Land Surface: Introduction to cold processes

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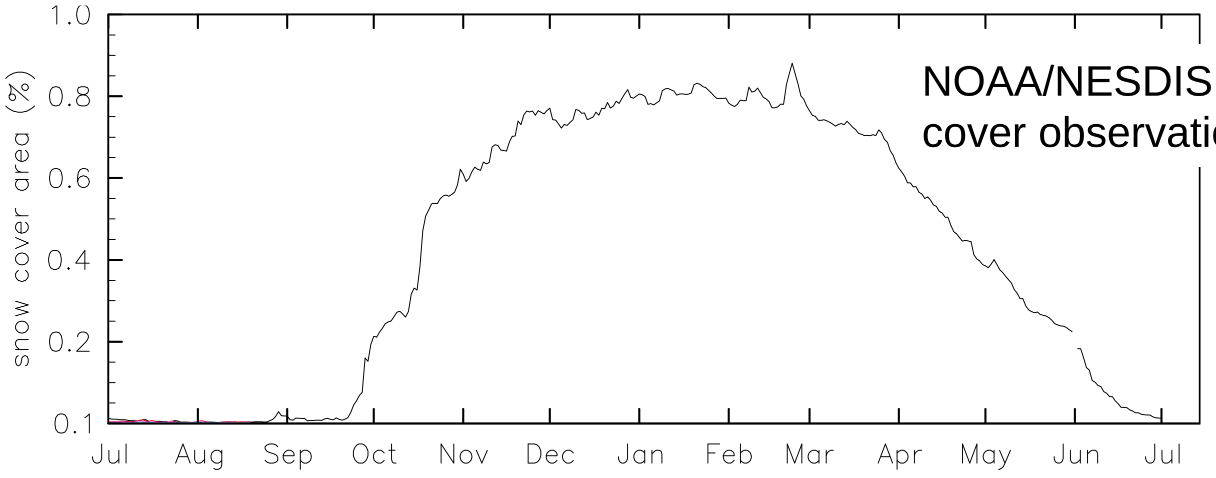
Outlines

- **Snow and its physical properties, an overview**
- Snow modelling at ECMWF
  - Complexity of snow modelling
  - Snow modelling at ECMWF
    - Water and energy balances
    - Snow properties (density, albedo)
    - Snow in global NWP models (snow cover fraction, impacts)

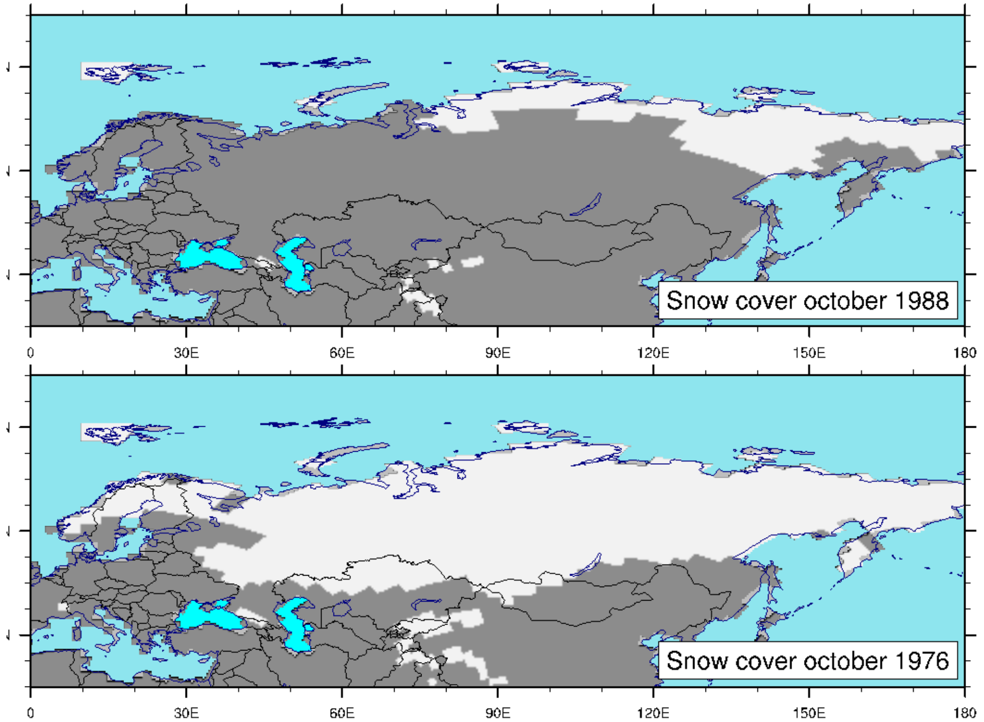
Snow in the climate system

- Snow cover is one of the main component of the Cryosphere (with seasonally frozen ground, ice sheets, sea-ice) with a mean maximum areal extend of 47 million km^2 (98% in the Northern Hemisphere);
- Several fundamental physical properties of snow largely affect the energy/water exchanges between the surface and the atmosphere
- Implications for all forecasts ranges (medium to seasonal)

Annual cycle of snow cover extent over Eurasia (northward 30N) between 2011-2018

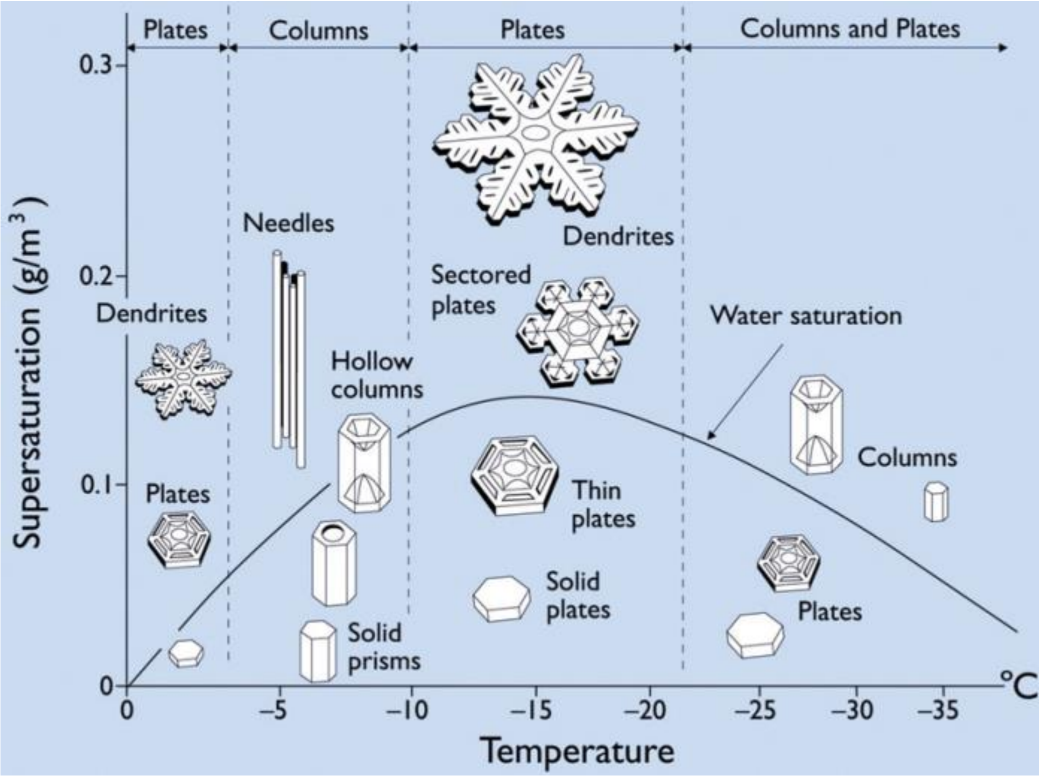


NOAA/NESDIS snow cover observations



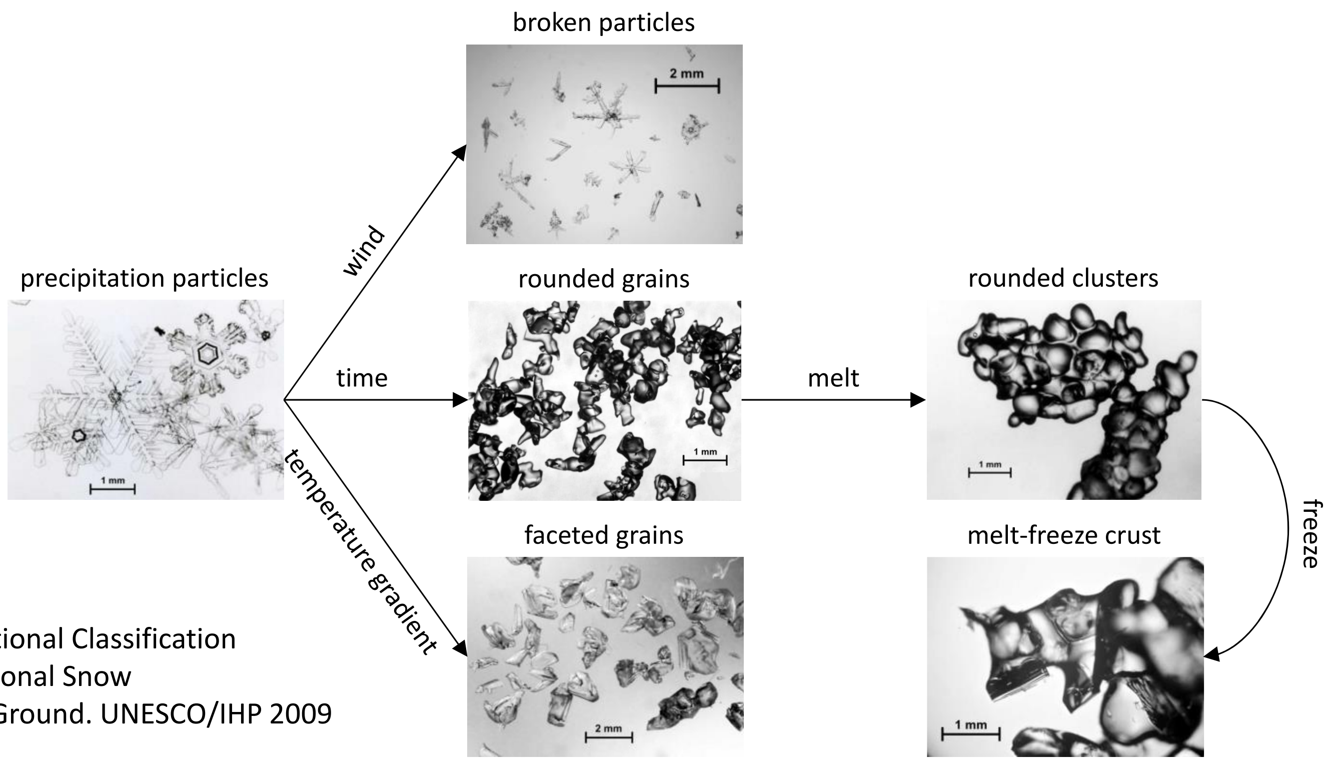
Snow properties – grain shape and its change with time

Snow grain shapes



Libbrecht (2005)

Snow grain metamorphosis



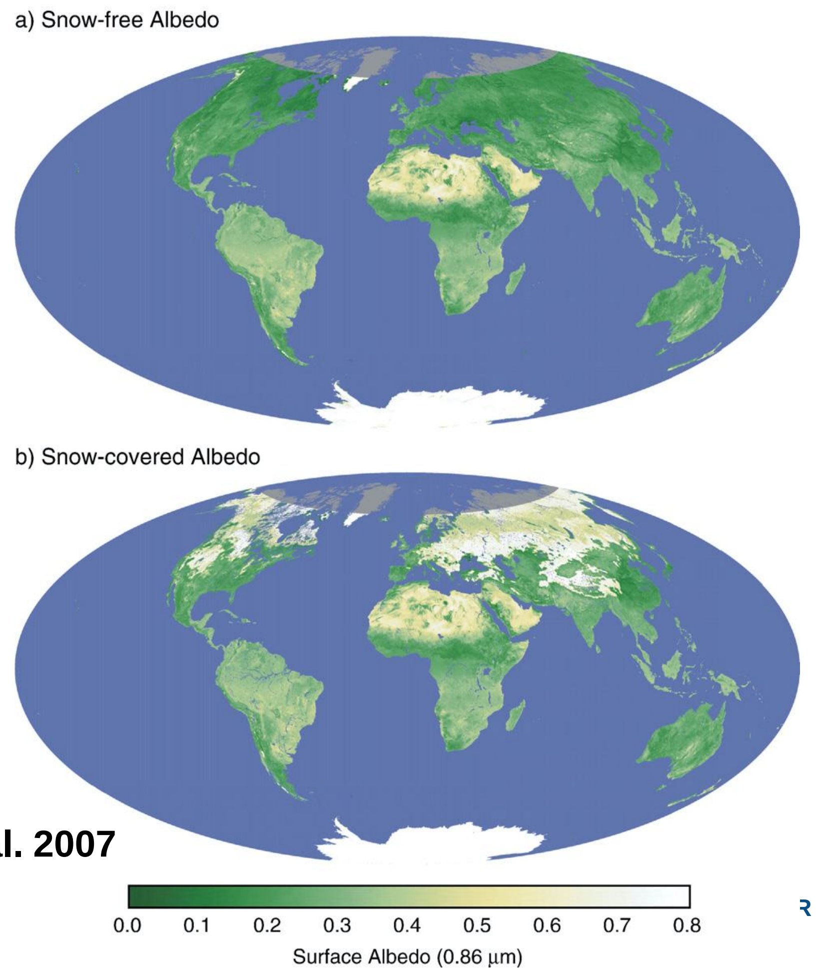
International Classification
for Seasonal Snow
on the Ground. UNESCO/IHP 2009

Adapted from Essery (2018)

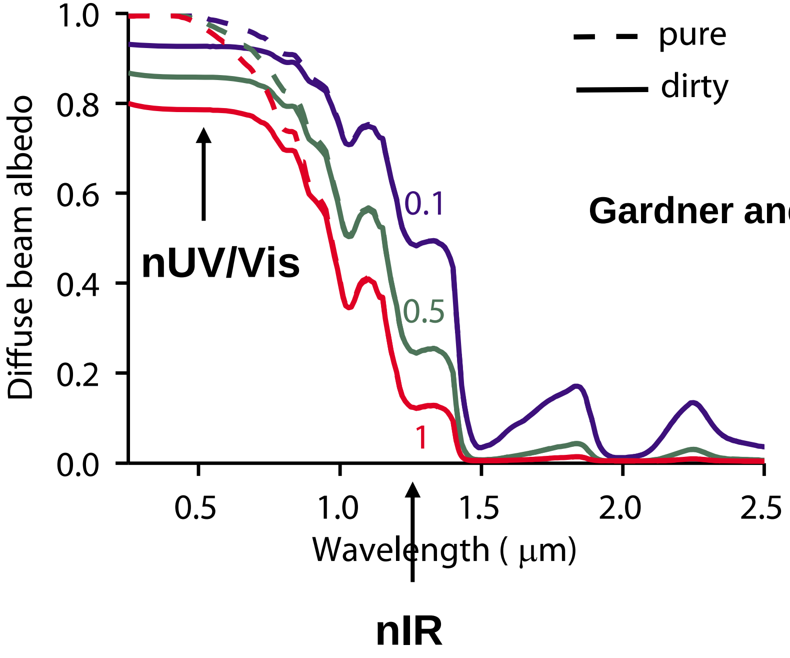
Snow is characterized by different type of grains undergoing a metamorphism with time, varying its shape and size
→ direct impact on physical properties

