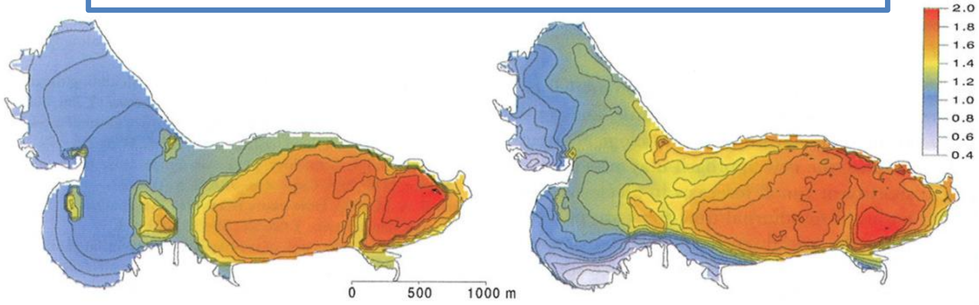
Classical degree-day factor

$$M = \begin{cases} \frac{1}{n} DDF(T_d - T_0), & T_d > T_0 \\ 0, & T_d \le T_0 \end{cases}$$

Temperature index including potential clear-sky direct solar radiation (Hock, 1999)

$$M = \begin{cases} \left(\frac{1}{n} MF + a\, I\right) T, & T_d > T_0 \\ 0, & T_d \le T_0 \end{cases}$$

MF	melt factor (mm d^{-1}°C^{-1})
a	radiation coefficient different for snow and ice surfaces (-)
I	potential clear-sky direct solar radiation at the ice or snow surface (W m^{-2})
T	air temperature (°C)
n	number of time steps per day, e.g. n=24 with a time step of 1 h



Classical degree-day factor

including potential clear sky direct solar radiation

Simulated cumulative areal meltwater equivalent (m) 5 July - 25 August on Storglaciären (Hock, 1999)

