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Surface-radiation interaction in Polar regions: challenges and future perspectives

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Outline

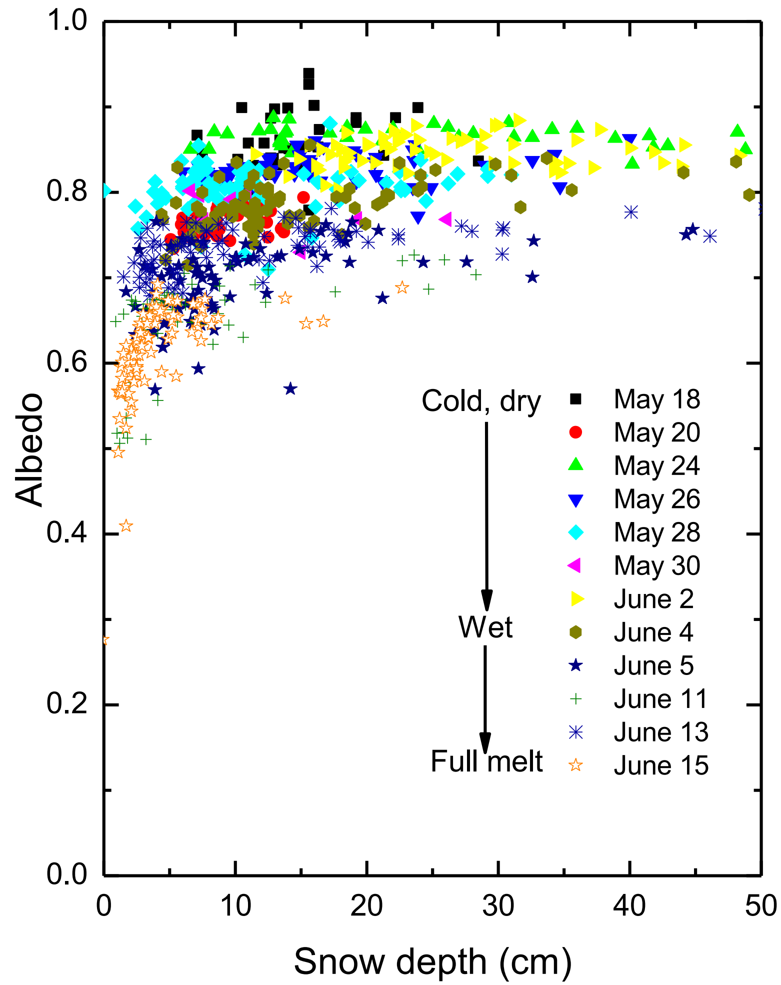
1. The nature: factors affecting the snow albedo
2. The modelling:
   - Challenges in the parameterizations of snow albedo
   - Proposed solutions (prognostic equation of snow grain size, RT in snow)
   - Expected impact in NWP
3. Again the nature: effect of snow on boreal forest albedo

Factors affecting snow albedo

- Snowpack thickness
- Snow grain size
- Snow grain shape
- Solar zenith angle
- Cloud cover fraction
- Snow impurity content
- Surface patterns

