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Interaction of air flow with snow and ice

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Outline

Part I: continental snow

- Properties of the snow cover
- Snow metamorphosis
- Fluxes at the snow surface
- Snow models of different complexity

Part II: sea ice

- Sea ice as a surface for the atmosphere
- Surface exchange processes
- Typical ABL structure over sea ice
- ABL in various flow conditions: off-ice and on-ice flows and flows parallel to the ice edge
- Cloud-top radiative cooling
- Summary and challenges

Part I: Continental snow

Why to study air-snow interaction?

- **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**

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 Ts = 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
- Snow metamorphosis

Snow metamorphosis

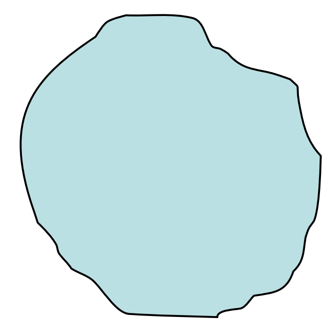
<u>Dry snow metamorphosis</u>

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

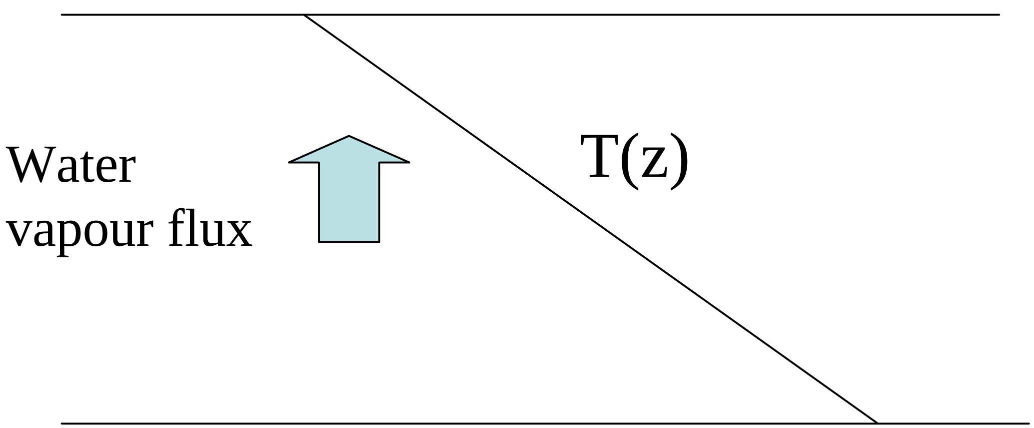
$\rightarrow$

$\rightarrow$



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

water vapour sublimates 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

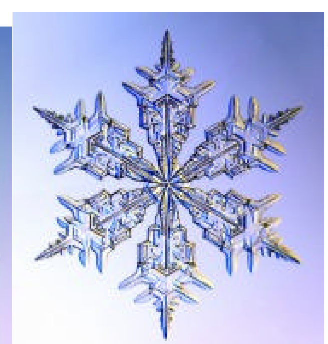
Funicular regime: liquid occupies continuous paths throughout the pore space, and snow is cohesionless

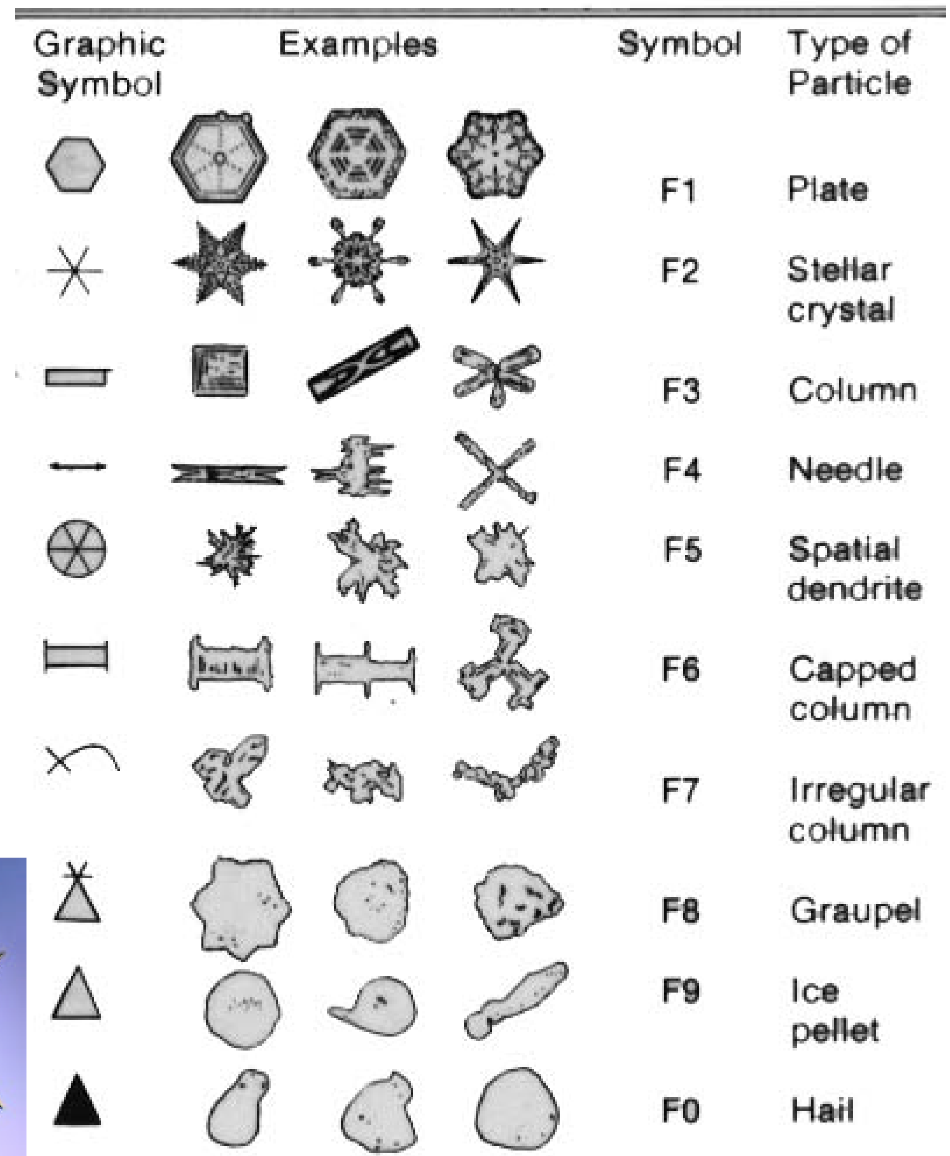
Melting temperature depends on the radius of curvature: sharp parts melt while concave parts freeze $\rightarrow$ flux of water is rounding the crystals.

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.

Practical problems: wind & solar radiation





INTERNATIONAL SNOW CLASSIFICATION FOR SOLID PRECIPITATION

Graphic Symbol	Examples	Symbol	Type of Particle
		F1	Plate
		F2	Stellar crystal
		F3	Column
		F4	Needle
		F5	Spatial dendrite
		F6	Capped column
		F7	Irregular column
		F8	Graupel
		F9	Ice pellet
		F0	Hail

Air-snow turbulent heat exchange

$$H = r c_p C_H (q_s - q_a) V \qquad LE = r\, l\, C_E (q_s - q_a) V$$

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

For snow on a flat surface: $z_T = f(z_0 u_* / \nu)$ (Andreas, 1987; 2002)

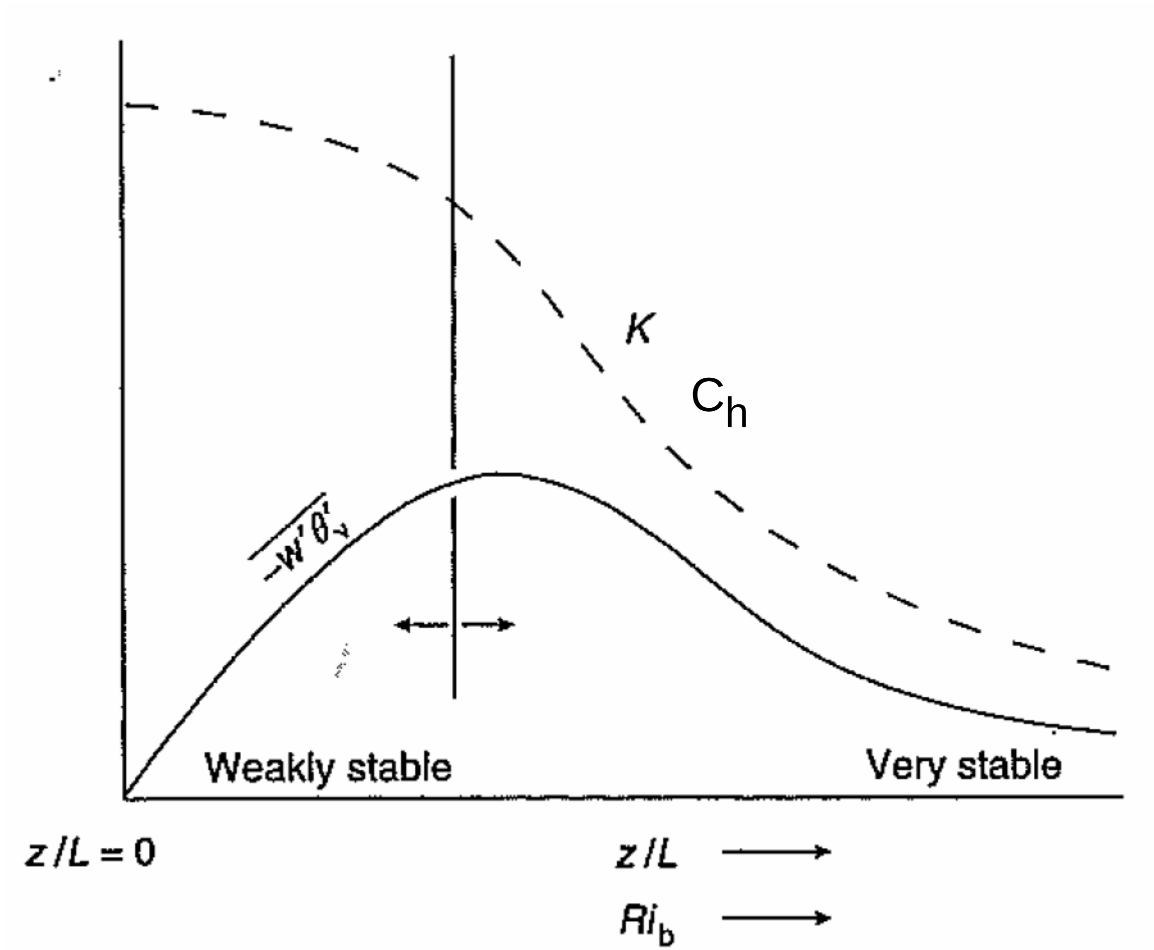
For snow-covered forest $z_T \approx 0.1 \times z_0$

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.

This can yield to decoupling between the air and snow.

$$H = \rho\, c_p\, C_h\, (\theta_a - \theta_S)\, V$$



The modelled air-surface temperature difference can be extremely sensitive to the quantitative formulation of stability dependence of C_h.

Radiative fluxes at the snow surface

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:

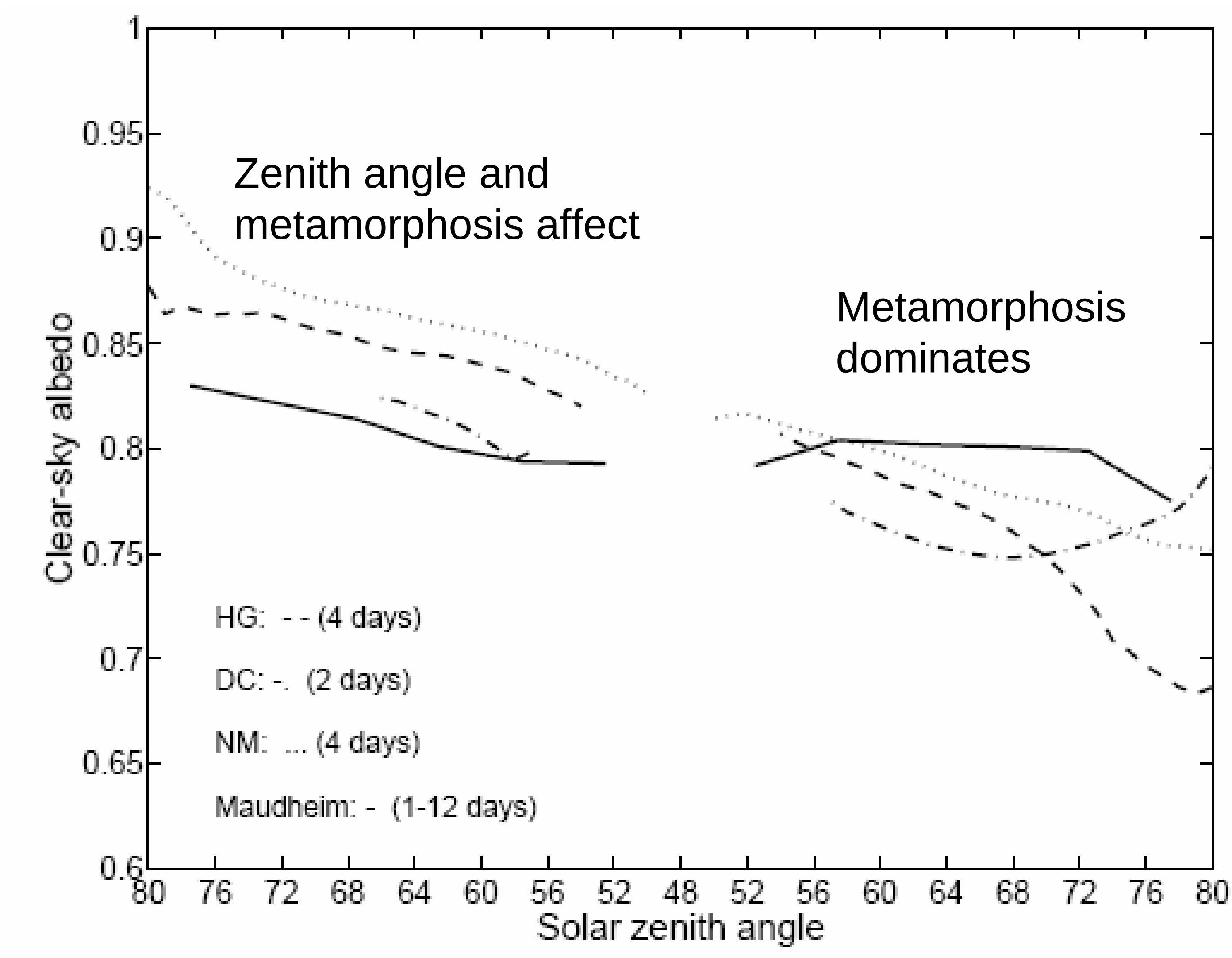
Factors increasing albedo

- Snow fall (usually)
- Snow drift
- Clouds (diffuse radiation)
- Increasing zenith angle
- 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
- High albedo of surrounding areas (multiple reflections)

