Cloudy boundary layer: Radiation-PBL interactions (including aerosol and diamond dust aspects)

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- **Brief review of rad schemes and PBL cloud effects**
- **Warm fog: obs and ECMWF model experience**
- **Cold fog (ice mist): obs and UH 1D model experiments**

- Clear and cloudy PBL

- Solar (SW) and thermal (LW) radiation

- Fog = cloud at ground; horizontal visibility VIS = -ln(0.02)/β, where β = volume extinction coefficient (1/km) of fog (or of aerosol).

- (Empirically $\beta = 144.7 \cdot (\rho \cdot q_l)^{0.88}$ in fogs (Kunkel 1984), where ql is the liquid water mixing ratio, observed or modeled).

Solar radiation in clear-sky PBL:

- is scattered by molecules and various aerosols

- is absorbed by water vapour in six near-infrared bands and by black aerosol

- Reduced direct + diffuse flux at the surface
- Weak solar heating, ~ (3 K/day) μ_o in clear PBL, where μ_o is cosine of the solar zenith angle.

In models (NWP, GCM):

Direct radiation: extinction along rays by Beer's law

Diffuse radiation: **two-stream methods**: analytic solutions of coupled diff. equations for the up and down diffuse fluxes (various assumptions for hemispheric integrations)

Delta-two-stream-adding methods: put the forward delta-peak scatter of large particles to direct radiation; this improves accuracy. E.g. ”delta-Eddington” (DE) and ”delta-discrete-ordinates” (DD) methods for multiple scattering.

Input: μ_o and **air layer particle optics** (optical depth δ, single-scattering albedo ω, asymmetry parameter g) at each wavelength or wavelength band.

Output: Layer spectral transmittance, reflectance and absorptance. Finally,

’Add’ all layer fluxes including multiple reflections between them; **Add spectral fluxes** to total (broadband) SW fluxes up and down. **Heating = vertical convergence of net flux**.

Cloud/aerosol layers: Delta-TSA with cloud optics:

For spheres (radius r), Mie theory gives the scattering and absorption efficiencies Qsc, Qa. Dropsize distribution N(r) defines cloudlayer liquid water content LWC and effective radius re, and δ, ω and g are integrals of Q(λ,r)N(r) over r. (For a typical PBL sc, LWC ~ 0.3 g/m3 and re 7-8 μm.)

Liquid water path $LWP = LWC \cdot \Delta z = q_l \Delta p / g$

For **water clouds**, $r_e \approx 5 + 8 \cdot LWC$. Mie calculations suggest

$$\delta \approx LWP \cdot (a / r_e + b / r_e^{\,2}), \quad \omega \approx c + d \cdot r_e, \quad g \approx e + f \cdot r_e$$

a,…,f for four SW bands are given in Slingo (1989 JAS) and Savijärvi et al. (1997 QJ). For **ice clouds**, see Ebert and Curry (1992 JGR), for **raindrops**, Savijärvi (1997 Tellus).

For ‘European’ **industrial aerosol**: δ ~ 0.1, ω ~ 0.9, g ~ 0.8.

These methods produce e.g. the cloud SW albedo R and absorptance A as functions of LWP, re and μ_o:

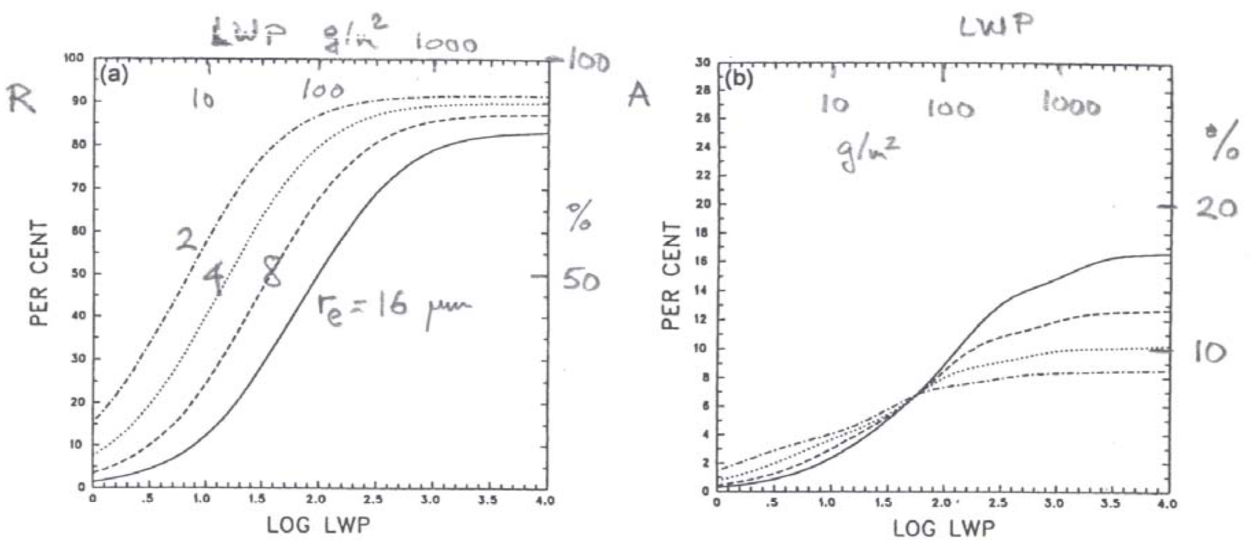


Figure 7. Cloud layer broad-band (a) reflectivity and (b) absorptivity for solar radiation (zenith angle 60°) as functions of the liquid-water path (LWP) ($1\text{–}10^4$ g m^{-2}), for effective radius of 16 μm (solid line), 8 μm (dashed line), 4 μm (dotted line) and 2 μm (dash-dotted line), using the parametrizations of Table 3.

(Savijärvi et al., 1997 QJ)

<u>Cloud/fog SW effects</u>:

Extinction of solar radiation in the cloud, and **solar heating** in a deep layer (~ 500 m).