Dependence of snow albedo on snowpack thickness

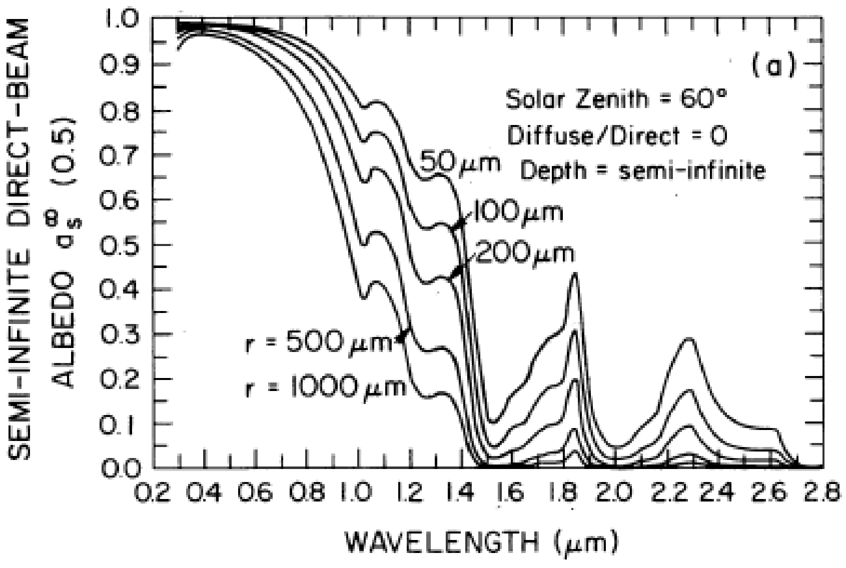
Optically thick at 5 to 10 cm



Perovich, Snow albedo workshop, 2016

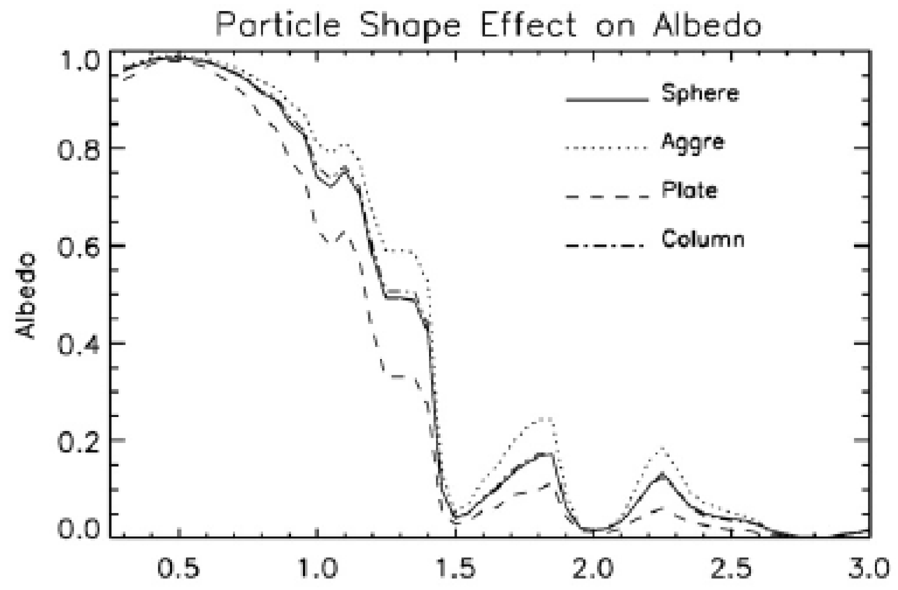
Dependence of snow albedo on:

Snow grain size



(Wiscombe and Warren, 1980)

Snow grain shape

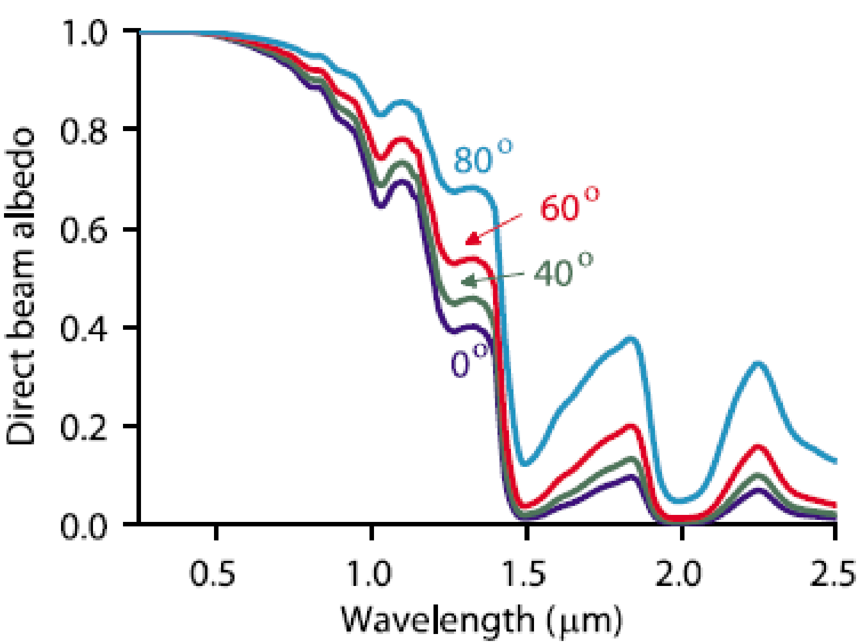


(Jin et al, 2008)

Dependence of snow albedo on solar zenith angle

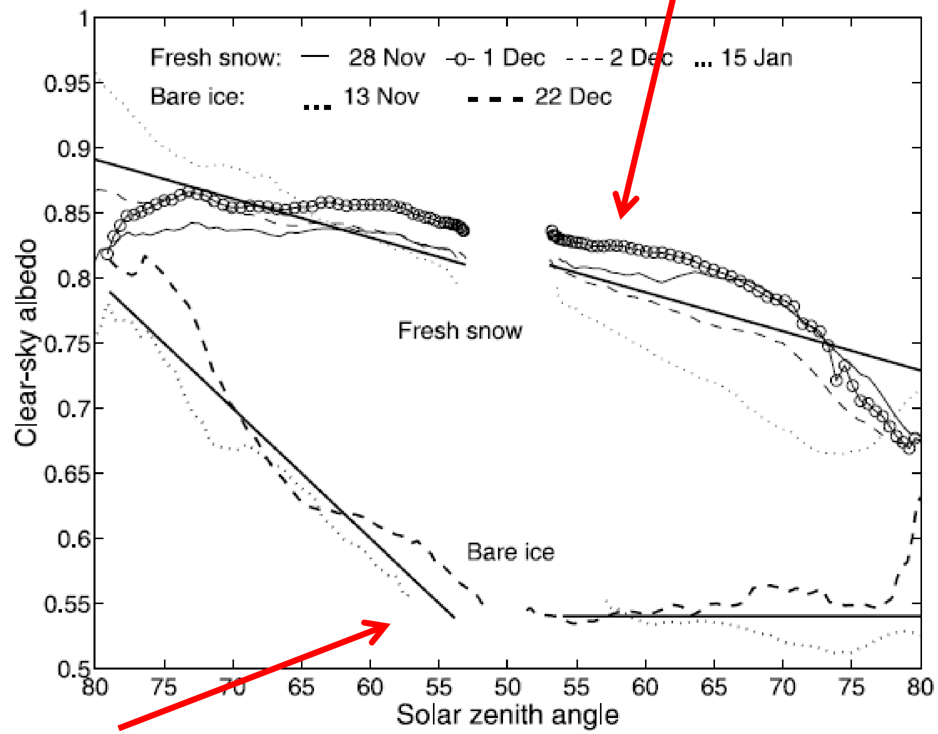
Snow metamorphism and diamond dust precipitation

Model sensitivity test



(Gardner and Sharp, 2010)