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



Night: crystal formation

Notes on albedo:

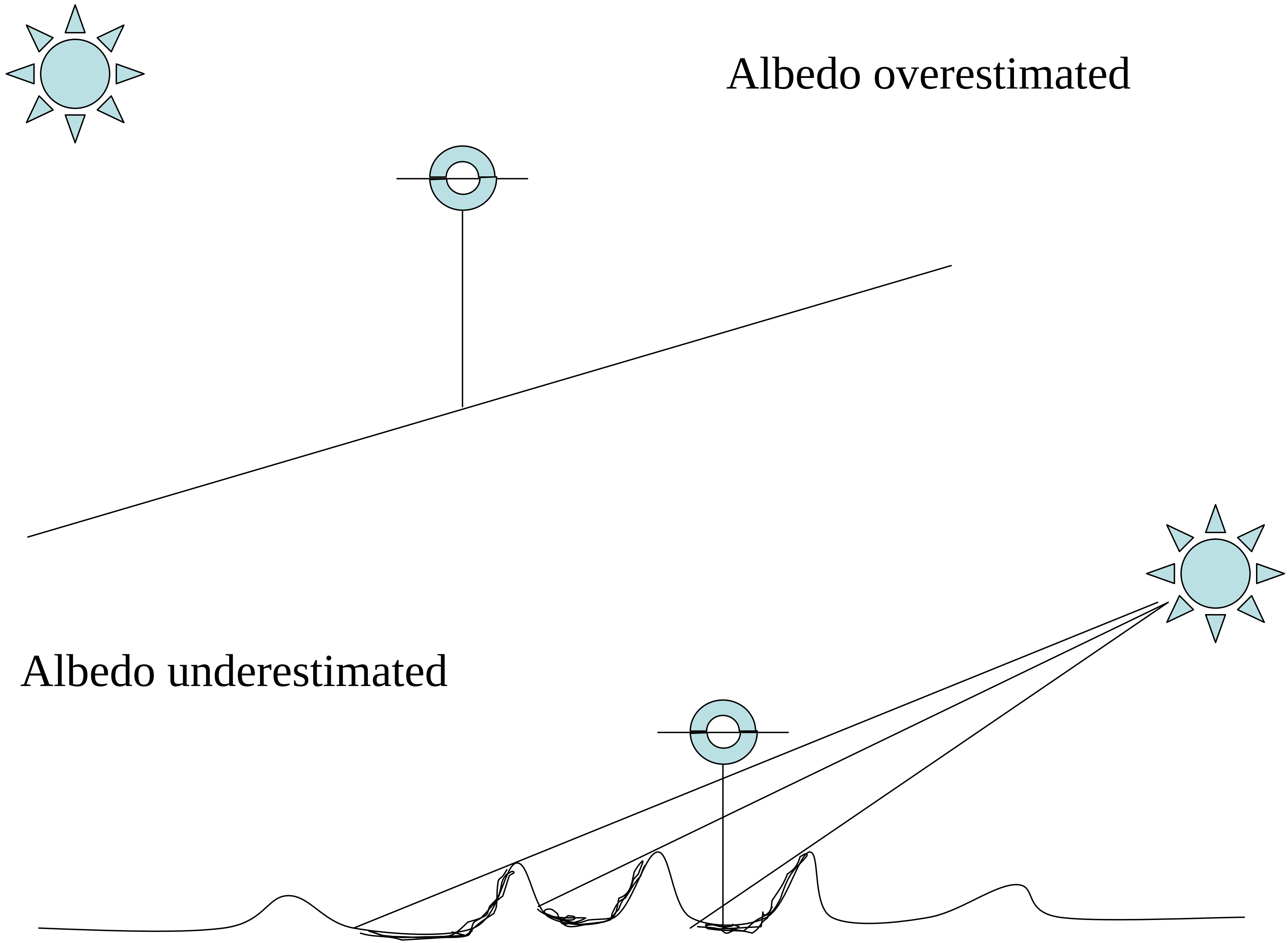
- Over slopes and surfaces with microstructures (such as sastrugi), the true and apparent albedo may differ a lot from each other
- Shadows have a small effect on the true albedo, but a large effect on the radiative fluxes: important in forest and mountains.
- Albedo of the snow surface may drastically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

Fluxes 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)

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


Albedo overestimated
Albedo underestimated

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Het flux at the bottom of the snow pack

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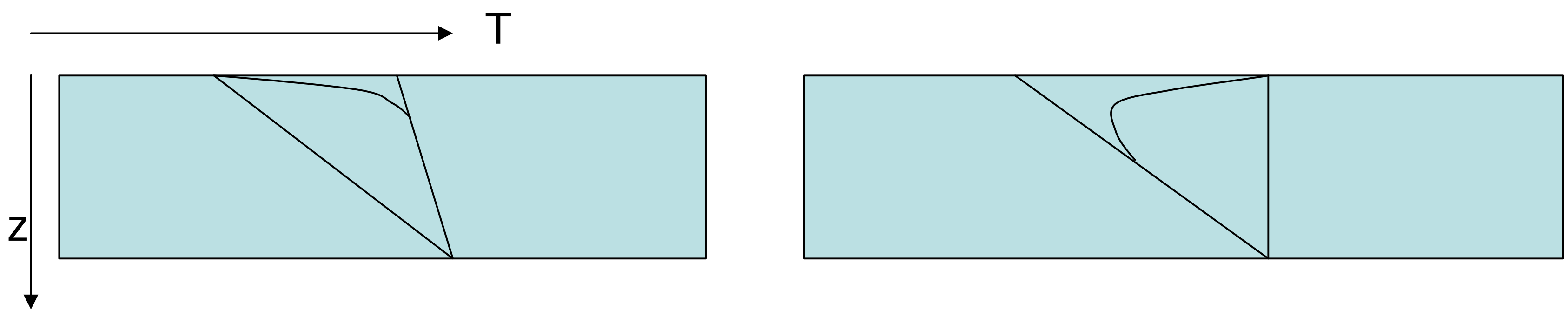
Snow models of different complexity

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

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

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:

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

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.)

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

$$q(z) = (1-a)\,q_S\,e^{-kz}$$

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

Effects of vertical resolution

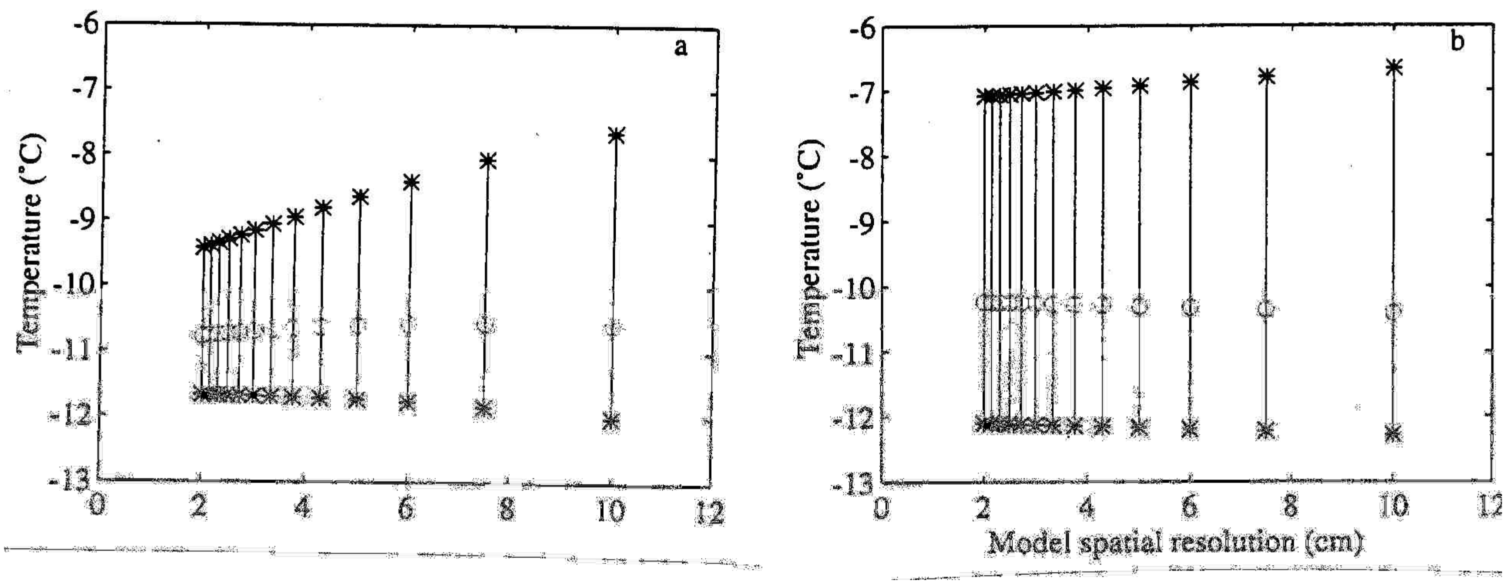
Important when (a) solar radiation is large or (b) forcing conditions change rapidly.

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})



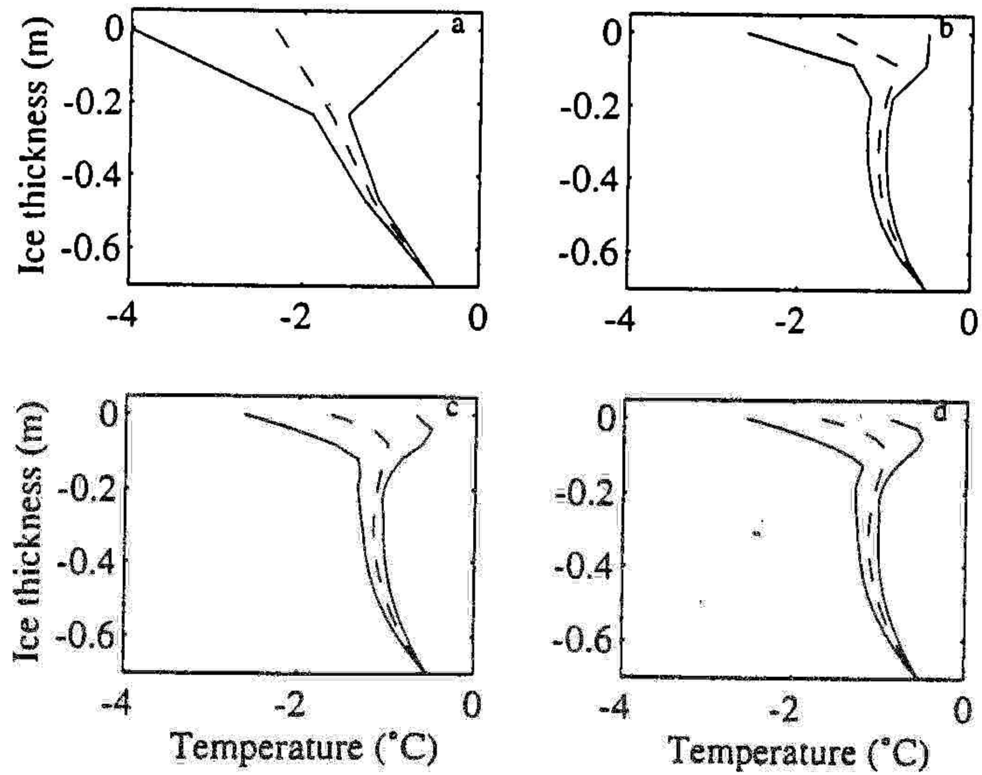


Figure 12. Modelled in-ice temperature profile using spatial resolutions of (a) 23 cm, (b) 8.7 cm, (c) 3 cm and (d) 2.5 cm. The corresponding values of N_i are: 3, 8, 18, and 2: respectively. In each panel, the three lines indicate respectively the daily minimum (left), dail average (middle) and daily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