This may dissolve PBL st, sc and fog during daytime.

Aerosol effect: Is basically similar but depends on how absorptive the particles are. White particles ($\omega = 1$) scatter only, black particles ($\omega << 1$) scatter and absorb, and heat the layer by their absorption.

<u>Problem</u>: For overcast sky and low sun, global radiation G is overestimated by ~15% by all delta-two-stream methods. In them the direct radiation scattered into the upward hemi-sphere is $\beta(\mu_o) = 0.5 - a\mu_o$ (e.g. a = 3g/(4(1+g)) in DE).

Solution:

For low sun ($\mu_o \approx 0$) some of the now nearly horizontal sun-rays scatter slightly up in the large particle deltapeak, not directly forward as is assumed. They thus escape upward. A simple correction for this, tested for dust particles in Mars' lower dusty atmosphere, is to define

$$\beta(\mu_o) = 0.5 + b - (a + b)\mu_o$$

where b ~ 0.1 for Martian mineral dust. This greatly improves all delta methods: error in G drops from 15% to < 3.5% (Savijärvi, Crisp and Harri, 2005 QJ).

The same can be applied to Earth cloud and aerosol particles. This provides the delta-two-stream methods with the typical accuracy of the more expensive four-stream methods.

<u>Thermal radiation in clear-sky PBL:</u>

Strong line emission and absorption by water vapour, CO2, O3,…, set into overlapping absorption bands. ’Continuum’ emission by H2O, notably in the LW window (8-12 μm).

<u>Models</u> (NWP, GCM): T(z) in grid points (only) with wideband emissivity schemes for gases tuned by line-by-line references. Continuum: ’Robert’ or ’Clough’ scheme.

<u>LW clear-sky PBL effects</u>: Above ~ 50 m: weak cooling (~ -2 K/day), Below: Strong LW cooling during moist, calm summer nights (~ -12 K/day); LW warming during day (Savijärvi, 2006 QJ).

If wind < 3 m/s, LW cooling dominates over turbulence (Savijärvi, 2006 QJ; Steeneveld et al., 2006 JAS). Radiation fog patches are then likely in moist conditions.

<u>Cloud/fog/aerosol layers</u>: <u>**A**bsorption approximation</u> is valid in the PBL. The **cloud emissivity** is then

$$\varepsilon_{cloud} = 1 - \exp(-k \cdot LWP)$$

where k ~ 0.15 m2/g is <u>mass absorption coefficient</u> and LWP the liquid water path of an air layer.

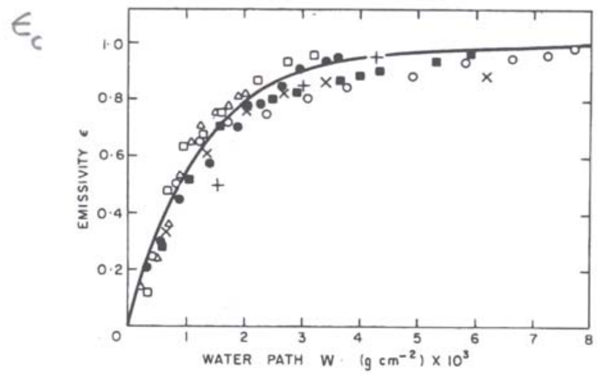


Fig. 8.8. The measured emissivity at 11 μm of layers of stratocumulus cloud decks versus liquid water path through the layers. Clouds sampled were the same as in Fig. 8.7.

(Paltridge and Platt, 1976)

Mie calculations again produce detailed parametrizations for k = k(λ, re), e.g. $k_{ice} \approx 1.35 / r_e(\mu m)$ (Savijärvi and Räisänen, 1998 Tellus).

For PBL ice clouds/fogs/mists, re ~ 35 μm so k is small, k ~ 0.04 m2/g: ice clouds are **semitransparent** to LW radiation.

Aerosols are even more transparent in LW and are therefore often neglected

<u>PBL cloud LW effects</u>:

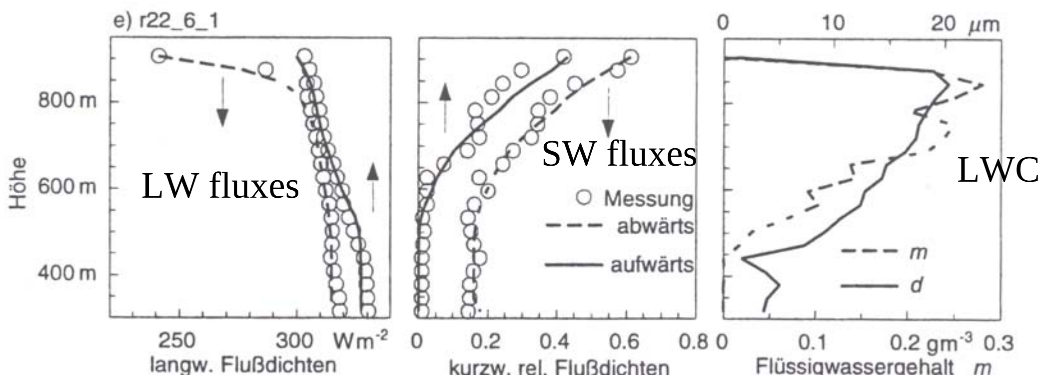
• **Strong LW cooling at cloud top in a shallow ~50 m layer** (clouddrops emitting to space and cooling)

• **Warming at the cloud base** (clouddrops net absorbing emission from the warmer ground beneath).

This drives **turbulence within the cloud during daytime**, and may drive **'top-down' turbulence down to the surface during the night**. Turbulence creates entrainment.

Models should be able to describe the thin-layer LW cooling/ heating peaks realistically. Strong LW cooling peak may need implicit treatment numerically. All existing rad schemes are explicit, however.

