Radiative heat transfer and its role in stable PBLs

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- **Brief review of rad schemes and rad effects on PBL**
- **Warm fog: obs and ECMWF model experience**
- **Cold fog: obs and UHel 1D model experiments**

(fogs are common in a stable PBL, they are harmful for traffic, and are not always well-predicted)

- Clear and cloudy planetary boundary layer (PBL)

- Solar (SW) and thermal (LW) radiation

- Fog = cloud at ground; horizontal visibility VIS < 1 km, VIS = -ln(0.02)/β, where β = volume extinction coefficient

- Empirically $\beta = 144.7 \cdot (\rho \cdot q_l)^{0.88}$ in fogs (Kunkel 1984), where ql is the <u>liquid water mixing ratio</u> (predicted by models)

$$q_l \leftrightarrow VIS$$

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 global radiation at the surface
- Weak solar heating, $\sim$ (3 K/day) μ_o in clear PBL, where μ_o is cosine of the solar zenith angle.

In models (NWP, GCM):

Direct radiation: exponential extinction along rays (Beer's law)

Diffuse radiation: **two-stream methods**: analytic solutions of coupled differential equations for diffuse fluxes up and down (with assumptions for hemispheric integration)

Delta-two-stream-adding methods: put the strong forward deltapeaked scatter of large particles to direct radiation; this improves accuracy (E.g. the popular ”delta-Eddington” method for multiple scattering).

Delta-TSA input: μ_o and **spectral 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.

’Add’ layer fluxes including multiple reflections between all layers; **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 drops (= spheres), Mie theory gives scattering and absorption efficiencies Q. Dropsize distribution N(r) defines cloudlayer liquid water contents LWC and effective radii re. δ, ω and g are integrals of $Q(\lambda,r)N(r)$ over r.

Vertical 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). Savijärvi et al. (1997 QJ) added small-drop effects (b), which are important for PBL clouds/fog/aerosol. For **ice clouds**, see Ebert and Curry (1992 JGR), for **raindrops**, Savijärvi (1997 Tellus).

For 'European' **industrial aerosol**: $\delta \sim 0.1$, $\omega \sim 0.9$, $g \sim 0.8$.

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

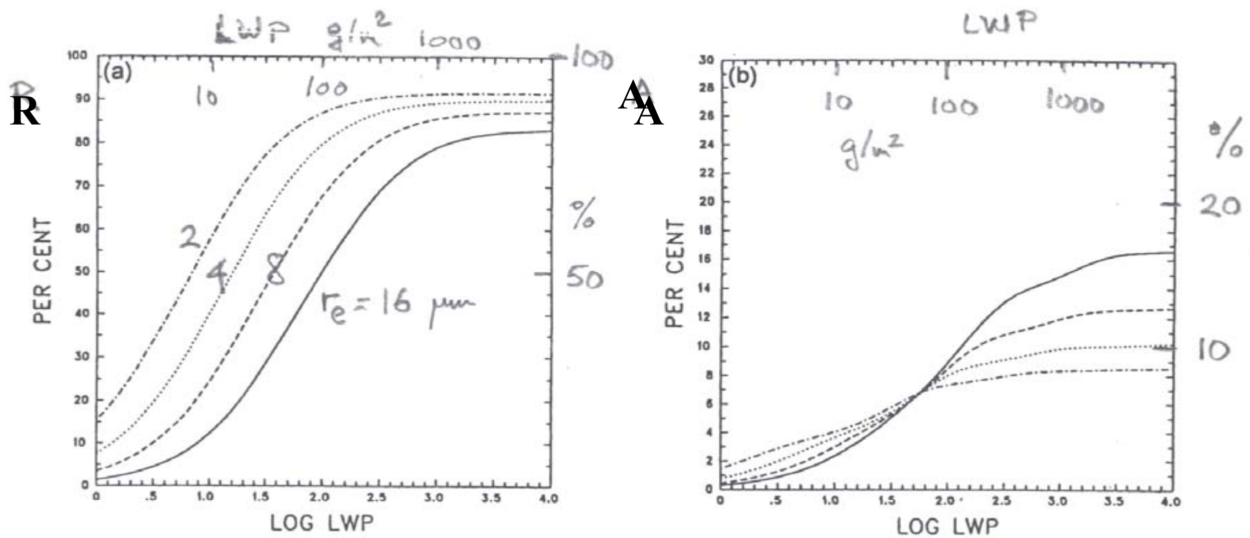


Figure 7. Cloud layer broad-band (a) reflectivity and (b) absorptivity for solar radiation (zenith angle 60°) as functions of the liquid-water path (LWP) ($1–10^4$ g m^{-2}), for effective radius of 16 μm (solid line), 8 μm (dashed line), 4 μm (dotted line) and 2 μm (dash-dotted line), using the parametrizations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

(Savijärvi et al., 1997 QJRMS)

(note the' Twomey effect': small-drop cloud is whiter than large-drop cloud for the same cloudwater amount)

Cloud/fog effects on solar radiation:

Extinction of solar radiation by the cloud,

Solar heating at the top, in a deep layer (~ 500 m).

This may dissolve st, sc and fog during daytime.

Aerosol effects: Basically the same but depend on how absorptive the particles are. White particles ($\omega = 1$) scatter only, black particles ($\omega < 1$) scatter and absorb, and heat the aerosol layer by their absorption.

