Modelling of the Snow Cover

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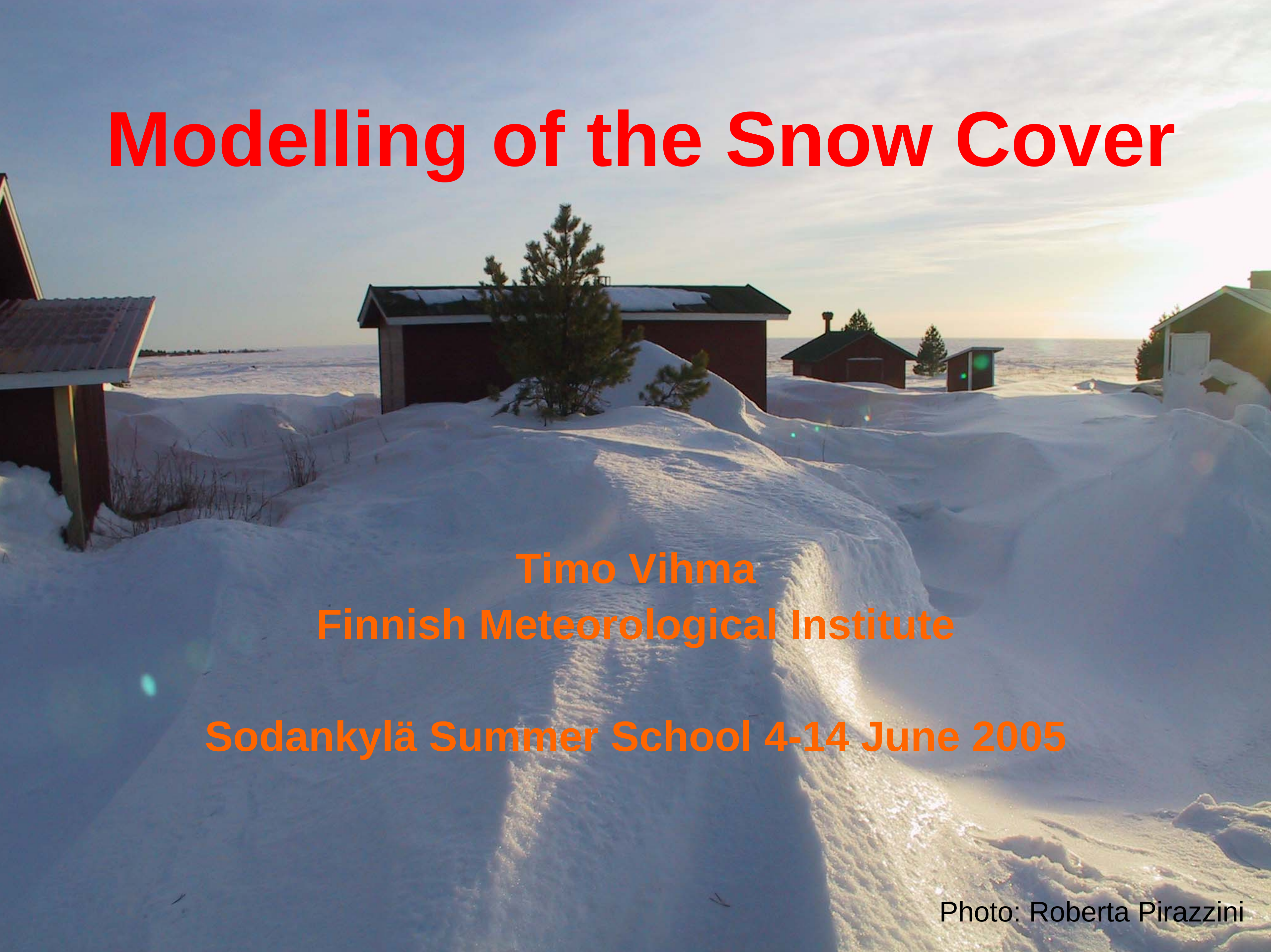


Photo: Roberta Pirazzini

Why to model the snow cover?

- Climate: snow cover is important due to its high albedo, strong insulation capability, and seasonal water storage
- Numerical weather prediction: Ts needs to be modelled
- Hydrological modelling: distribution of snow water equivivalent and timing of snow melt essential
- Avalanche forecasting: snow layering and visco-plastic properties essential
- Road weather service: snow and ice thermodynamics on roads
- Boreal botanics, zoology, and ecology: snow structure and layering
- Glasiology: firnification
- Oceanography: snow on sea ice

Properties of the snow cover

- Snow = ice crystals and bonds (= ice skeleton) + humid air + liquid water
- Evolution affected by both thermodynamics and dynamics (snow drift)
- Metamorphosis of snow crystals
- High and highly variable albedo
- Low and highly variable heat conductivity
- Variable thickness and density
- Part of the solar radiation penetrates into the snow pack
- Highly variable extinction coefficient
- Max $T_s = 0°C$
- Horizontal heterogeneity