AWI aircraft observations through a PBL stratus:



Typical SW and LW fluxes in a sc-topped PBL (Stull, 1988):

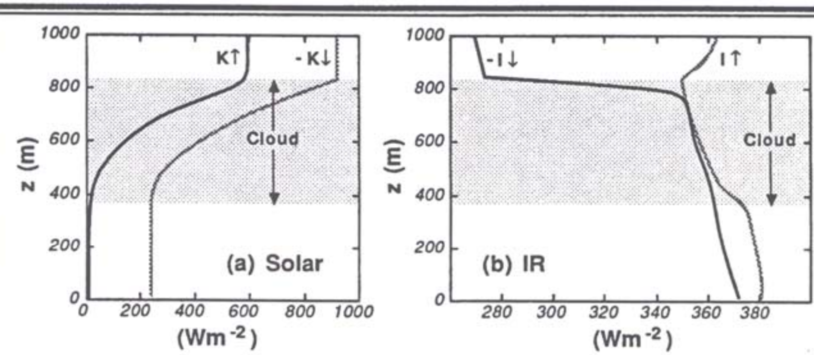


Fig. 13.4 Average (a) solar, K, and (b) IR, I, radiation fluxes. The extent of the stratocumulus cloud layer is shown by the shading. (After Nicholls, 1984).

Typical net fluxes and heating rates in a sc-topped PBL:

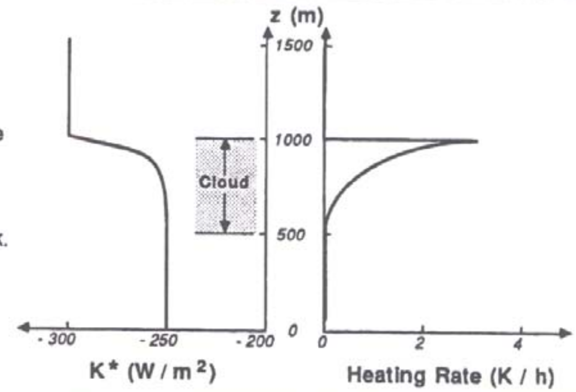


Fig. 13.5
Profiles of net shortwave radiative flux and the corresponding heating rate in an idealized stratocumulus deck. (After Hanson, 1987).

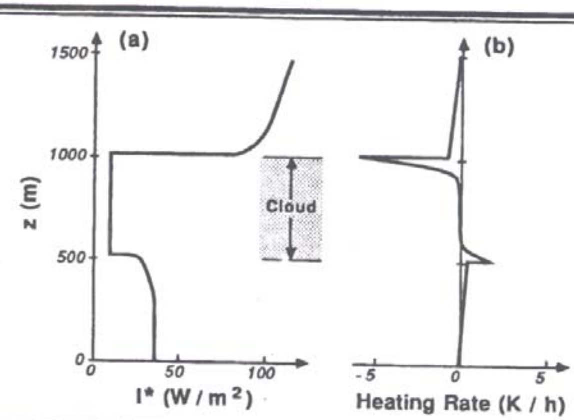


Fig. 13.7
Profiles of net longwave radiative flux and corresponding heating rate in an idealized stratocumulus deck. (After Nicholls and Leighton, 1986).

Physical processes in a sc-topped PBL:

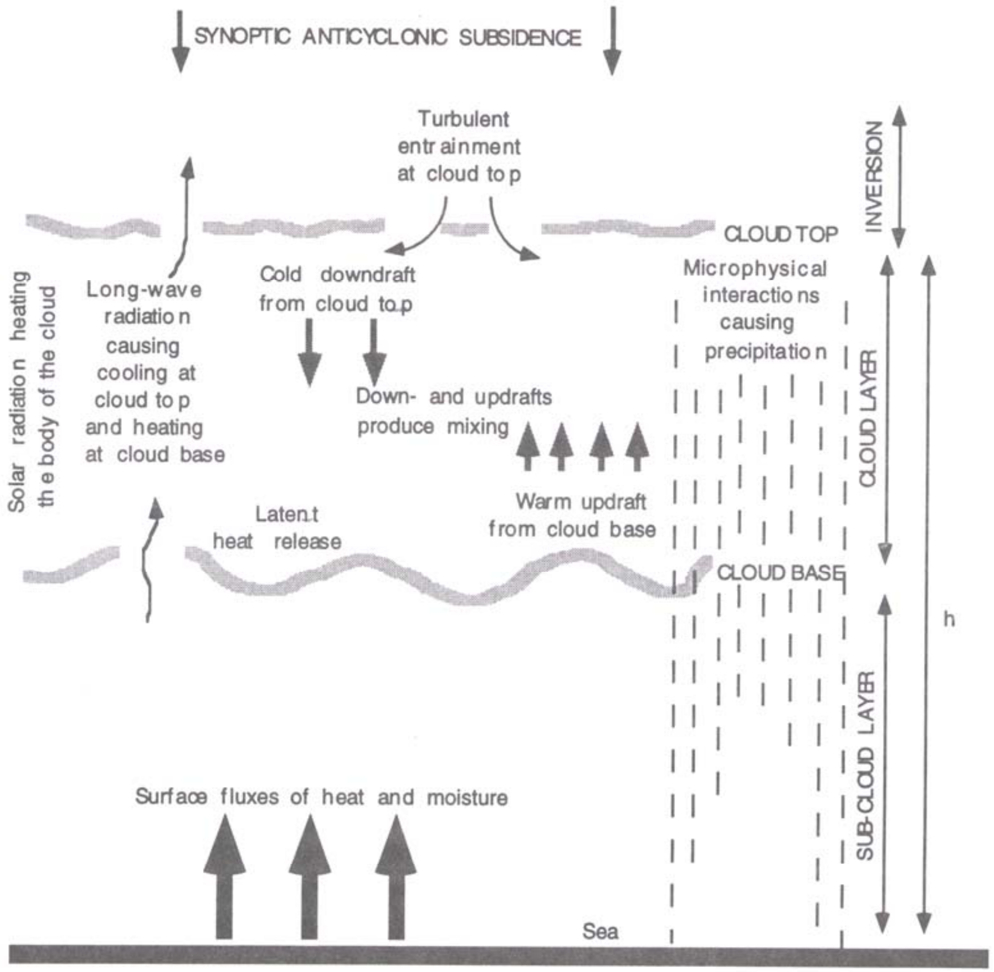


Fig. 2.1 - The fundamental physical processes in the development of marine stratocumulus (from Nieuwstadt Duynkerke 1996).