Snow properties - albedo

- Snow is characterized by very high values of albedo in the near-ultraviolet (nUV) and visible (Vis) range of the spectrum
- Snow albedo greatly varies with wavelength: in the near-infrared (>~1.5µm) snow is relatively opaque



Moody et al. 2007



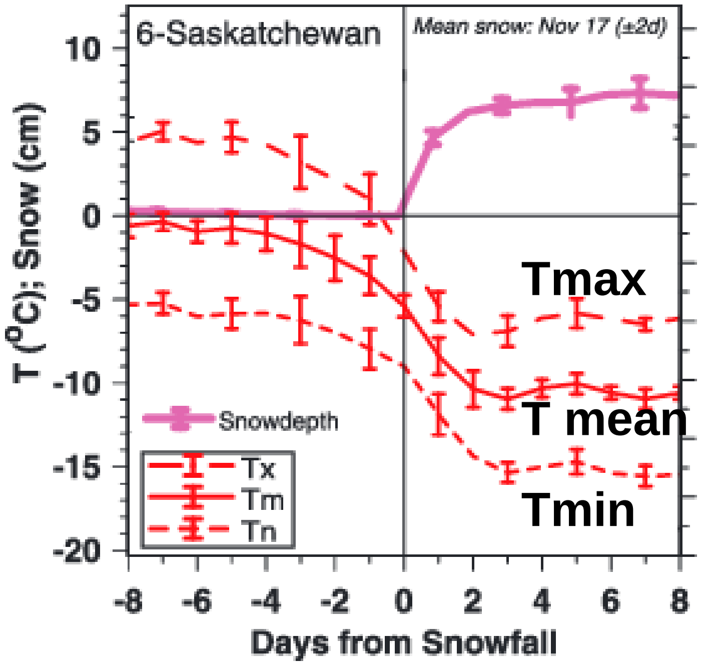
Gardner and Sharp, 2010

- Snow albedo is influenced by aerosol deposition in the **nUV/Vis range and by the snow grain size in the nIR**.
- Other factors influencing albedo: solar angle of incidence, roughness…

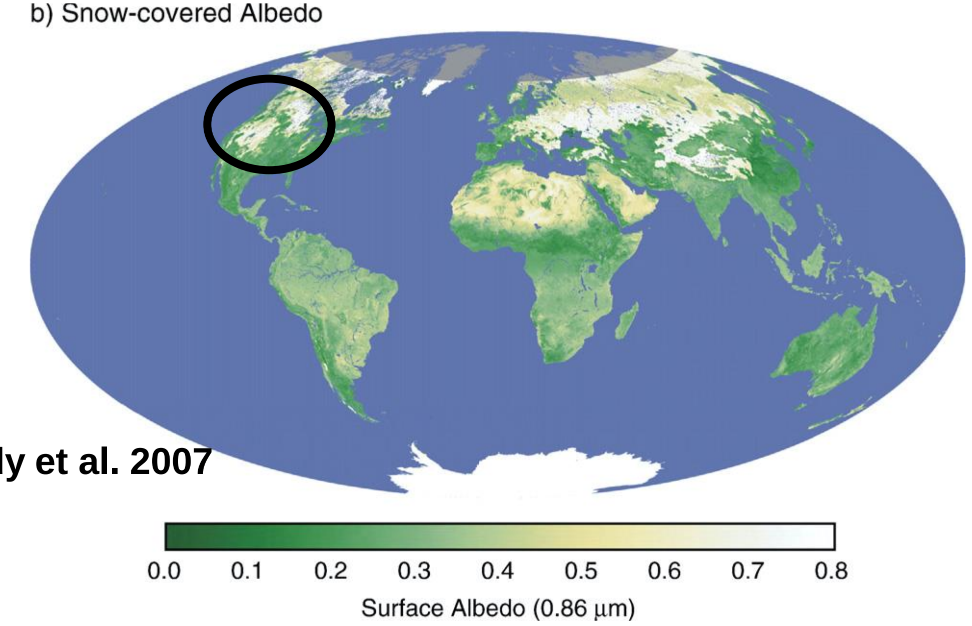
Impact of the high albedo of snow (Canadian Prairies)

- The change in the net shortwave solar radiation at the surface is the main responsible of the “climate-switch” when snow deposits on the ground (Betts et al. 2014)

Adapted from Betts et al 2014

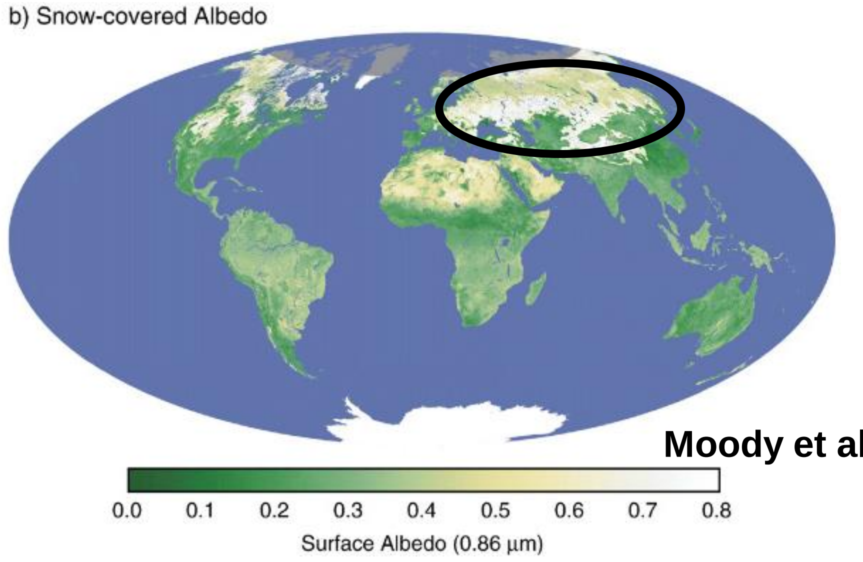


Climatology for six climate stations in the Canadian Prairies all having more than 30 years of data



Moody et al. 2007

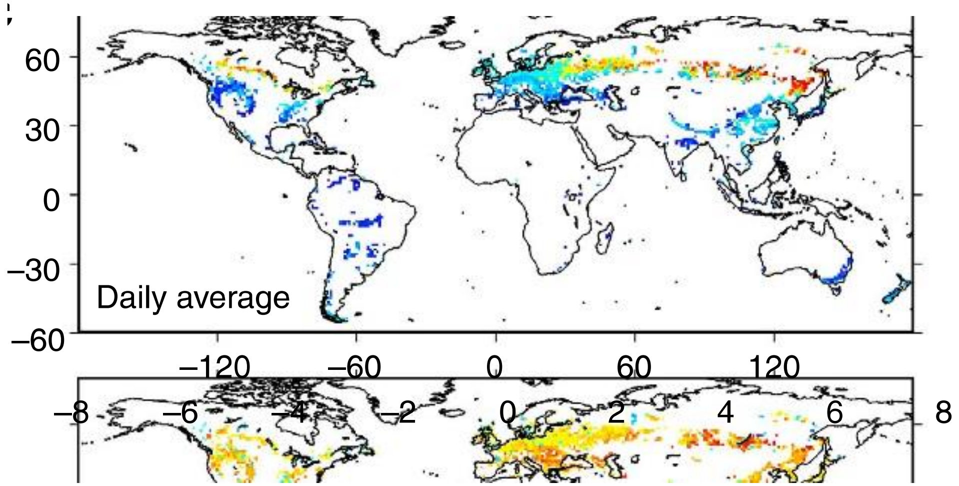
Role of boreal forest on albedo



Moody et al 2007

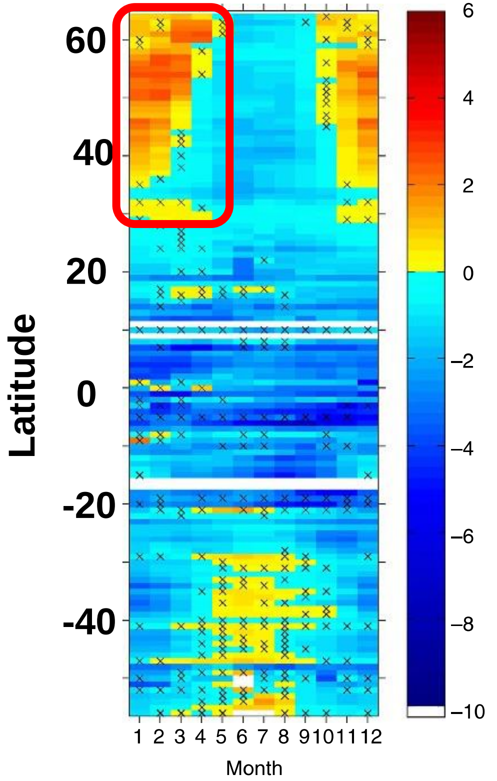
- The presence of boreal forests has a direct control on the climate of high-latitudes.
- Warmer surface temperature than on snow accumulating on grass/soil, with a possible effect also on boundary layer

Surface temperature difference forest-open_land (C) from MODIS satellite

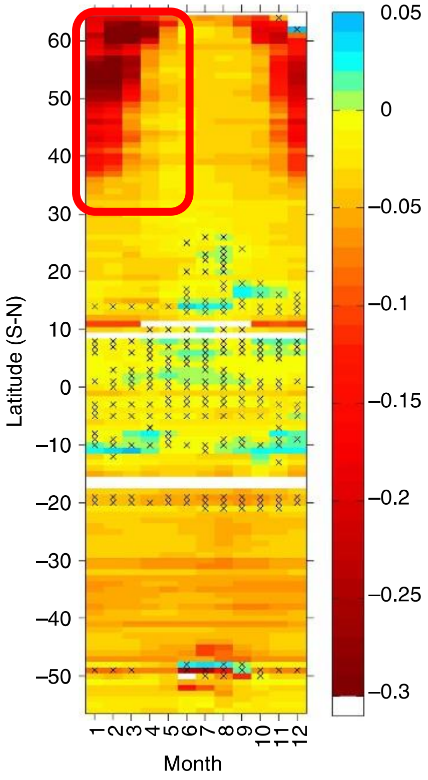


Li et al. 2015

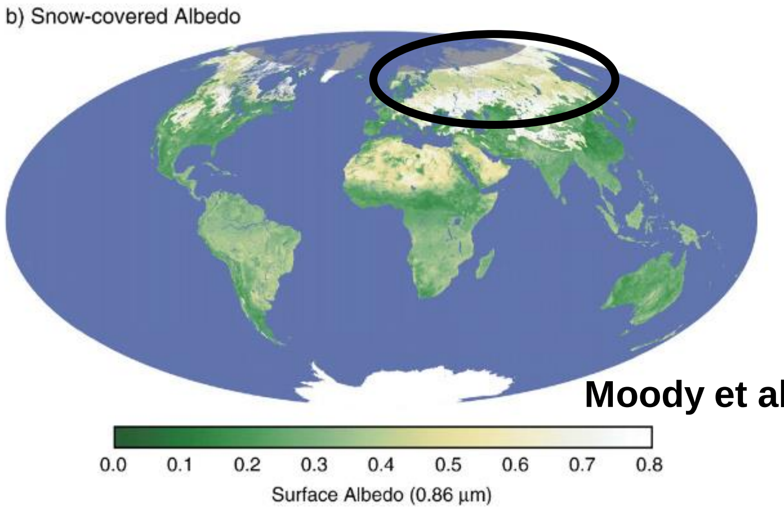
Daily average LST forest -open



Albedo forest – open_land



Boreal forest albedo and impact on NWP



Moody et al 2007

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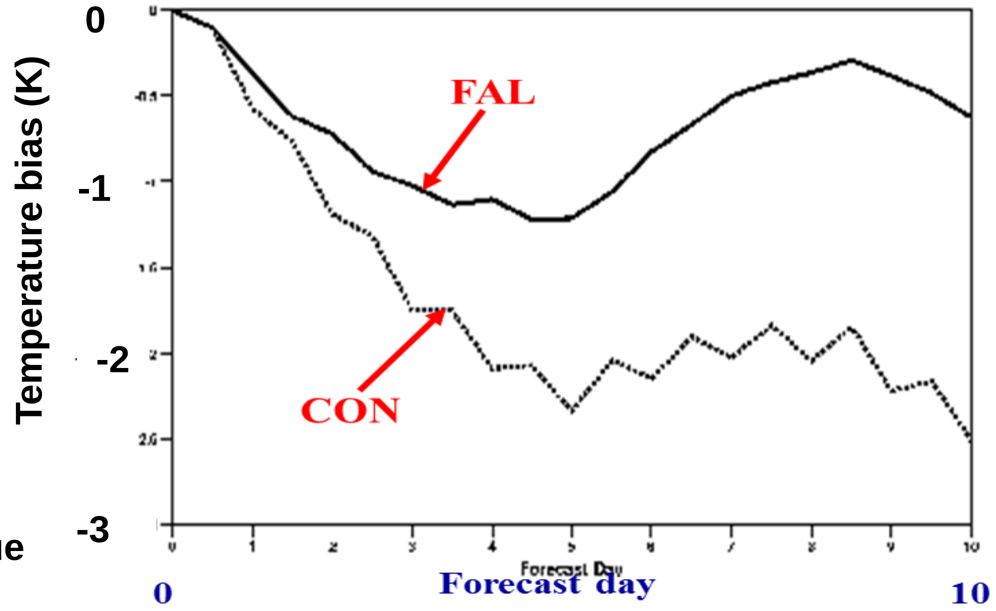
Bias of temperature at 850hPa
For east Asia region, March to April



