Introduction to the cloud physics module of PALM

–Amendments to the dry version of PALM–

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1 Introduction

The dry version of PALM does not contain any cloud physics. It has been extended to account for a nearly complete water cycle and radiation processes:

Water cycle

- evaporation/condensation
- precipitation
- transport of humidity and liquid water

Radiation processes

- short-wave radiation
- long-wave radiation

The dynamical processes are covered by advection and diffusion and they are described by the implemented methods. For the consideration of the thermodynamical processes modifications are necessary in the thermodynamics of PALM . In doing so evaporation and condensation are treated as adiabatic processes whereas precipitation and radiation are treated as diabatic processes. In the dry version of PALM the thermodynamic variable is the potential temperature θ. The first law of thermodynamics provides the prognostic equation for θ. The system of thermodynamic variables has to be extended to deal with phase transitions:

$$
\begin{aligned}
q_v &= \text{specific humidity to deal with water vapour} \\
q_l &= \text{liquid water content to deal with the liquid phase}
\end{aligned}
$$

Additionally, dependencies between these variables have to be introduced to describe the changes of state (condensation scheme).
In introducing the two variables liquid water potential temperature θ_l and total liquid water content q the treatment of the thermodynamics is simplified. The liquid water potential temperature θ_l is defined by Betts (1973) and represents the potential temperature attained by evaporating all the liquid water in an air parcel through reversible wet adiabatic descent. In a linearized version it is defined as

$$\theta_l = \theta - \frac{L_v}{c_p}\left(\frac{\theta}{T}\right) q_l. \tag{1}$$

For the total water content it is valid:

$$q = q_v + q_l. \tag{2}$$

The usage of θ_l and q as thermodynamic variables is based on the work of Ogura (1963) and Orville (1965). The advantages of the θ_l-q system are discussed by Deardorff (1976):

- Without precipitation, radiation and freezing processes θ_l and q are conservative quantities (for the whole system).
- Therewith, the treatment of grid volumes in which only a fraction is saturated is simplified (sub-grid scale condensation scheme).
- Parameterizations of the sub-grid scale fluxes are retained.
- The liquid water content is not a separate variable (storage space is saved).
- For dry convection θ_l matches the potential temperature and q matches the specific humidity when condensation is disabled.
- Phase transitions do not have to be described as additional terms in the prognostic equations.

2 Model equations

In combining the prognostic equations for dry convection with the processes for cloud physics the following set of prognostic and diagnostic model equations is gained:

Equation of continuity

$$\frac{\partial \overline{u}_j}{\partial x_j} = 0 \tag{3}$$

Equations of motion

$$\frac{\partial \overline{u}_i}{\partial t} = -\frac{\partial \left(\overline{u}_j \overline{u}_i\right)}{\partial x_j} - \frac{1}{\rho_0}\frac{\partial \overline{\pi}^*}{\partial x_i} - \varepsilon_{ijk} f_j \overline{u}_k - \varepsilon_{i3k} f_3 u_{\mathrm{g}_k} + g\frac{\overline{\theta}_v - \langle\overline{\theta}_v\rangle}{\theta_0}\delta_{i3} - \frac{\partial\,\tau_{ij}}{\partial x_j} \tag{4}$$

with

$$\overline{\pi}^* = \overline{p}^* + \frac{2}{3}\rho_0\,\overline{e} \tag{5}$$

$$\tau_{ij} = \overline{u_j^{\prime} u_i^{\prime}} - \frac{2}{3}\overline{e}\,\delta_{ij} \tag{6}$$

First law of thermodynamics

$$\frac{\partial \overline{\theta}_l}{\partial t} = -\frac{\partial \left(\overline{u}_j \overline{\theta}_l\right)}{\partial x_j} - \frac{\partial\,\overline{u_j^{\prime}\theta_l^{\prime}}}{\partial x_j} + \left(\frac{\partial \overline{\theta}_l}{\partial t}\right)_{\mathrm{RAD}} + \left(\frac{\partial \overline{\theta}_l}{\partial t}\right)_{\mathrm{PREC}} \tag{7}$$

Conservation equation for the total water content

$$\frac{\partial \overline{q}}{\partial t} = -\frac{\partial \left(\overline{u}_j \overline{q}\right)}{\partial x_j} - \frac{\partial\,\overline{u_j^{\prime}q^{\prime}}}{\partial x_j} + \left(\frac{\partial \overline{q}}{\partial t}\right)_{\mathrm{PREC}} \tag{8}$$