A basic problem in snow modelling is that the properties of snow cover essential for thermodynamics change a lot in space and time

Snow density $\rho = f(\rho_i, \rho_w, \rho_a, V_i, V_w, V_a) \approx 100 – 400$ kg m^{-3}

Thermal conductivity $k = f(\rho_i, \rho_w, \rho_a, V_i, V_w, V_a, T_i, T_w, T_a$, grain structure and bonding$) \approx 0.1 – 0.4$ W m^{-1} K^{-1}

Extinction coefficient $\kappa = f(\rho_i, \rho_w, \rho_a, V_i, V_w, V_a$, grain size, grain shape, impurities, wavelength, incidence angle$) \approx 5 – 50$ m^{-1}

The temporal changes are due to:

- New snowfall $\rightarrow$ increasing thickness, albedo, and extinction coefficient, decreasing density and heat conductivity (exceptions occur)
- Snow melt and sublimation $\rightarrow$ decreasing thickness, albedo, and extinction coefficient, increasing density and heat conductivity (exceptions occur)
- Snow metamorphosis

Snow metamorphosis (e.g. Rasmus, 2005)

Dry snow metamorphosis

Small $\partial T/\partial z$ through the snow pack (< 5 K/m): equilibrium growth metamorphosis:

Water vapour pressure varies between the differently shaped parts: higher in the convex grain boundaries, lower in the concave air pores $\rightarrow$ sublimation from the convex grain surfaces and transport of vapour to the concave parts between the the grains: originally stellar particles $\rightarrow$ rounded grains

Large $\partial T/\partial z$ (> 10-20 K/m): kinetic growth metamorphosis:

water vapour sublimats from the bottom layers of the snow pack, is transported upward, and recrystallizes on the bottom surfaces of ice crystals in the upper snow layers. The extensions of these grains grow downwards $\rightarrow$ hexagonal crystals.

Wet snow metamorphosis

Wet snow is macroscopically isothermal, but temperature variations occur in the grain scale

LWC 8 %

Pendular regime: air occupies continuous paths throughout the pore space, and snow grains are well bonded together

Funicular regime: liquid occupies continuous paths throughout the pore space, and snow is cohesionless

Melting temperature depends on the radius of curvature: sharp parts melt while concave parts freeze $\rightarrow$ flux of water is rounding the crystals.

Process is very fast compared to dry snow metamorphosis (hours instead of days)

Surface or sub-surface melt $\rightarrow$ downward percolation of meltwater $\rightarrow$ refreezing to hard layers of large crystals.

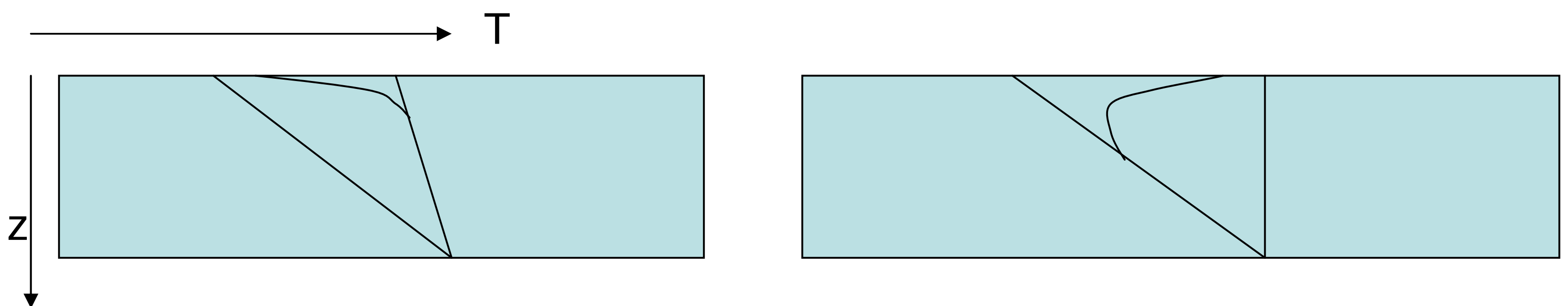
Models of different complexity

Force Restore Method (Bhumralkar, 1975; Deardorff, 1978)

$$\frac{\partial T_S}{\partial t} = \frac{(1-\alpha)SWR + LWR_{NET} + H + LE}{c_S} - \frac{T_S - T_{deep}}{\tau}$$

Limitations:

- Penetration of SWR not taken into account
- No good for a layered snow pack nor a thin snow cover
- No good under rapid changes in surface energy budget



Multi-layer models

a) Only heat conduction taken into account:

$$\rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(k \frac{\partial T}{\partial z} \right)$$