(Dust effects on Mars' climate as the function of δ, ω, g: see Savijärvi, Crisp and Harri, 2005 QJRMS)

<u>A 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 hemisphere is defined $\beta(\mu_o) = 0.5 - a(g)\mu_o$

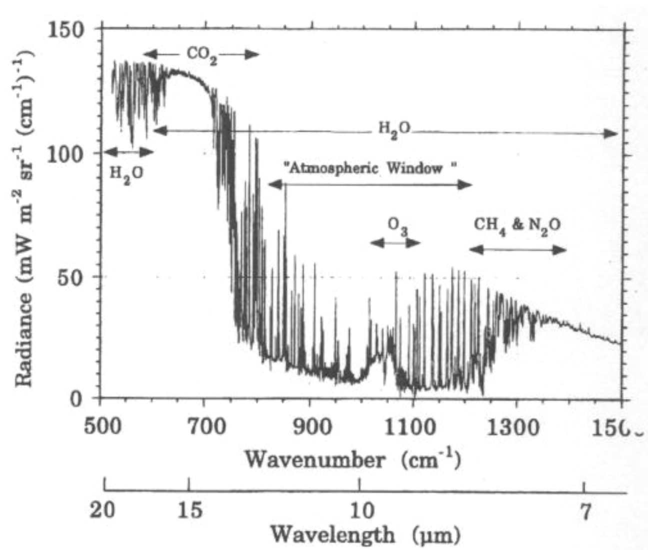
For low sun ($\mu_o \approx 0$) some of the nearly horizontal sunrays do scatter **slightly upward** in the foreward peak (and escape); **not directly forward** as is assumed by delta scaling. A simple compensation (tested for dust particles in Mars' lower quite dusty atmosphere) is to redefine $\beta(\mu_o) = 0.5 + b - (a+b)\mu_o$ where b = 0.1 for Martian mineral dust. This greatly improves all delta methods: max errors in G drop from 15% to < 3.5% (Savijärvi, Crisp and Harri, 2005 QJ).

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

Thermal radiation in clear-sky PBL:

Strong line emission and absorption by water vapour, CO2, O3,…, in **overlapping absorption bands**

’Continuum’ emission by H2O, notably in the LW window (8-12 μm).



Part of the observed LW spectrum from clear-sky zenith atmospheric emission, measured by a Wisconsin spectrometer at 2000 UT on 6 December 1991 (from Ellingson and Wiscombe, 1996).

Models (NWP, GCM): T(z), q(z) in grid points, wideband LW emissivity schemes for gases (tuned by line-by-line references), H2O continuum: 'Robert' or 'Clough' scheme. -> Downwelling LW radiation (DLR), LW cooling rates.

LW clear-sky PBL effects:

Above about 50 m: weak cooling (~ -2 K/day)

Below about 50 m: 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 therefore likely in moist, calm conditions, due to LW cooling.

<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 is mass absorption coefficient (~ 0.15 m2/g for water clouds) and LWP the liquid water path.

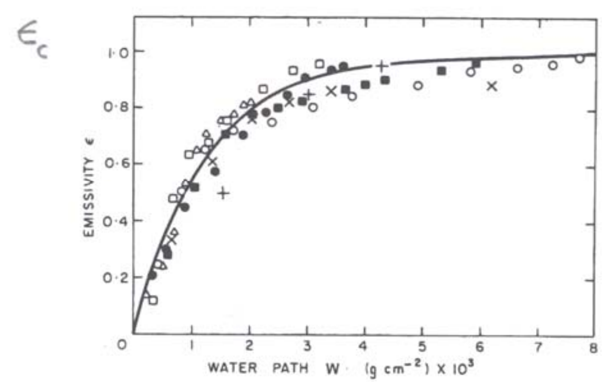


Fig. 8.8. The measured emissivity at 11 μm of layers of stratocumulus cloud decks versus liquid water path through the layers. Clouds sampled were the same as in Fig. 8.7.

(Paltridge and Platt, 1976)

Mie calculations produce more detailed LW-paramet-rizations for k = k(λ, re). E.g. $k_{ice} \approx 1.35 / r_e(\mu m)$ for ice clouds (Savijärvi and Räisänen, 1998 Tellus).

For PBL ice clouds/fogs/mists, re ~ 35 μm, so k is small, (k ~ 0.04 m2/g). Thin, large-particle ice clouds are therefore **semitransparent** to LW radiation.

Thin aerosols are even more transparent in LW. They are therefore often neglected (rightly or wrongly).

PBL cloud LW effects:

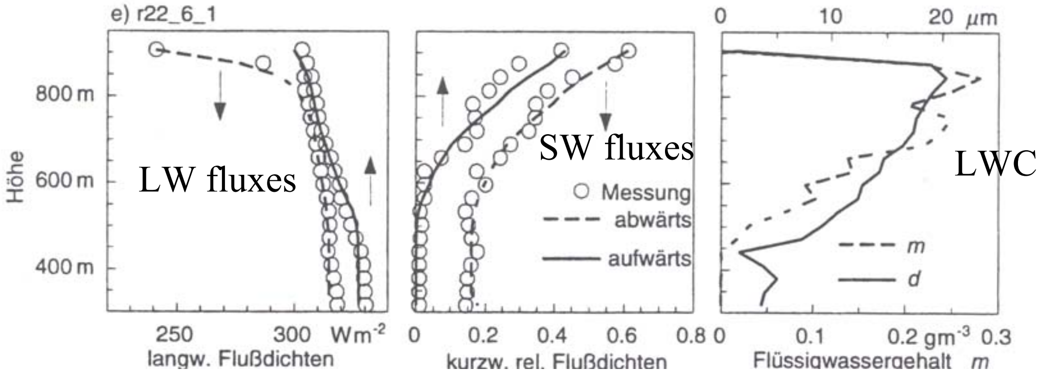
• **Strong LW cooling at cloud top in a shallow ~50 m layer** (clouddrops emitting to space and cooling)

• **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 peaks may need implicit treatment numerically. All existing rad schemes are explicit, however.)

Aircraft observations through PBL stratus:



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

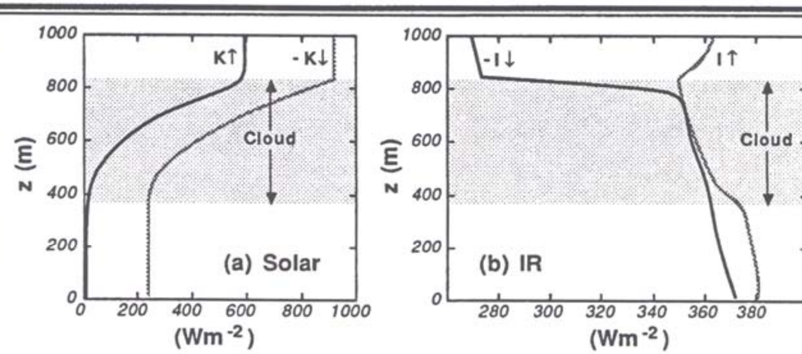


Fig. 13.4 Average (a) solar, K, and (b) IR, I, radiation fluxes. The extent of the stratocumulus cloud layer is shown by the shading. (After Nicholls, 1984).