Problems in radiation when modeling the cloud-covered BL

- Cloud property inputs for radiation!!
- Vertical resolution!
- Cloud inhomogeneities!
  (need for prognostic cloud cover and its variance)
- Sloping ground (in high horizontal resolution)
- Aerosol concentrations and types
  (aerosol optics input is often the largest source of uncertainty in SW, especially in clear conditions)

Empirical rules by Finnish duty forecasters for the very common stratus-topped wintertime PBL with inversion:

- if wind is weak, coldest spot is at the surface and fog results

- if wind is stronger, a mixed layer results, with st/sc formed at its top by mixing. Cloudbase (in m) is at ~ 125(T-Td).

- even if wind then relaxes, st/sc once created by mixing survives (it is now driven by its top LW cooling)

- if windy PBL air is at about -12C, water vapor condenses to ice crystals, which sediment down and a thin stratus may disappear -> more outradiation, colder -> again fog or low icy st which sediments down -> more outradiation and really cold.

- also, if thick mixed st top is at around -12C, snowbands are initiated (by the difference between ice and water saturation).

These can be used e.g.to test models

Fog: Saturation by cooling (radiation fogs); by moistening (advection fogs); by mixing (e.g. sea smokes)

An example: Warm radiation fog in Cabauw, 3 Aug 1977 (Teixeira, 1999 QJ; ECMWF 1-D model simulation)

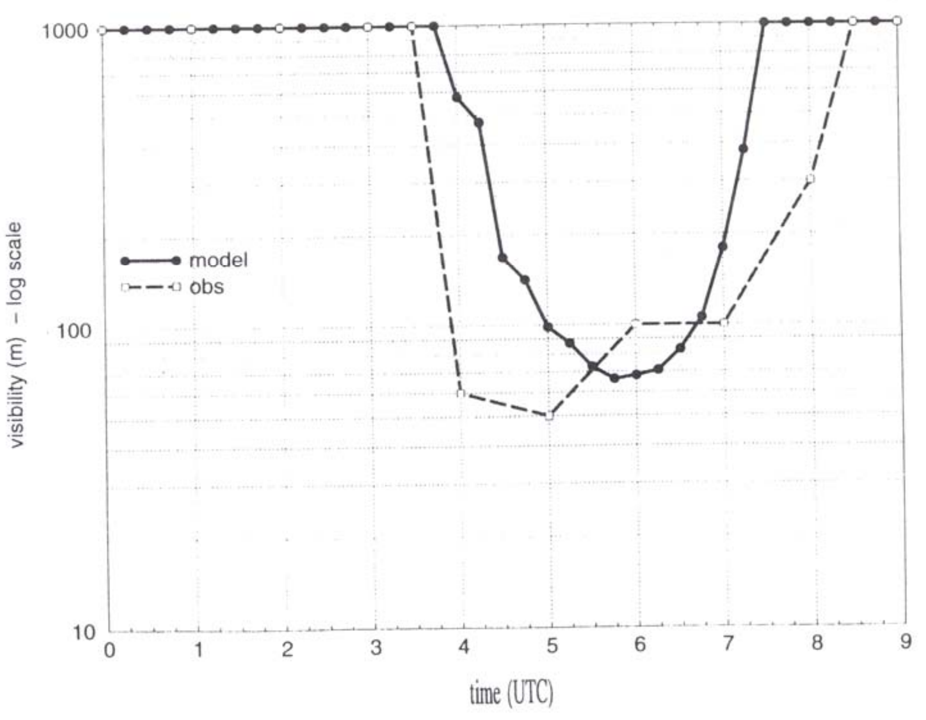


Fig.5.2 - The time evolution of the simulated visibility at the lowest model level together with the visibility observations from the Cabauw tower at the 30 m level.

q evolution 00-05h at the lowest level (30 m) in three 1-D experiments: control, no rad, no rad + no cold advection

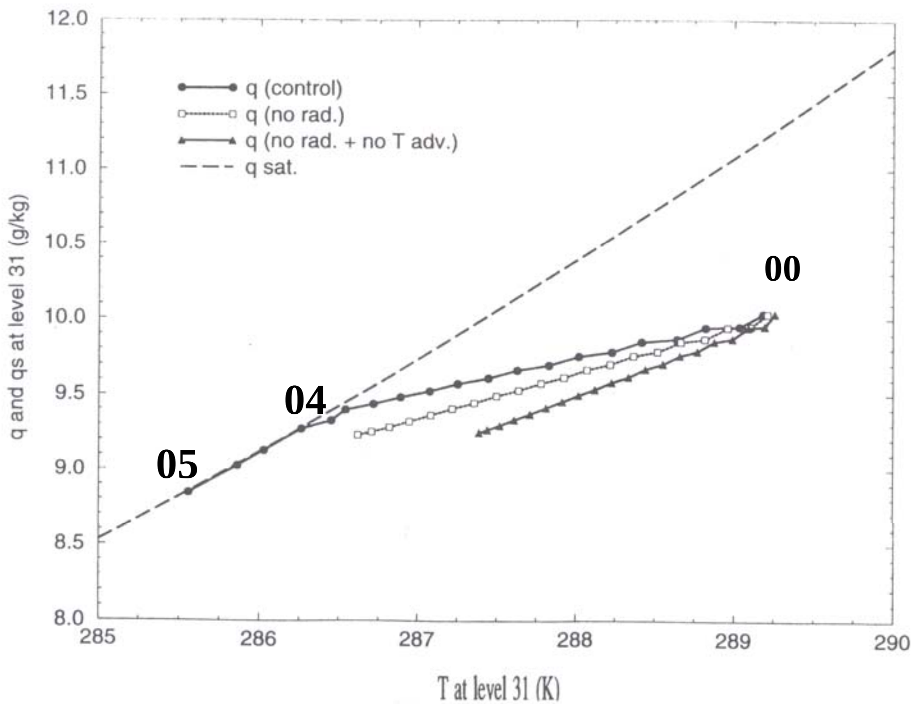


Fig.5.6 - Clausius-Clapeyron diagram, of the lowest model level, for the first 5 hours of 3 different experiments: control, control without radiative cooling of the atmosphere and control without radiative cooling of the atmosphere and horizontal cold advection.