Compared to Force-restore, a significant improvement in rapidly changing conditions

Limitations: no good when SWR plays a major role

b) Heat conduction, penetration of SWR, and release of freezing heat taken into account (Cheng, 2002, J. Glac.)

$$\rho c \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(k \frac{\partial T}{\partial z} \right) - \frac{\partial q}{\partial z} + C_{MF}$$

$$q(z) = (1-\alpha) q_S \, e^{-\kappa z}$$

- Essential in spring (and summer)
- Temperature maximum often at the depth of ~ 10 cm
- Sub-surface melting possible under surface temperatures downto ~ -5°C

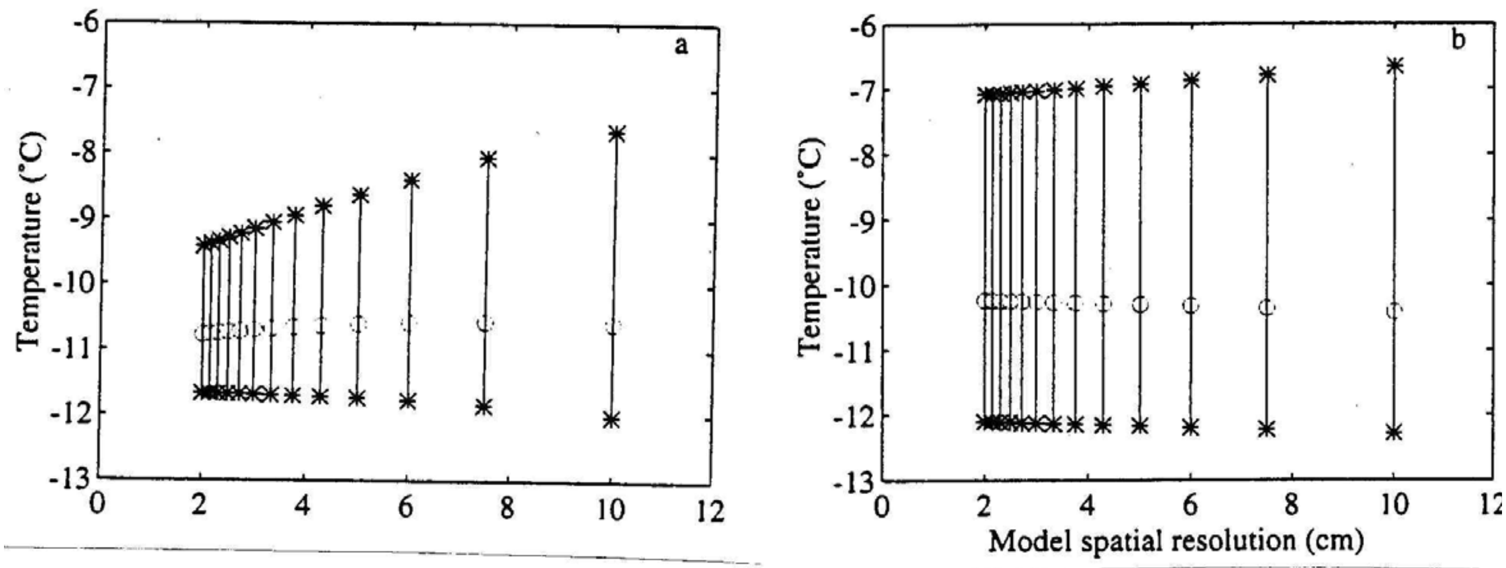
Effects of vertical resolution

Important when (a) solar radiation is large or (b) forcing conditions change rapidly.