The new soil-snow-forest schema for HIRLAM

Developed by Stefan Gollvik and others

Based on the Rossby Centre Regional Climate Model Land Surface Scheme

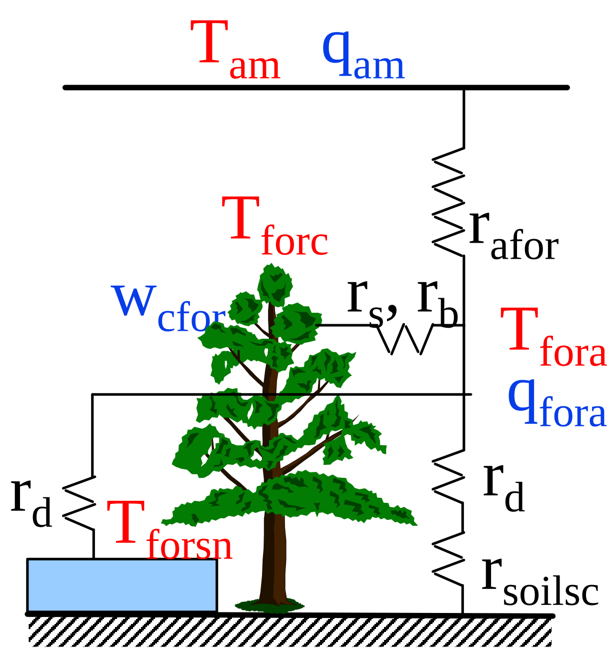
New compared to reference HIRLAM:

- insulation effects of snow
- exchange of heat and moisture between forest and air
- forest effects on the albedo

Partick Samuelsson, SMHI

The forest tile sensible heat flux

Characterized by low tree heat capacity & small r_b



Sensible heat flux

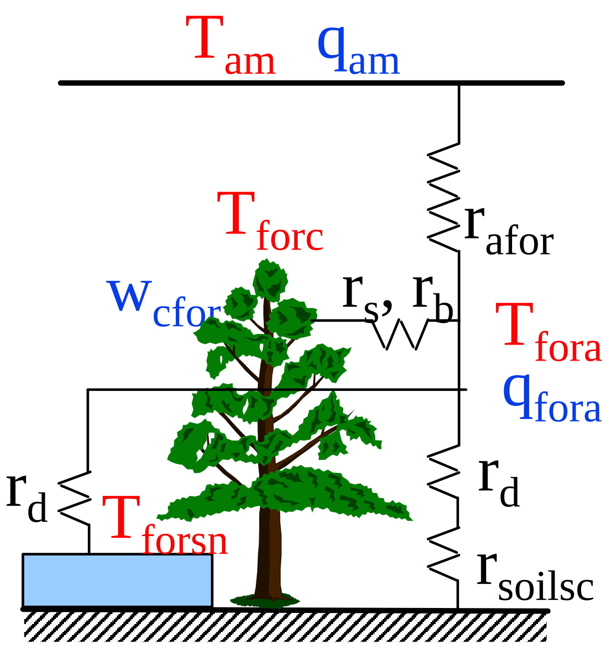
Forest canopy	$H_{forc} = \rho c_p \dfrac{T_{forc} - T_{fora}}{r_b}$
Forest floor (soil)	$H_{fors} = \rho c_p \dfrac{T_{fors} - T_{fora}}{r_d}$
Forest floor (snow)	$H_{forsn} = \rho c_p \dfrac{T_{forsn} - T_{fora}}{r_d}$
Canopy air - Atmos.	$H_{for} = \rho c_p \dfrac{T_{fora} - T_{am}}{r_{afor}}$

T_{fora} q_{fora} are canopy air temperature and humidity

where T_{fora} is solved from the relationship

$$H_{for} = H_{forc} + (1 - frsnfor)H_{fors} + frsnforH_{forsn}$$

The forest tile aerodynamic resistances r_b and r_d



The aerodynamic resistance

$$r_b = f\,(LAI^{-1}, u_{for}^{-1}, (T_{forc} - T_{fora})^{-1})$$

Choudhury and Monteith (1988)

Sellers et al. (1986)

The aerodynamic resistance

$$r_d = f\,(z_{for}, u_{for}^{-1}, (T_{fors} - T_{fora})^{-1})$$

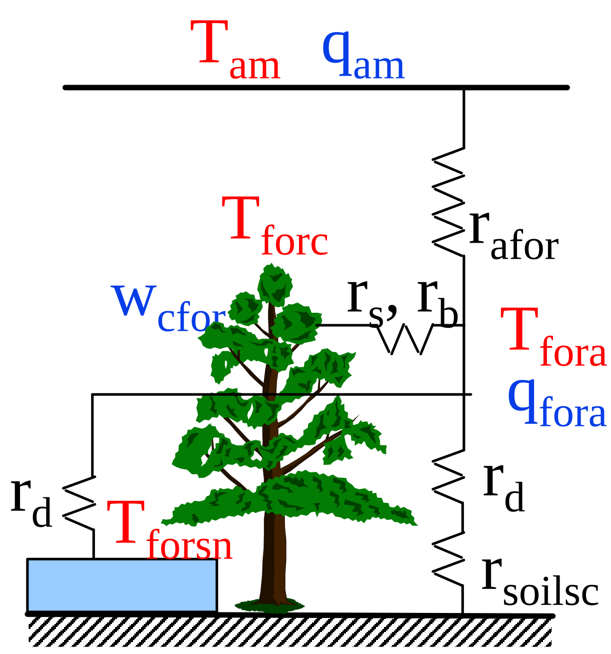
Choudhury and Monteith (1988)

Sellers et al. (1986, 1996)

$r_b \approx 10\%$ of r_d

Forest temperatures

Characterized by low tree heat capacity & small r_b



$$\begin{cases} \dfrac{\partial T_{forc}}{\partial t} = \dfrac{1}{C_{forc}}\Phi_{forc} \\ \dfrac{\partial T_{fors}}{\partial t} = \dfrac{1}{(\rho C)_{fors} z_s}[\Phi_{fors} + \Lambda_s(T_{fors2} - T_{fors})] \\ \dfrac{\partial T_{forsn}}{\partial t} = \dfrac{1}{(\rho C)_{forsn} z_{forsn}}[\Phi_{forsn} + \Lambda_{forsn}(T_{forsns} - T_{forsn})] \end{cases}$$

where

$$\phi_{forx} = Rn_{forx} + H_{forx} + E_{forx}$$

C_{forc} defined according to Verseghy et al., (1993)

T_{fora} q_{fora} are canopy air temperature and humidity

Conclusions from Part I

- From the point of view of the atmosphere, snow is a particular surface type because it is a very good insulator and its thermal properties change quicly
- Stable stratification is very common over snow surface and decoupling between air and snow can take place bot in nature and in numerical models. The low heat conductivity of snow favours decoupling.
- 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 is 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
- Considering the impacts of the climate change, the evolution of snow layers is important e.g. from the point of view of botanics and zoology

Part II: Air flow over sea ice

Characteristics of sea ice as a surface for the atmosphere

Consists of undeformed ice cover, rafted and ridged ice floes of variable thickness, in places separated by leads

Surface relatively homogeneous with respect to roughness, but often very heterogeneous with respect to Ts and albedo

Interaction of dynamics and thermodynamics:

- surface type may change rapidly due to ice motion, slowly due to ice and snow thermodynamics

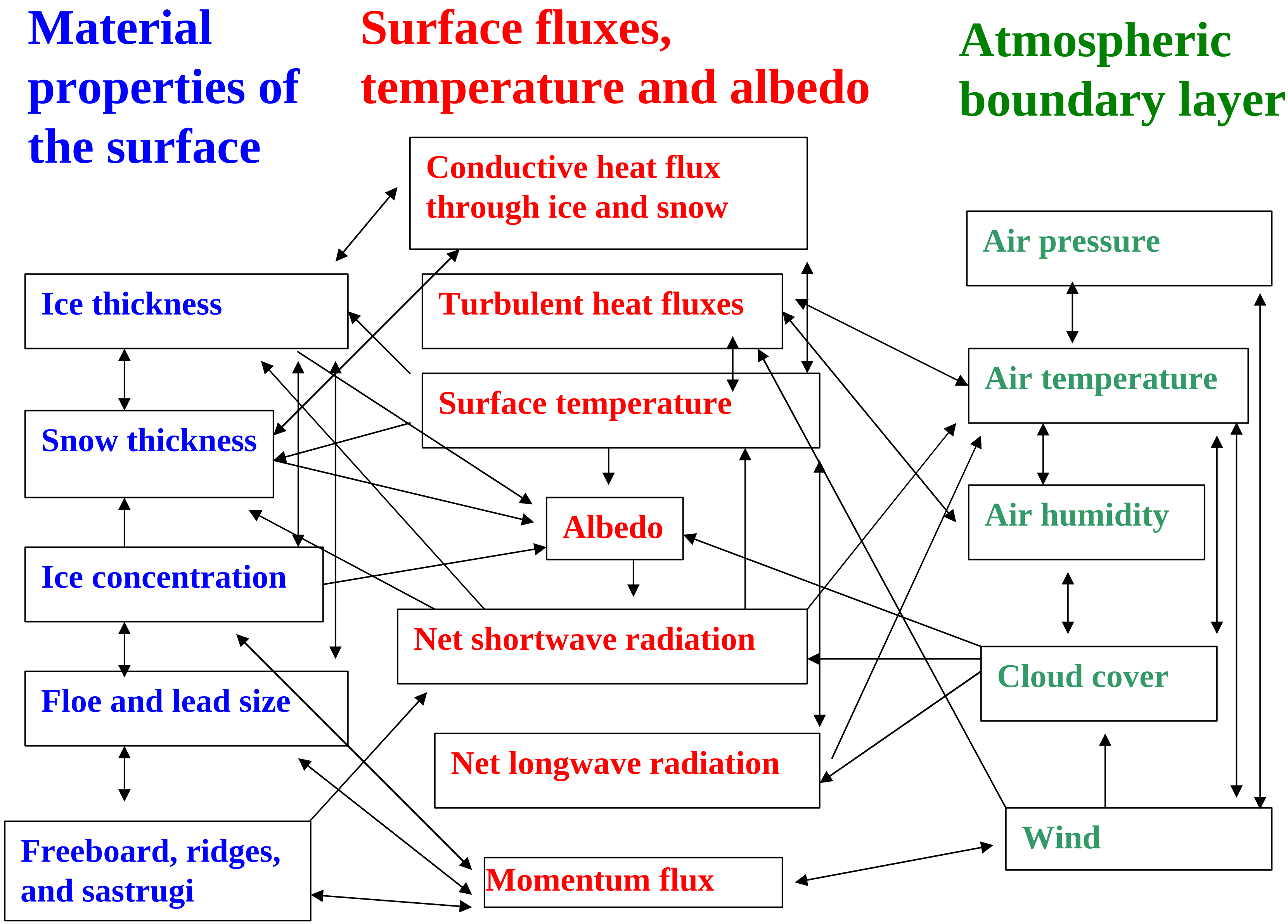
- due to the low heat conductivity of snow, Ts responds rapidly to radiative and turbulent forcing, and ?T/?z through snow/ice can be very large, up to 30 K/m

Over the marginal ice zone, the ABL often under modification

Snow on sea 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.)

Interactive surface exchange and ABL processes



Sea ice roughness and air-ice momentum flux

$$t = rC_{Dz}V_z^2 \qquad C_{Dz} = k^2\left(\ln\frac{z}{z_0} - \Psi_M(z/L)\right)$$

- z_0 is affected by ice ridges, floe edges, and sastrugi

Photo Global Expeditions Adventures, Inc.

Photo Henry Söderman, FIMR

Photo Denis Evans

- observations over various ice types (Guest and Davidson, 1991):
  very smooth first-year ice: z_0 = 0.3 mm – very rough multiyear ice: z_0 = 30 mm
- sastrugi are often overlooked, but may be even more important than ridges (Andreas & Claffey, 1995), and sometimes even have a detectable effect on the ice drift (Vihma et al., 1996)
- various schemes presented for the aggregation of τ
- sea ice models are more sensitive to the stratification effect than to z_0 (Uotila, 2000)

Heat fluxes over sea ice

Solar radiation: albedo is ~ 0.1 for open water and 0.4-0.9 for sea ice. Dominating surface heat balance component in summer.

Longwave radiation: in winter, usually the dominating component in surface heat balance; net longwave radiation is usually negative (in winter some 20 Wm^{-2}**)**, but close to zero or slightly positive under a thick cloud cover in summer.