Obs fog vs. ECMWF op. 24h fcsts in Europe: LWC vs V10m

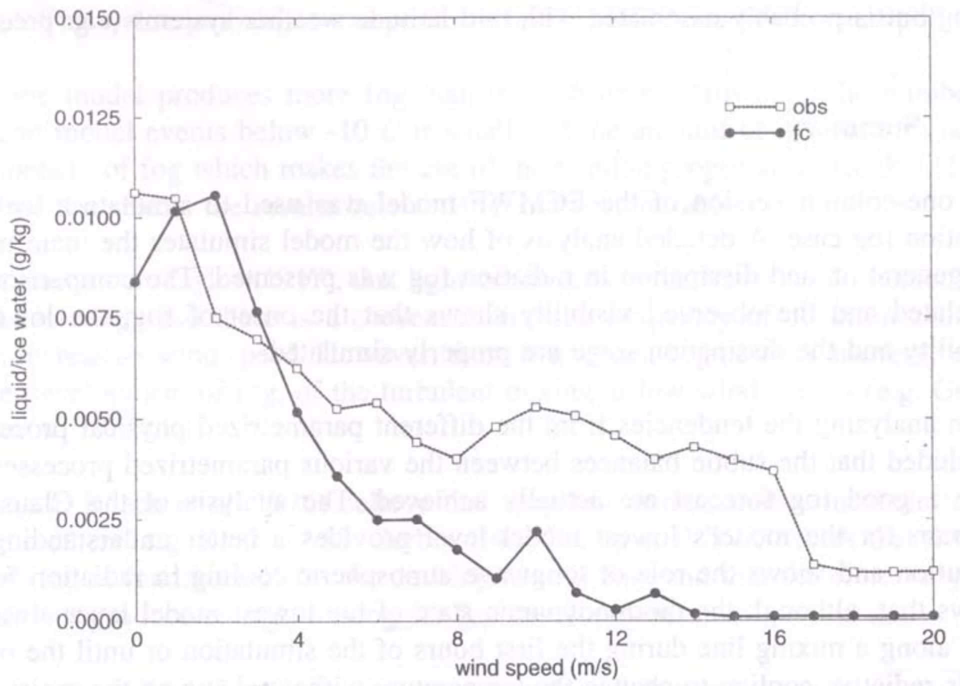


Fig.5.26 - The average liquid/ice water content versus the wind speed at 10 m (forecasts and observations). T
average is performed dividing the wind speed in intervals of 1 m/s and uses the observations as in fig.5.25.

The Cabauw fog and other case studies (e.g. Savijärvi and Kauhanen, 2001 TAC) stress the delicate balance between several physical processes in fog formation and evolution. LW cooling of moist air is necessary in initiating radiation fogs, and LW cooling from the fogtop then starts to drive it. SW heating or wind increase may dissolve it.

<u>**Ice fogs**</u>: Lasse Makkonen has shown, using heated hygro-meters, that **supersaturations with respect to ice** are common in Northern Finland during winter. They are often reached in clear, calm weather via cooling of air with q conserved.

In cold conditions (T < -18C) such supersaturations are often associated with good vertical visibility but **’diamond dust’** (tiny ice crystals) is falling from clear sky. Horizontal visibility is 1-10 km, i.e. thin and shallow **ice mist** prevails.

Makkonen and Laakso (2005 BLM): RHi measured during winter in Northern Finland, using a heated humicap (Vaisala HMP243) at 4 m height:

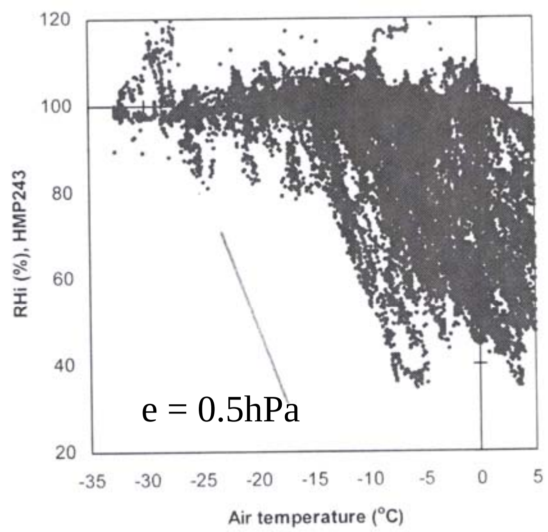


Figure 7. Relative humidity with respect to ice RH_i as measured by HMP243 versus air temperature (Pyhätunturi).