CON: forecasts with albedo of snow for all surfaces

FAL: forecasts with lower value of snow albedo for forest area

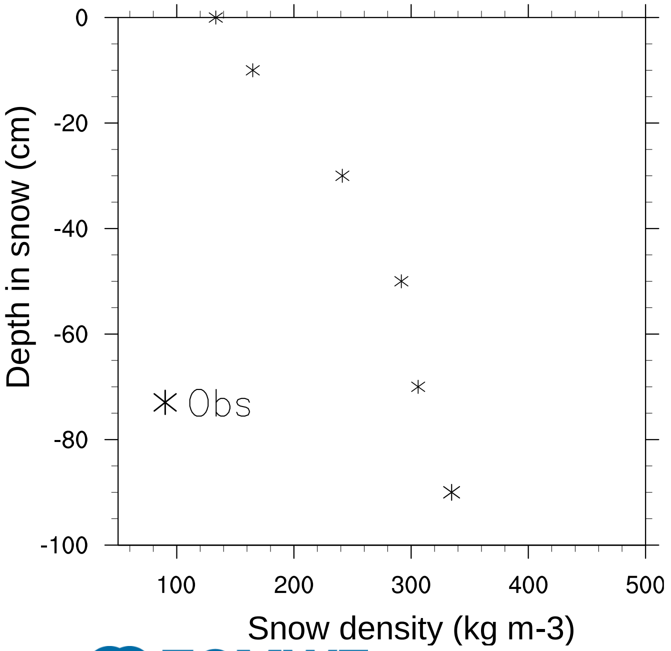
Viterbo and Betts 1999

Snow properties – snow density and thermal insulation

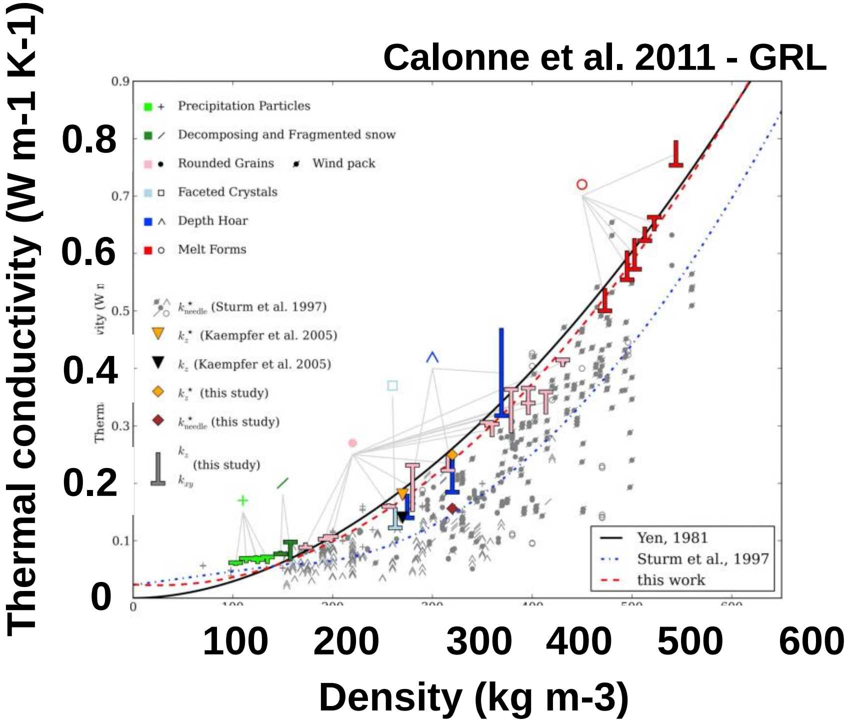
Thermo-insulation properties of snow:

- Snow is a porous medium composed of (frozen/liquid) water and air, the latter being a very effective thermal insulator:
    - thermal conductivity of air: ~0.023 W m^{-1} K^{-1} (at -15 C)
    - Thermal conductivity of ice: ~2.2 W m^{-1} K^{-1} (at -15 C)
- Thermal conductivity depends on the grain shape and size, but can be related to density
- Density of snow ranges between ~50 kg m^{-3} (fresh) to ~500 kg m^{-3} (compacted)

Snow density profile (Dec average 1994-2011) at Col de Porte, France



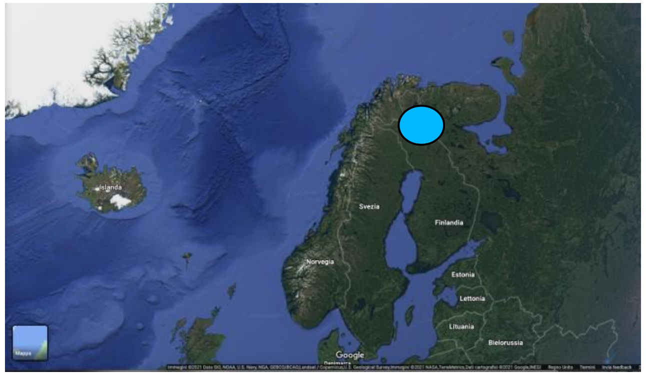
Calonne et al. 2011 - GRL



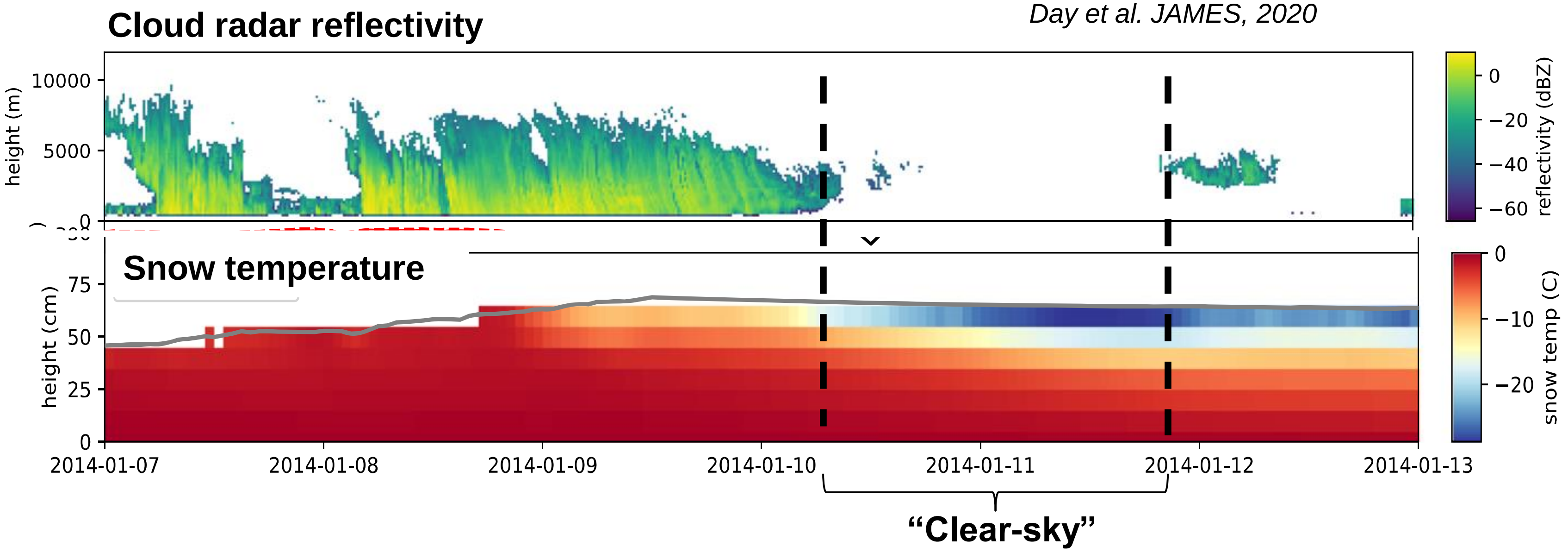
Snow insulation during wintertime at a high latitude site

In the Arctic during wintertime, the main radiative forcing at the surface is longwave radiation from clouds

Observations at Sodankyla, Finland, 67N
From 7th Jan 2014 to 13th Jan 2014.



Day et al. JAMES, 2020



- Snowpack is relatively isothermal during cloudy conditions.
- During periods of low cloud radiative forcing (e.g., clear-skies), **large temperature gradients** are established within the snowpack.

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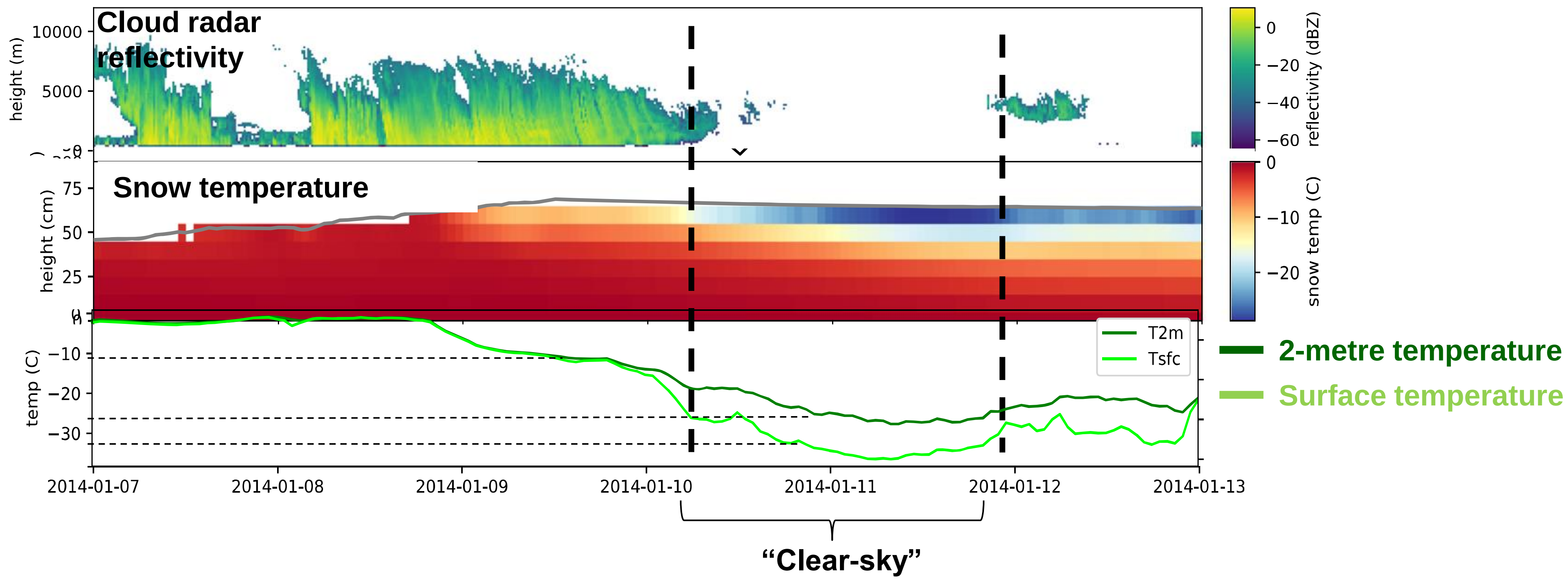
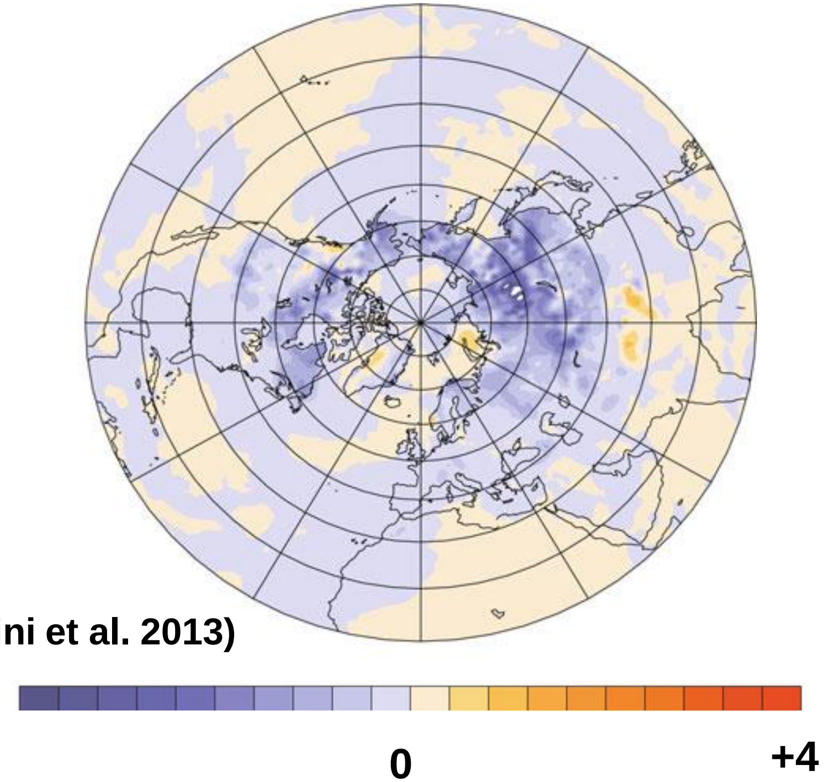


Effective decoupling between the atmosphere and snow-soil below:

- very cold 2-metre temperature over snow surfaces
- Soil remains relatively warm (e.g. ~0 C)
- Strong surface-based inversions

Snow insulation and impact in medium-range forecasts

Mean 2-metre temperature difference after 15 days between **ensemble** of simulations initialized **with lower values** of snow depth and **ensemble** initialized **with larger values** of snow depth.



(Orsolini et al. 2013)

A **thicker** snowpack (of same density) is associated to **colder** near-surface temperatures (up to 4K in the example), because of the stronger insulation of the lower atmosphere from the warmer soil below

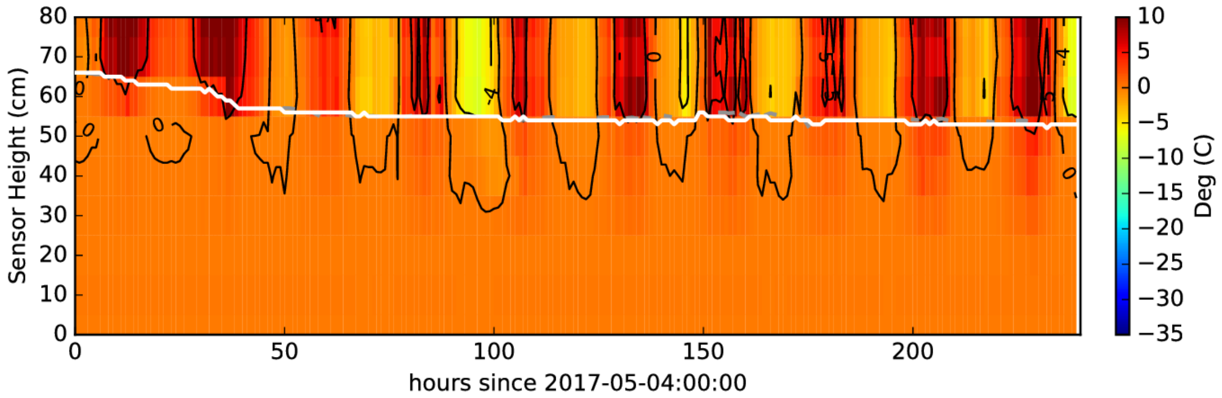
Snow properties – phase changes (melting)

Snow is characterized by a large latent heat of fusion:

- Latent heat of fusion snow = ~ 334 kJ kg-1
- Specific heat capacity of ice = ~2.1 kJ kg-1 K-1

Observations at Sodankyla, Finland, 67N, 4th May 2017 to 14th May 2017.

Measured **temperature profiles within the snowpack** (below white line) and **air temperature** (above white line)



especially below about 100 m. As the result, the upper limit at which the influence of melting surface reaches could be estimated about 100 m above the ground surface at Moshiri.