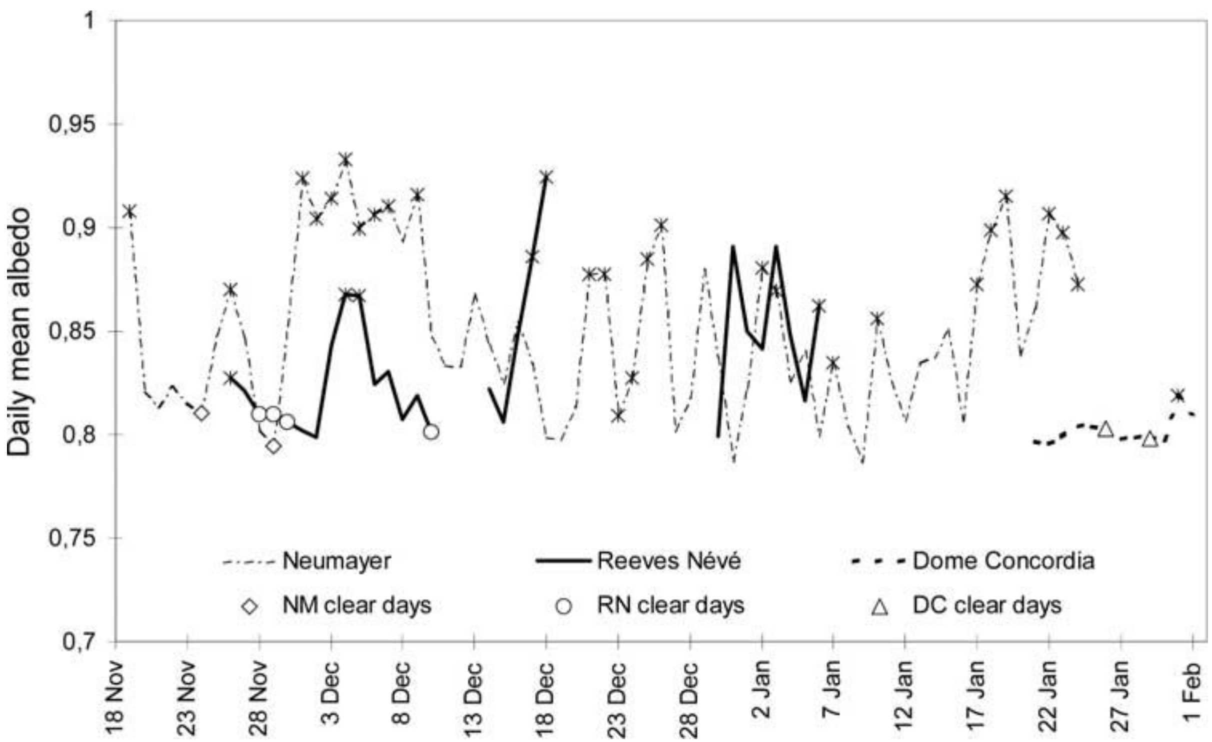
Real world…



Melt-freeze cycles and surface hoar formation

(Pirazzini, 2004)

Dependence of snow albedo on cloud cover



- Clouds convert direct radiation into diffuse radiation thus changing θ_{eff} (for diffuse radiation $\theta_{eff} \sim 50^{\circ}$)
- Clouds increase the fraction of visible light and reduce the fraction of near-infrared light. Thus, **clouds increase the spectrally integrated albedo**

Clouds are the main drivers of the albedo variability over the Antarctic continent (Pirazzini 2004; Van den Broeke et al. 2006)

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Dependence of albedo on impurity content

Impurity parameters which affect snow albedo:

- **Concentration**
- **Composition (spectral imaginary refractive index)**
- **Impurity particle size**

➢ Impurities decrease the snow albedo in the visible but have no effect in the near-infrared region, where snow strongly absorbs infrared radiation and mask the impurity absorption.

➢ For fixed concentration, smaller impurity particles are more effective in lowering the albedo than larger ones

(Warren and Wiscombe, 1980)