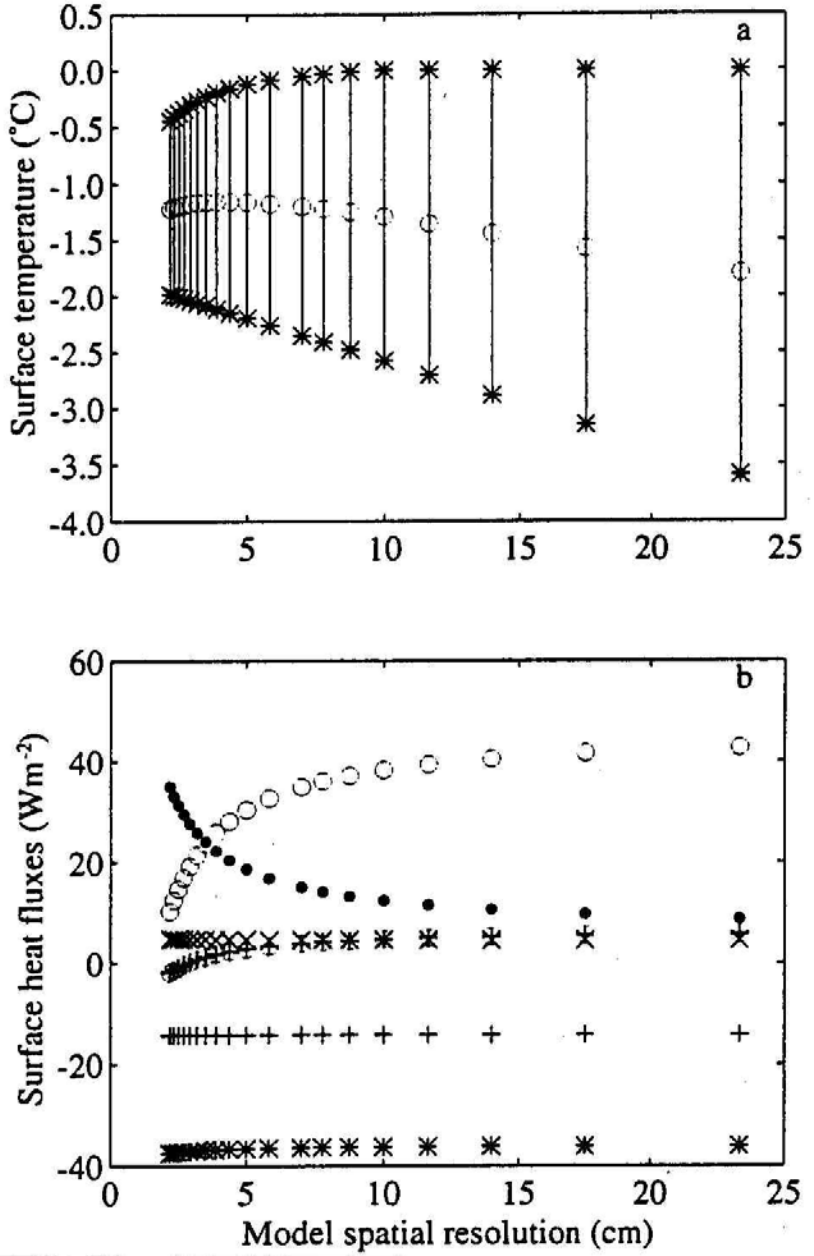
Coarse spatial resolution:

- Conductive heat flux just below the surface may have a wrong direction
- In extreme cases, the magnitude may have an error of 20-30 W m^{-2}
- all SWR is absorbed in the uppermost snow layer $\rightarrow$ too large diurnal cycle in Ts (Cheng, 2002, J. Glac.)

Dependence of diurnal Ts range on model vertical resolution for
(a) compact snow (κ = 5 m^{-1}) and (b) new snow (κ = 25 m^{-1})



Same for bare ice in typical April conditions in Finland:



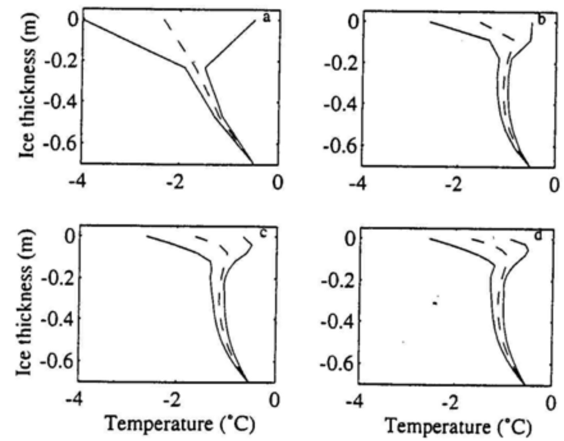


Figure 12. Modelled in-ice temperature profile using spatial resolutions of (a) 23 cm, (b) 8.7 cm, (c) 3 cm and (d) 2.5 cm. The corresponding values of N_i are: 3, 8, 18, and 2: respectively. In each panel, the three lines indicate respectively the daily minimum (left), dail average (middle) and daily maximum (right) in-ice temperature.

Snow pack structure models; some examples

SNOWPACK (Swiss, Lehning et al., 1998)

- operational for avalanche forecasting
- Input: V, Ta, Ts, RH, SWR, snow depth, Tg, T(z)
- Energy balance, mass balance, phase changes, movement of water and water vapour, wind drift calculated
- Most calculations based on snow microstructure: crystal size and form, bond size, number of bonds per crystal
- Snow = porous medium including ice, water and air
- Snow fall or drift $\rightarrow$ finite elements added to the grid
- Snow melt, sublimation or drift $\rightarrow$ elements removed from the grid
- Snow treated as viscoelastic material
- Snow layers defined by their sizes, bulk density, temperature, grain size and shape, and bond size.
- Output: New snow amount, settling rate, formation of surface hoar, T(z), ρ(z), metamorphic development in the layers