The Scale Effect of the Melting Surface on the Lower Atmosphere

The sensitivity of climate to melting and snow-free surfaces at various regions is compared in Figure 4. The values for the snow-free period were obtained just after the snow cover disappeared in the seasonal snow cover regions such as Moshiri and Spitsbergen. In the case of the glaciers or the snowpatch the values were obtained on the bare surface close to them. Since β depends on geographical and climatic conditions of the region as well as the surface conditions, β values are different among regions on even the similar melting surface. However, it is clear as shown in Figure 4, that the value of β are smaller for the

Takeuchi et al., 2002

Snow in melting conditions:

- Delaying the warming of the near-surface atmosphere: energy is used for melting and snow temperature bounded at 0C
- Refreezing can effectively warm the snowpack
- Co-existence of ice and liquid water in the snowpack: melting/refreezing cycles

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  - Complexity of snow modelling
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    - Water and energy balances
    - Snow properties (density, albedo)
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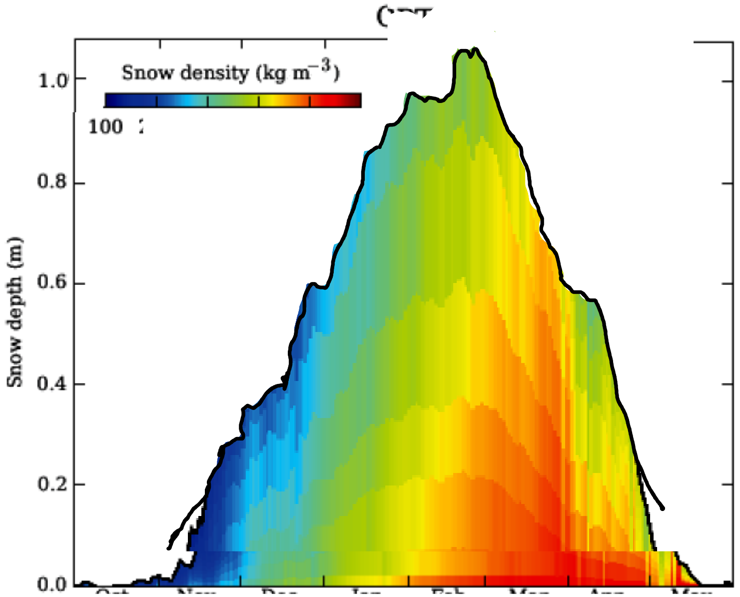
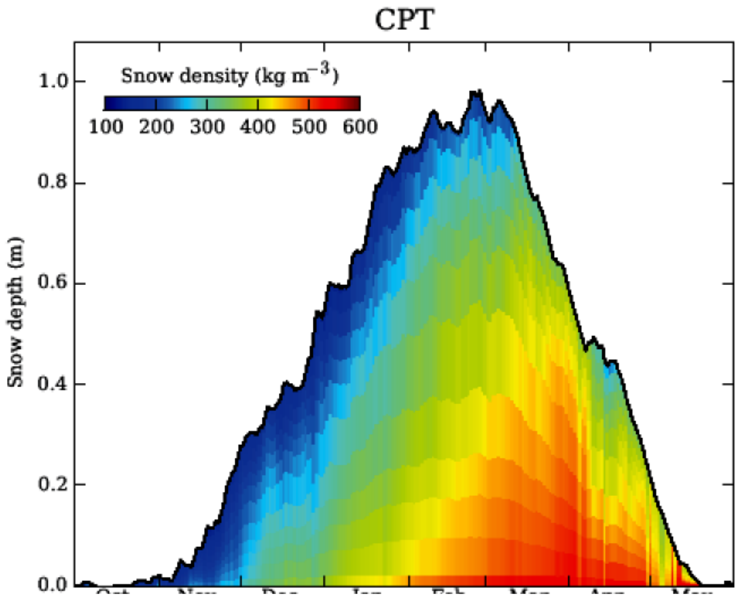
Complexity of snow modelling

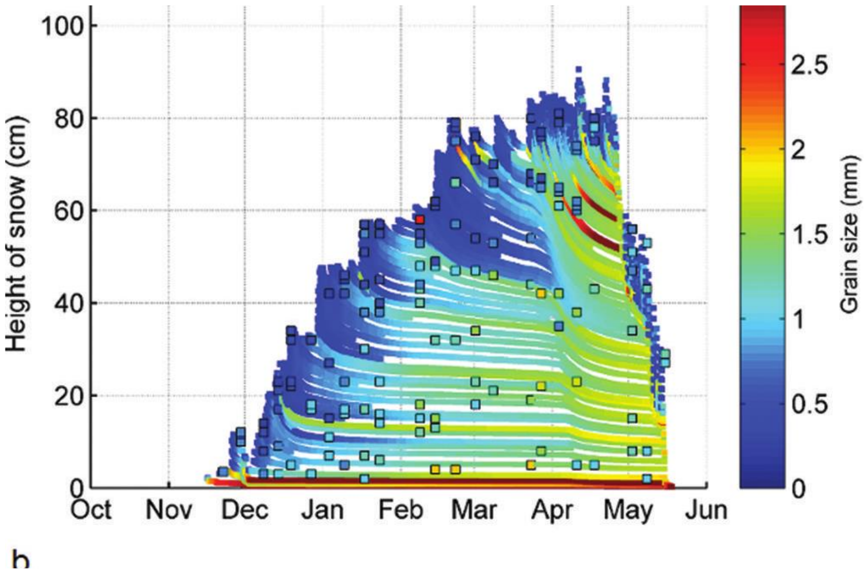
- Single layer:
  - Simpler and quicker
  - Empirical or physical
  - Bulk quantities (total mass, average temperature)

- Intermediate complexity:
  - Multiple layers, limited number
  - Simplified microstructure parametrizations
  - Representing layer-average macrophysical quantities (density, temperature…)

- Detailed snow physics models
  - High vertical resolution (e.g. each snowfall forms a layer)
  - Snow microstructure properties (grain size, shape etc.)

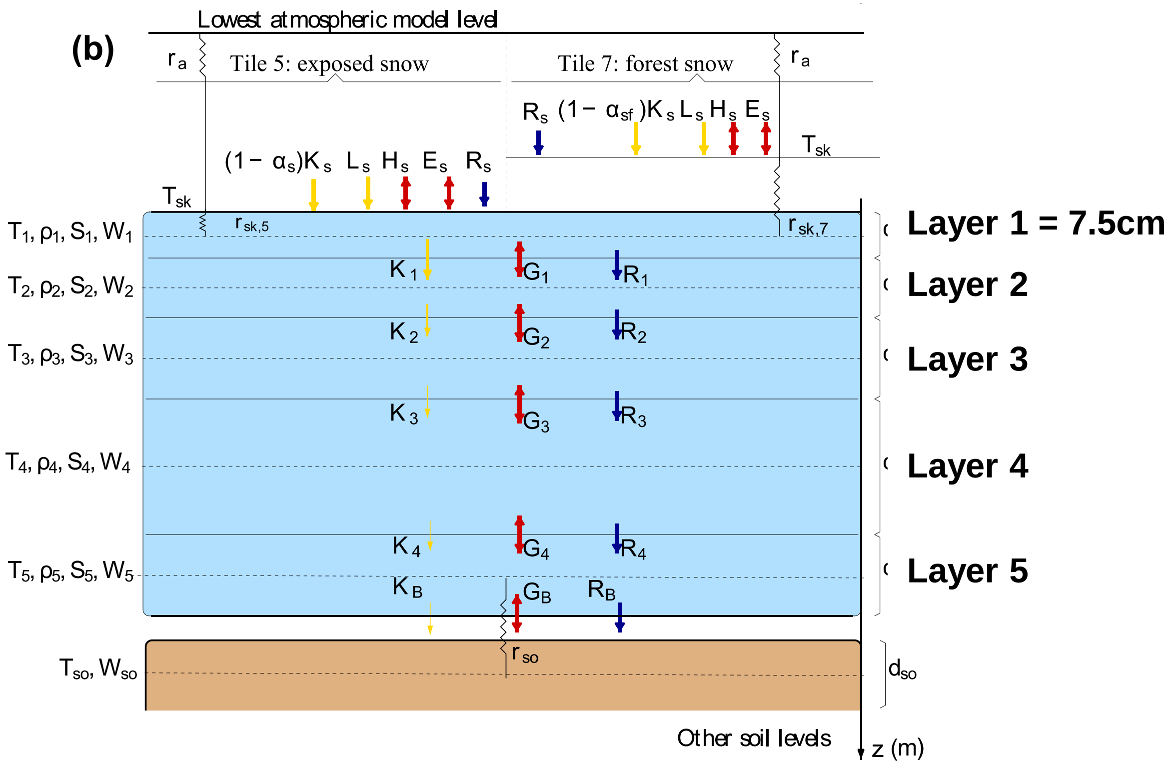


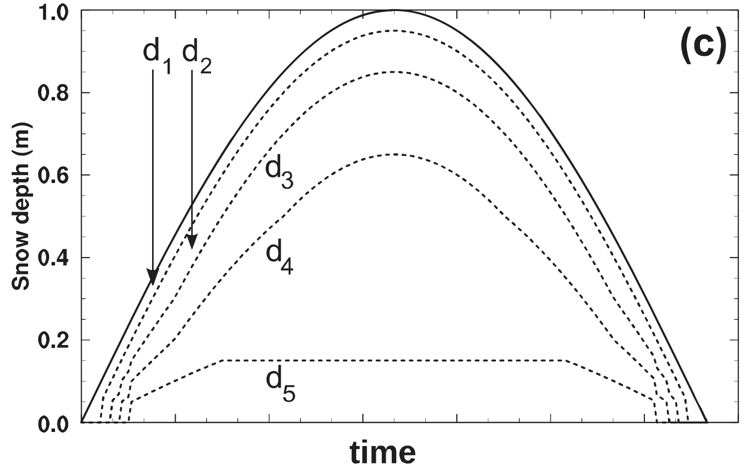




Snow modelling at ECMWF

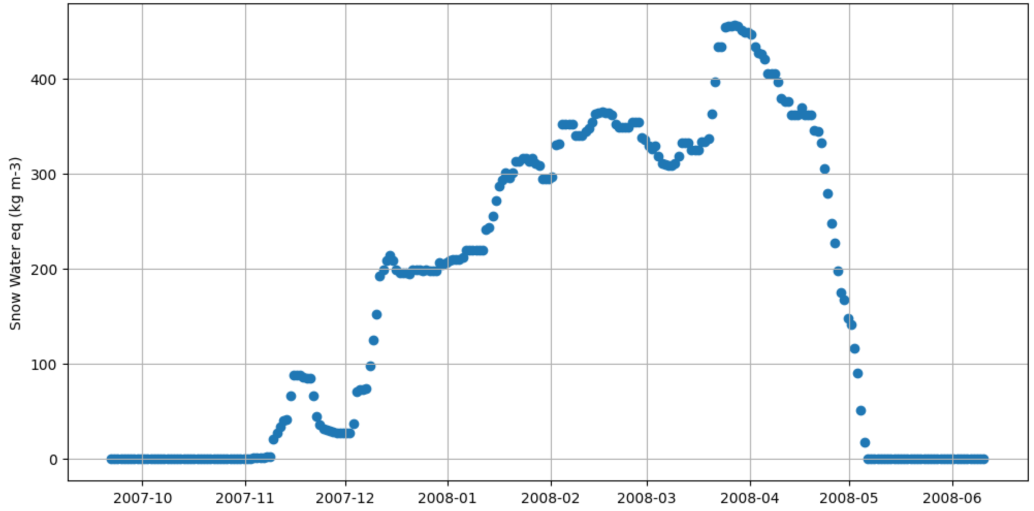
- Snow is an additional module with dedicated number of layers, sitting on top of the soil column.
- Multi-layer intermediate complexity:
  - **Dynamic layering up to 5 layers →** limited "memory" of snowfalls
  - **<u>no</u> microstructure** properties evolution (grain size and shape etc.)
  - Physically based water and energy balance
  - 5 prognostic variables: **snow water eq, snow density, snow temperature, snow liquid water and snow albedo**

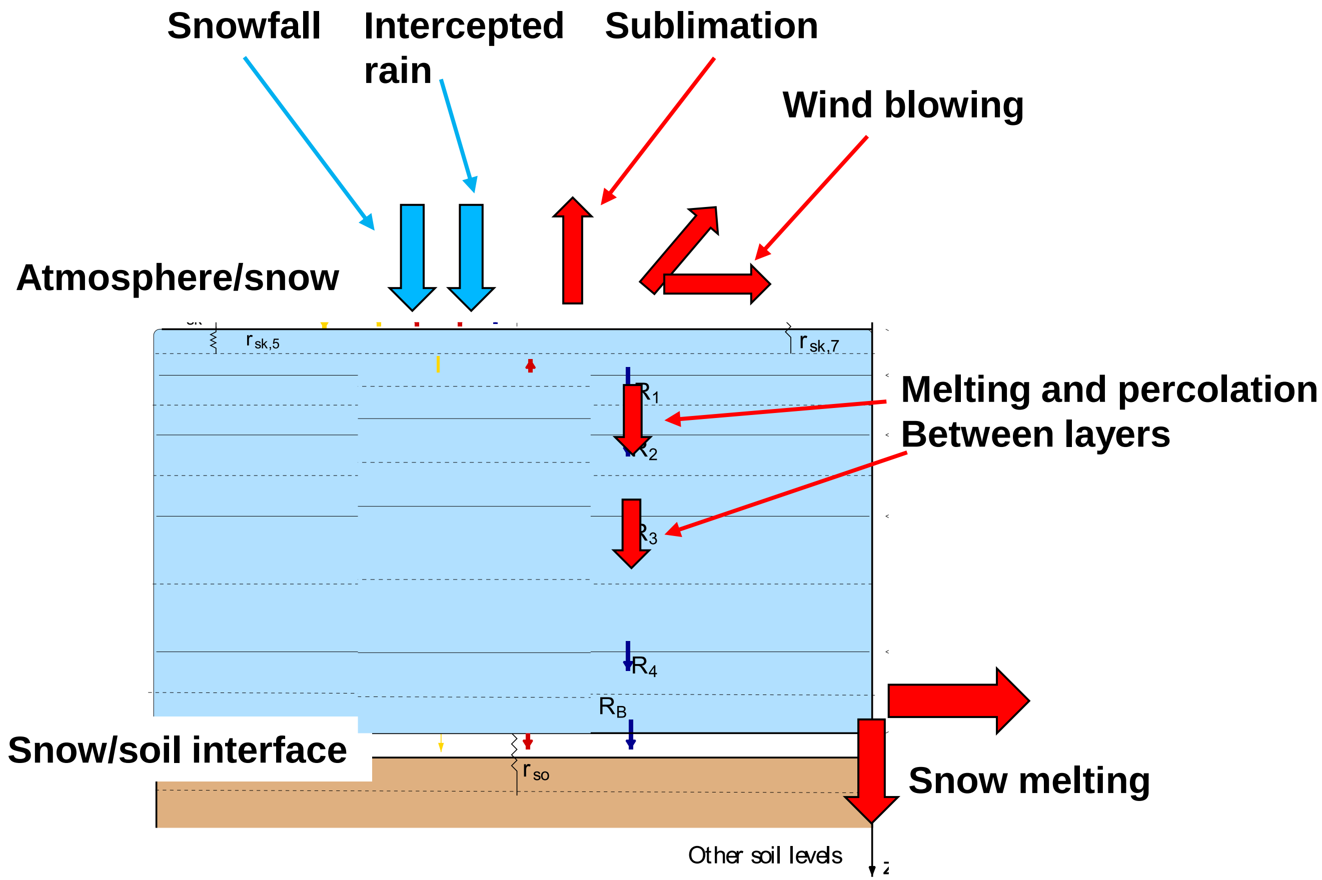




Snow accumulation and water balance

Snow water equivalent (snow mass) evolution at Col de Porte, French alps





- **Snow melt at the bottom layer** percolate into the soil or leave the grid cell as total runoff
- **How Does snow melt?** Coupling between the energy and water balance