Conductive heat flux through ice and snow: typically ~ 5 Wm^{-2}

Sensible heat flux (H): typically from air to ice ~ 10-20 Wm^{-2}
Latent heat flux (LE): typically close to zero

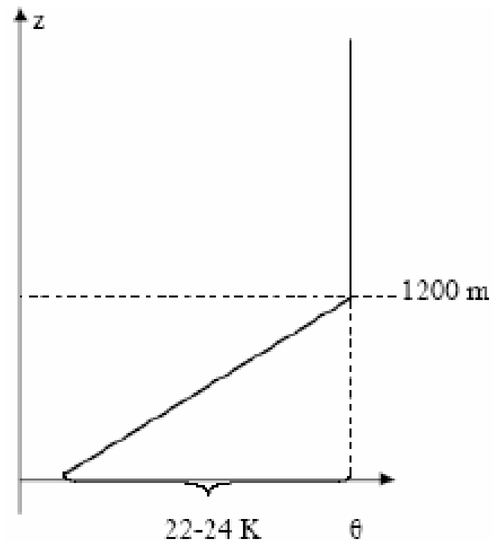
$z_T = z_q = f\,(z_0 u_*/?)$, Andreas (1987; 2002)

Over leads in winter, H + LE can reach several hundreds of Wm^{-2}
→ Localized convection over leads; usually only reaches heights less than 100 m, but in rare cases can penetrate through the Arctic inversion.
→ parameterization of grid-averaged heat heat fluxes is a problem in models

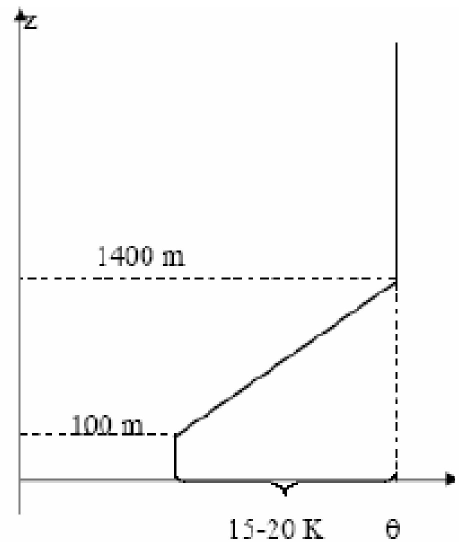
Typical potential temperature and wind profiles in the Arctic

Winter

Clear skies, weak winds, high ice concentration

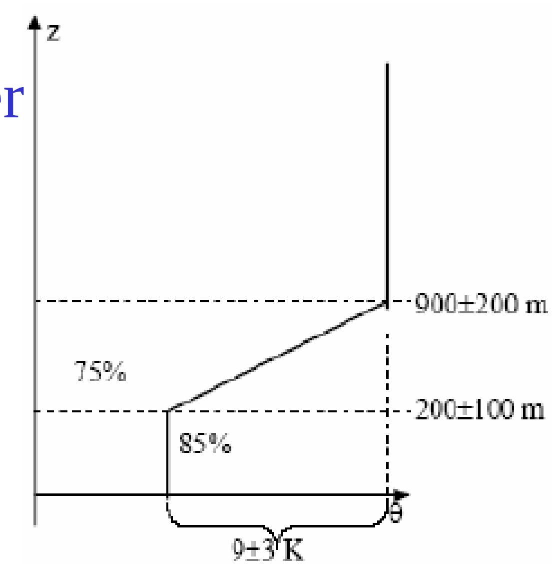


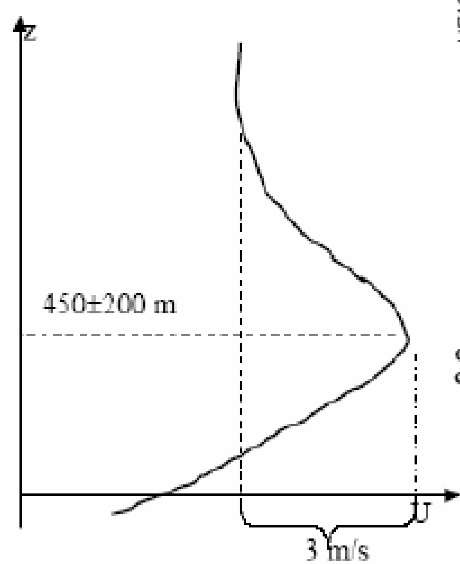
Cloudy, strong winds, low ice concentration



Summer

Almost continuous Stratus in summer





Serreze et al., 1992, J. Clim

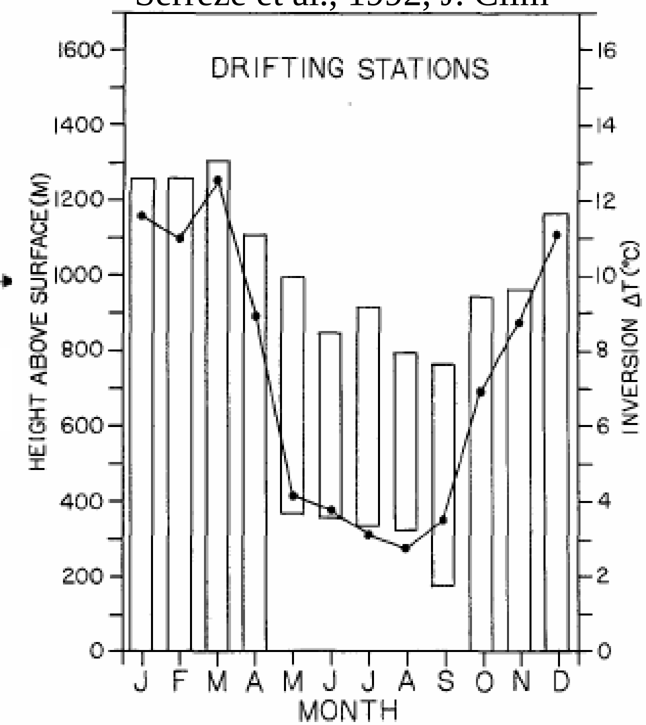


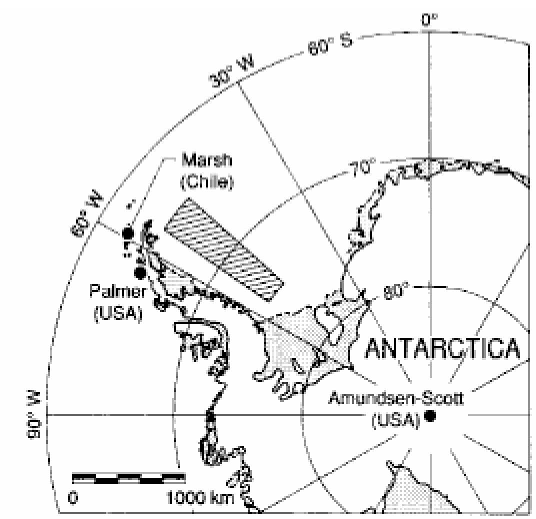
FIG. 7. Monthly median inversion top (top of bars), base (bottom of bars), and temperature difference (solid line) from drifting station data.

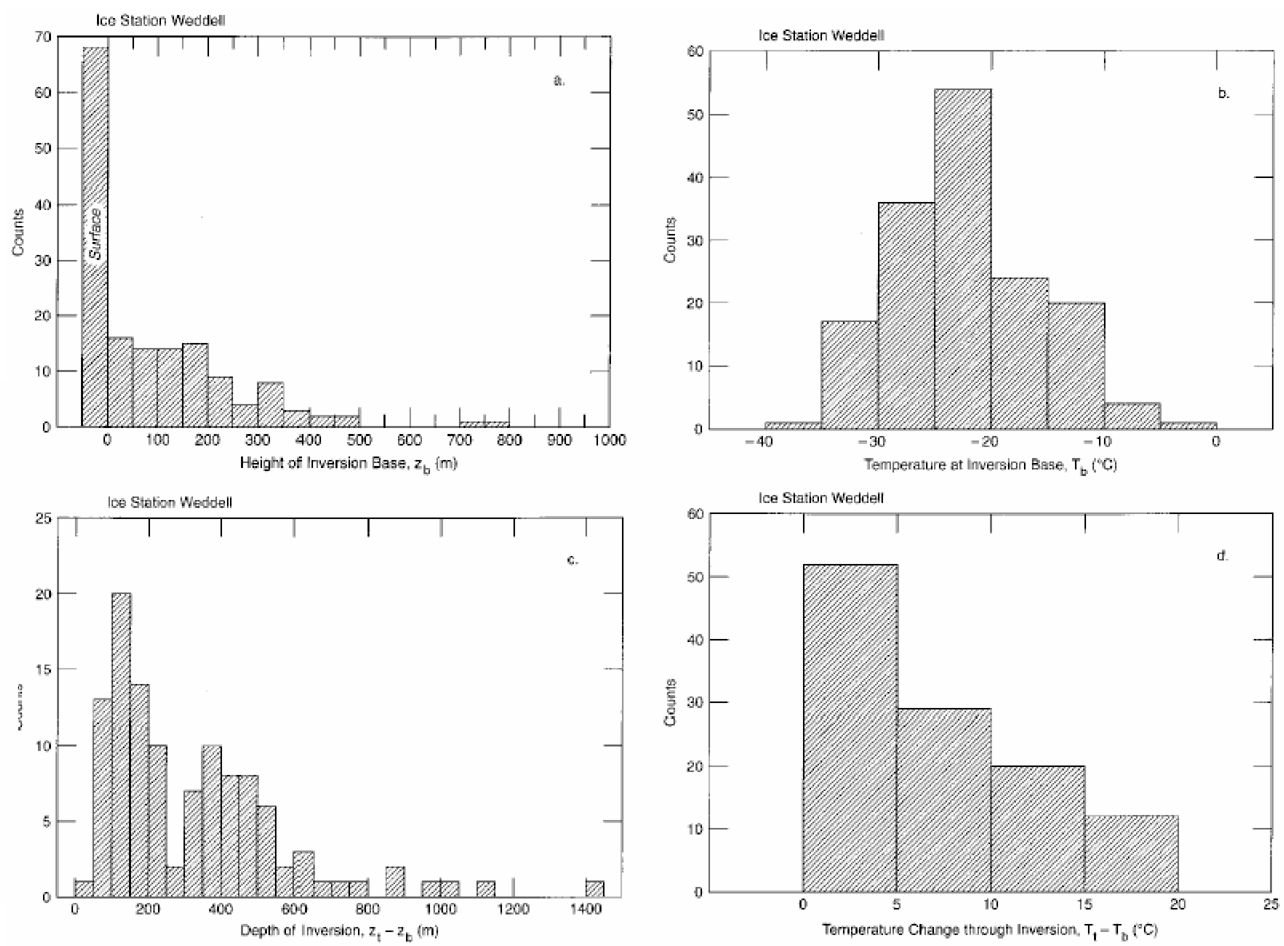
In the Antarctic

- sea ice is mostly seasonal and thinner (0.6-1 m)
- sea ice concentration lower
- sea ice locates in lower latitudes on average

→Temperature inversion is assumed to be weaker and unstable ABL more common than in the Arctic, but there are not much data to prove this.

In the western Weddell Sea in autumn – early winter, with compact multiyear ice cover, the inversions are roughly as those in the Arctic (Andreas et al., 2000,



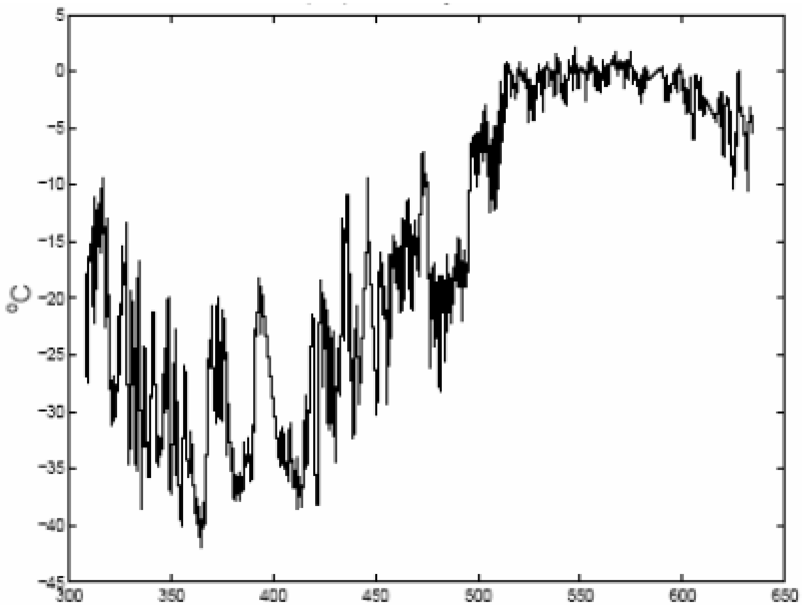


Temperature variations

Large synoptic-scale variations related to cyclones; three mechanisms can cause a rapid increase in temperature : (a) warm-air advection, (b) mixing of the SBL due to strong winds, and (c) cloud radiative heating

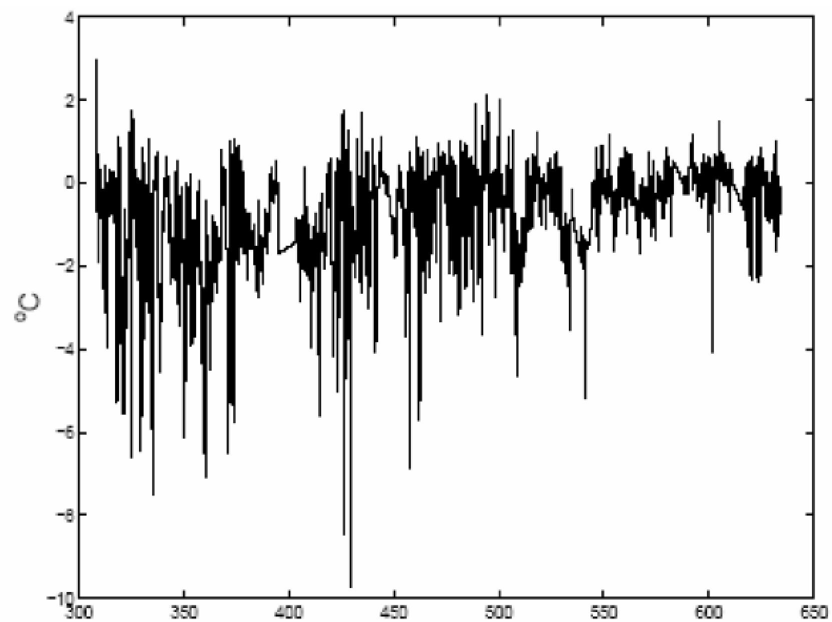
Slightly stable surface-layer stratification prevails; but during clear skies and weak winds, the surface inversion is strong and during cyclone passages the stratification is near-neutral