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).

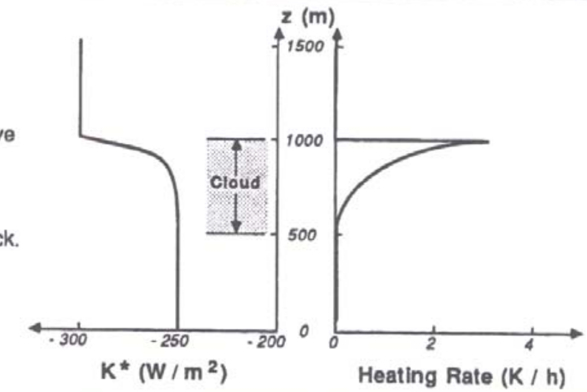
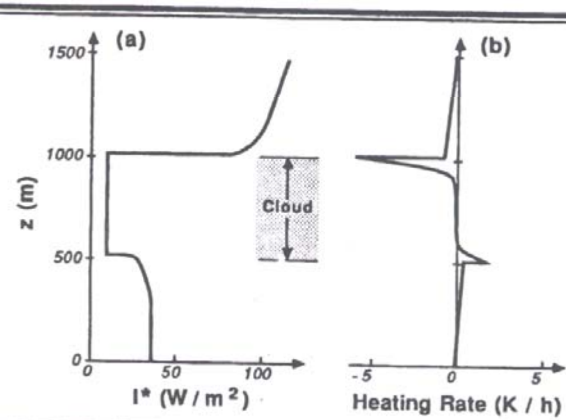


Fig. 13.7
Profiles of net longwave radiative flux and corresponding heating rate in an idealized stratocumulus deck. (After Nicholls and Leighton, 1986).



Physical processes in a sc-topped PBL:

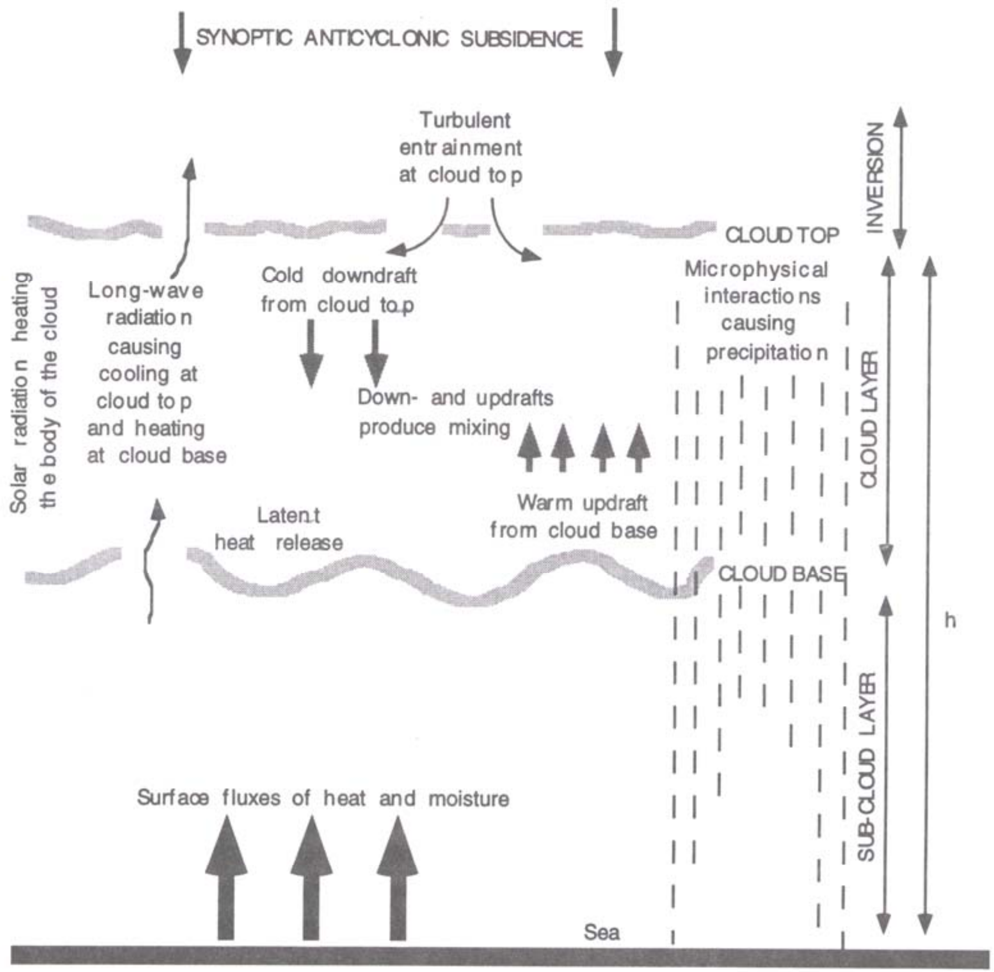


Fig. 2.1 - The fundamental physical processes in the development of marine stratocumulus (from Nieuwstadt Duynkerke 1996).

Problems in radiation when modeling the BL

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

Empirical rules by Finnish duty forecasters for stratus-topped wintertime very stable PBL (these can be used e.g. to validate models):

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

- If wind is stronger, a mixed layer results near the surface, with sc formed at its cold top by mixing. Cloudbase is at ~ 125(T-Td).

- Even if wind now relaxes, st/sc once created by mixing survives since it gets driven by LW cooling at its top.