Conservation equation for the sub-grid scale turbulent kinetic energy $\overline{e} = \frac{1}{2}\overline{u_i'^2}$

$$\frac{\partial \overline{e}}{\partial t} = -\frac{\partial \left(\overline{u}_j \overline{e}\right)}{\partial x_j} - \overline{u_j' u_i'}\frac{\partial \overline{u}_i}{\partial x_j} + \frac{g}{\theta_0}\overline{u_3' \theta_v'} - \frac{\partial}{\partial x_j}\left\{\overline{u_j'\left(e' + \frac{p'}{\rho_0}\right)}\right\} - \epsilon \tag{9}$$

The virtual potential temperature is needed in equation (4) to calculate the buoyancy term. It is defined by e.g. Sommeria and Deardorff (1977) as

$$\overline{\theta}_v = \left(\overline{\theta}_l + \frac{L_v}{c_p}\left(\frac{\theta}{T}\right)\overline{q}_l\right)\left(1 + 0.61\,\overline{q} - 1.61\,\overline{q}_l\right). \tag{10}$$

Therewith, the influence of changing in density due to condensation is considered in the buoyancy term.
The closure of the model equations is based on the approaches of Deardorff (1980):

$$\overline{u_j' u_i'} = -K_m\left(\frac{\partial \overline{u}_i}{\partial x_j} + \frac{\partial \overline{u}_j}{\partial x_i}\right) + \frac{2}{3}\overline{e}\,\delta_{ij} \tag{11}$$

$$\overline{u_j' \theta_l'} = -K_h\left(\frac{\partial \overline{\theta}_l}{\partial x_j}\right) \tag{12}$$

$$\overline{u_j' q'} = -K_h\left(\frac{\partial \overline{q}}{\partial x_j}\right) \tag{13}$$

$$\overline{u_j'\left(e' + \frac{p'}{\rho_0}\right)} = -2K_m\frac{\partial \overline{e}}{\partial x_j} \tag{14}$$

$$\overline{u_3' \theta_v'} = K_1\,\overline{u_3' \theta_l'} + K_2\,\overline{u_3' q'} \tag{15}$$

$$K_m = 0.1\,l\,\sqrt{\overline{e}} \tag{16}$$

$$K_h = \left(1 + 2\frac{l}{\Delta}\right)K_m \tag{17}$$

$$\epsilon = \left(0.19 + 0.74\,\frac{l}{\Delta}\right)\frac{\overline{e}^{\frac{3}{2}}}{l} \tag{18}$$

with

$$l = \begin{cases} \min\left(\Delta,\, 0.7\,d,\, 0.76\,\sqrt{\overline{e}}\left(\frac{g}{\theta_0}\frac{\partial \overline{\theta}_v}{\partial z}\right)^{-\frac{1}{2}}\right) & , \quad \frac{\partial \overline{\theta}_v}{\partial z} > 0 \\ \min\left(\Delta,\, 0.7\,d\right) & , \quad \frac{\partial \overline{\theta}_v}{\partial z} \leq 0 \end{cases} \tag{19}$$

and

$$\Delta = \left(\Delta x \Delta y \Delta z\right)^{1/3} \tag{20}$$

At the lower boundary Monin-Obukhov similarity theory is valid ($\overline{w'q'} = u_* q_*$).
Cuijpers and Duynkerke (1993) for example define the coefficients K_1 and K_2 as follows:
in unsaturated air:

$$K_1 = 1.0 + 0.61\,\overline{q} \tag{21}$$

$$K_2 = 0.61\,\overline{\theta} \tag{22}$$

in saturated air:

$$K_1 = \frac{1.0 - \overline{q} + 1.61\,\overline{q}_s\left(1.0 + 0.622\frac{L_v}{RT}\right)}{1.0 + 0.622\frac{L_v}{RT}\frac{L_v}{c_pT}\overline{q}_s} \tag{23}$$

$$K_2 = \overline{\theta}\left(\left(\frac{L_v}{c_pT}\right)K_1 - 1.0\right) \tag{24}$$

The saturation value of the specific humidity comes from the truncated Taylor expansion of $q_s(T)$:

$$q_s(T) = q_s = q_s\left(T_l\right) + \left(\frac{\partial q_s}{\partial T}\right)_{T=T_l}(T - T_l). \tag{25}$$

Using the Clausius-Clapeyron equation

$$\left(\frac{\partial q_s}{\partial T}\right)_{T=T_l} = 0.622\frac{L_v q_s(T_l)}{RT_l^2} \tag{26}$$

with

$$T = T_l + \frac{L_v}{c_p}q_l \qquad \text{respectively} \qquad q_l = q - q_s \tag{27}$$

gives

$$\overline{q}_s = \overline{q}_s(\overline{T}_l)\frac{(1.0 + \beta\,\overline{q})}{1.0 + \beta\,\overline{q}_s(\overline{T}_l)}. \tag{28}$$