A typical weakly supersaturated scene, T4m -19.4C, RHi 105.9%, good vertical visibility and clear sky, horizontal visibility ~ 5 km, diamond dust falling:

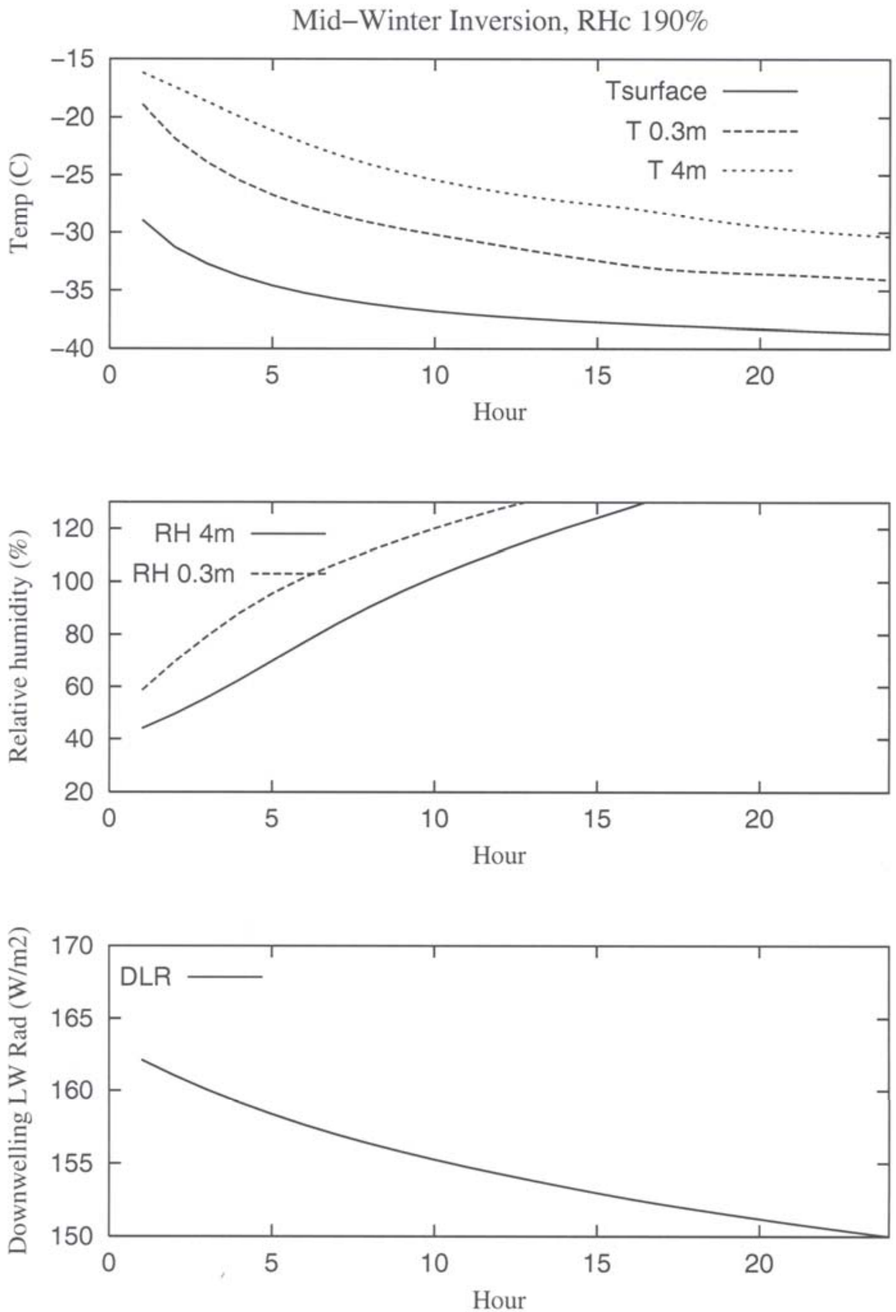
Is the cooling due to LW radiation or turbulence? Is further cooling of ground and air stopped by formation of mist? If so, is it due to latent heating or rad effects of ’diamond dust’?

To answer these questions (by Makkonen) we use the UH 1-D model in a typical N Finland midwinter case:

Model: - M-O surface layer, Blackadar-type turbulence
- Narrowband LW scheme, Roberts continuum, k = 0.04 m2/g
- Ice formation at RHcrit with latent heating and rad feedbacks
- High vertical resolution (lowest model level at 30 cm)
- 5-level optimized snow scheme (Savijärvi 1992 BPA)

Case: - snow and sfc air initially at –15C, lapse rate 2 C/km and RH 30%, Vg 0.1 m/s. No solar radiation (midwinter).