T2m at SHEBA Ice Station



November 1997 – November 1998

Ts-T2m at SHEBA Ice Station



November 1997 – November 1998

Surface exchange and ABL processes strongly depend on the flow conditions

1. Cold-air outbreaks (off-ice flows)

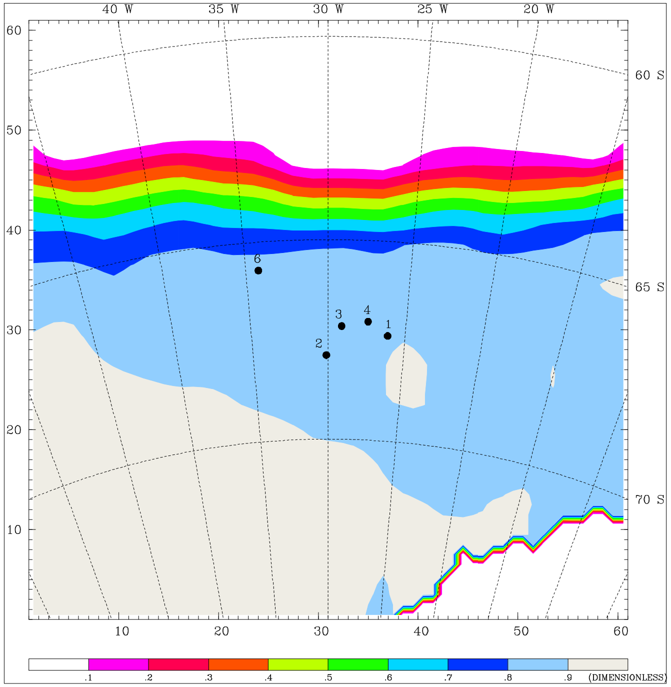
- development of a convective boundary layer with increasing distance from the ice edge
- diffuse ice margin
- <u>Essential in modelling:</u> counter-gradient transport in CBL (Lüpkes and Schlünzen, 1996), cloud physics, heat fluxes from open water upwind of the ultimate ice edge (Vihma & Brümmer, 2002), information on the ice concentration

An example from the Weddell Sea (Tenhunen, Vihma and Doble, submitted to MWR, 2007)

Environment: Mixture of ice floes and open water → large heat fluxes from the open water to the atmosphere.

Data: Wind and temperature data from six buoys in the marginal ice zone of the Weddell Sea. Satellite data on sea ice concentration based on various processing algorithms.

Modelling: evolution of the ABL applying the Polar version of the mesoscale model MM5. Simulation period: 23-25 May 2000: first northerly and then southerly winds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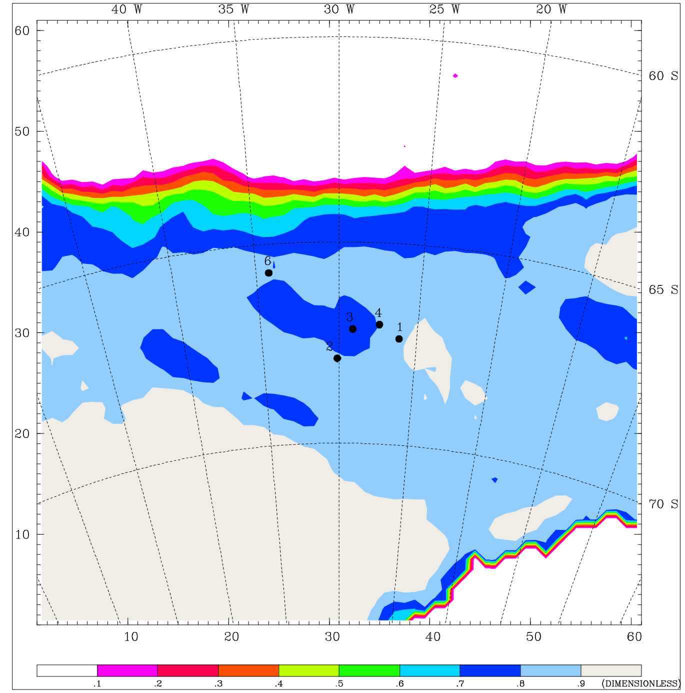
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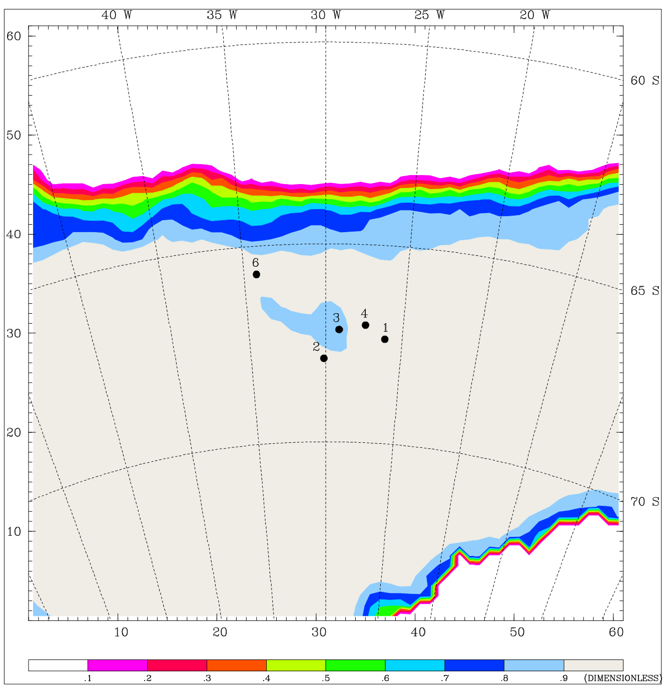
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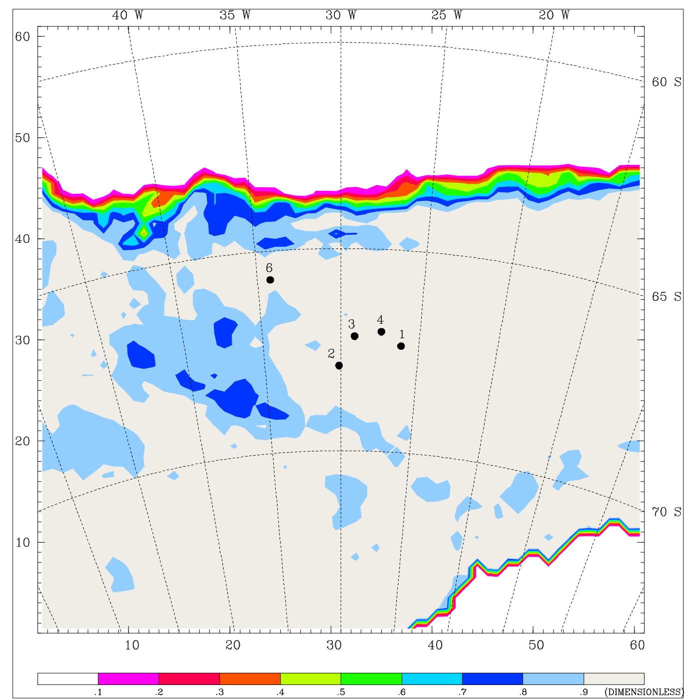
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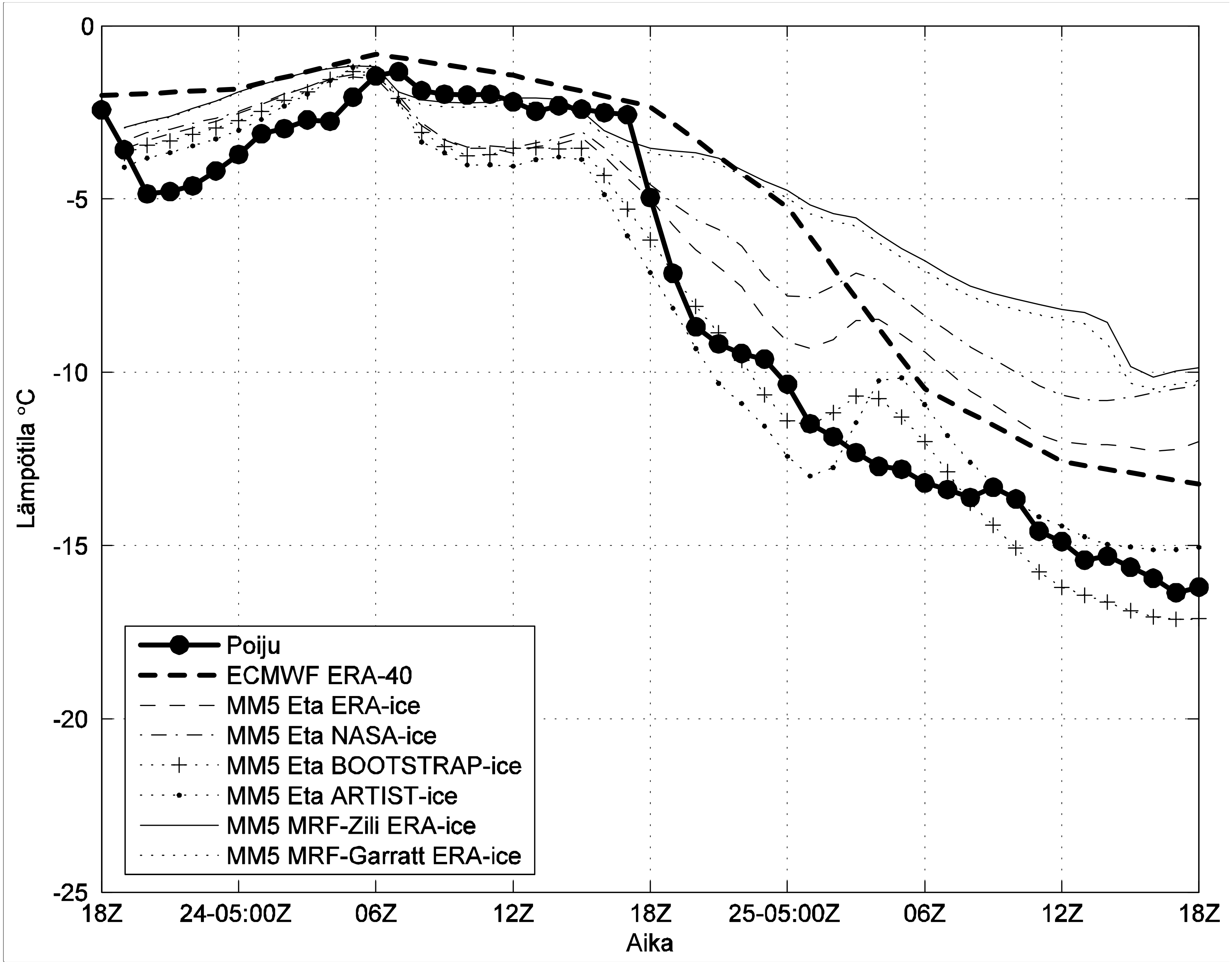
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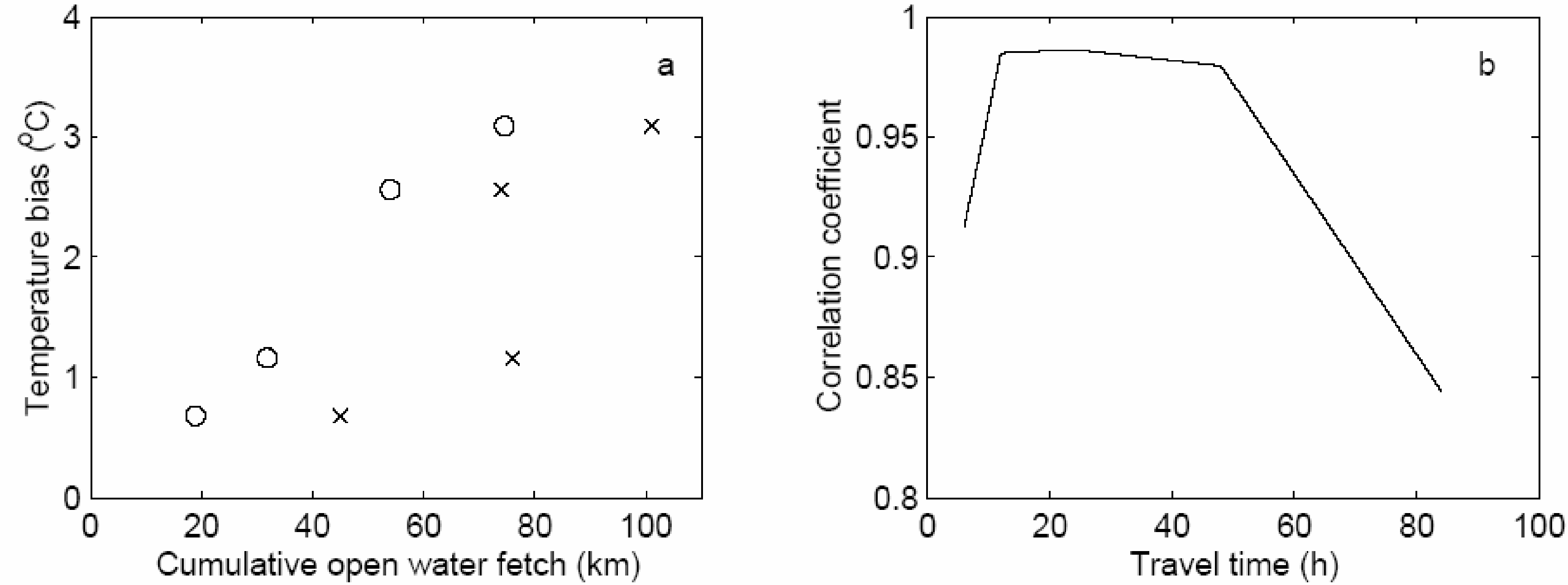
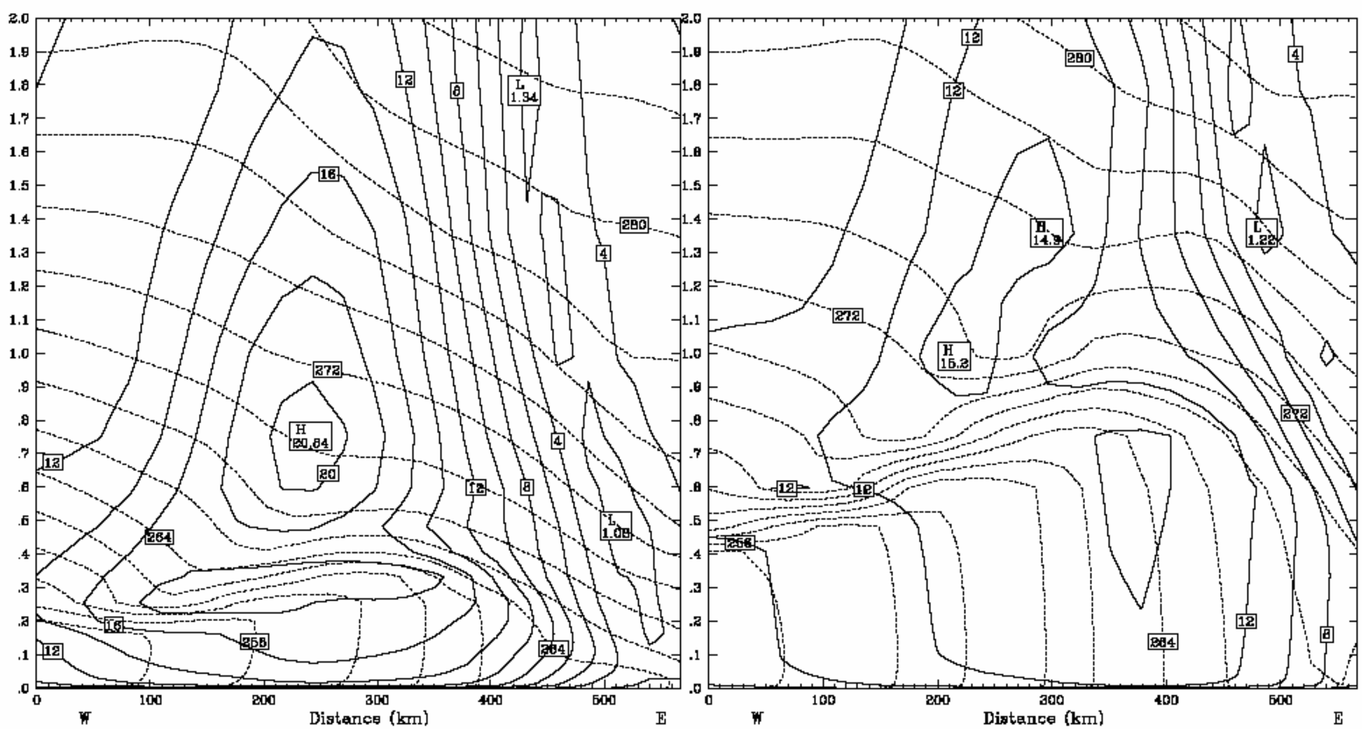