Energy balance of the snowpack, step 1

Snow energy equation, in absence of melting, resulting from heat diffusion and absorption of shortwave radiation by the snowpack:

$$\rho\, c_p \frac{\partial T_{sn}}{\partial t} = -\frac{\partial}{\partial z}\left(\kappa_{sn} \frac{\partial T_{sn}}{\partial z}\right) - \frac{\partial K}{\partial z}$$

Heat storage evolution in a snow layer

Snow thermal conductivity, for each layer:

$$k_{sn,i} = f(\rho_{sn,i}, T_{sn,i}, P_{atm})$$

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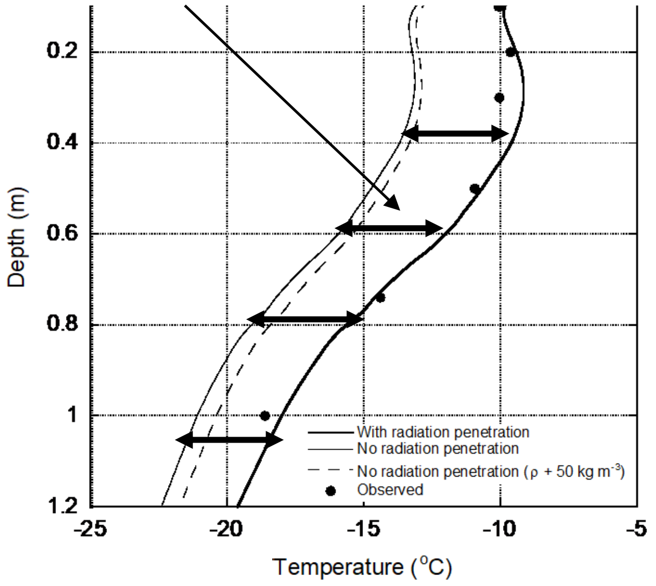
Shortwave radiation penetration

Snow thermal conductivity, for each layer:

$$k_{sn,i} = f(\rho_{sn,i}, T_{sn,i}, P_{atm})$$

Observations

Temperature difference between model with and without shortwave radiation penetration



From Pomeroy et al. 2016

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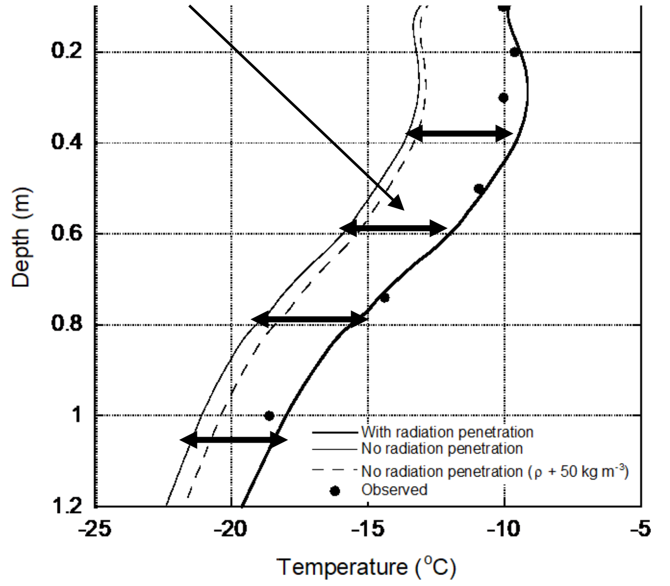
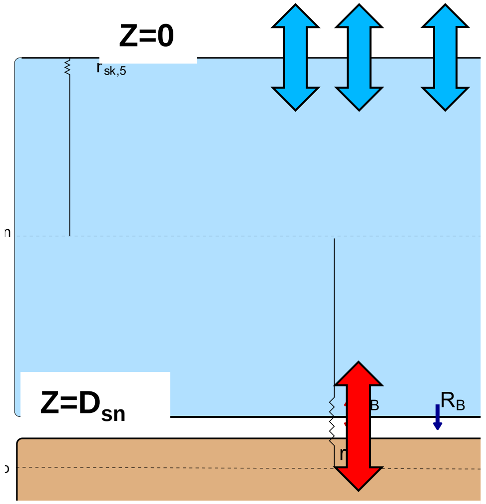


Flux boundary conditions (atmospheric and soil forcing):

F(z=0) = $R^N_{sn} - L_s E_{sn} - H_{sn}$

Atmosphere/snow interface



F(z=D$_{sn}$) = G^B_{sn}

Snow/soil interface

R^N_{sn} : heat flux due to net radiation

$L_s E_{sn}$: Latent heat flux due to snow sublimation

H_{sn} : Sensible heat flux due to turbulent processes

G^B_{sn} : basal heat flux in/from the soil due to conduction

Energy balance of the snowpack, step 2, melting

Melting and freezing are diagnosing using the “cold” (Heat) content of the snowpack, for each layer i:

$$H_i = \underbrace{c_p}_{\text{Heat capacity}} S\,(\underbrace{T_i}_{\text{Snow temperature}} - \underbrace{T_0}_{\text{Freezing temperature}}) \qquad \begin{cases} H_i > 0 \;:\; \text{melting occurs} \\ H_i < 0 \;:\; \text{freezing occurs, if liquid water available,} \end{cases}$$

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Heat capacity

Snow temperature

Freezing temperature

$$H_i > 0 \ :$$

$$L_f M_i = \min(\,H_i, L_f S_i\,) \longrightarrow$$

“If the amount of available energy for melting, H, is larger than the latent heat associated with the frozen water in the layer, all the snow in that layer is melted”

Snow melt

Latent heat of the frozen snow

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Snow melt

Latent heat of the frozen snow

$$H_i < 0 \;:$$

$$L_f F_i = \max(\,H_i, -L_f W_i\,) \longrightarrow$$

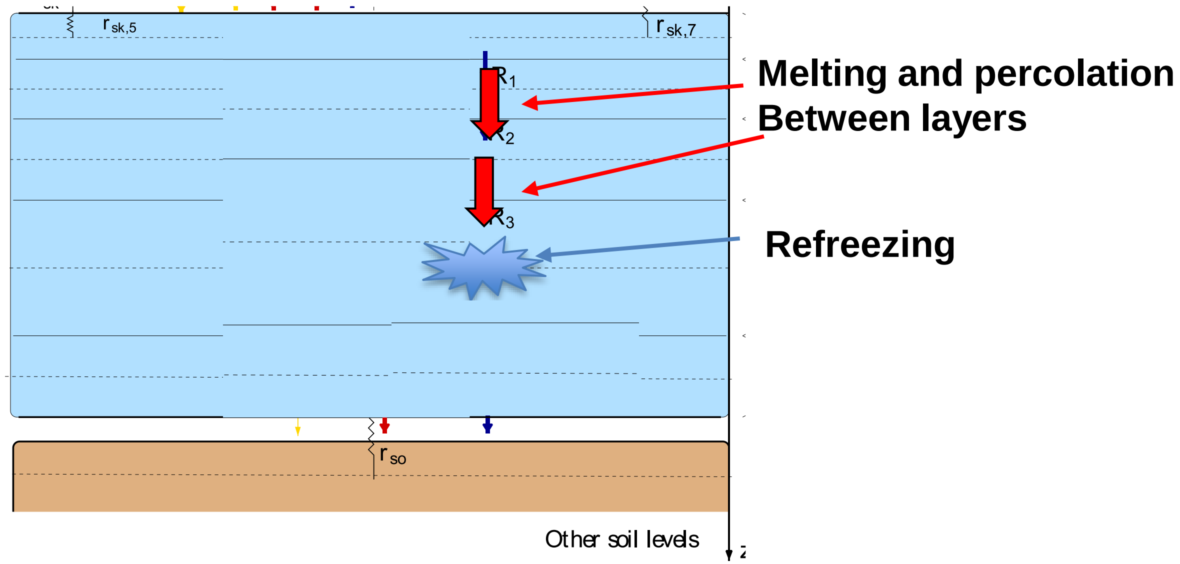
“If the amount of available energy for refreezing, H, is larger (in absolute sense) than the latent heat of the liquid water in the layer, all the water refreeze”

refreezing

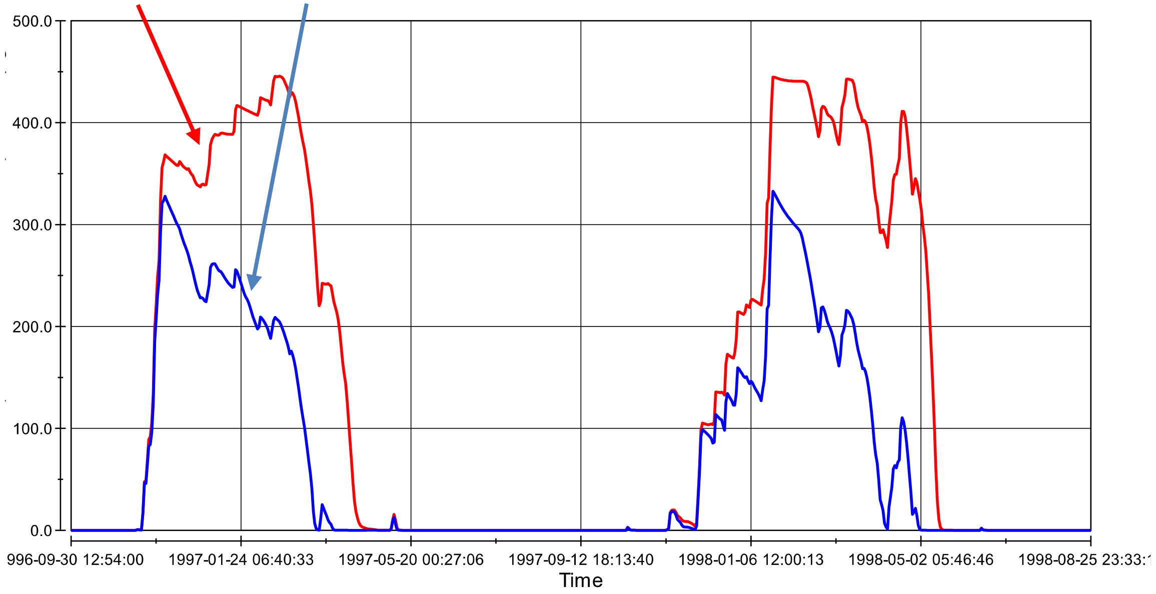
Latent heat of the liquid water

Is the snow melt water retained in the snowpack and refrozen important?

Liquid water retaining within the snowpack is important to correctly simulate the snow water balance through the winter and spring season.



Snow water equivalent ("mass") for simulation **with and without liquid water retention and refreezing**

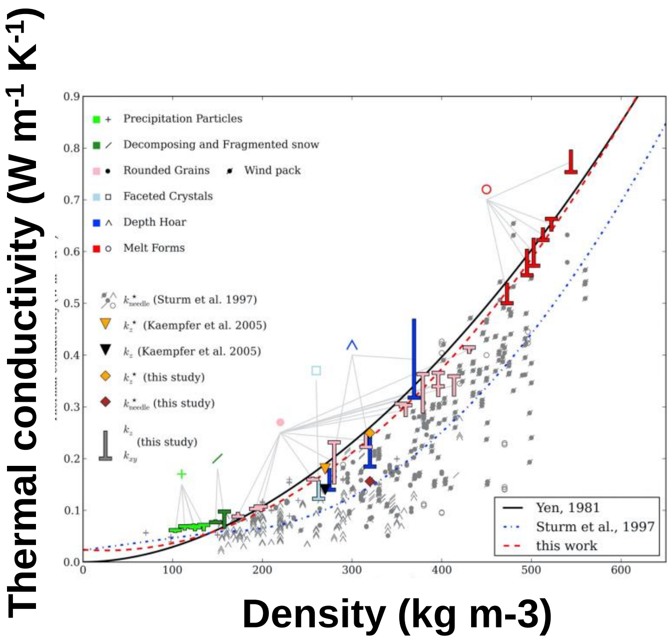


Snow density evolution

Snow density is important for calculating:

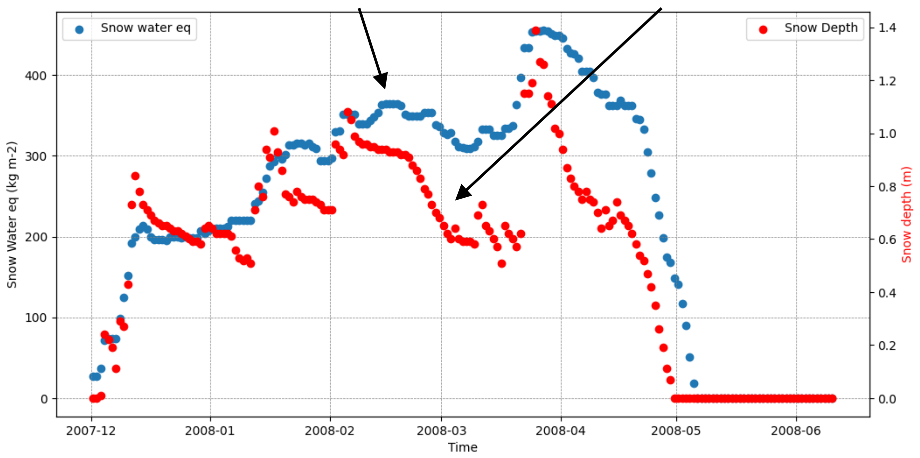
- The snow depth variation with time: $$D_{\mathrm{sn}} = \frac{S}{\rho}$$

- The thermal conductivity of the snowpack, e.g. Yen et al.: $$k_{sn} = k_{ice} \left(\frac{\rho_{sn}}{\rho_{ice}} \right)^{1.88}$$

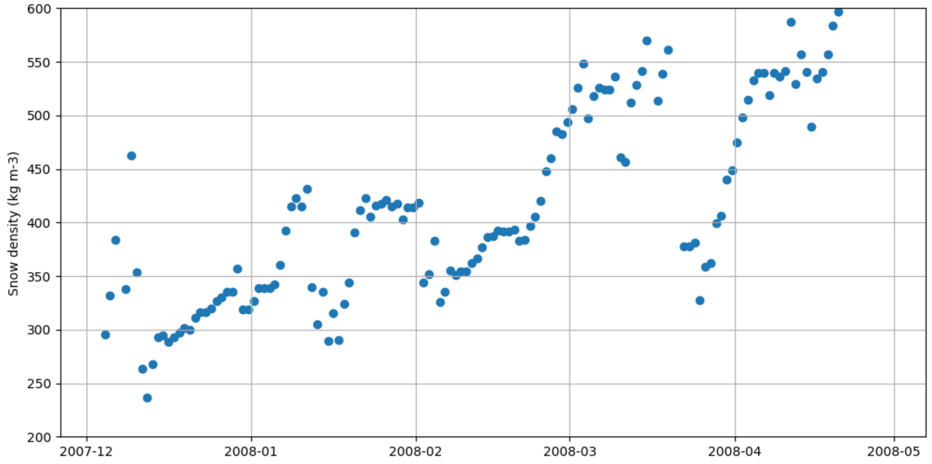


Snow water equivalent

Snow depth



Snow density evolution



Snow density evolution, processes

Snow density evolves in each snow layer due to different processes and time-scales:

- **overburden**: increase of density due to the snow weight above

Snowpack pressure
= SWE * g

$$\frac{1}{\rho_{\mathrm{sn}}}\frac{\partial \rho_{\mathrm{sn}}}{\partial t} = \frac{\sigma_{\mathrm{sn}}}{\eta_{\mathrm{sn}}(T_{\mathrm{sn}}, \rho_{\mathrm{sn}})}$$

Newtonian
viscosity

Snow density evolution, processes

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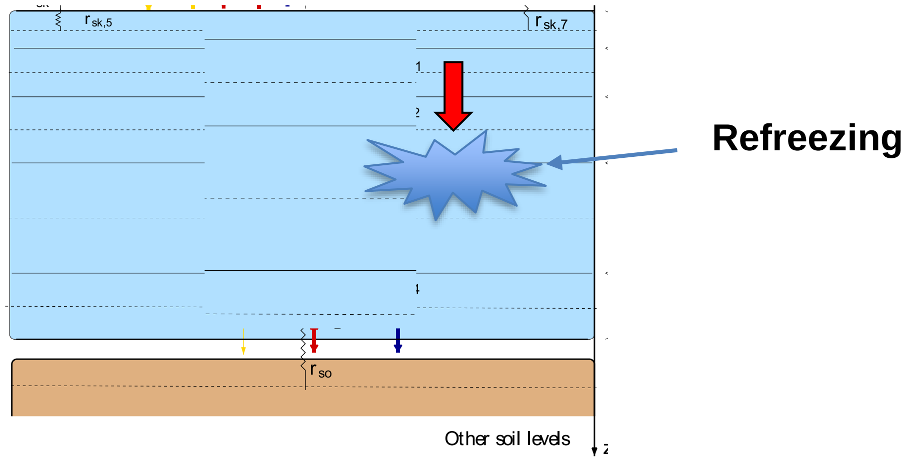
- **overburden**: increase of density due to the snow weight

- **Increase of density due to refreezing:**
  Meltwater retained into the snowpack can refreeze at ice density (~920 kg m-3), filling air space between snow grains



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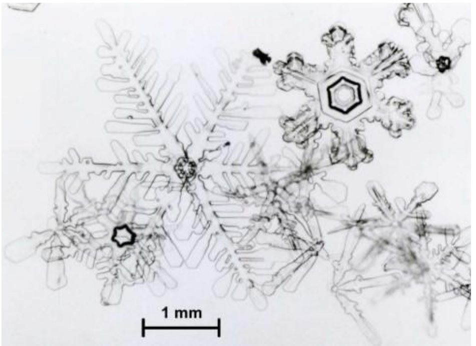
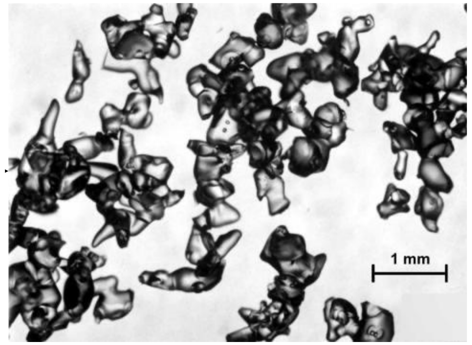
- **Equi-temperature metamorphism**:
  change of shape and size of snow crystal with time. Mainly active for fresh snow.

precipitation particles



time

rounded grains



$$\frac{1}{\rho_{\mathrm{sn}}}\frac{\partial \rho_{\mathrm{sn}}}{\partial t} = a \exp\left[-b\,(T_0 - T_{sn}) - c\max(0, \rho_{sn} - 150.0)\right]$$

Empirical function of T and density

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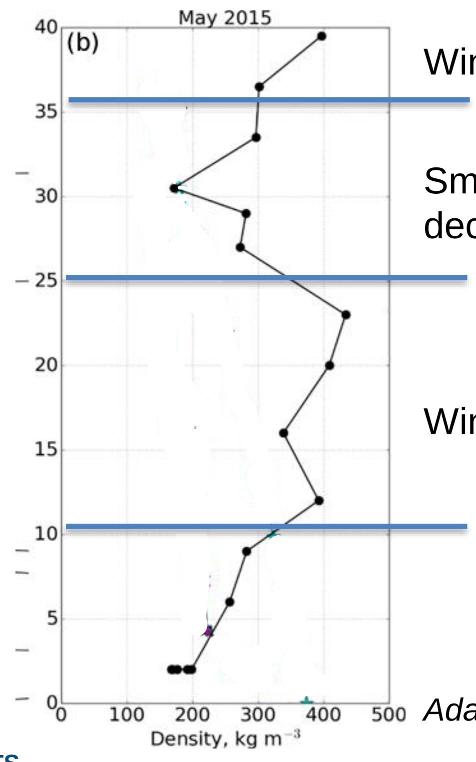
  Meltwater retained into the snowpack can refreeze at ice density (~920 kg m-3), filling air space between snow grains

- **Equi-temperature metamorphism**:
  change of shape and size of snow crystal with time. Mainly active for fresh snow.

- **Compaction induced by wind**:
  Wind can blow, redeposit and crash snow crystals, effectively increasing the packing between crystals
  → increasing density



Wind-compacted snow layer

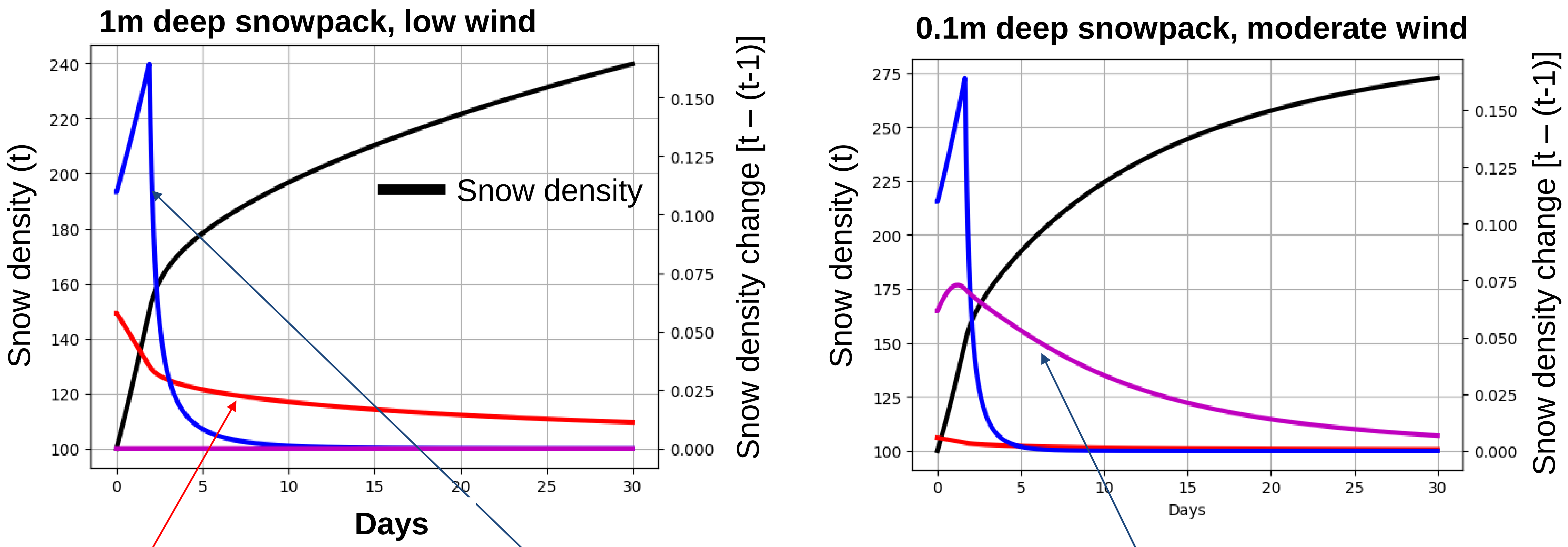
Small rounded and decomposing crystals

Wind-compacted snow layer

Adapted from Barrere et al. 2017

Snow density evolution, time-scales

Two examples of evolution with time of overburden, metamorphism and wind densification processes



overburden:
increase of density due to the snow weight

Equi-temperature metamorphism:
change of shape and size of snow crystal with time.

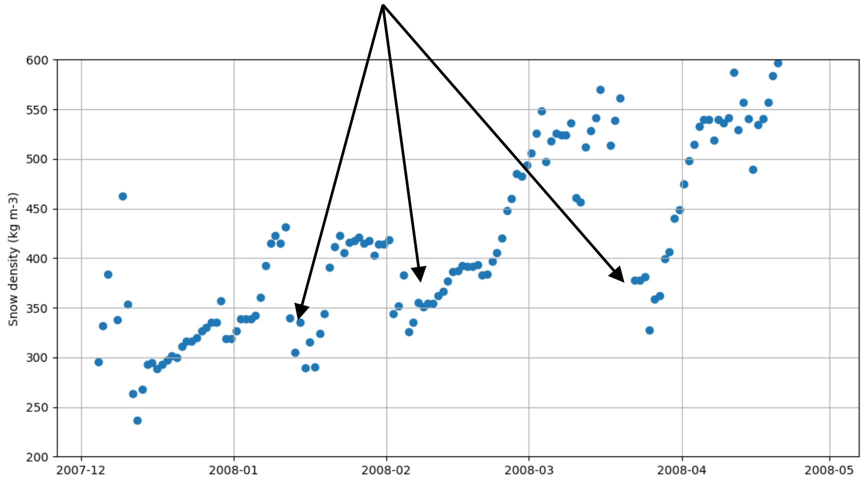
Compaction induced by wind:
effective on top snow layers (or low snow amounts)

Snow density evolution, new snow

What about new snow after a snowfall?

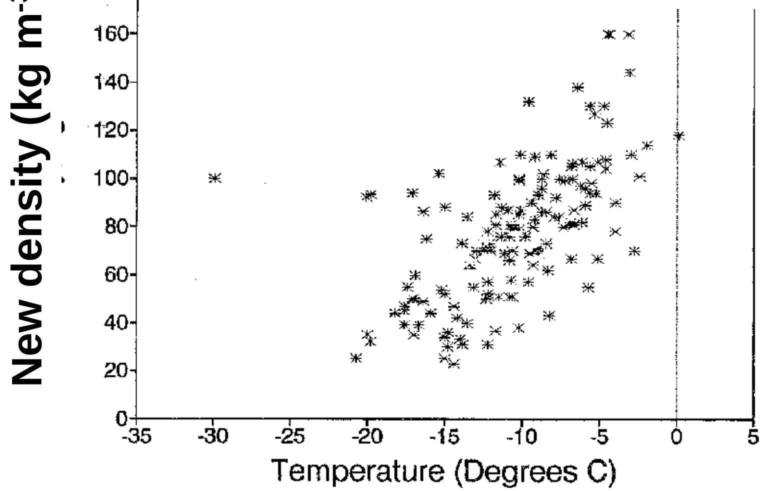
- New snow density updated after snowfall;
- Snowfall density (ρ_{new}) as a function of wind speed and air temperature.

Snowfall

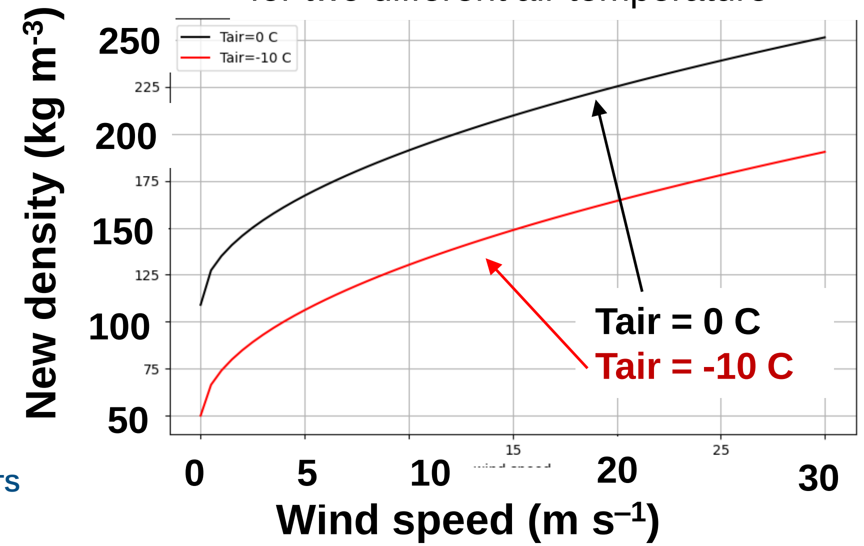


Col de Porte (French Alps)

Snow density of freshly fallen snow
And the dependency on air temperature
From observations



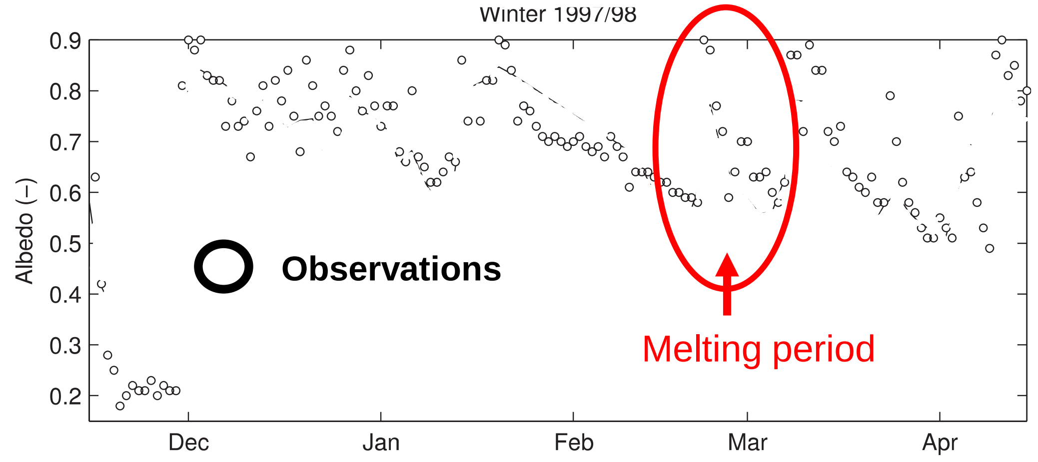
Model relationship between density
Of freshly fallen snow and wind speed,
for two different air temperature



Snow albedo evolution over open/exposed area

For snow in exposed area snow albedo is a prognostic field, computed using an empirical parametrization for snow aging

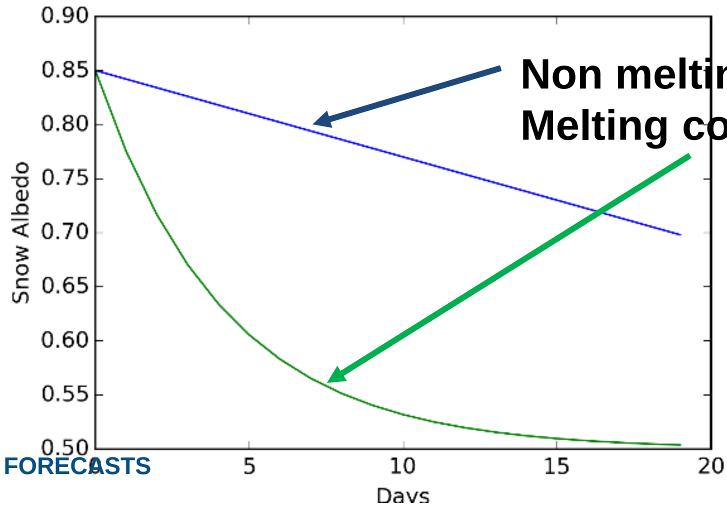
Snow albedo from Winter 97/98 at col de Porte (France)