- If windy cloud air is at about –12 C, water vapor condenses to ice crystals, which sediment down and a <u>thin</u> cloud may disappear -> colder -> again thin fog or low icy stratus which may sediment -> more LW-outradiation and really cold.

- If <u>thick</u> mixed windy stratus has its top at around –12 C, snowbands are often initiated.

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 as a case study)

VIS (m)

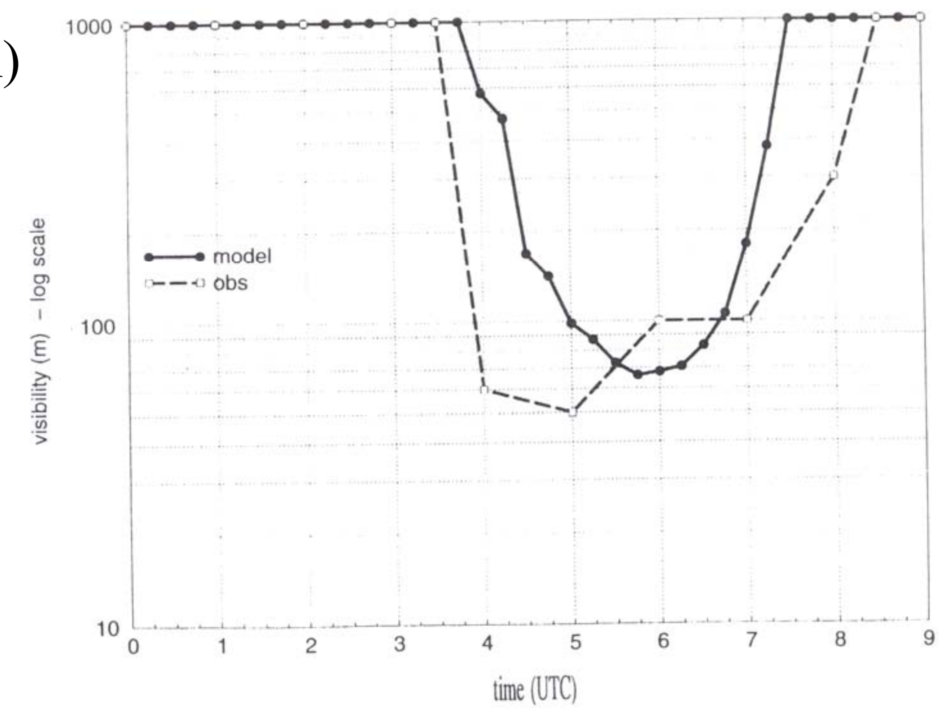


Fig.5.2 - The time evolution of the simulated visibility at the lowest model level together with the visibilit observations from the Cabauw tower at the 30 m level.

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

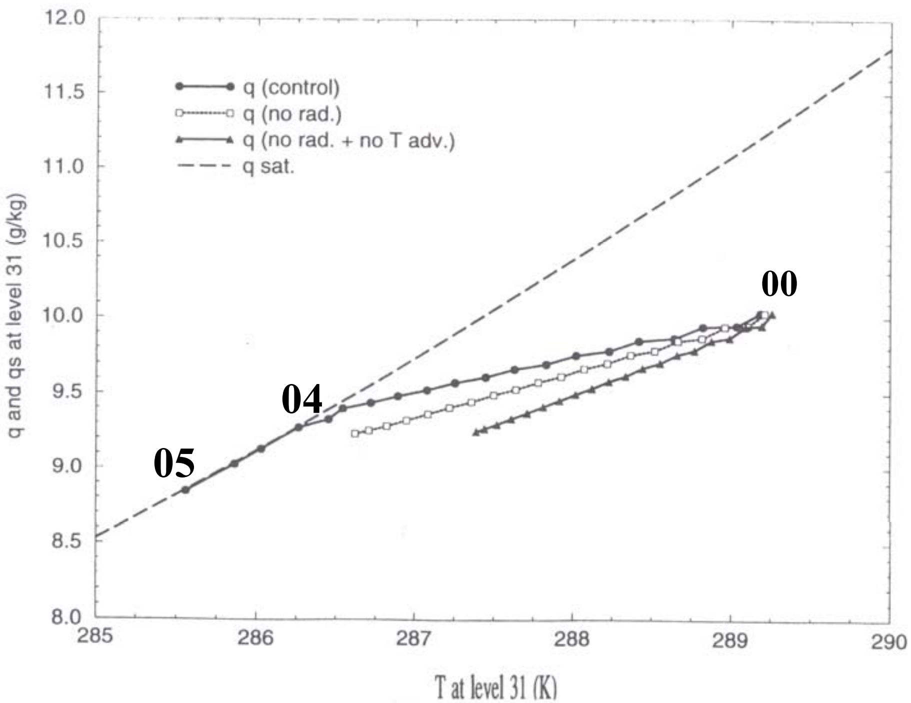
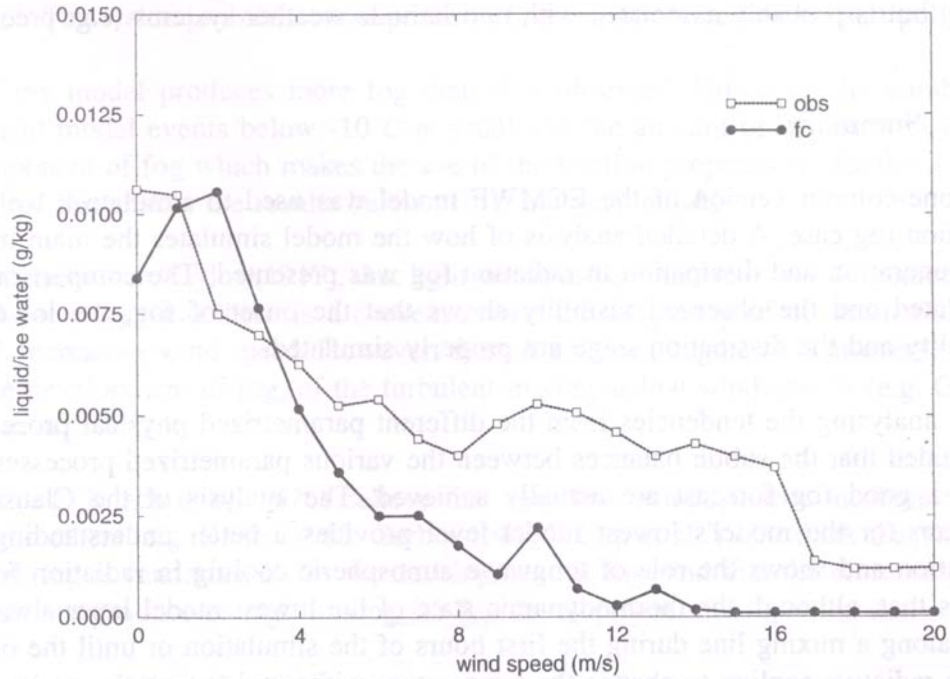