SNTHERM (CRREL, Jordan, 1991)

- snowmelt hydrology, polar meteorology
- various phases of precipitation
- Melt-refreeze cycles
- Transport of liquid water and water vapour included in the heat balance
- Effects of snow accumulation, melting, packing and metamorphosis on the heat conductivity and optics of snow
- Snow described as ice skeleton with air or liquid wate in the pores
- Output: snow depth, water equivalent, T(z), ρ(z), liquid water content, grain size

SAFRAN - CROCUS - MEPRA: operational chain in France since 1992

SAFRAN: analyses and forecasts of the atmospheric forcing on the snow pack

CROCUS: uses output of SAFRAN and simulates snow depth, T(z), ρ(z), LWC, bottom runoff, and stratigraphy

MEPRA: avalanche model, uses output of CROCUS and calculates mechanical indexes of the strength of the snow pack.

Boundary conditions

Air-snow turbulent heat exchange

$$H = \rho c_p C_H (\theta_s - \theta_a) V \qquad LE = \rho \lambda C_E (q_s - q_a) V$$

C_H and C_E depend on z, z_0, z_T, and on the thermal stratification usually expressed by the Monin-Obukhov theory: ψ_M(z/L), ψ_{HE}(z/L).

For snow on a flat surface (e.g. sea ice): $z_T = f(z_0 u_* / \nu)$ (Andreas, 1987)

For snow on forest $z_T \approx 0.1 \times z_0$

The turbulent surface fluxes are usually smaller than the radiative ones, except under cloudy skies with a strong wind, or during Föhn.

From the point of view of modelling of Ts, turbulent echange under very stable stratification is critical, and the results can be very sensitive to ψ-functions and z_T.

Radiative fluxes

In most applications, the radiative fluxes can be taken from the output of an atmospheric model.

In general, the most important parameter related to the boundary conditions for radiation is the snow surface albedo, which depends on various factors:

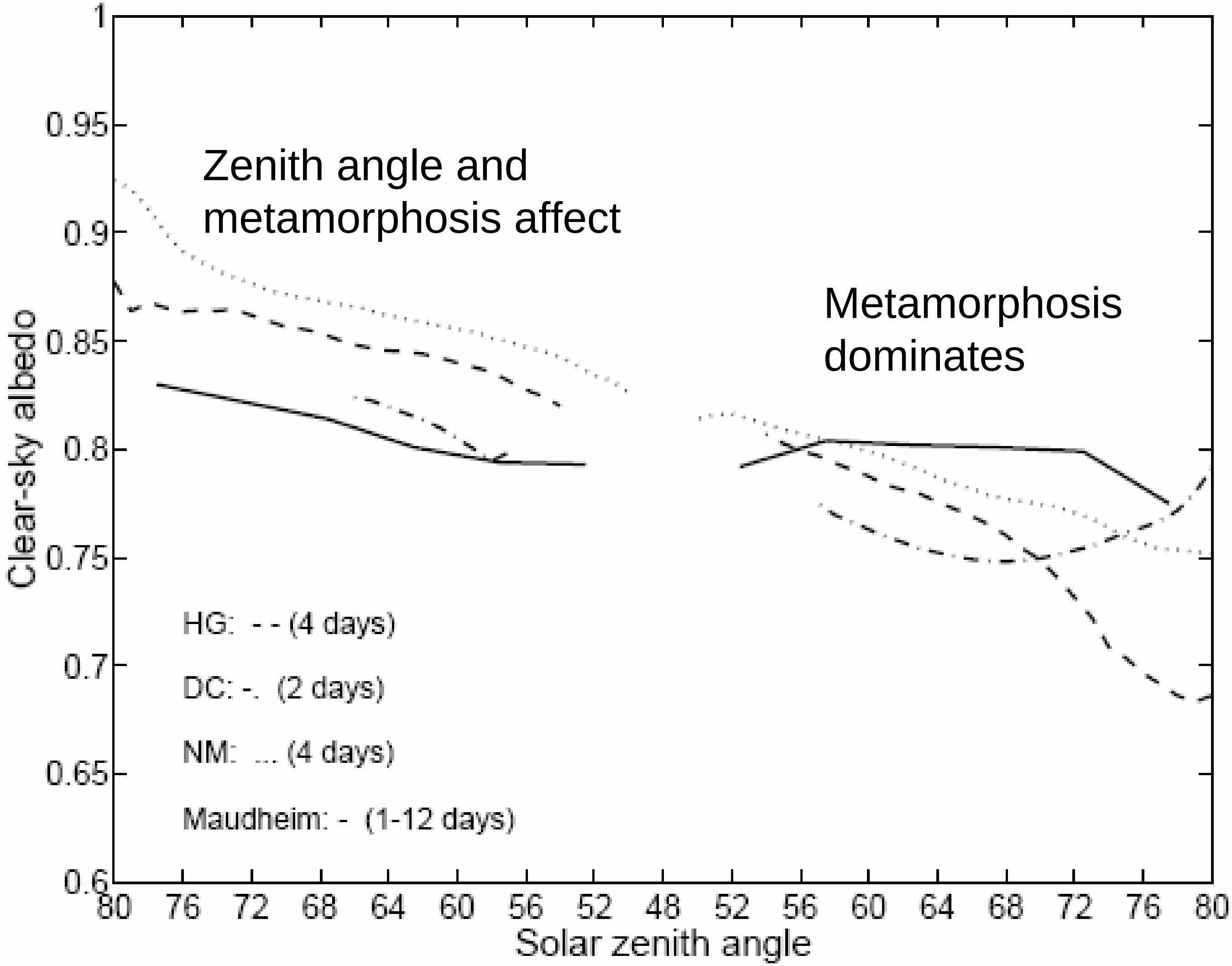
Factors increasing albedo

- Snow fall (usually)
- Snow drift
- Clouds (diffuse radiation)
- Increasing zenith angle
- High albedo of surrounding areas (multiple reflections)
- Air humidity (water vapour absorbs in near-infrared, for which snow albedo is low)

Factors decreasing albedo

- snow impurities
- destructive metamorphosis
- sastrugi
- decreasing zenith angle

Pirazzini (2004, JGR): average diurnal cycle of albedo at four Antarctic sites



Night: crystal
formation

Notes on albedo:

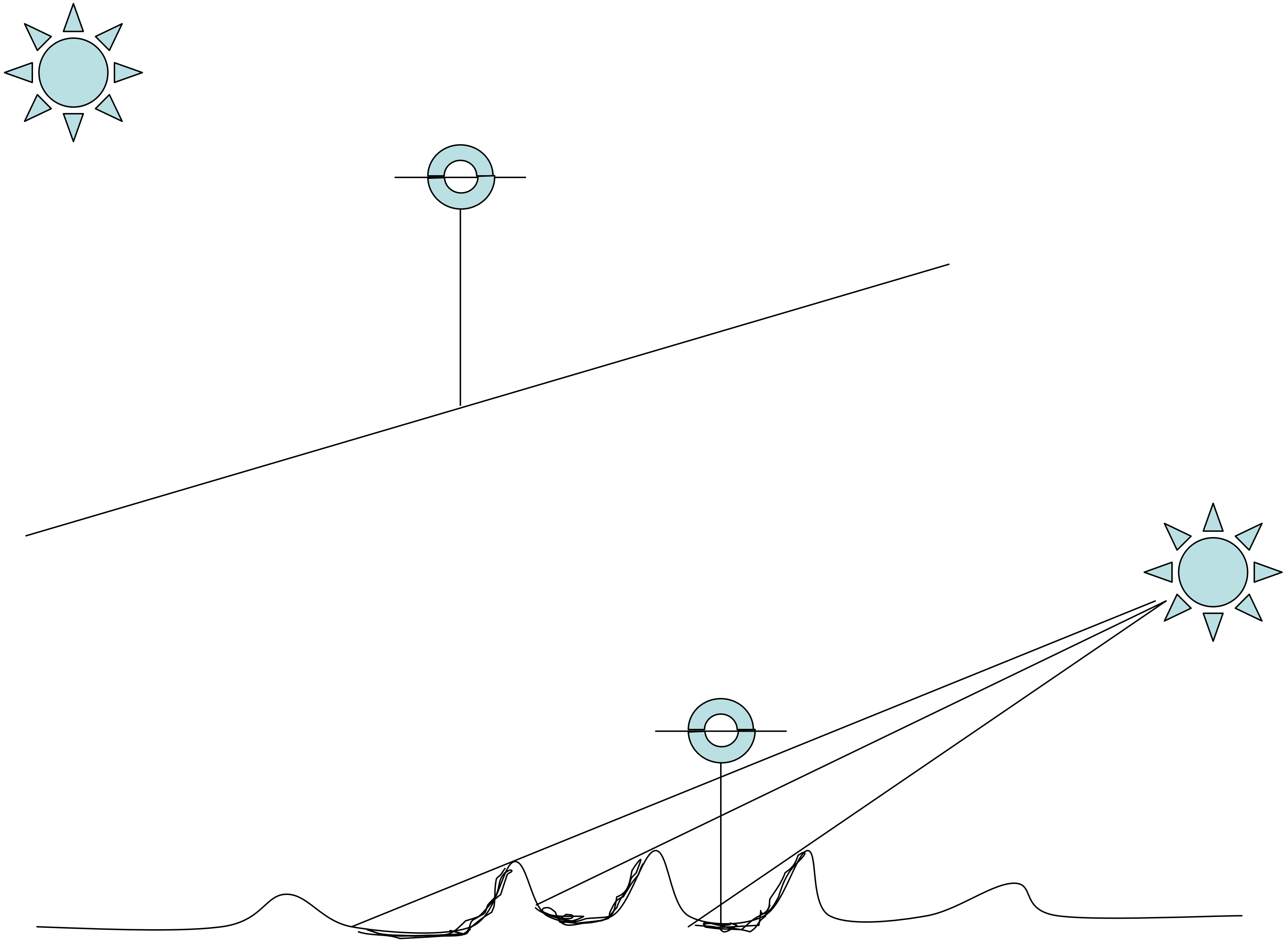
- Over slopes and surfaces with microstructures (such as sastrugi), the true and apparent albedo may differ a lot from each other
- Shadows have a small effect on the true albedo, but a large effect on the radiative fluxes: important in forest and mountains.
- Albedo of the snow surface may drastically differ from the regional albedo of a boreal forest