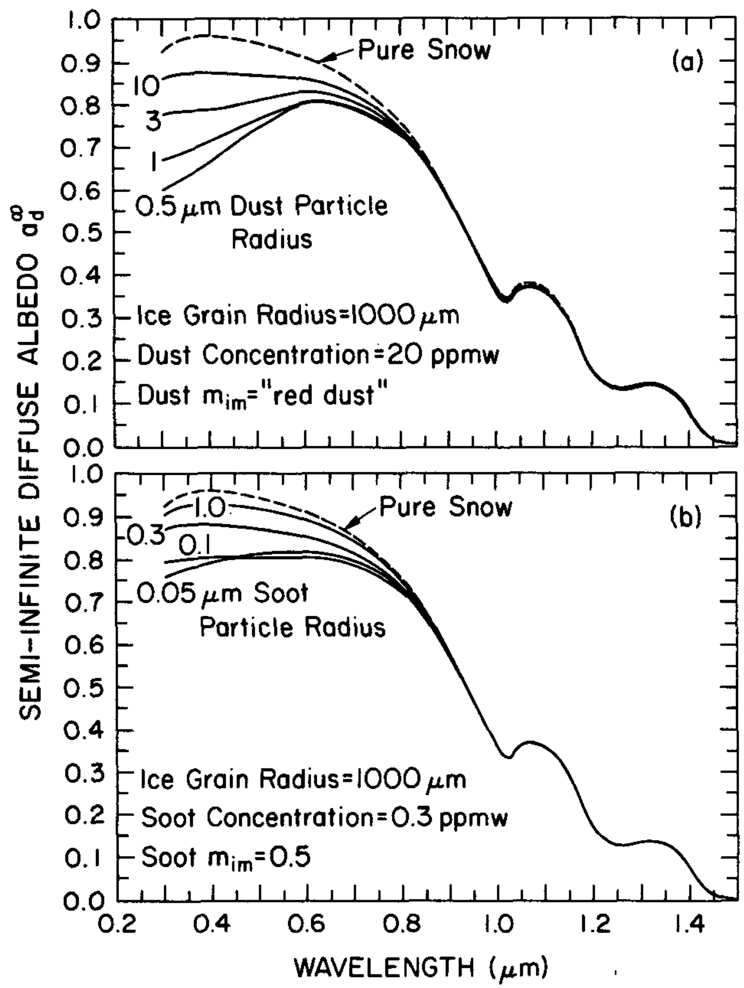
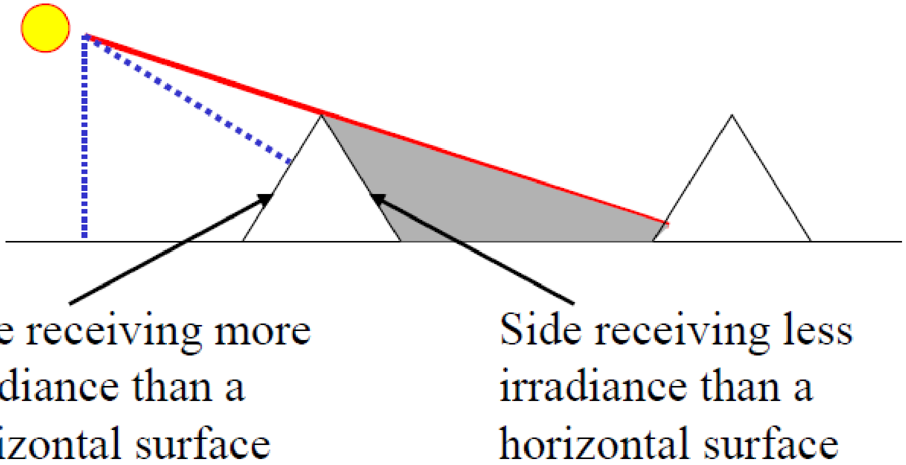


FIG. 4. Effect of (a) dust or (b) soot particle size on snow albedo a_d^∞.

Dependence of snow albedo on surface patterns



- The average of the incident angles, weighted with the amount of incoming flux, is lower than the solar zenith angle, thus **the albedo is lower**.
- Radiation reflected from one side may be absorbed by the facing side of the neighbouring feature ("light-trapping"), **reducing the albedo**.

⇨ **The daily albedo reduction due to sastrugi is ~2-4%.**

Snow albedo parameterization in climate and NWP models

Albedo is expressed as a function of the bulk snow/ice properties that are considered to be more related to its variability:

- **Surface temperature**
- **Snow age**
- **Snow density**
- **Snow/ice depth**

They are somewhat correlated to snow grain metamorphysm

Difficulties encountered when parameterizing snow/ice albedo:

- Parameters that are found to be significantly affecting the daily mean snow albedo (using a 95% confidence level) are different in different sites. For example:

	Surface temperature	Cloud cover fraction	Wind speed	Snow depth	Positive Accumulated degrees days	Dummy snowfall
Svalbard	X			X	X	X
French Alps (1340m)	X	X		X	X	X
Russian Stations (48.5N-57.8N)	X	X		X	X	X
Barrow, Alaska	X		X	X		

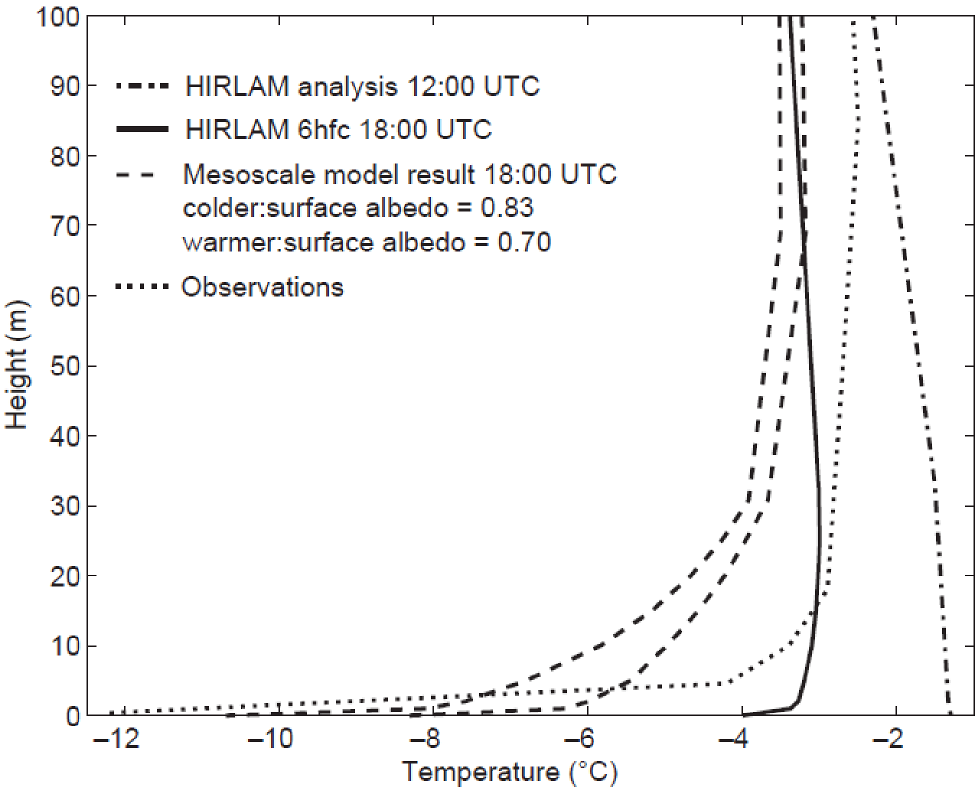
- Because of the differences in the local weather regimes, surface types, annual precipitation, different parameterizations work better in different seasons, different latitudes, and different years.