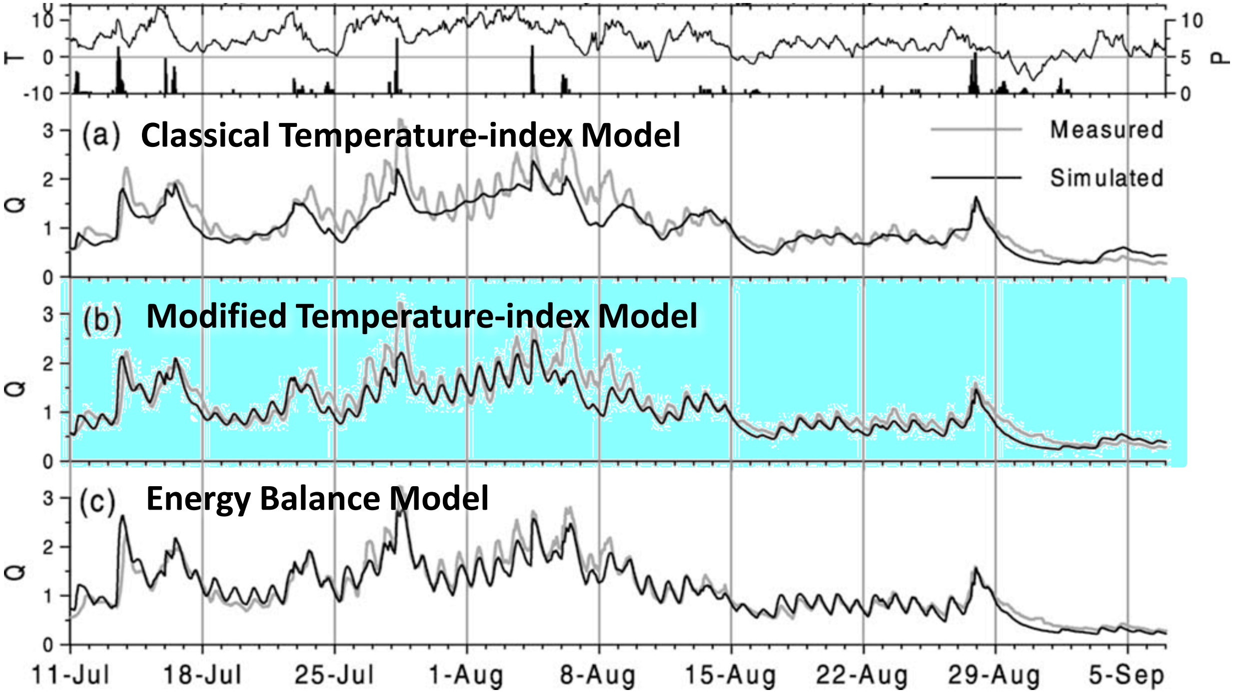
3.2. Temperature-index model

improved temperature-index models

Hock (1999) included a radiation index

Not included:
cloud cover, albedo

Included:
Shading of the surrounding mountains



Hourly data of air temperature, T (°C), precipitation P (mm h^{-1}), simulated and measured hourly discharge, Q (m^3s^{-1}) of Storglaciären, Sweden, from July 11 to September 6, 1994. (Hock, 1999)

Improvements of the Hock (1999) model compared to the classical temperature-index model:
- **improved modelled melt on a glacier**
- **improved diurnal runoff from the glacier** (see Figure)

3.2. Temperature-index model

application

What is the application of temperature-index models?

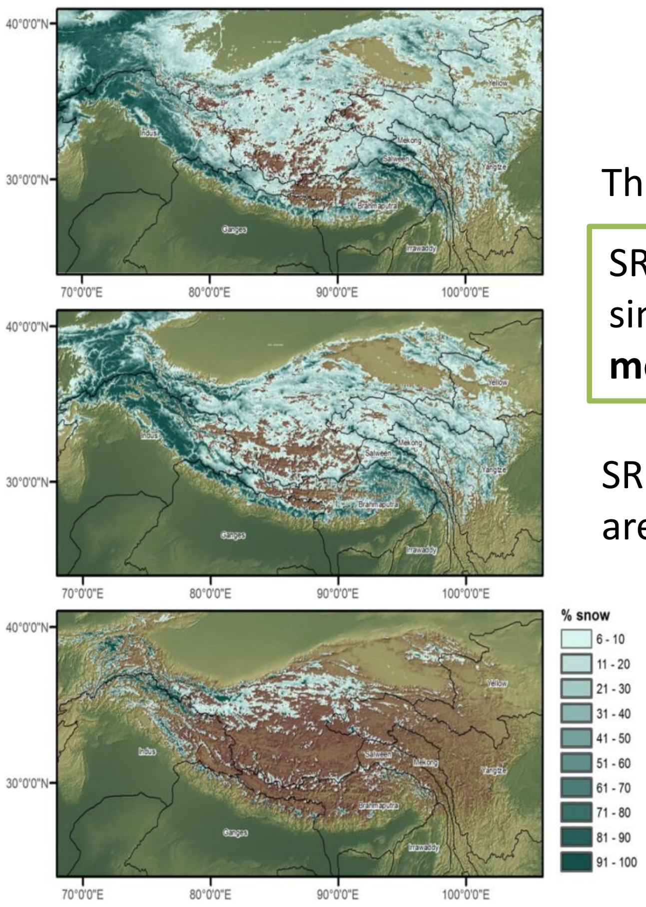
Applications cover a wide range - including:

- operational runoff modelling (e.g. HBV-, SRM-, UBC-, HYMET-model)
- flood forecasting
- water balance assessment
- glacier mass balance modelling
- assessment of the response of snow and ice to predicted climate change

flash flood in the Swiss Alps

3.2. Temperature-index model

SRM



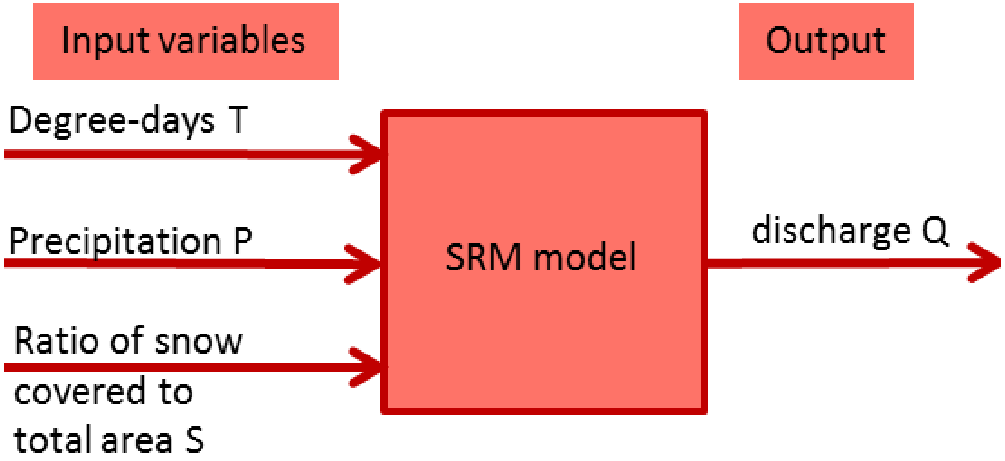
Seasonal snow cover extent (winter, spring, summer) based on MODIS from March 2000 to February 2008. (Immerzeel et al., 2009)

An example for a model where the DDF is applied: Snowmelt-Runoff Model SRM

The model was originally developed by Martinec (1975)

SRM is a conceptual model **based on degree-days** and used to simulate **daily runoff resulting from snowmelt and rainfall in mountainous regions**.

SRM requires daily air temperature, precipitation and snow covered area values as input parameters



Input variables

Output

Degree-days T

Precipitation P

Ratio of snow covered to total area S

SRM model

discharge Q

3.2. Temperature-index model

SRM

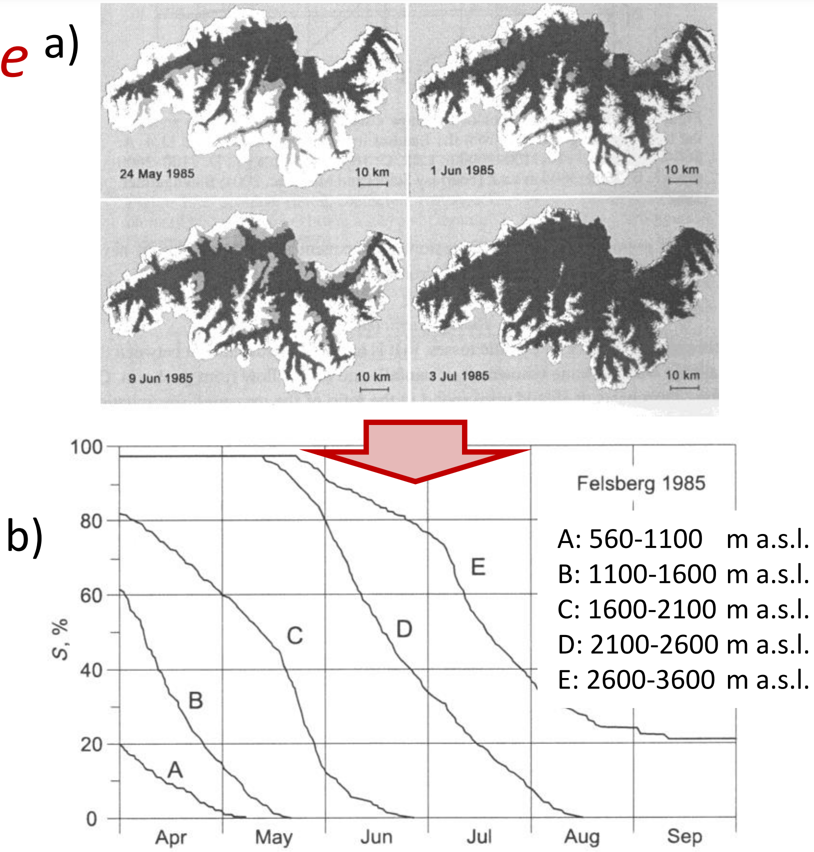
Depletion curves of the snow coverage can be interpolated from periodical snow-cover mapping.