Boundary conditions at the bottom of the snow pack

Usually poorly known; climatologically there is a small conductive heat flux from the gound to the snow pack

In the case of snow on sea ice, the boundary conditions at the ice bottom can be applied: $T_{bot} = T_{fr}$(salinity)

$$-\left(k_i \frac{\partial T_i}{\partial z}\right)_{bot} + F_{ice/water} = -\rho_i L_{fr} \frac{dh_i}{dt}$$



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Horizontal heterogeneity of the snow cover

The heterogeneity is present in all seasons and simultaneously over different spatial scales

Factors generating the heterogeneity:

- Redistribution of snow by wind
- Effects of topography on the amount and phase of precipitation
- Effects of vegetation on the local turbulence and snow fall
- Effects of topography and vegetation (shadows) on snow melt

In boreal zone, largest effects typically occur in spring: e.g. trees free of snow while the surface of forests and fields are covered by snow

How to parameterize the grid-averaged surface fluxes <H> and <LE>?

Mosaic method $$\langle H\rangle = c_p \rho \langle V\rangle \left\{ \left(1 - fr\right) C_H^s \left(\theta_S^s - \langle\theta_a\rangle\right) + fr\, C_H^f \left(\theta_S^f - \langle\theta_a\rangle\right) \right\}$$

Large spatial scale of heterogeneity $\rightarrow$ <Ta> not representative for the local Ta $\rightarrow$ extended mosaic method with estimates for the local Ta

Modelled and parameterized H and LE over a patchy snow cover (Arola, 1999, JAS)

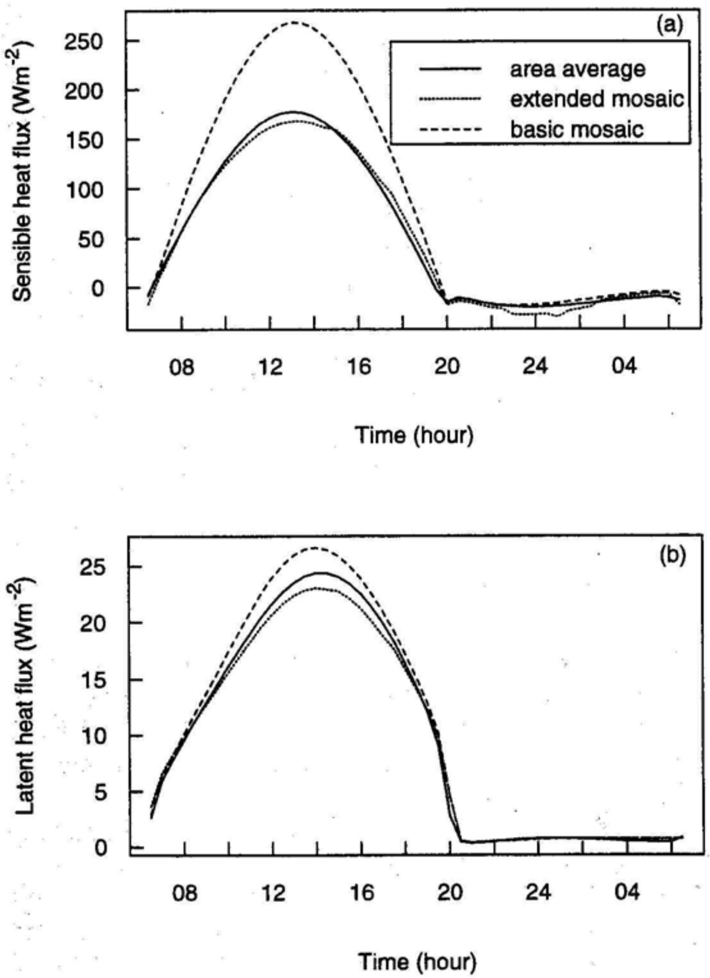

FIG. 2. Sensible (a) and latent heat flux (b) predicted by: basic mosaic method, extended mosaic method, and area-averaged fluxes for patchy snow cover. Patch size = 80 km and U = 7 m s^{-1}.