a
Temperature bias (°C)
Cumulative open water fetch (km)
b
Correlation coefficient
Travel time (h)


W
Distance (km)
E
W
Distance (km)
E

Conclusions from the Weddell Sea case

- first mesoscale meteorological modeling experiments for the Antarctic sea ice zone with detailed validation at mesoscale spatial resolution.
- in the ABL, temperatures sensitive to ice concentration: differences up to 13 K in T_{2m}, which is more than observed in previous studies in the Arctic and the Baltic Sea

- above the ABL, temperature and wind field sensitive to the turbulence scheme applied in Polar MM5

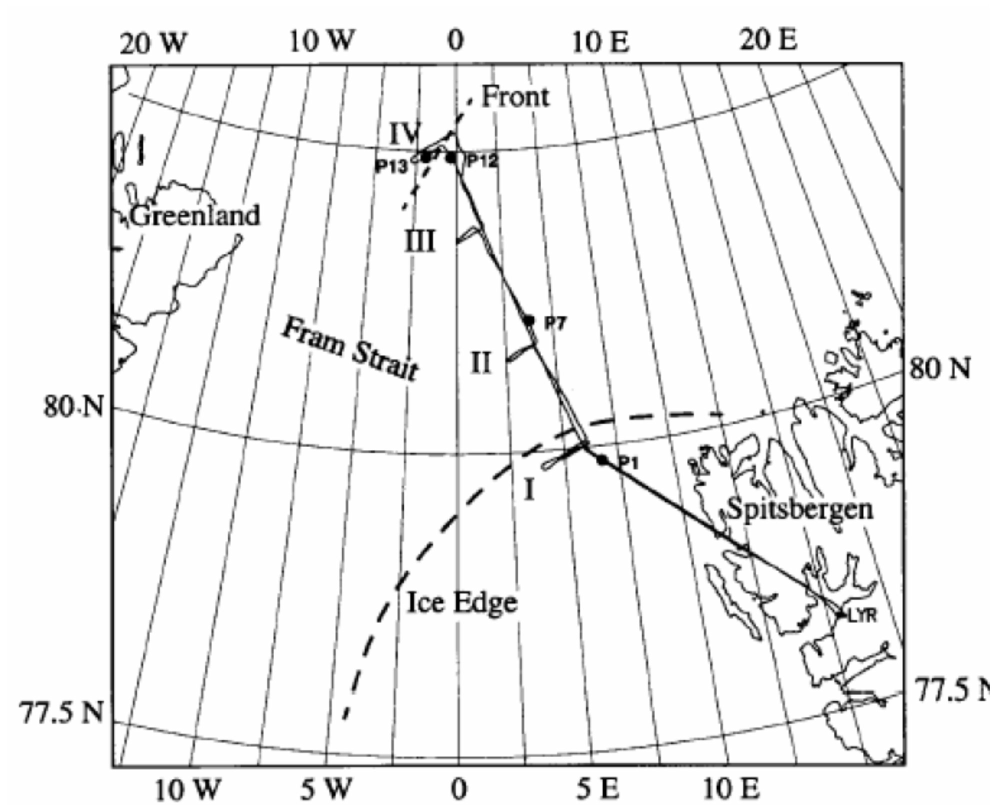
- effects of analyses nudging good for pressure fields, bad for temperature fields

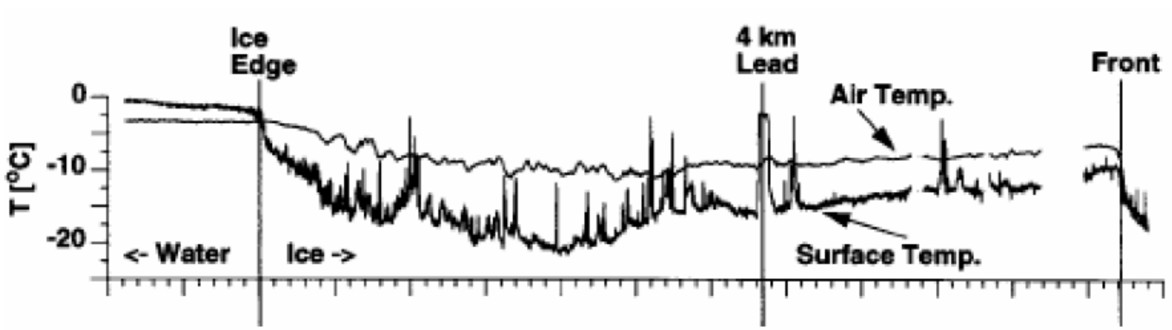
2. Warm air advection from the open sea over sea ice

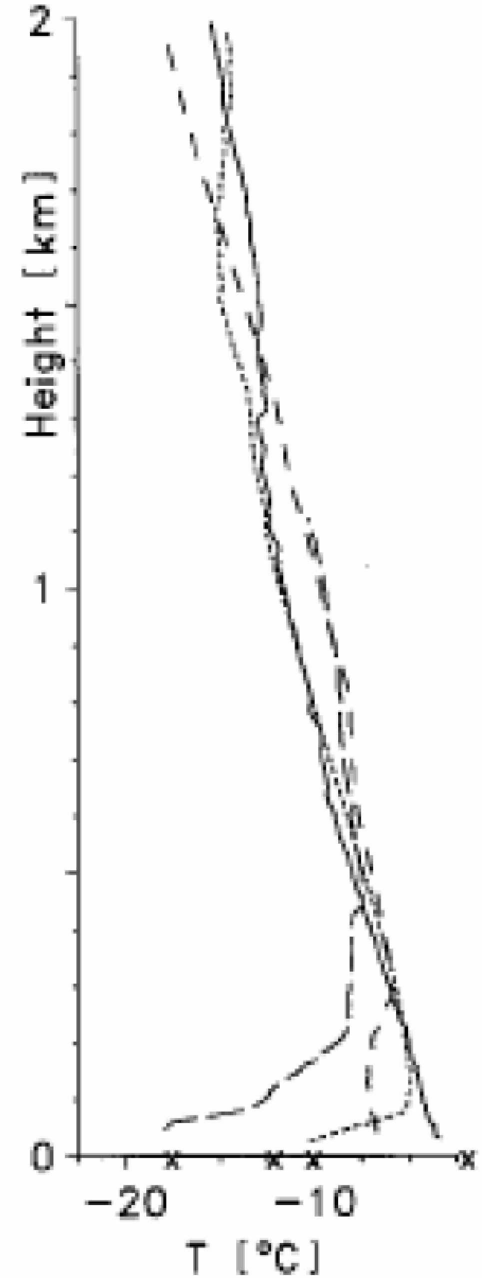
- shallow, stable boundary layer with downward turbulent fluxes
- distinct ice margin

- Essential in modelling: good vertical resolution in the ABL, accurate information on sea ice roughess (Vihma, Hartmann, Lüpkes, 2003), thermal interaction between the snow surface and the ABL with good resolution in snow and ice (Cheng and Vihma, 2002);

An example from the Arctic (Brümmer and Thiemann, 2002)