Fig.5.6 - Clausius-Clapeyron diagram, of the lowest model level, for the first 5 hours of 3 different experiments: control, control without radiative cooling of the atmosphere and control without radiative cooling of the atmosphere and horizontal cold advection.

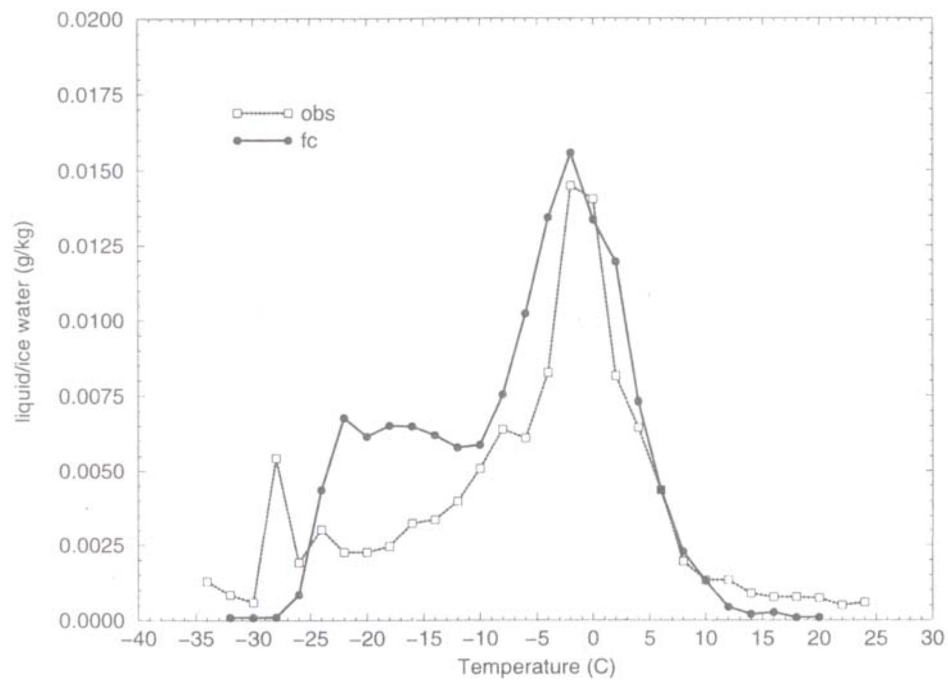
Obs fog and ECMWF 24h op.fcsts in Europe: LWC vs V10m



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

(weak wind (i.e. radiation) fogs are fairly well simulated,
stronger wind (i.e. advection) fogs not as well)

Obs fog and EC 24h fcsts: LWC vs. T 2m



5 - The average liquid/ice water content versus the 2m temperature for the model and the observations. The avera is performed dividing the temperature in intervals of 2°C and uses all the observations over Europe for the perioc January 1996, which is around 100000 observations.

(warm fogs are well simulated, observed thinning beyond –12C not so well)

The Cabauw 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. LW cooling from the fogtop then starts to drive it. SW heating or increase of wind may dissolve it.

Ice fogs: Lasse Makkonen has shown, using heated hygro-meters, that **supersaturations with respect to ice** are common in a cold and moist climate (Northern Finland during winter).

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

Makkonen and Laakso (2005 BLM): RHi vs. T

Observations during winter in Northern Finland, using a heated humicap (Vaisala HMP243) at 4 m height:

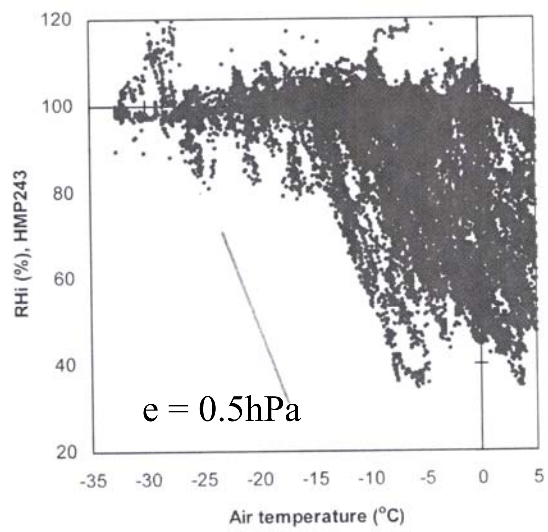


Figure 7. Relative humidity with respect to ice RH_i as measured by HMP243 versus air temperature (Pyhätunturi).