Effect of snow albedo bias on the forecasted near surface temperature

A 10% error in the springtime snow albedo can cause a bias of several Kelvins in the 6-hour forecast of 2m temperature (Pirazzini et al, 2002):

23.03.1999, over sea ice in the Bay of Bothnia, in the cloud-free evening:

- **Observed α=0.83**
- **HIRLAM α=0.7** →6h forecast of surface temperature was **8.2 K too warm.**
- **2D mesoscale model** (with better snow and ice thermodynamic than HIRLAM) **α=0.7** →6h forecast of surface temperature was **4 K too warm.**
- **2D mesoscale model α=0.83** →6h forecast of surface temperature was **1.5 K too warm**



Prognostic equations for snow microstructure metamorphism (grain size)

Coupled to optical radiative transfer models

- SNICAR (Flanner and Zender, 2006)

Coupled to microwave emission models

- SNTHERM (Jordan, 1991)
- MOSES (Essery et al., 2001)

Simulations against SSA derived grain sizes (Sandells et al., 2017):

- SNTHERM simulated snow grains are too large (mean error 0.12 to 0.16 mm),
- MOSES simulated grains are too small (mean error −0.16 to−0.24mm)
- SNICAR simulate grains are too small (mean error−0.14 to−0.18 mm)

SNTERM (Jordan, 1991)

The growth of snow grain diameter d is based on the rate of vapour transport through the snow (and therefore temperature gradient).

Dry snow: $\frac{\partial d}{\partial t} = \frac{g_1}{d} D_{eos} \left(\frac{1000}{P_a}\right) \left(\frac{T_s}{T_m}\right)^6 C_{kT_s} \left|\frac{\partial T_s}{\partial z}\right|$

d = snow grain diameter
P_a = **atmospheric pressure**
T_s = **snow temperature**
C_{kTs}= variation of saturation vapor pressure with snow temperature T_s
T_m = 273.15 K
$\frac{\partial T_s}{\partial z}$ is the **temperature gradient**

Grain growth under wet conditions is more rapid, with empirical constant g2 and is dependent on the **liquid fractional volume** f_1 by

Wet snow:

$$\frac{\partial d}{\partial t} = \frac{g_2}{d}(f_1+0.05) \quad f1 < 0.09$$

$$\frac{\partial d}{\partial t} = \frac{g_2}{d}(0.14) \quad f1 \geq 0.09$$

MOSES (Essery et al., 2001)

$$r(t+\Delta t) = \left[r(t)^2 + \frac{G_r}{\pi}\Delta t\right]^{1/2} - [r(t) - r_0]\frac{S_f \Delta t}{d_0}$$

Gr = empirical **temperature-dependent grain area growth rate**

Sf = **snowfall rate** in Δt

SNow, ICe, and Aerosol Radiative model (SNICAR) (Flanner and Zender, 2006)

Applied in the version 4 of the Community Land Model (CLM4) part of the Community Earth System Model CESM-UCAR (Oleson et al., 2010)

Contributions to grain size evolution are from:

1. vapor redistribution (dry snow): SNICAR (Flanner and Zender, 2006)
2. liquid water redistribution (wet snow): Brun (1989)
3. re-freezing of liquid water (constant $r_{e,rfrz}$ = 1000 μm)

The equation for r_e at time step t is (Flanner and Zender, 2006; Oleson et al., 2010)

$$r_e = \left[r_e(t-1) + dr_{e,dry} + dr_{e,wet}\right] f_{old} + r_{e,0} f_{new} + r_{e,rfrz} f_{rfrz}$$

f_{old} = fraction of old snow
f_{new} = fraction of freshly fallen snow
f_{rfrz} = fraction of refrozen liquid water
r_{e0} = 54.5 μm

SNICAR

Dry snow : The method is to retrieve 3 best-bit parameters (τ, k, $\left(\frac{dr}{dt}\right)_0$) from a look-up table as a function of **snow temperature**, **temperature gradient**, and **density**. The look-up table is derived from the microphysical model described in Flanner and Zender (2006) :

$$\frac{\partial r_{dry}}{\partial t} = \left(\frac{\partial r}{\partial t}\right)_0 \left(\frac{\tau}{dr_{fresh}+\tau}\right)^{(1/\kappa)}$$

r = effective radius

τ, k = best fit parameters

$\left(\frac{dr}{dt}\right)_0$ = initial rate of change of *r*

$dr_{fresh} = r_{current} - r_{fresh}$

Computationally efficient approximation to a model based on physics!