Whereas

$$\overline{q}_s(\overline{T}_l) = 0.622\frac{\overline{e}_s(\overline{T}_l)}{p_0(z) - 0.377\,\overline{e}_s(\overline{T}_l)} \tag{29}$$

and

$$\beta = 0.622\left(\frac{L_v}{R\overline{T}_l}\right)\left(\frac{L_v}{c_p\overline{T}_l}\right). \tag{30}$$

The actual liquid water temperature is defined as

$$\overline{T}_l = \left(\frac{p_0(z)}{p_0(z=0)}\right)^{\kappa}\overline{\theta}_l \tag{31}$$

with $p_0(z=0) = 1000\,\text{hPa}$. The value of the saturation vapour pressure at the temperature $\overline{T}_l$ is calculated in the same way as in Bougeault (1982):

$$\overline{e}_s(\overline{T}_l) = 610.78\exp\left(17.269\frac{\overline{T}_l - 273.16}{\overline{T}_l - 35.86}\right). \tag{32}$$

The hydrostatic pressure $p_0(z)$ is given by Cuijpers and Duynkerke (1993):

$$p_0(z) = p_0(z=0)\frac{T_{\text{ref}}(z)^{c_p/R}}{T_0} \tag{33}$$

with

$$T_{\mathrm{ref}}(z) = T_0 - \frac{g}{c_p} z. \tag{34}$$

The pressure is calculated once at the beginning of a simulation and remains unchanged. For the reference temperature at the earth surface T_0 the initial surface temperature is applied. The ratio of the potential and the actual temperature is given by:

$$\frac{\theta}{T} = \left(\frac{p_0(z=0)}{p_0(z)}\right)^{\kappa}. \tag{35}$$

The liquid water content q_l is needed for the calculation of the virtual potential temperature (eq. (10)). It is calculated from the difference of the total water content at a single grid point and the saturation value at this grid point:

$$\overline{q}_l = \begin{cases} \overline{q} - \overline{q}_s(\overline{T}_l) & \text{if} \quad \overline{q} > \overline{q}_s(\overline{T}_l) \\ 0 & \text{else} \end{cases} \tag{36}$$

With this approach a grid volume is either completely saturated or completely unsaturated. The values of the cloud cover of a grid volume can only become 0 or 1 (*0%-or100% scheme*).

3 Parameterization of the source terms in the conservation equations

3.1 Radiation model

The source term for radiation processes is parameterized via the scheme of effective emissivity which is based on Cox (1976):

$$\left(\frac{\partial \overline{\theta}_l}{\partial t}\right)_{\mathrm{RAD}} = -\frac{\theta}{T}\frac{1}{\rho\, c_p\, \Delta z}\left[\Delta F(z^+) - \Delta F(z^-)\right] \tag{37}$$

ΔF describes the difference between upward and downward irradiance at the grid point above (z^+) and below (z^-) the level in which θ_l is defined.
The upward and downward irradiance $F\uparrow$ and $F\downarrow$ are defined as follows:

$$F\uparrow(z) = B(0) + \varepsilon\uparrow(z,0)\,(B(z) - B(0)) \tag{38}$$

$$F\downarrow(z) = F\downarrow(z_{\mathrm{top}}) + \varepsilon\downarrow(z, z_{\mathrm{top}})\,(B(z) - F\downarrow(z_{\mathrm{top}})) \tag{39}$$

$F\downarrow(z_{\mathrm{top}})$ describes the impinging irradiance at the upper boundary of the model domain which has to be prescribed. $B(0)$ and $B(z)$ represent the black body emission at the ground and the height z respectively. $\varepsilon\uparrow(z,0)$ and $\varepsilon\downarrow(z, z_{\mathrm{top}})$ stand for the effective cloud emissivity of the liquid water between the ground and the level z and between z and the upper boundary of the model domain z_{top} respectively. They are defined as

$$\varepsilon\uparrow(z,0) = 1 - \exp\left(-a \cdot LWP(0,z)\right) \tag{40}$$

$$\varepsilon\downarrow(z, z_{\mathrm{top}}) = 1 - \exp\left(-b \cdot LWP(z, z_{\mathrm{top}})\right) \tag{41}$$

$LWP(z_1, z_2)$ describes the liquid water path which is the vertically added content of liquid water above each grid column:

$$LWP(z_1, z_2) = \int_{z_1}^{z_2} \mathrm{d}z\, \rho \cdot \overline{q}_l. \tag{42}$$

a and b are called mass absorption coefficients. Their empirical values are based on Stephans (1978) with $a = 130\,\mathrm{m^2kg^{-1}}$ and $b = 158\,\mathrm{m^2kg^{-1}}$.
The assumptions for the validity of this parameterization are:

- Horizontal divergences in radiation are neglected.
- Only absorption and emission of long-wave radiation due to water vapour and cloud droplets is considered.
- The atmosphere is assumed to have constant in-situ temperature above and below the regarded level except for the earth surface.