Basic idea:
The areal extent of the seasonal snow cover generally decreases during the snowmelt season in mountainous basins.

Depletion curves describe the snow-covered area S and can be interpolated from periodical snow-cover mapping

Mapping by terrestrial observation, aircraft photography and most efficiently by remote sensing

Daily values of snow-covered area can then be interpolated and used as an important input variable to SRM.



a) Sequence of snow-cover maps from Landsat b) Depletion curves of the snow coverage for five elevation zones of the Rhine basin, derived from Landsat imagery

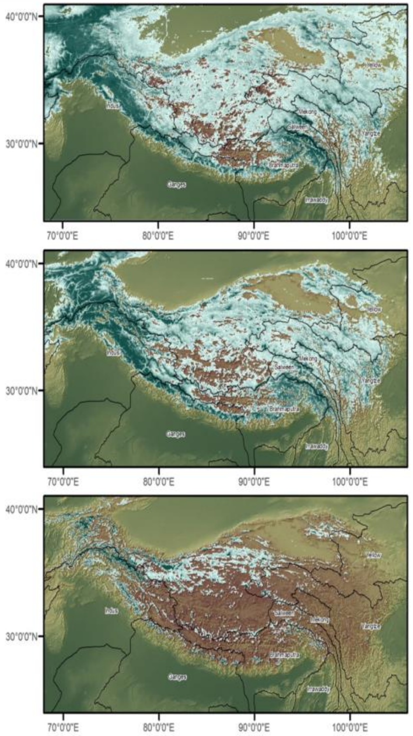
(Seidel and Martinec, 2004; Baumgartner, 1987, taken from DeWalle & Rango, 2008)