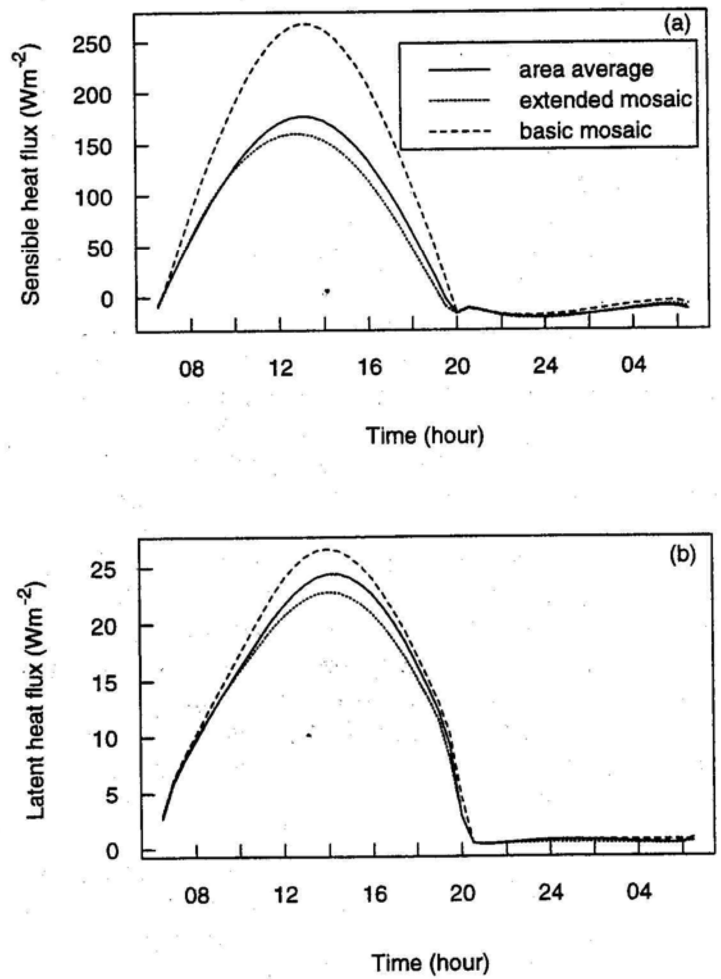

FIG. 3. Same as Fig. 2 except that the extended mosaic method uses the temperature adjustment of Eq. (18).

Snow on sea ice and lake ice

- Heavy snow load on thin ice $\rightarrow$ flooding of sea water on the ice $\rightarrow$ freezing to snow ice; can be calculated on the basis of Archimedes law. Flooding requires, however, also pathways for the water. Important in the Baltic Sea (up to 30% of the total ice thickness) and in the Antarctic.
- Surface or subsurface meting $\rightarrow$ percolation of the meltwater to snow-ice interface $\rightarrow$ refreezing to superimposed ice. Important in spring; e.g. Umeå 2004: 20 cm snow pack $\rightarrow$ 7 cm layer of superimposed ice (Granskog et al., 2005)
- Sastrugi common $\rightarrow$ effect on z_0 $\rightarrow$ effect on the ice drift ratio (Vihma et al., 1996; JGR)
- Lake ice: thick snow cover $\rightarrow$ thin ice
- Arctic sea ice: thick snow cover on thick ice, thin snow cover on thin ice.
- Complex $T_{snow}(x,z)$ during warm-air advection over the ice margin (Cheng and Vihma, 2002; J. Glac.)

Conclusions

The most essential aspects in modelling of the snow cover depend on our objectives:

- In short-term NWP, it is usually enough to simulate the surface temperature accurately: multi-layer model with heat conduction and penetration of solar radiation can yield benefit.
- Realistic description of snow albedo as well as grid-averaging of surface fluxes are essential both in NWP and climate models
- In hydrological applications, snow melt simulated in seasonal time scale is essential, and accurate parameterization of solar radiation is needed.
- For avalanche forecasting, snow layering and visco-elastic properties are important; requires snow pack structure modelling
- For botanical and zoological applications, the evolution of snow layers is important. This is also related to the impacts of the climate change.
- For treatment of snow cover in NWP and climate models (snow anlyses, intercepted snow, comprehensive land surface modelling), see the presentations by Bazile, Rodriguez and Samuelsson
- Important aspects of snow not addressed during the summer school: drifting and blowing snow, firnification, snow melt and floods, snow climatology, numerical methods (Lagrangian and Eulerian grids).

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