Wet snow: Apply the grain growth function from Brun (1989):

$$\frac{\partial r_{wet}}{\partial t} = \frac{C_1 f_{liq}^{\;3}}{4\pi r^2}$$

f_{liq} = **mass fraction of liquid water** in the snowpack

Snow density and its coupling to grain size

At constant grain sizes, there is very little dependency of albedo on snow density alone (Kaemper et al., 2007)

However,

- Radiative transfer and diffusion of heat into the snowpack affect both snow grain metamorphism and density, thus, in a physically based model, the two parameterizations should be coupled.
- Snow density affect the penetration depth of light: in a snowpack that is vertically inhomogeneous with respect to grain size, albedo depends on snow density.
- Parameterized snow density is crucial in the simulation of thermal conductivity and snow mass: direct effect on snow thickness, snow cover fraction, ALBEDO

How is snow density parameterized?

Snow density parameterizations

- Constant value: **old HIRLAM, old Noah** (**old WRF**)
- Prognostic equation function of snow age, constant in depth: **ISBA-FR**, **old HTESSEL**, **old Noah, HIRLAM 6.4.2, HIRLAM_newsnow**
- Prognostic equation based on Anderson 1976 (function of SWE, T_s, density of new snowfall): **CLM3**, **Noah-MP** (**WRF 3.4**), **ECMWF** (**new HTESSEL**), **ISBA-ES**, **SNTHERM**

➢ Compared to observations, **more consistent results are obtained from those snow schemes that have prognostic representation of snow density** (Essery et al., 2012; Dutra et al., 2012)

RT in snow

A prognostic equation of snow effective grain size allows the implementation of a RT scheme in the snow:

- SNICAR (Flanner and Zender, 2006)
- TARTES (Libois, 2013)

They express **albedo** and the **absorption of radiation** in snow as a function of the **snow single scattering properties**.

The generally used approach to model light scattering in snow is to consider its grains as **independent scatterers** of different shapes (spheres, hexagonal prisms, cylinders, fractals, etc.) as a “thick ice cloud on ground”.

> **HYPOTHESIS**: *a collection of scatterers with the same r_e as the true snow grain population possesses the same scattering properties as the snow*

Different solutions can be used to compute the **single scattering** by a small volume element containing a representative size distribution of snow grains, and then the **multiple scattering** can be calculated.

Snow single scattering properties

Each layer of the snowpack is characterized by the **phase function**, **extinction coefficient σ_e** and **single-scattering albedo ω** which are determined from **snow grain size and shape** (extinction also from **snow density**).
Scattering occurs each time a photon reaches the grain surface while absorption is due to the photon path within the grain.

Complex index of refraction $\boldsymbol{m} = m_{re} + i m_{im}$

EXTINCTION COEFFICIENT: $\boldsymbol{\sigma_e}\left(\frac{2\pi r_{eff}^2}{\lambda}, shape, m, density\right) = \sigma_s + \sigma_a$

describes the extent to which the radiant flux of a beam is reduced as it passes through the snow crystal [m^{-1}]

SINGLE-SCATTERING ALBEDO: $\boldsymbol{\omega} = \frac{\sigma_s}{\sigma_e}$ $0 \le \omega \le 1$

A measure of the relative strength of scattering vs. absorption

PHASE FUNCTION: P gives the angular distribution of the scattered light, i.e. the probability of scattering in any given direction. It depends on grain shape

ASYMMETRY PARAMETER: $\boldsymbol{g} = \langle cos\theta_s \rangle$ $-1 \le g \le 1$

It is a measure of the P asymmetry:

g = 1 : complete forward scattering

g = 0 : as much scattering forward and backward

Snow single scattering properties

The parameterization of snow single scattering properties allows a physically consistent treatment of pure and dirty snow:

Pure snow Dirty snow

ABSORPTION COEFFICIENT: $\sigma_a = n_{snow} C_a + \sum_i n_i C_a^i$

$n = \frac{\rho_{snow}}{\rho_{ice} V}$
Number density of snow grains

$n = \frac{\rho_i}{\rho_i^0 \rho_i V_i}$
Number density of impurity i particles

Absorption cross section of the i particles

EXTINCTION COEFFICIENT and **ASYMMETRY PARAMETER**: effect of impurities generally negligible.

SNICAR

In SNICAR applied to CLM4, the single-scattering properties of snow **are tabulated** for five spectral intervals (0.2–0.7, 0.7–1.0, 1.0–1.2, 1.2–1.5, and 1.5–5.0 μm) and for 1471 snow grain r_e ranging from 30 to 1500 μm, separately for direct and diffuse solar radiation.

Calculations depend on the **assumed grain shape**. Different tables corresponding to different shapes can be calculated (as in Räisänen et al., 2017)

Look-up tables → computationally efficient!

Multilayered snowpack: RT in snow is computed using a two-stream approximation for multiple scattering. The delta-Eddington approximation (Joseph et al., 1976) is applied in the visible region and the delta-hemispheric mean approximation (Toon et al.,1989) in the near-infrared bands.

Two-streAm Radiative TransfEr in Snow (TARTES) (Libois et al. 2013)

TARTES applies the asymptotic analytical radiative transfer theory presented in Kokhanovsky (2004):

Geometric optics approximation ($r_{eff} >> \lambda$)
Large optical depth
Weak absorption

The solution depends on grain shape through two parameters:

Absorption enhancement parameter **B** → f(grain shape)

Absorption coefficient:
$\sigma_a = n\, \boldsymbol{B}\, k_a\, V$

Asymmetry parameter $\boldsymbol{g} \cong \frac{1}{2}(g^G + 1)$ g^G→ f(grain shape, m_{re})

In the **spectral range 300–1350nm**:

B and g^G are nearly constant → independent of wavelength (Kokhanovsky and Zege, 2004)

TARTES (Libois et al. 2013)

For each snow layer, albedo and flux extinction coefficient are calculated from the asymptotic RT theory.