A typical supersaturated scene: T4m -19.4C, RHi 106%, good vertical visibility and clear sky, horizontal visibility ~ 5 km, diamond dust falling from clear sky:

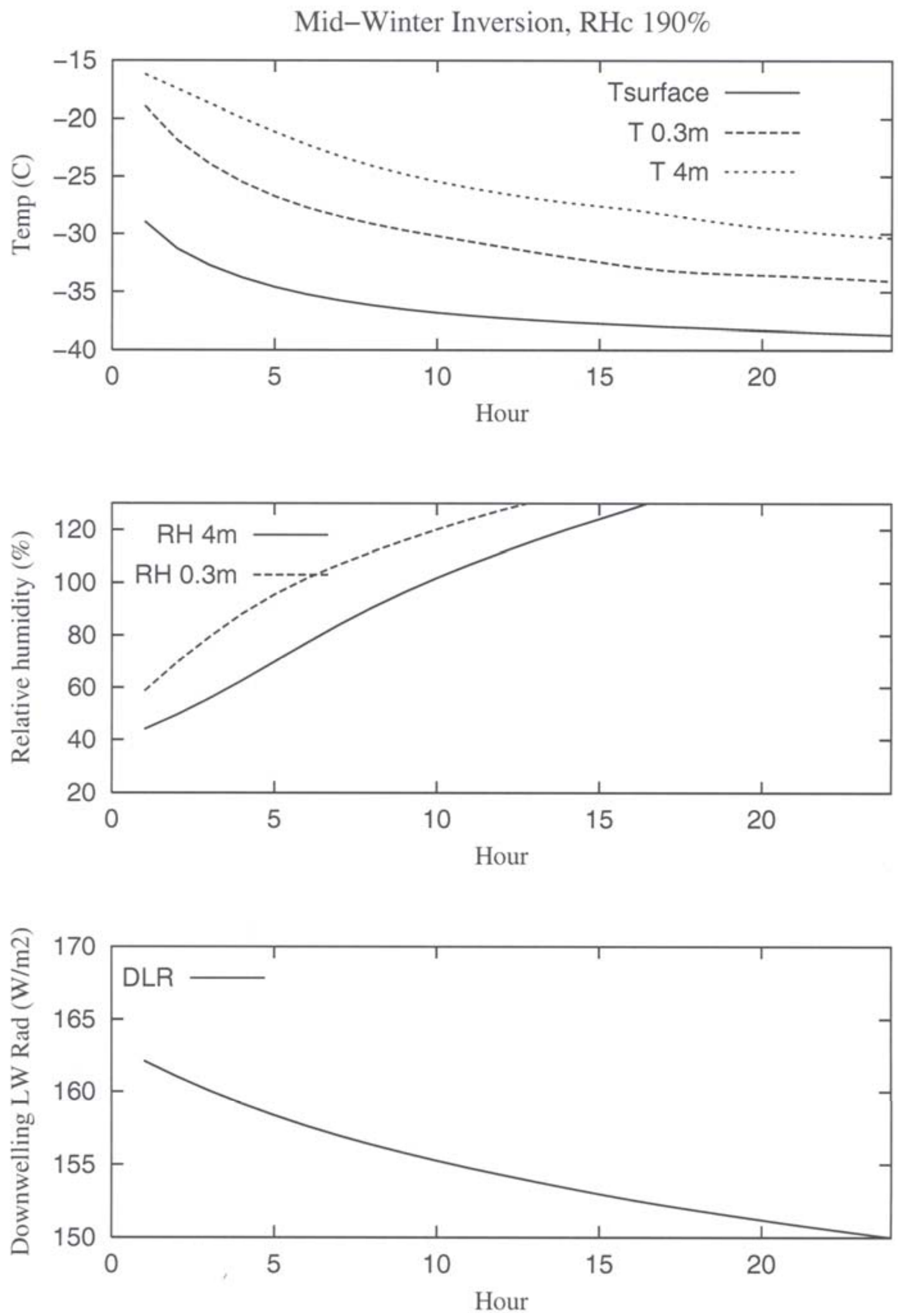
We now apply the UH 1-D model in a typical N Finland midwinter (very stable, misty) case:

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: - initially snow and sfc air at –15C, lapse rate 2 C/km, RH 30%, Vg is 0.1 m/s. No solar radiation (midwinter). The model then cools to a typical observed ~200 m high wintertime surface inversion in a few hours.

Questions to be answered: - Which factor dominates the cooling, turbulence or radiation? –What is the role of latent heating, if fog is formed? – What stops the cooling?

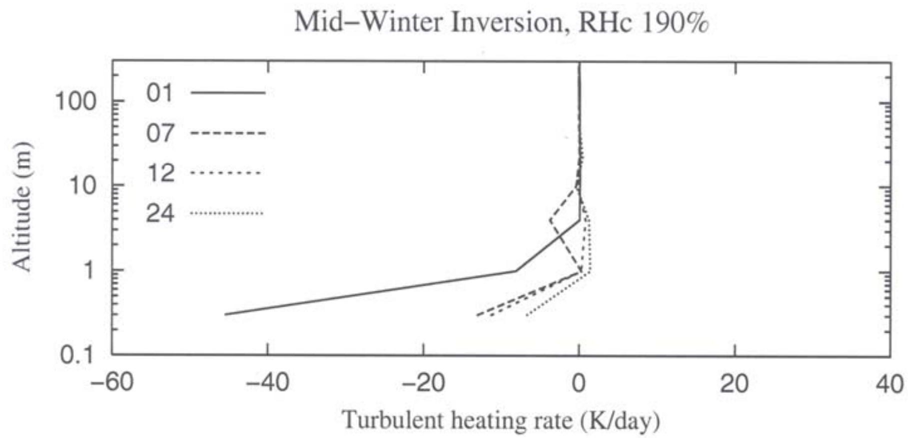
Time evolution, no fog allowed:



Temperatures decrease from the initial –15C. Snow surface is cooling via LW net loss to space (but gets heat by diffusion from warmer snow layers beneath: snow properties are important)

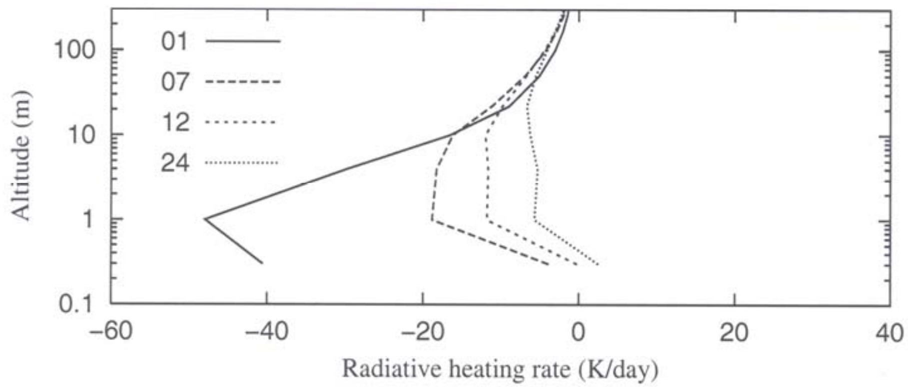
Relative humidities increase as air temperatures decrease. There is no fog as RHcrit for cloud formation is set here (artificially) to 190% (clean air, no CCNs)

Downwelling longwave radiation decreases monotonically as air temperatures decrease. Absolute humidity is nearly conserved.