Adapted from Dutra et al. 2012

- Snow albedo physically results from snow grain size and aerosol/soot deposition. A model of snow albedo requires a prognostic evolution of :
  - Snow grain properties
  - Aerosol deposition on the snowpack
- An empirical parametrization for snow aging is used within ECMWF snow model to capture these two processes

$$\alpha_{\mathrm{sn}}^{t+1} = \begin{cases} \alpha_{\mathrm{sn}}^{t} - \tau_a \Delta t/\tau_1, & M_{\mathrm{sn}} = 0 \\ (\alpha_{\mathrm{sn}}^{t} - \alpha_{\min}) \exp(-\tau_f \Delta t/\tau_1) + \alpha_{\min}, & M_{\mathrm{sn}} > 0 \end{cases}$$



Non melting: linear decay
Melting conditions: exponential

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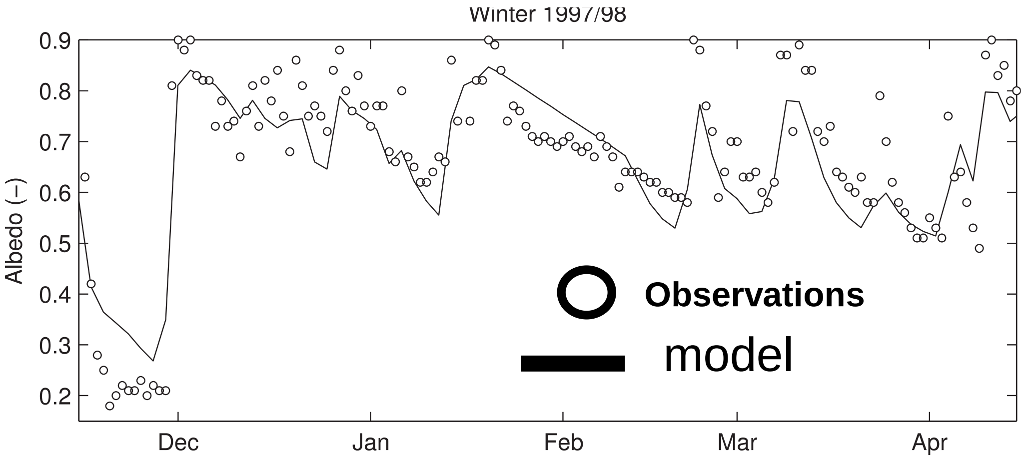

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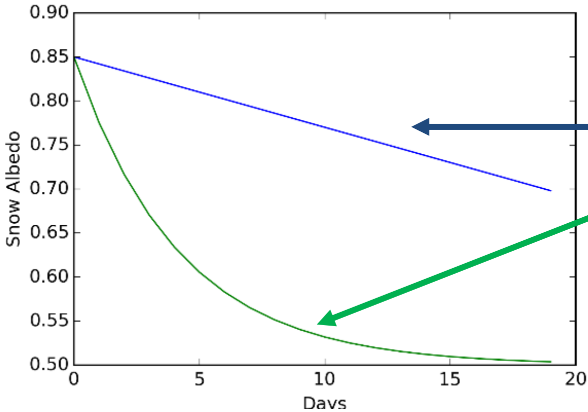
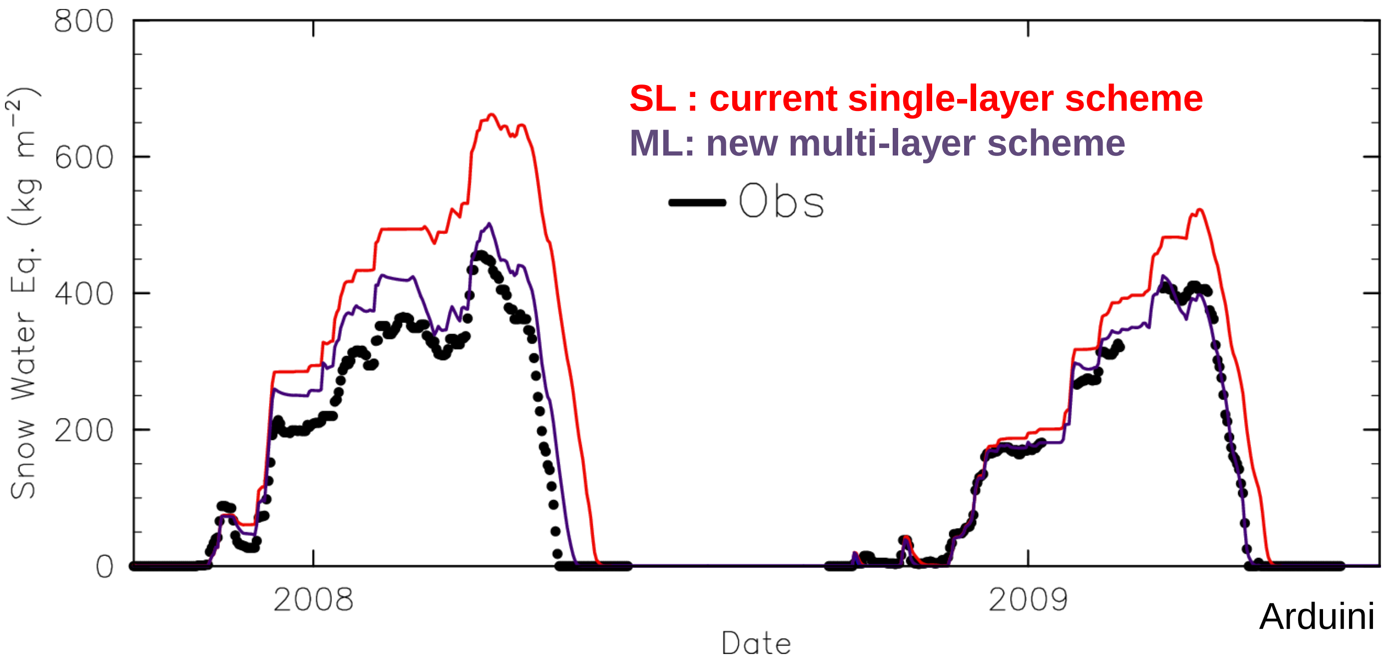


Non melting: linear decay
Melting conditions: exponential

Advantages of using multiple snow layers in NWP, snow depth



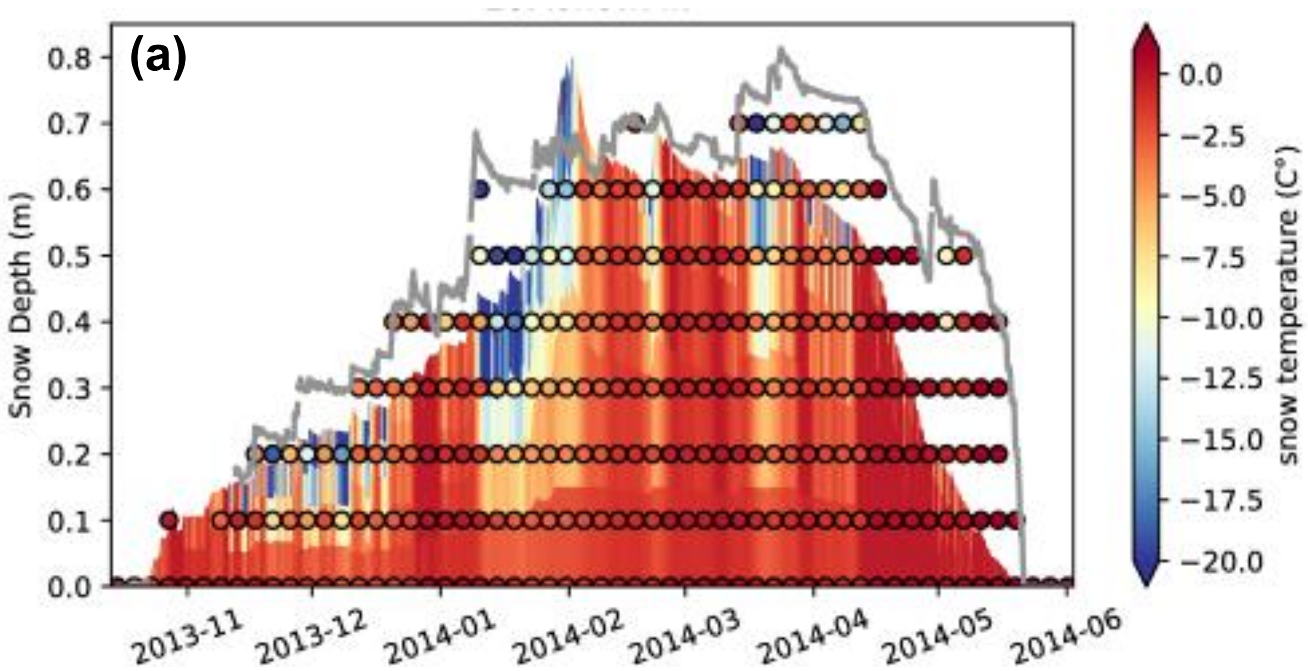
Arduini et al., JAMES, 2019

Better simulations of snow mass (and depth) using multiple vertical layers (up to 5 layers):

- Melting events during the season better represented because of lower thermal inertia of top snow layer
- Improved timing of ablation

Internal snow properties

Vertical profiles of **snow temperature** from model (contours) and observations (dots)

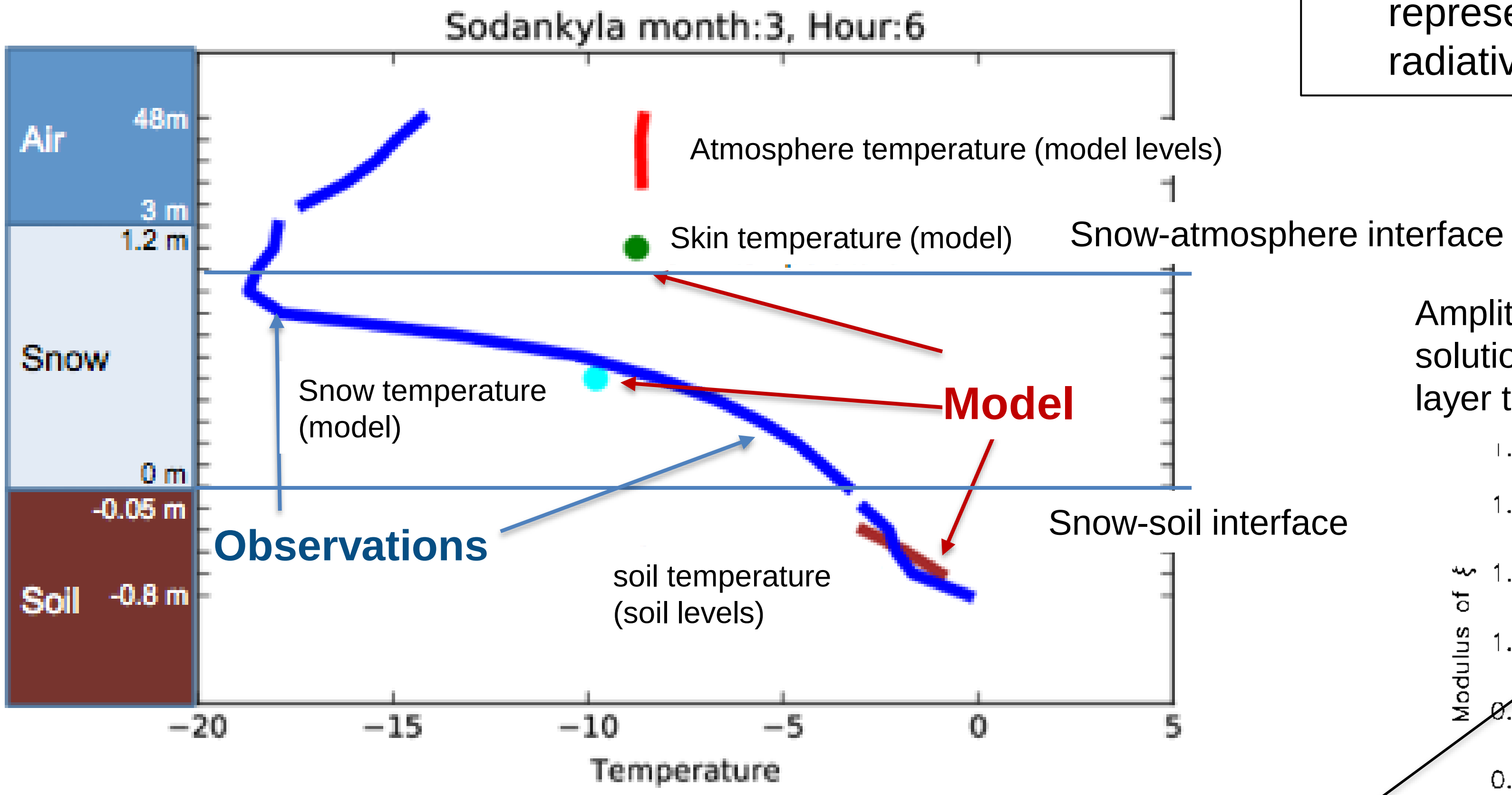


Arduini et al., JAMES, 2019

Realistic simulation of temperature gradients within the snowpack during cold spells

Why using multiple layers to represent the snowpack in NWP?

Average temperature profile atmosphere-snow-soil (March 2017) from observations and the ECMWF model

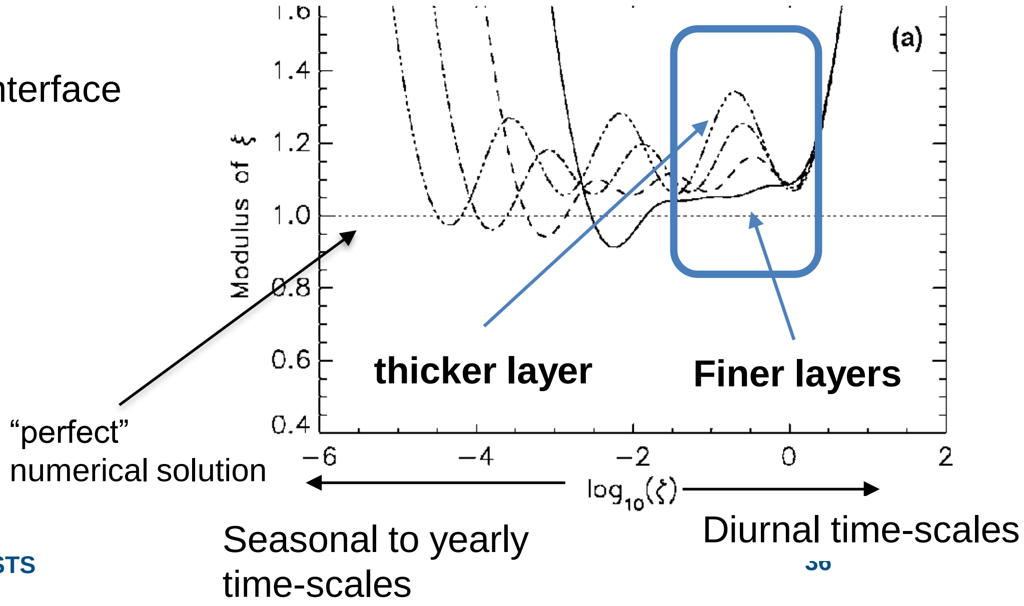


Thanks to Finnish Meteorological Institute's Arctic Research Centre (FMI-ARC) for observations and Linus Magnusson for the figure.