Multilayered snowpack: delta-Eddington approximation to solve the two-stream radiative transfer equation. The solution applies the albedo and flux extinction coefficient obtained for each layer.

TARTES is currently implemented in Crocus (snow physical model that is modularly included in SURFEX → HARMONIE)

Implications for NWP models

- Multilayered snow models and multiband RT schemes are needed to treat the penetration of light into the snowpack

  Five snow layers and more than three spectral bands in both the visible and near-infrared spectra are necessary to accurately simulate albedo and solar heating profile in the snowpack (Aoki et al. 2011).

- The use of a full radiative transfer model (SNICAR or TARTES) embedded within the snowpack model allows:

  - the coupling between radiative and thermal transfer into the snow. Micro-scale properties (grain size), bulk properties (density, temperature) and albedo become physically consistent.

  - the assimilation of the satellite reflectance data to constrain the snow properties, as demonstrated by Charrois et al. (2016)

Expected impact: improvement in the forecast of snow albedo, snow depth, snow water equivalent → improvement in the forecast of 2m temperature over snow
→ improvement in the forecast of hydrological parameters

Effect of snow on boreal forest albedo

Snow cover fraction (SCF) determines the contribution of snow cover and snow free albedo to the grid mean albedo:

$$\boldsymbol{\alpha_{grid-mean} = SCF \cdot \alpha_{snow-covered} + (1 - SCF) \cdot \alpha_{snow-free}}$$

What is $\boldsymbol{\alpha_{snow-covered}}$ in the presence of a forest?

Effect of snow on boreal forest albedo

- Snow-free coniferous forest albedo ~12 %
- Pure snow albedo in midwinter ~85%
- In snow covered conditions the albedo of forested area varies with the density of the forest and the snow properties. Albedo ~22% (LAI 2.5)
- In a case of crown snow-load ('tykky') even the structure of the canopy changes and snow dominates completely

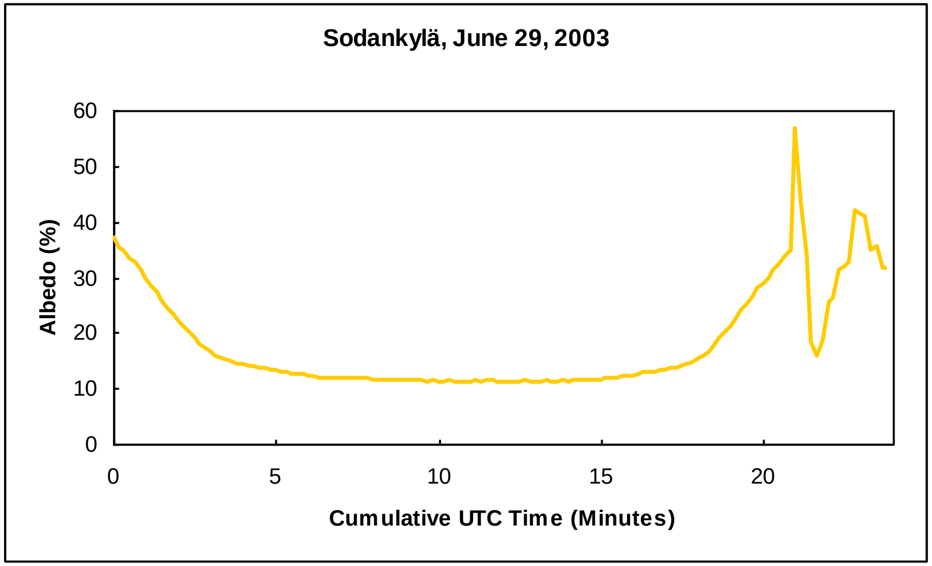
Wikimedia commons by **Muu-karhu**

Effect of snow on forest albedo

- Diurnal variation:
  - Without snow on the forest floor the midday albedo is smallest
  - With snow on the forest floor the midday albedo is a local maximum
  - The relative error of not taking the snow into account can be ~40%

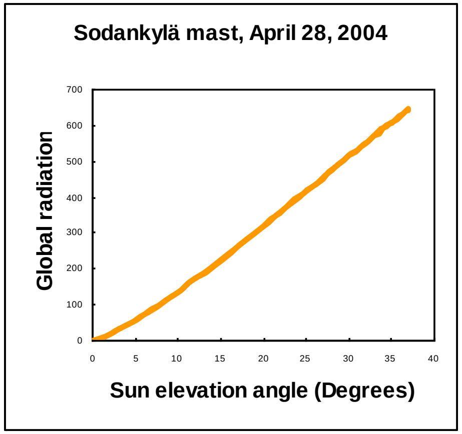
- Snow on the forest floor:
  - The snow on the forest floor causes single scattering of the incident light without canopy interaction and multiple scattering between canopy and forest floor.