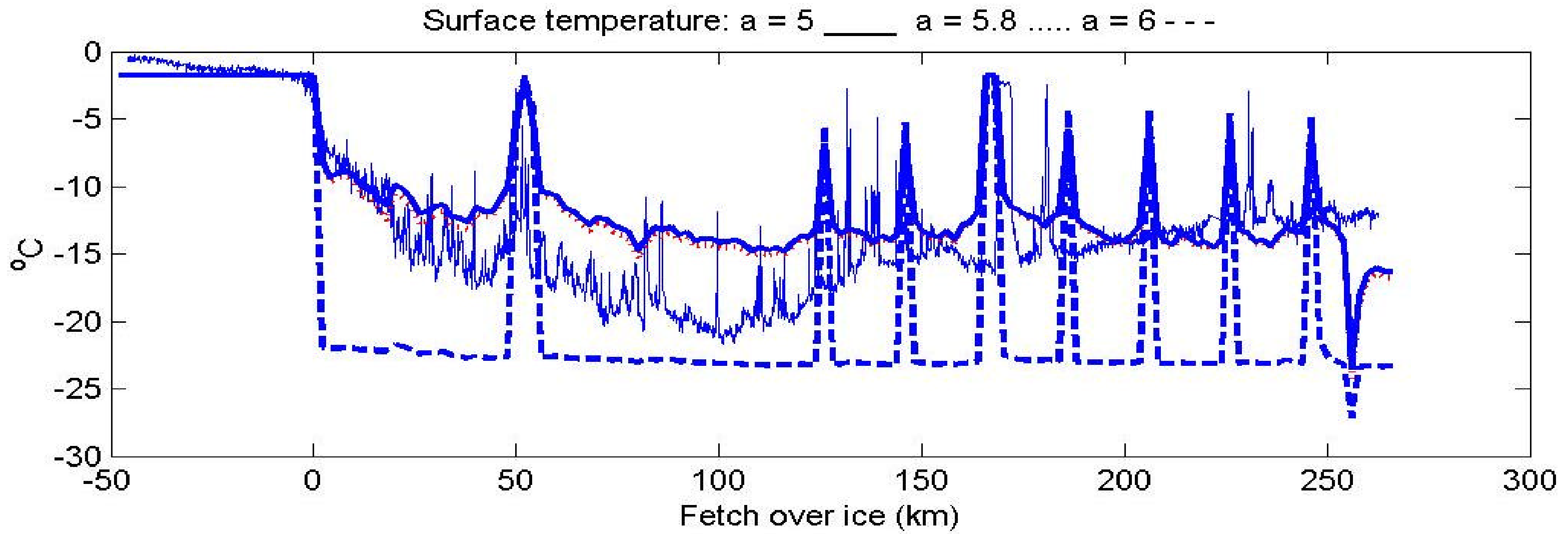


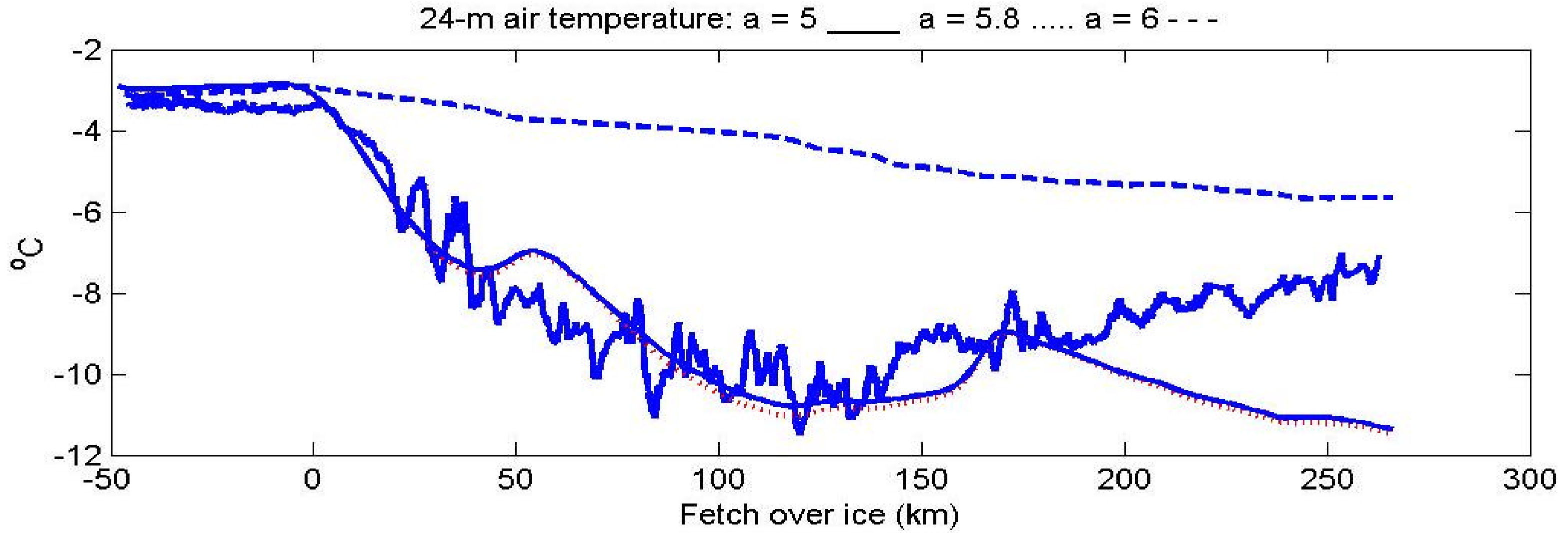




2D model results with different stability parameterizations

$C_H = C_{HN}\ f(Ri)\ ,\qquad f(Ri) = \max\ (\ 0.005,\ (1\text{-}a\ Ri)^2\ )$

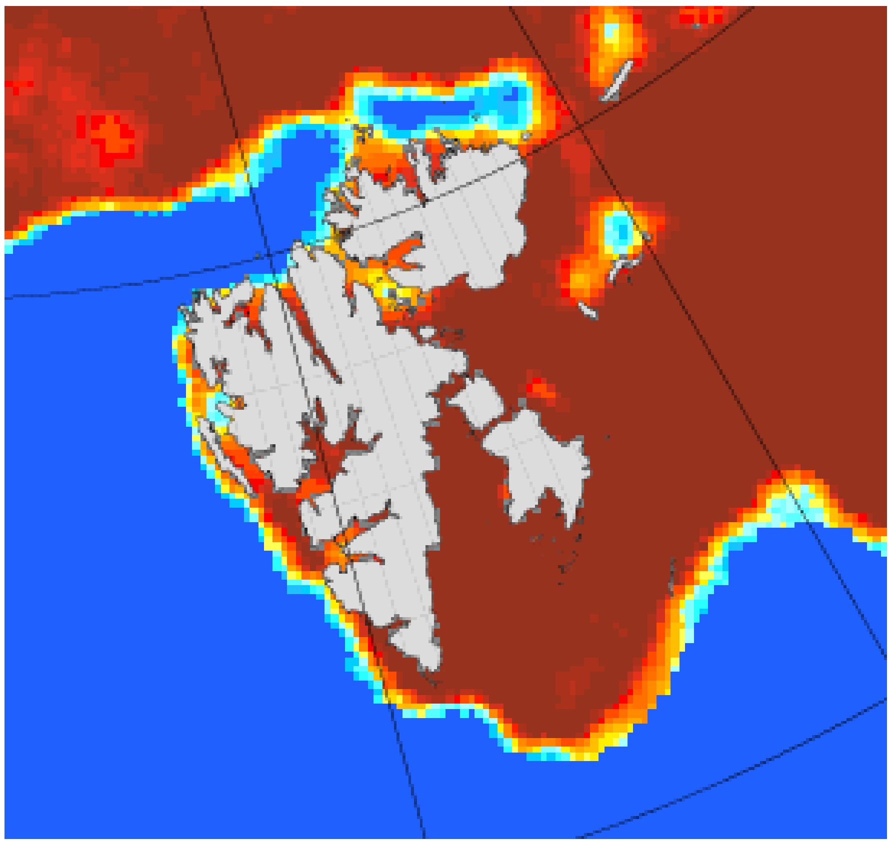




3. Flow parallel to the ice edge

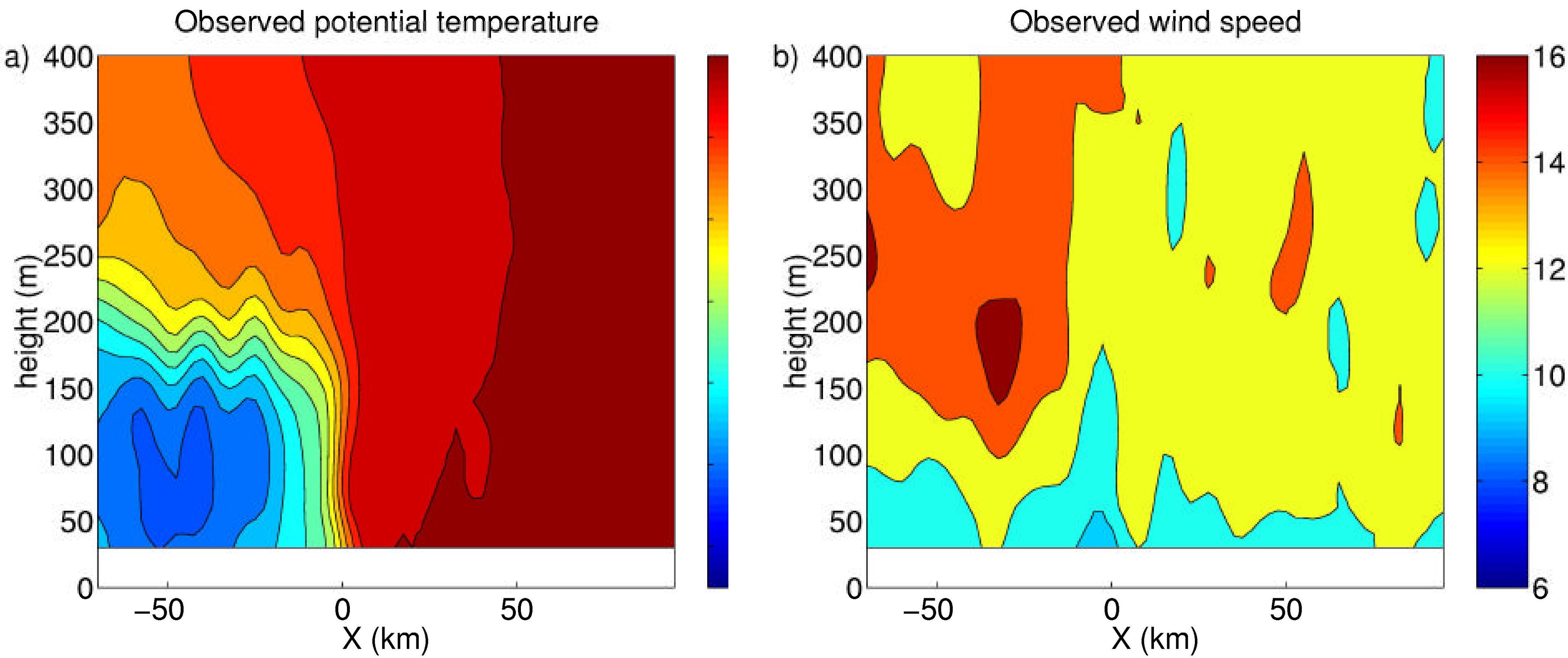
often a strong horizontal Ta gradient → very baroclinic ABL

Example: Svalbard, 30 March, 1998.



University of
Bremen ice chart

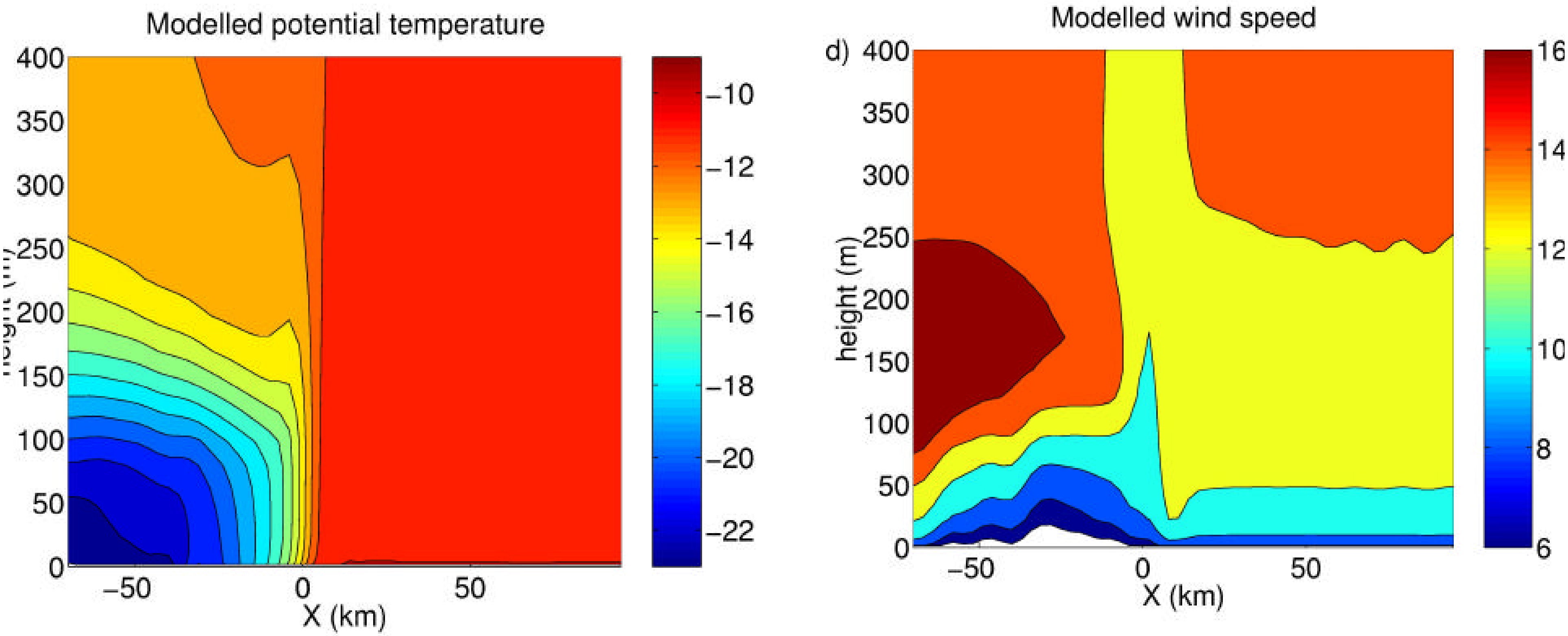
Aircraft observations by the Alfred Wegener Institute