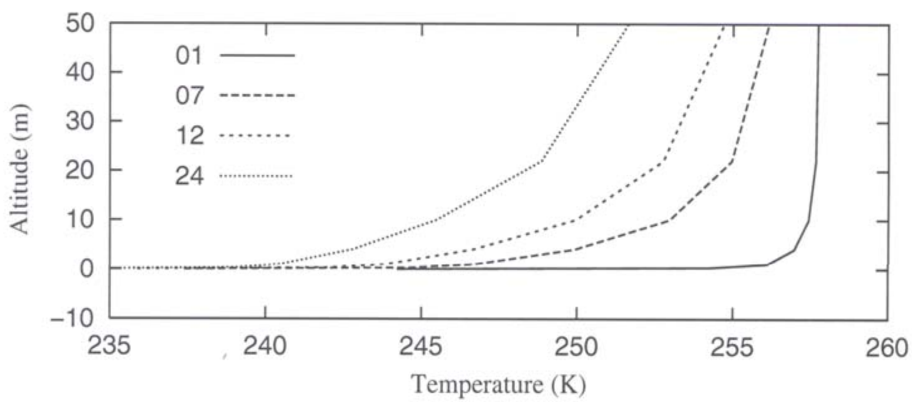


No fog allowed: Vertical profiles 1, 7, 12 and 24 hours after start:

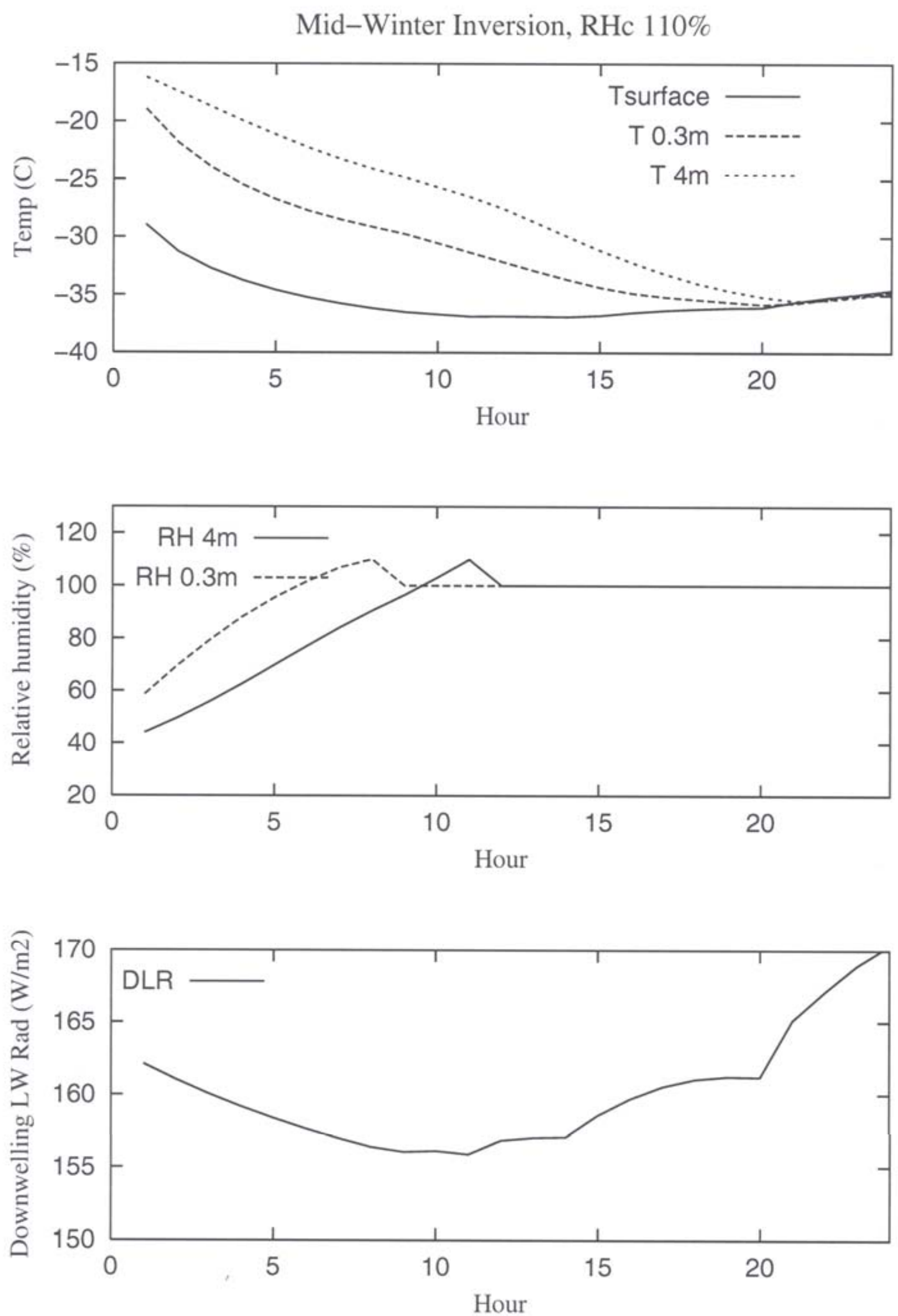
Turbulent cooling is weak (since wind is weak and air is extremely stable). It is concentrated to the shear layer next to the surface