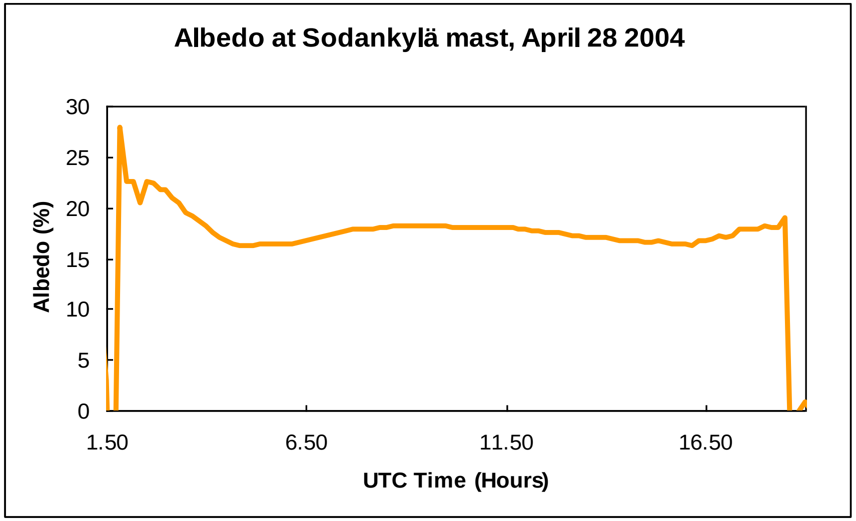
Diurnal variation of boreal forest albedo in summer



Observations from a 45 m mast in Sodankylä, Finnish Lapland

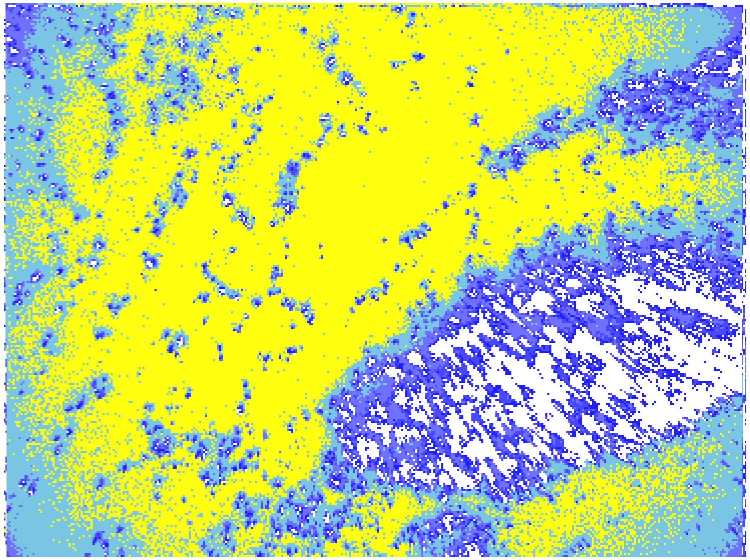
Diurnal variation of boreal forest albedo in winter





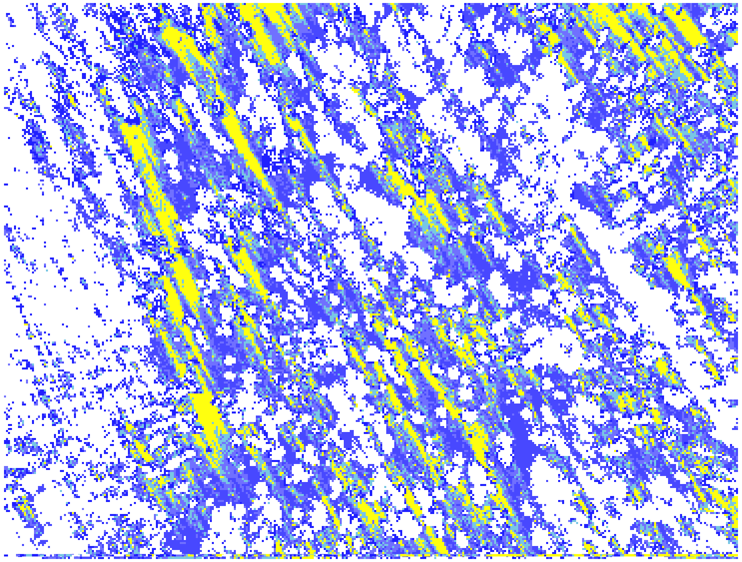
Unsupervised classification of the forest floor, cloudy sky

- 8 classes: three for canopy (white), 5 for forest floor
- Highest snow reflectance class
- Lowest snow reflectance class



Unsupervised classification of the forest floor, clear sky

- 8 classes: three for canopy (white), 5 for forest floor
- Highest snow reflectance class
- Lowest snow reflectance class



Black sky forest albedo

Only forest floor scattering α_{tt}

Only canopy scattering α_s

Multiple canopy and forest floor scattering, last reflection from the floor α_{st}

Multiple canopy and forest floor scattering, last reflection from the canopy α_{ss}

$$\alpha = \alpha_{tt} + \alpha_s + \alpha_{st} + \alpha_{ss}$$

$$\alpha_{tt} = k\alpha_b t_0^2 + (1-k)\alpha_b t_0 t_1$$

$$\alpha_s = q(1-t_0)\frac{\omega_L - p\omega_L}{1-p\omega_L}$$

ω_L = leaf single scattering albedo
p= photon's recollision probability

$$\alpha_{st} = \left[\alpha_b(1-q_b)\left(kt_0(1-t_0)+(1-k)t_0(1-t_1)\right)\right]\frac{\omega_L - p\omega_L}{1-p\omega_L}$$

$$\alpha_{ss} = \left\{\alpha_b\left[(1-q)(1-t_0)+q_b\alpha_b t_0\left(k(1-t_0)+(1-k)(1-t_1)\right)\right]\frac{t_1(1-\omega_L+q_b\omega_L-q_b p\omega_L)+(1-q_b)(\omega_L-p\omega_L)}{1-p\omega_L-q_b\alpha_b(1-t_1)(\omega_L-p\omega_L)}\right\} \cdot \frac{\omega_L - p\omega_L}{1-p\omega_L}$$

(Manninen and Stenberg, 2009)

Thank you for your attention!

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