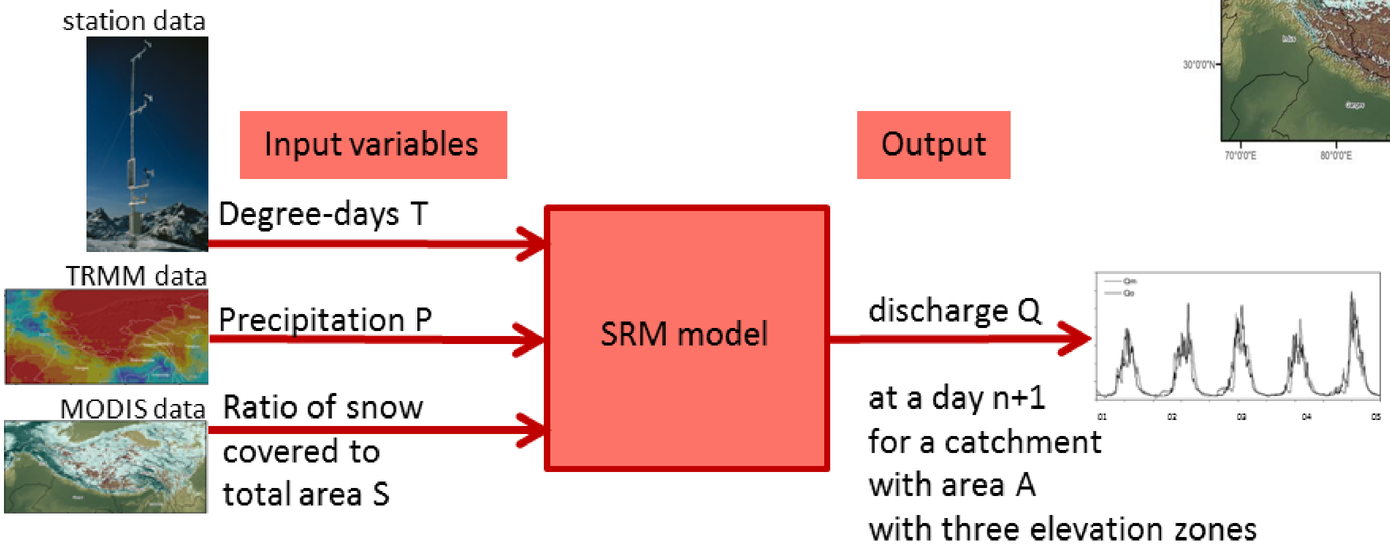
3.2. Temperature-index model

Example of a SRM model using depletion curves derived by satellite data

Immerzeel et al. (2009) used remote sensing products to force a hydrological model of the upper Indus basin

snow covered to total area S	-> from MODIS satellite product
Precipitation	-> from satellite based TRMM data
Degree-days	-> station data