LW cooling clearly dominates over turbulent cooling across the growing inversion



Temperature profile evolves into a typical Lappish winter inversion. Inversion top is at about 200 m by 24h and stabilizes there

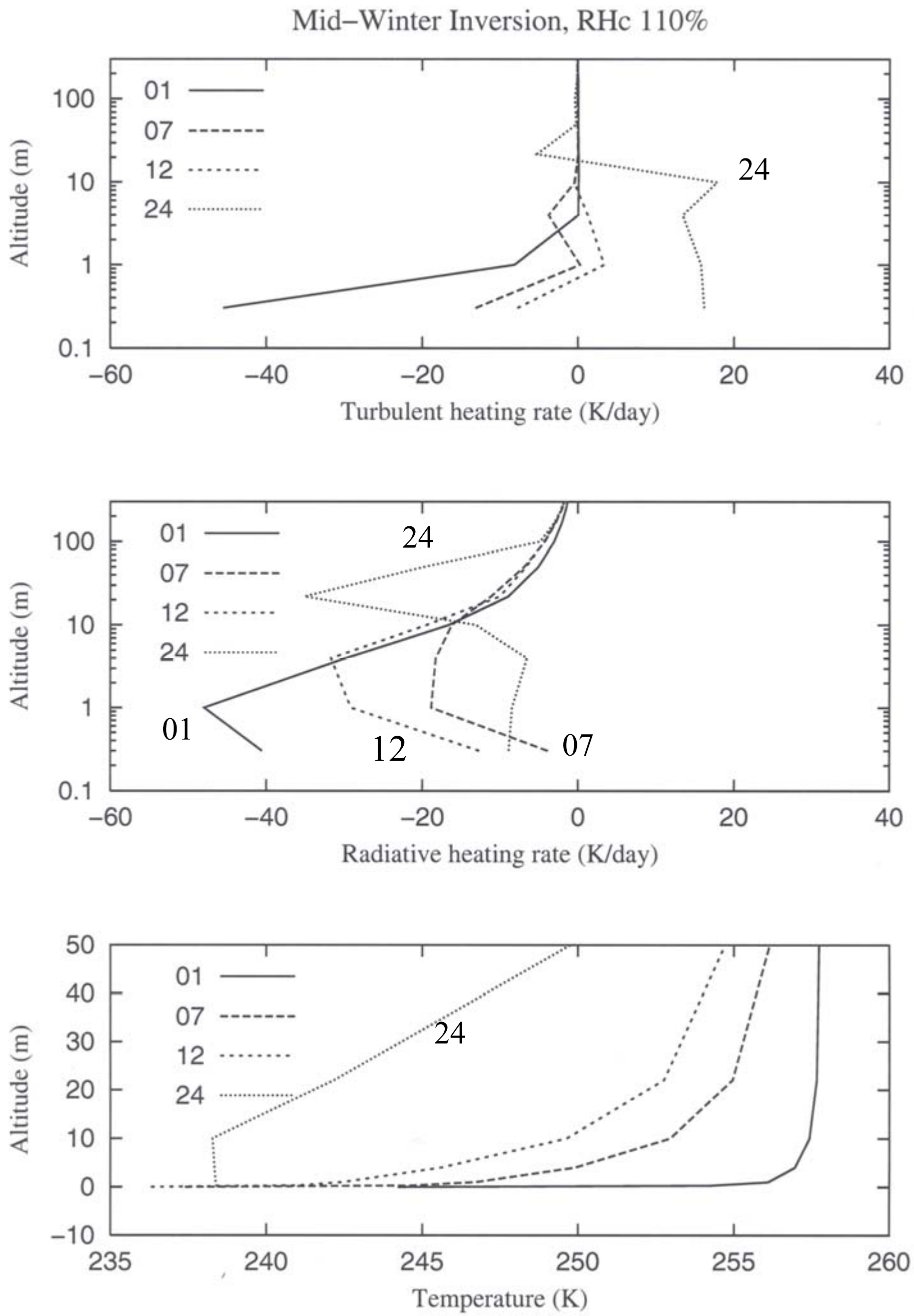


<u>Fog evolution **for RHcrit 110%**</u>:

Fog appears at 0.3 m by 08h. Surface temperature drop stops by 11h. By 21h surface becomes slightly warmer than air at 4 m.

Icefog is formed when RHi exceeds the critical value of 110%. (Foggy air stays at RHi = 100%, incorrect in models vs. observed?)

Downwelling LW radiation starts to increase after the formation of fog, by <u>ice crystal emission</u>. This stops further drop of surface temperature, and so <u>limits the cold extreme</u>



Fog case profiles:

Turbulent cooling changes to heating (convection!) within the well-established fog by 24h (as Tsfc is warmer than Tair due to droplet emission)

Current top of the fog LW-cools strongly, creating more fog

The temperature profile within thick fog evolves toward a quasi-isothermal state

Further experiments with the UH 1-D model:

-If k = 0 but latent heating is included, temperatures evolve similarly to the ’no fog’ case.

-If latent heating is shortcut but fog’s LW effect is included, 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 temperature drop is via increased LW emission by diamond dust to the ground.

-For extreme cold cases, a **dry**, clean and calm continental airmass is therefore needed so that condensation into ice crystals is delayed (Siberia, Antarctica).

- In cold and moist air ”it is snowing all the time”: diamond dust falls from clear sky.

Conclusions

- Small-drop SW effects are important for small-drop st, sc, fogs, aerosol. A simple improvement for delta-TSA during low sun is available.
- **LW cooling dominates sfc layer, if V < 3 m/s**
- RHi > 100% may be common in ice mists
- Cold extremes are obtained in **dry**, clear, calm airmasses, where DLR by thin 'diamond dust' finally stops the drop of surface temperature
- Latent heat release is unimportant in cold fogs
- Current snow properties are important for the in-snow heat diffusion and hence, for accurate prediction of Tg.