Using one layer to represent the snow implies that thick snow has a large thermal inertia

- ➢ Simulated skin temperature cannot represent very low temperatures during radiative cooling events

Amplitude error (Numerical wrt analytical solution) for different fixed layers but different layer thicknesses

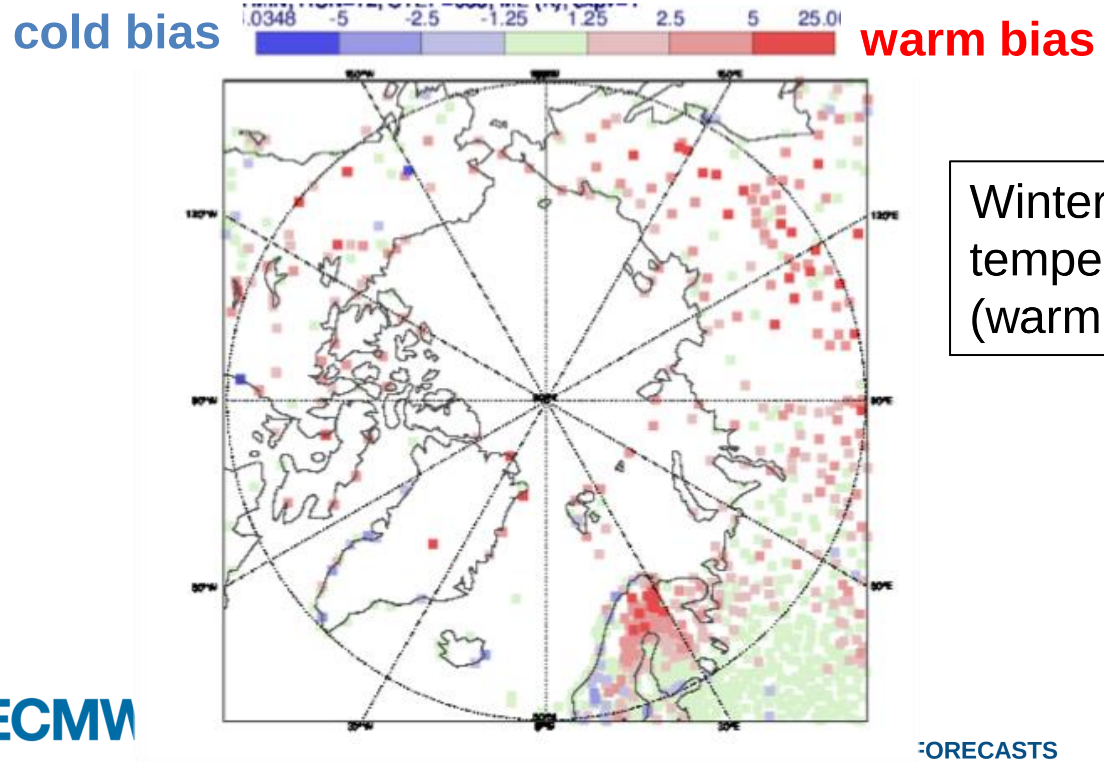


Challenges in the coupled model (Operational)

Using one vertical layer to represent the snow implies that thick snowpack have a large thermal inertia

- ➢ Simulated skin temperature cannot represent fast time scales and very low temperatures
- ➢ Direct impact on 2-metre temperature forecasts

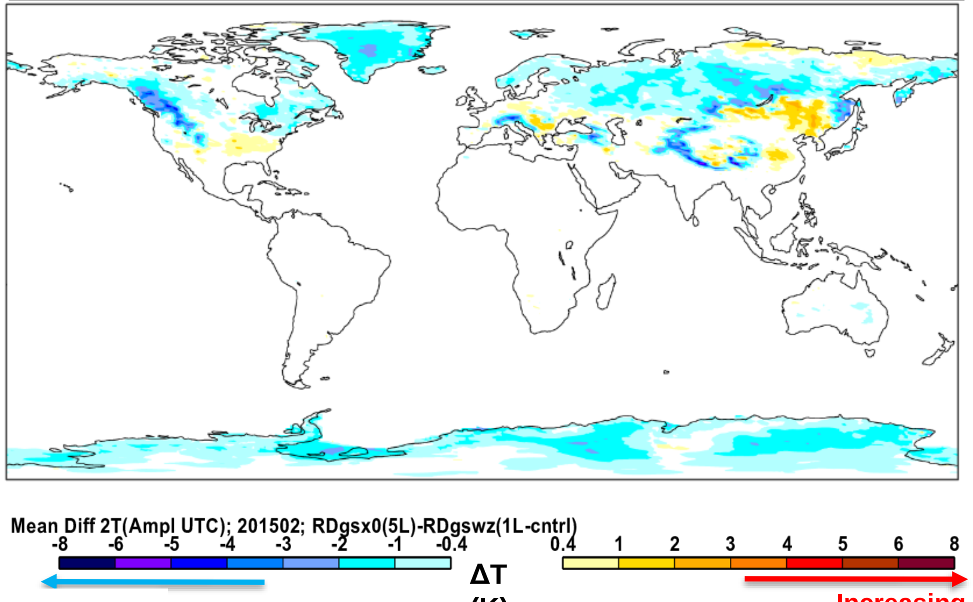
Mean error in daily minimum 2-metre temperature of operational forecasts January-March against surface observations (day 1)



Wintertime minimum 2-metre temperature is generally overestimated (warm bias) over the Arctic region

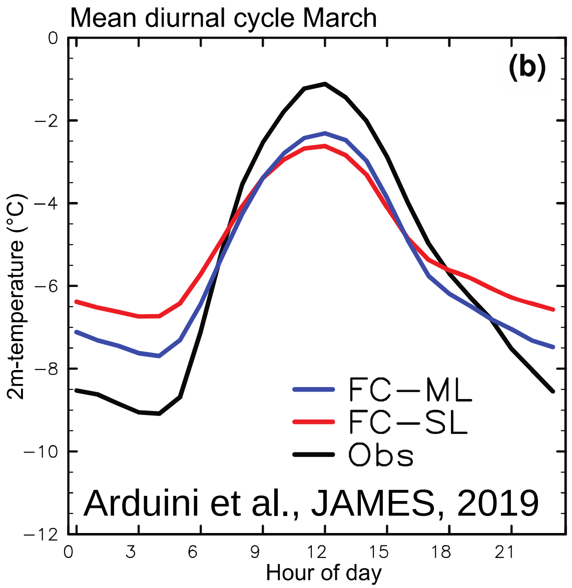
Simulations with single and multi-layer snow schemes (coupled land-atmosphere)

Difference of the minimum 2-metre temperature for Feb 2015 between **coupled** simulations performed using the **multi-layer snow** and **single-layer scheme**



1-year continuous forecast initialized in Jan 2015
Upper troposphere nudged towards ERA-I reanalysis

Mean diurnal cycle 2-metre temperature from observations and forecasts in day 2 with the single-layer and multi-layer snow schemes

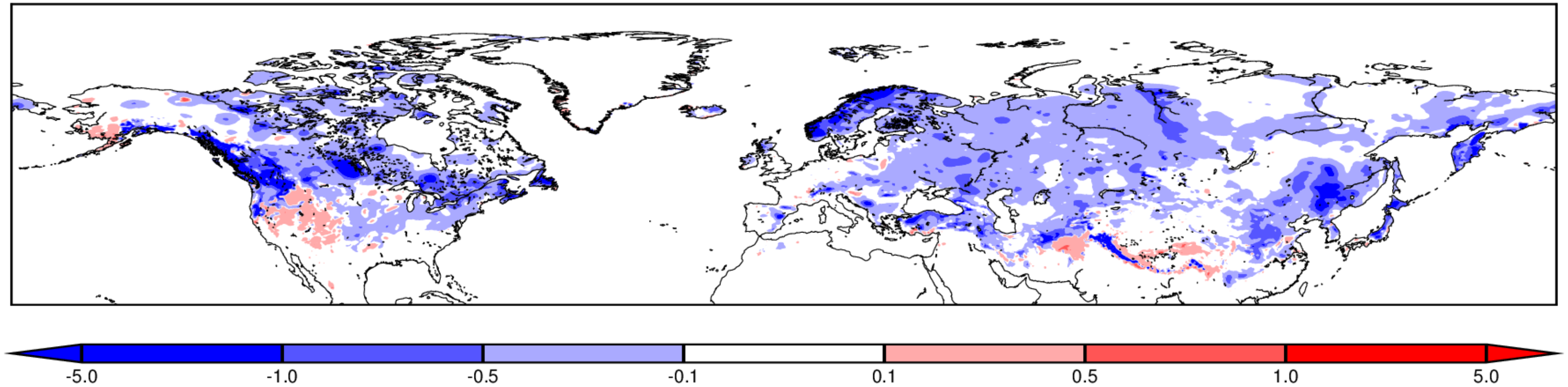


31 forecasts initialized everyday March 2017
FC lead time: t+27 to t+48
No data assimilation

Simulations with single and multi-layer snow schemes, analysis increments in data assimilation experiments

An improved simulation of the snowpack is foreseen to have a positive impact in data assimilation. By reducing errors in the first guess (FG), the "activity" of the snow analysis (as measured by the Analysis-FG increments) is also reduced.

RMSE diff in snow depth analysis increments for Jan 2020



cm

ML snow better than single-layer

Snow in NWP: snow cover fraction

The tiling approach allows to take into account sub-grid scale variability of the surface.

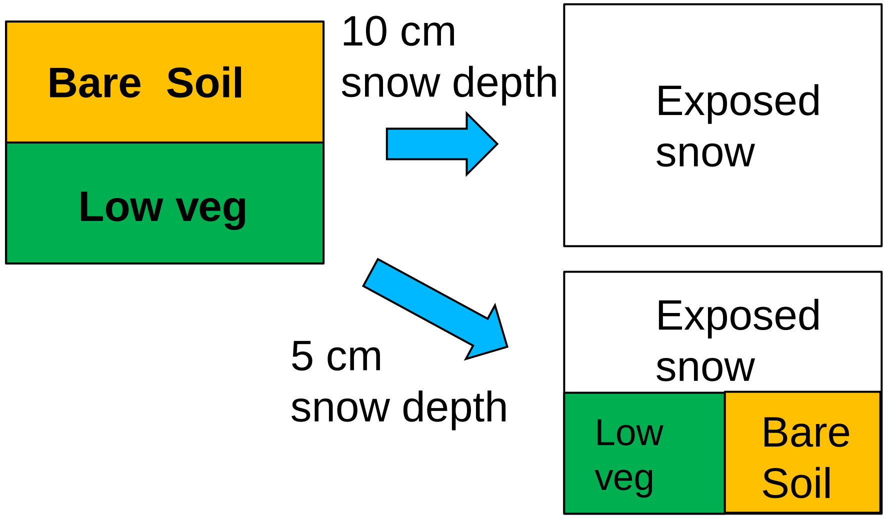
Snow cover fraction is the part of a grid cell of the model covered by snow.

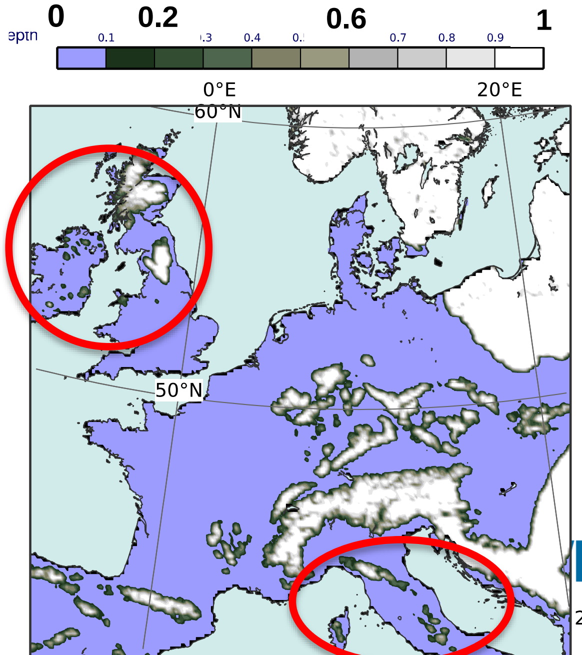
- "Dynamic" tile: it is parametrized as a function of snow mass (S) and density (ρ_{sn}):

$$c_{sn} = min(1, \frac{\boxed{S/\rho_{sn}}}{0.1})$$

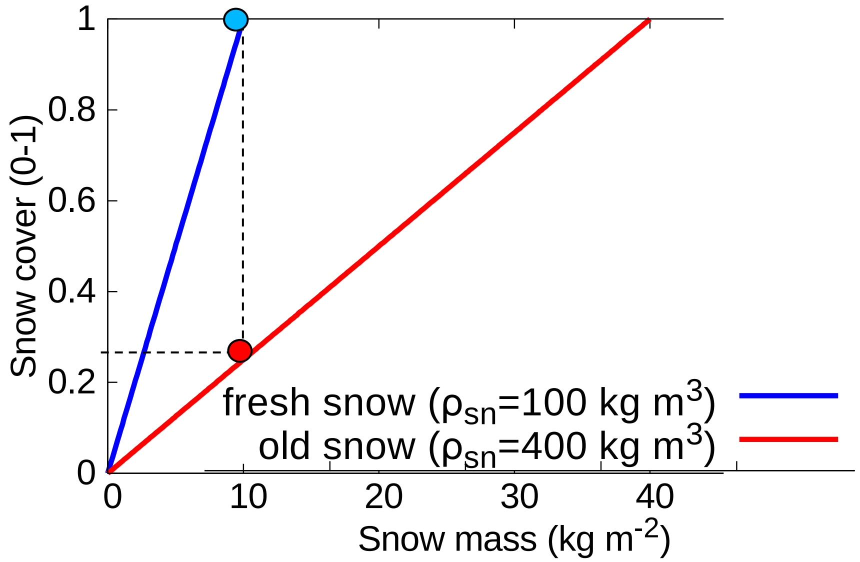
Snow depth

- **Particularly important near the snow-line or sporadic snow regions**





Snow cover "fraction" from satellite, aggregated to 9km IFS model gridbox



Summary

- Snow is a major component of the climate system. Because of its unique properties, snow impacts all forecast ranges.
- Physically-based snowpack schemes solving the energy and mass balances, are used in numerical weather predictions to simulate the space-time variability of snow and the coupling with the atmosphere.
- Multi-layer snow schemes enable to better represent temperature and density gradients within the snowpack with a direct impact on the simulation of snow and near-surface temperatures

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