station data

TRMM data

MODIS data

Input variables

Output

Degree-days T

Precipitation P

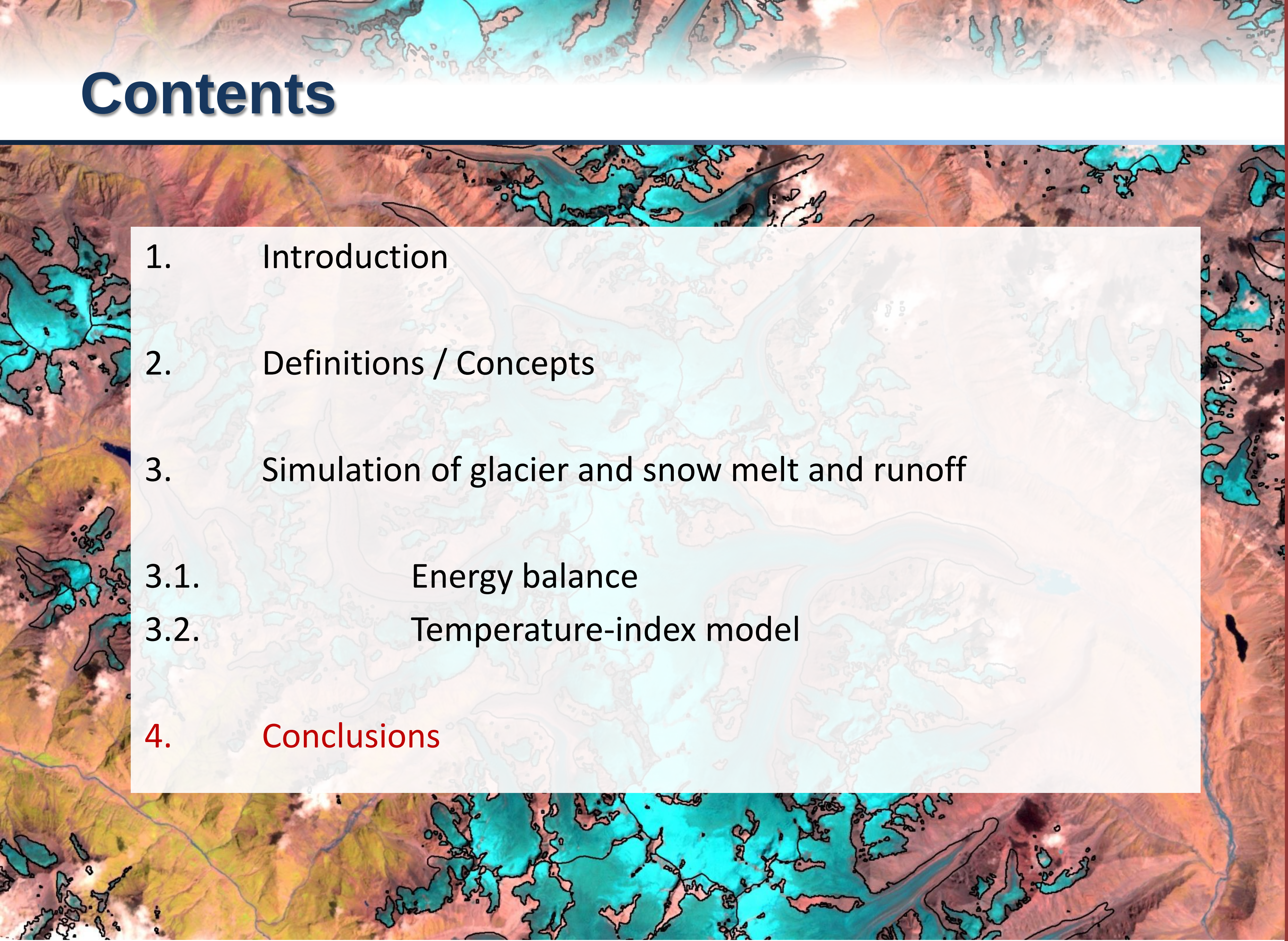
Ratio of snow covered to total area S

SRM model

discharge Q

at a day n+1
for a catchment
with area A
with three elevation zones

Contents



1. Introduction
2. Definitions / Concepts
3. Simulation of glacier and snow melt and runoff
   - 3.1. Energy balance
   - 3.2. Temperature-index model
4. Conclusions

4. Conclusions

Remote sensing data and techniques are useful tools and should be futher developed and applied

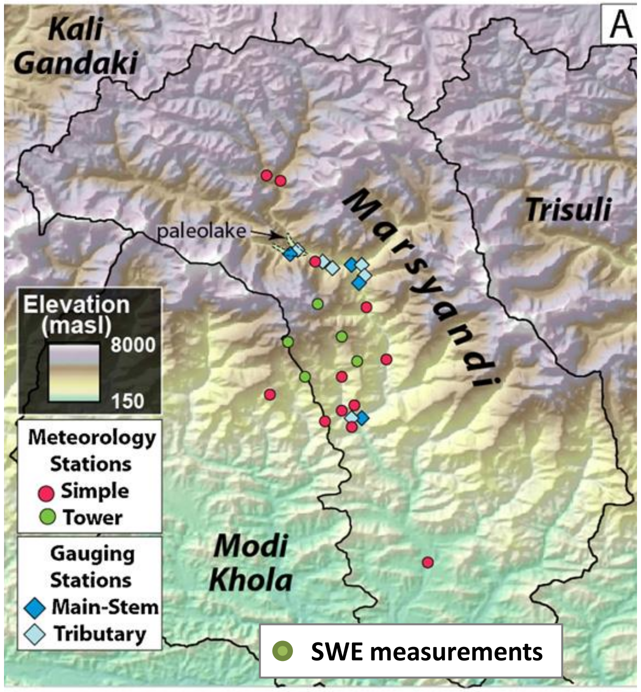
More in-situ measurement sites are advocated including

- daily measurements of SWE and snow cover
- energy balance components
- discharge measurements

Why designing a permanent high altitude network?

- calibration and validation of models e.g. SRM
- calibration and validation of remote sensing products
- components of the energy balance could change under a changing climate

A station concept similar to the former Marsyandi network in the Annapurna region of Nepal would be ideal (see figure)



Map of meteorological, snow and river-gauging network in Marsyandi, central Nepal, from 1999 to 2004.
(Burbank et al., 2012)