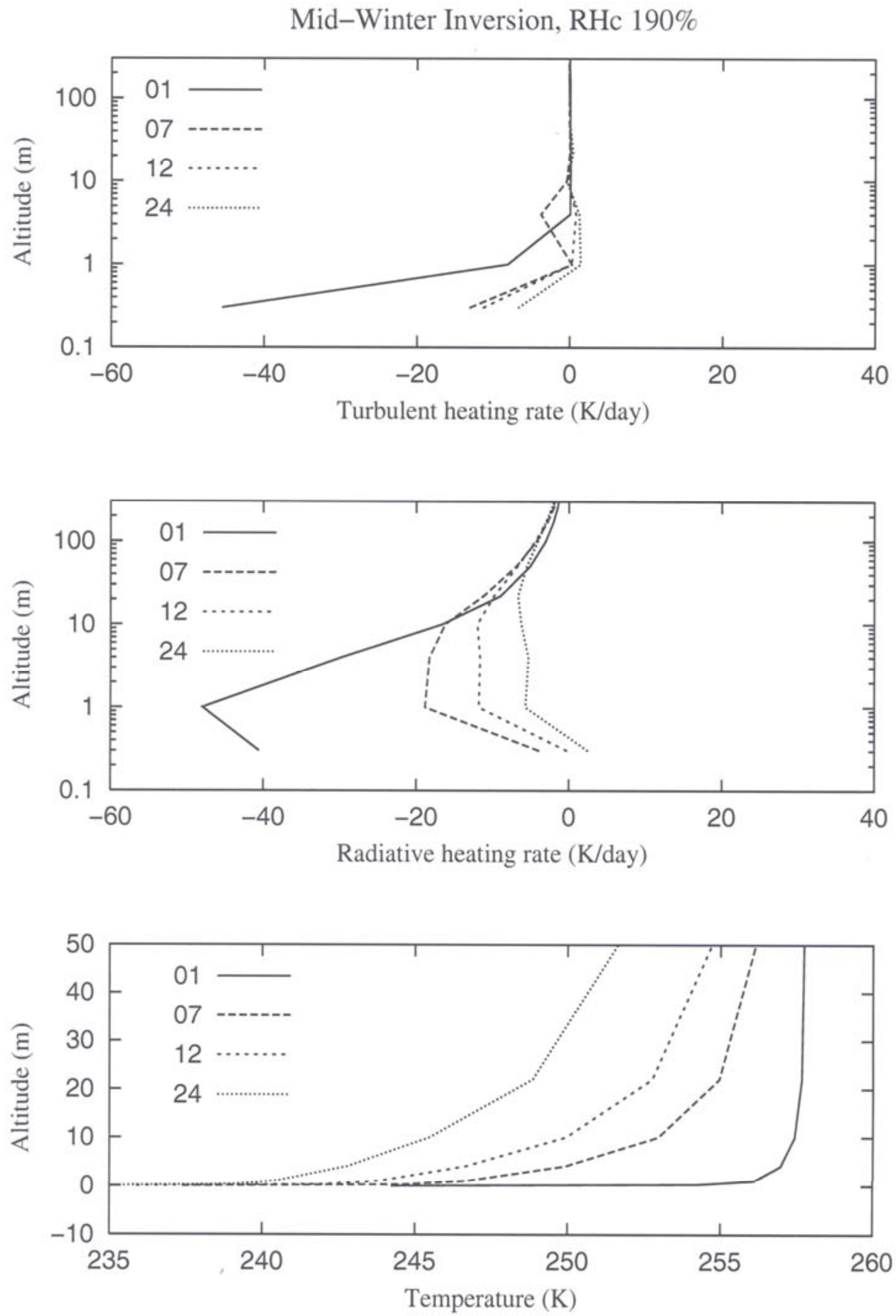
The model then cools to a typical ~200 m high wintertime surface inversion in a few hours.



<u>Time evolution, no fog case</u>: Temperatures decrease with time from the initial –15C. Snow surface is cooling via LW net radiation loss to space but gets some heat from warmer snow beneath

Relative humidities increase as air temperatures decrease. There is no fog as RHcrit for cloud formation is set here (artificially) to 190%

Downwelling longwave radiation decreases monotonically as air temperatures decrease

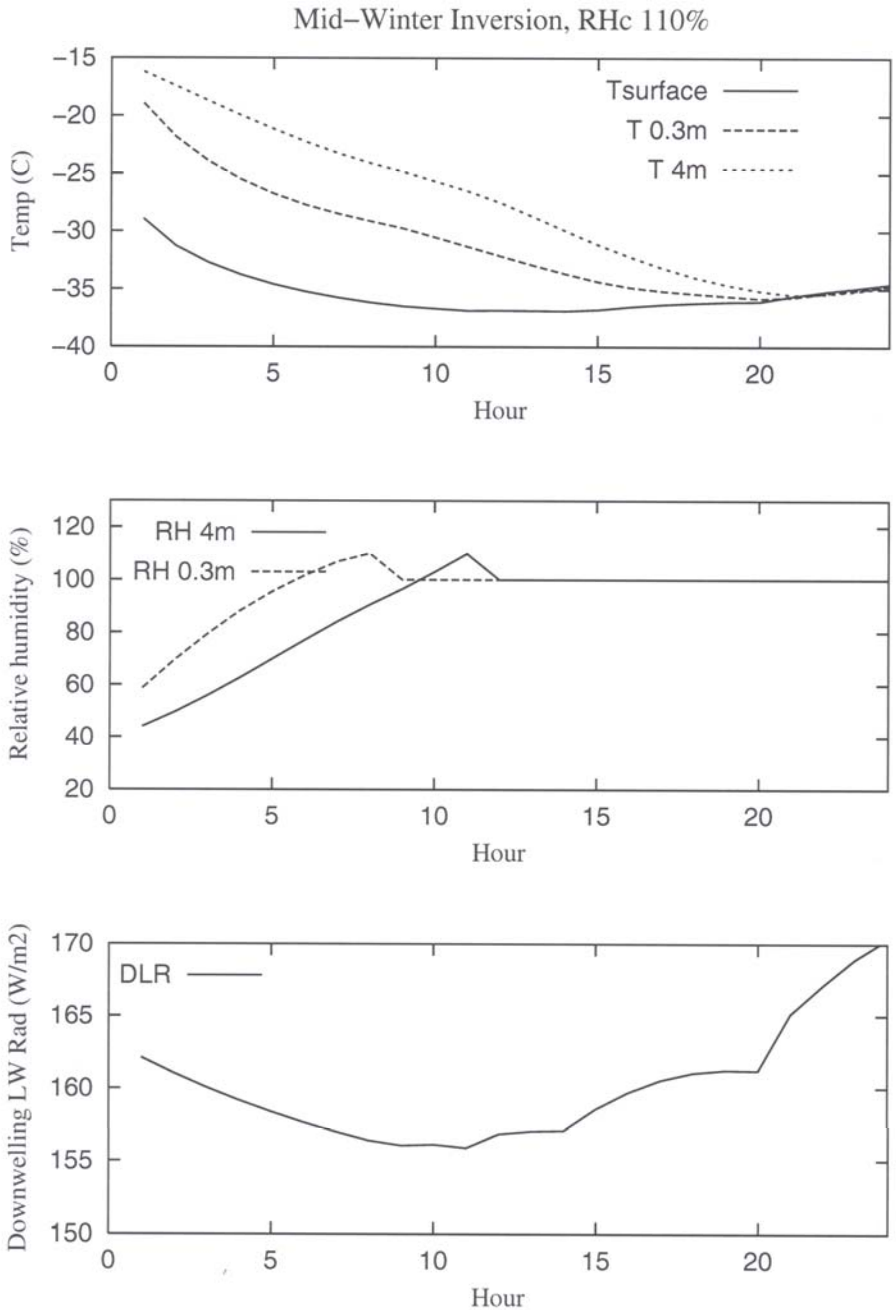


<u>No fog case, profiles,</u> 1, 7, 12 and 24 hours after start:

Turbulent cooling is weak (as wind is weak). It is concentrated to the shear layer next to the surface

LW cooling dominates over turbulent cooling across the growing inversion

Temperature profile evolves into a typical Lapland winter inversion. Inversion top is at about 200 m.