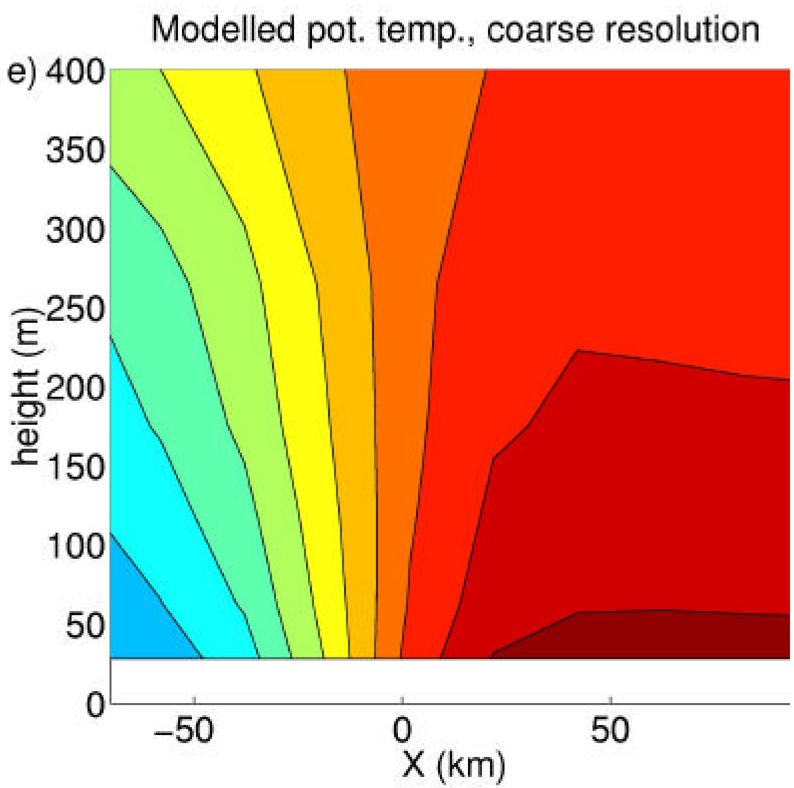
- ABL properties over sea ice and open ocean distinctly different from each other

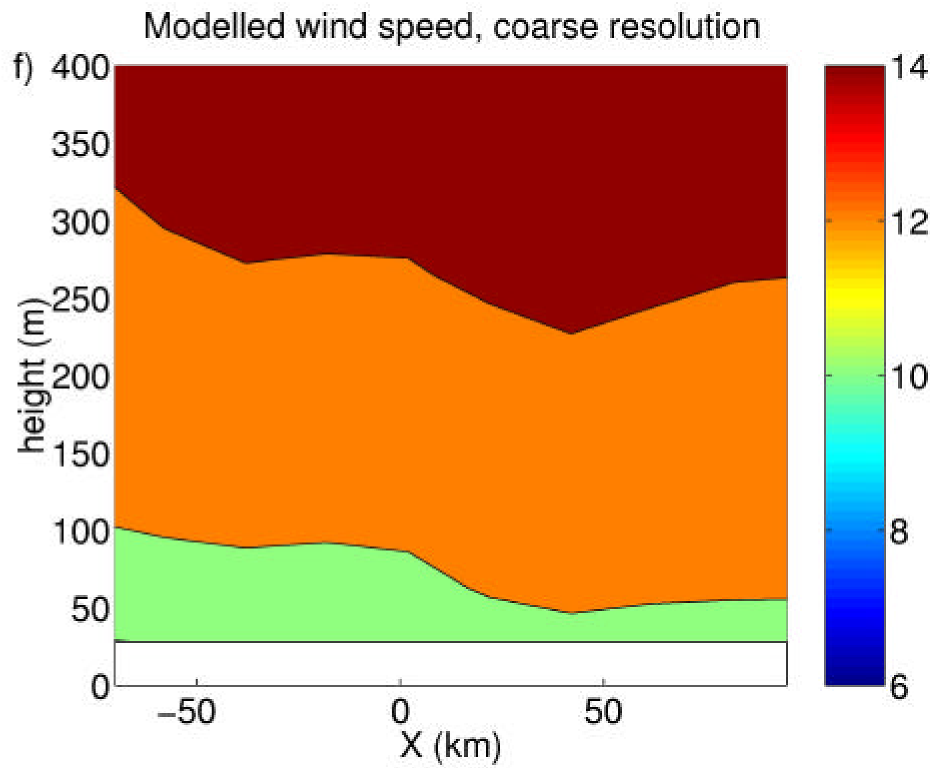
- much larger horizontal gradients than in flows perpendicular to the ice edge

Steady-state simulations with a 2D high-resolution model



- Strong inversion and a low-level jet due to baroclinicity well reproduced





- Results very sensitive to spatial resolution of the model and the ice concentration data

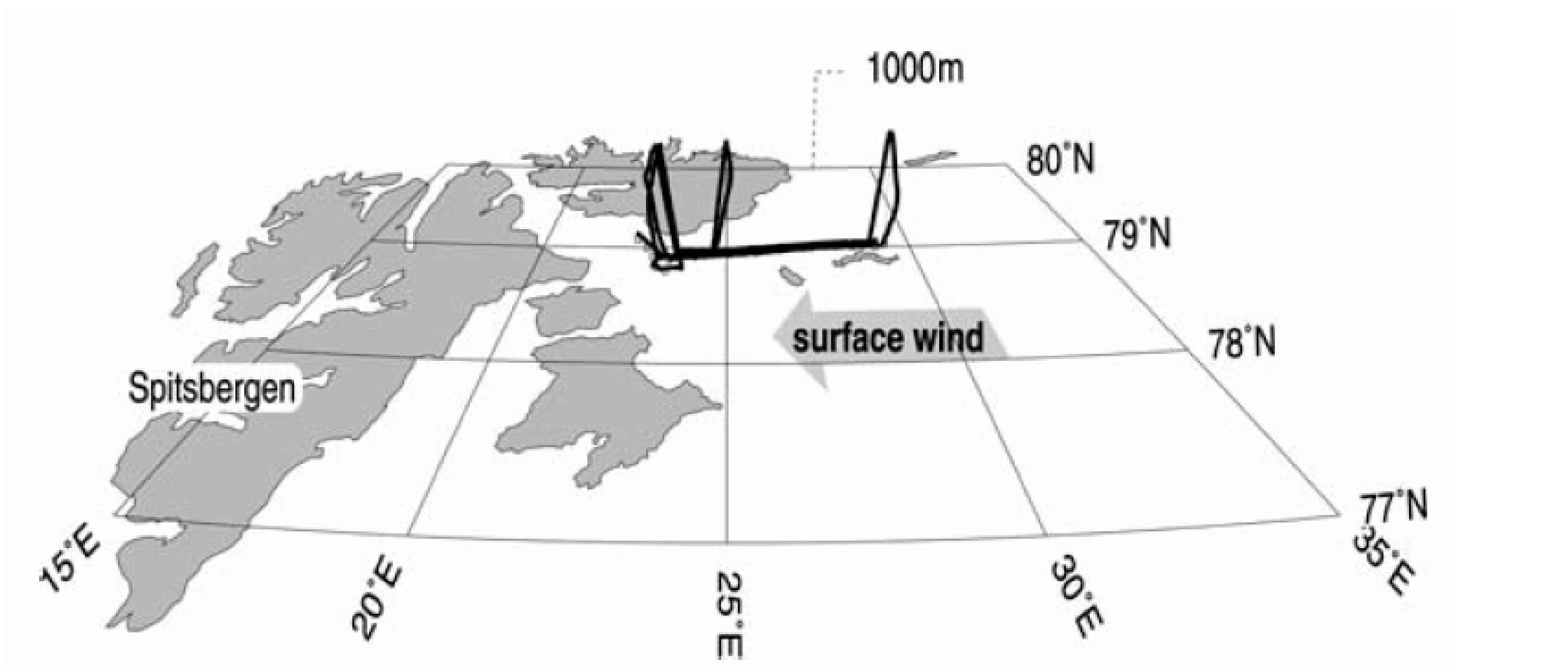
- Results not as sensitive to surface exchange as in the other cases, because ABL approximately in balance with the local surface

Cloud-top radiative cooling

Over sea ice (small z_0 and usually stable stratification), the principal source of turbulence may often be at the top of the ABL

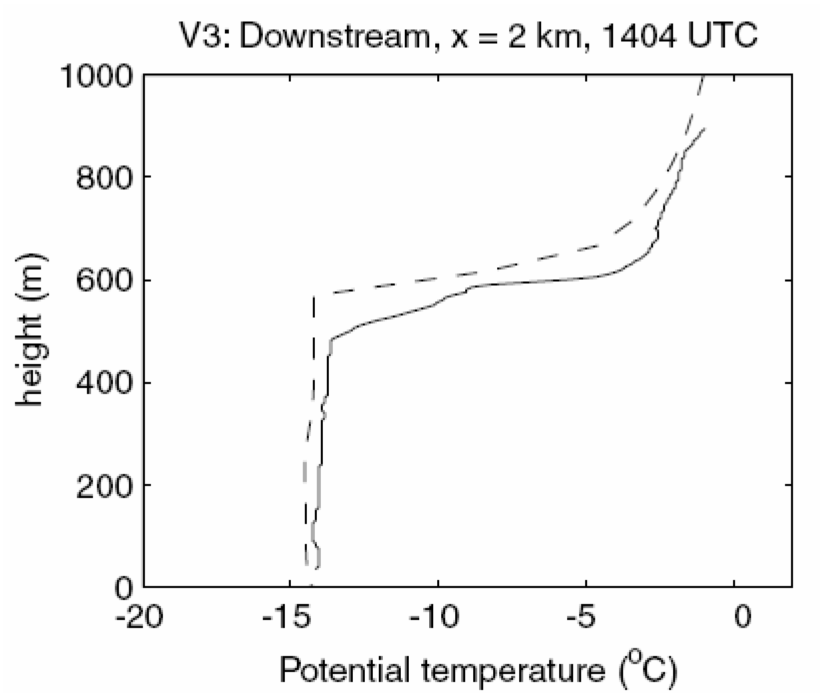
An example from the Barents Sea: AWI aircraft observations on 27 March, 1998.

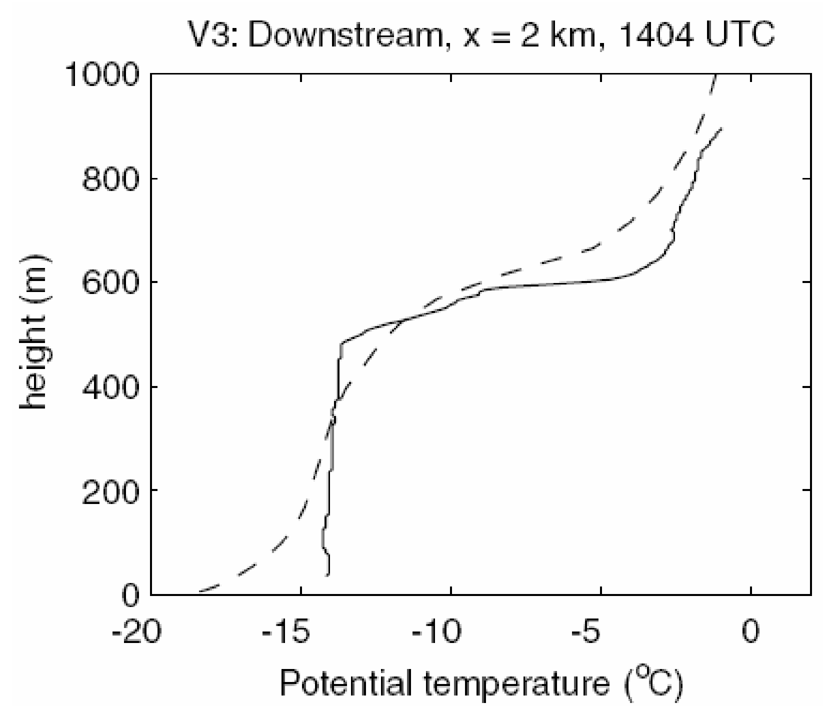
Upwind observations in the east were used as prescribed inflow boundary conditions for a 2D model.



If we prescribe the inflow boundary conditions and cloud cover according to the observations, we can reasonably well reproduce the ABL evolution

If we remove clouds from the model, the evolution of the ABL is very different.





Further sensitivity tests demonstrated that the mixed layers were maintained by the combined effects of cloud-top radiative cooling (→ top-down mixing), cloud forcing on the surface heat balance, and leads (Vihma, Lüpkes, Hartmann, Savijärvi, 2005).

Summary from Part II: Atmosphere over sea ice

1. Sea ice is a heterogeneous and rapidly changing surface

2. Sea ice – atmosphere interaction processes are usually most vigorous in the ice-edge zone, and strongly depend on the wind direction with respect to the ice-edge orientation

3. **Pre-requisites for successful modelling of the ABL processes include:**

- accurate information on the ice concentration and sometimes also on the surface roughness

- high horizontal and vertical model resolution

- observations and data assimilation: if we know the inflow boundary conditions, we can often reasonably well model the ABL evolution even in the case of a heterogeneous surface (leads) and interacting processes (radiation & turbulence)

Remaining problems in numerical modelling

Physics

Physics of the stably-stratified ABL, and its representation with a coarse-resolution model

Localized convection over leads and aggregation of surface fluxes. Even in the simplest case with two surface types (thick ice and open water), there is no perfect solution for the problem. Further, several ice types with different surface temperatures and albedos may co-exist.

Cloud and radiative processes, and their interaction with ABL turbulence: Errors in the modelled cloud cover often prevent succesfull presentation of the ABL.

Operational

1. **Inaccurate information on the ice concentration**
2. **Over thin ice, Ts is sensitive to the ice and snow thickness, but we do not have accurate information on these.**
3. **Inadequate information on the cloud cover**
4. **Boundary conditions may change rapidly, and informaton in models is sometimes too old: ice can drift up to 50 km / day**
5. **Polar lows and other mesoscale cyclones are not always detected by regular observations**

Suggested reading:

Boundary-Layer Meteorology, vol. 117, no. 1 & 2, 2005: Special Issues on the ABL over sea ice