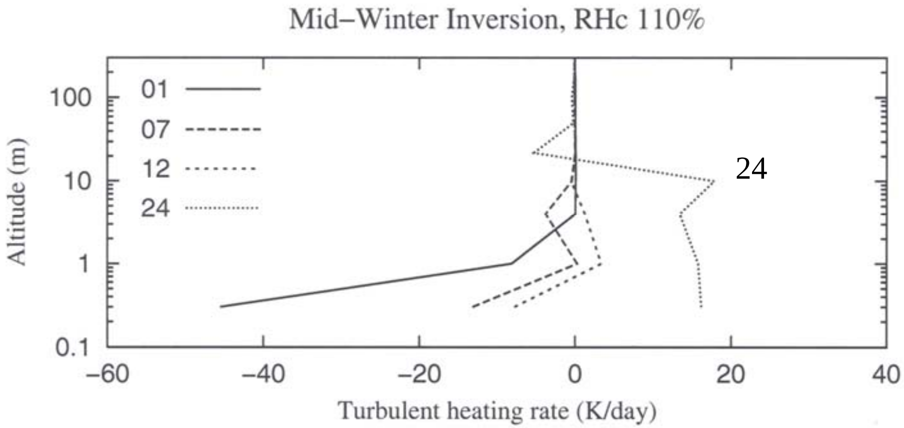


<u>Fog case evolution</u>:
Fog forms first at 07h at 0.3 m

The surface temperature decrease stops at 11-12h. At 20h surface becomes slightly warmer than air.

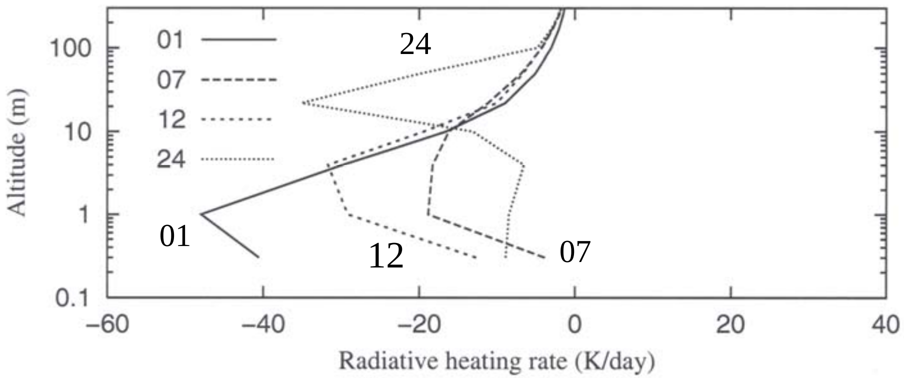
Fog is formed once RHi exceeds the critical value of 110%. (Foggy air stays at RHi = 100%, incorrect in our models?)

Downwelling LW radiation starts to increase soon after fog formation (by ice crystal emission). This stops further sfc temperature decrease

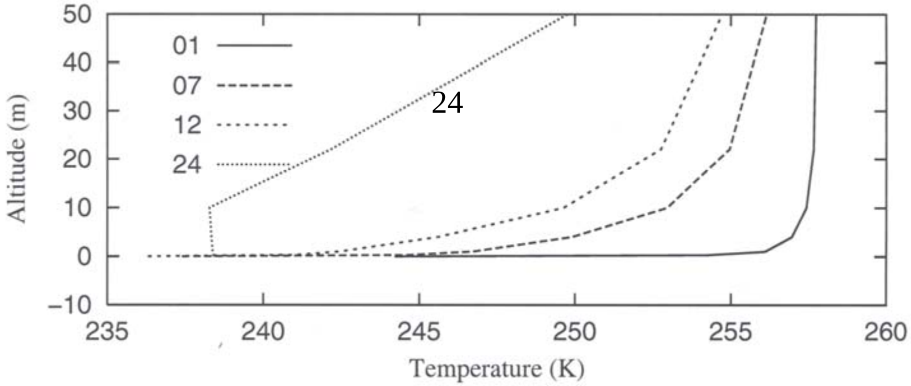


Fog case profiles:

Turbulent cooling changes to heating (convection) in the well-established fog by 24h, as Tsfc then is warmer than Tair



Top of the fog LW-cools strongly, creating more fog



The temperature profile within thick fog (24h) evolves toward moist adiabat, with fog-DLR increasing the sfc temperature

Further experiments with the UH 1-D model:

-If k = 0 (i.e. fog has no LW effect), temperatures drop and fog forms with latent heating effects, but the temperature evolution is rather similar to the no fog case.

-If all latent heating effects are shortcut but the fog LW effect is included (k = 0.04 m2/g), the temperature evolution is similar to the ’full fog’ case.

-Latent heating thus plays only a minor role in thin ice mist. The major role in limiting the T decrease is in the increased LW emission by diamond dust to the ground.

-For cold extremes, a **dry**, cold and calm continental airmass is needed, as condensation into ice crystals is then delayed.

Conclusions

- Small-drop SW effects are important for small-drop clouds, fogs, aerosols $(\delta = LWP \cdot (... + b / r_e^2))$
- Simple delta-TSA improvement for low sun
- RHi > 100% may be common in ice mists
- Cold extremes are obtained in <u>dry</u> airmasses, where DLR by thin ’diamond dust’ finally limits the decrease of surface temperature
- Latent heat release is insignificant for cold fogs