3.2 Precipitation model

The source term for precipitation processes is parameterized via a simplified scheme of Kessler (1969):

$$\left(\frac{\partial \overline{q}}{\partial t}\right)_{\mathrm{PREC}} = \begin{cases} \left(\overline{q}_l - \overline{q}_{l_{\mathrm{crit}}}\right)/\tau & \overline{q}_l > \overline{q}_{l_{\mathrm{crit}}} \\ 0 & \overline{q}_l \leq \overline{q}_{l_{\mathrm{crit}}} \end{cases} \tag{43}$$

The precipitation leaves the grid volume immediately if the threshold of the liquid water content $\overline{q}_{l_{\mathrm{crit}}} = 0.5\,\mathrm{g/kg}$ is exceeded. Hence, evaporation of the rain drops does not occur.
τ is a retarding time scale with a value of 1000 s.
The influence of the precipitation on the temperature is as follows:

$$\left(\frac{\partial \overline{\theta}_l}{\partial t}\right)_{\mathrm{PREC}} = -\frac{L_v}{c_p}\left(\frac{\theta}{T}\right)\left(\frac{\partial \overline{q}}{\partial t}\right)_{\mathrm{PREC}} \tag{44}$$

List of symbols

Variable	Description	Value
B	black body radiation	
c_p	heat capacity for dry air with p=const	$1005\,\mathrm{J\,K^{-1}kg^{-1}}$
d	normal distance to the nearest solid surface	
$\overline{e}$	sub-grid scale turbulent kinetic energy	
e_s	saturation vapour pressure	
f_i	Coriolis parameter $i \in \{1, 2, 3\}$	
$F\uparrow$	upward irradiance	
$F\downarrow$	downward irradiance	
i, j, k	integer indices	
K_h	turbulent diffusion coefficient for momentum	
K_m	turbulent diffusion coefficient for heat	
K_1	coefficient	
K_2	coefficient	
l	mixing length	
L_v	heat of evaporation	$2.5 \cdot 10^6\,\mathrm{J\,kg^{-1}}$

LWP	liquid water path	
R	gas constant for dry air	$287\,\mathrm{J\,K^{-1}kg^{-1}}$
T	actual temperature	
T_l	actual liquid water temperature	
u, v, w, u_i	velocity components, $i \in \{1,2,3\}$	
p_0	hydrostatic pressure	
q	total water content	
q_l	liquid water content	
$q_{l_{\mathrm{crit}}}$	threshold for the formation of precipitation	
q_s	specific humidity in case of saturation	
q_v	specific humidity	
x, y, z, x_i	Cartesian coordinates, $i \in \{1,2,3\}$	
Δ	characteristic grid length	
ϵ	dissipation of sub-grid scale turbulent kinetic energy	
$\varepsilon\uparrow$	upward effective cloud emissivity	
$\varepsilon\downarrow$	downward effective cloud emissivity	
κ	R/c_p	0.286
ρ	air density	
τ	time scale for the Kessler scheme	
θ	potential temperature	
θ_l	liquid water potential temperature	
θ_v	virtual potential temperature	
θ_0	reference value for the potential temperature	
$\overline{\psi}$	resolved scale variable	
ψ'	sub-grid scale variable	
ψ^*	departure from the basic state (Boussinesq approximation)	
$\langle\psi\rangle$	horizontal mean	

References

Betts, A. K., 1973: Non-precipitating cumulus convection and its parameterization. *Quart. J. R. Meteorol. Soc.*, **99**, 178-196.

Bougeault, P., 1982: Modeling the trade-wind cumulus boundary layer. Part I: Testing the ensemble cloud relations against numerical data. *J. Atmos. Sci.*, **38**, 2414-2428.

Cox, S. K., 1976: Observations of cloud infrared emissivity. *J. Atmos. Sci.*, **33**, 287-289.

Cuijpers, J. W. M. and P. G. Duynkerke, 1993: Large eddy simulation of trade wind cumulus clouds. *J. Atmos. Sci.*, **50**, 3894-3908.

Deardorff, J. W., 1976: Usefulness of liquid-water potential temperature in a shallow-cloud model. *J. Appl. Meteor.*, **15**, 98-102.

Deardorff, J. W., 1980: Stratocumulus-capped mixed layers derived from a three-dimensional model. *Boundary-Layer Meteorol.*, **18**, 495-527.

Kessler, E., 1969: On the distribution and continuity of water substance in atmospheric circulations. *Met. Monogr.*, **32**, 84 pp.

Müller, G. A. and A. Chlond, 1996: Three-dimensional numerical study of cell broadening during cold-air outbreaks. *Boundary-Layer Meteorol.*, **81**, 289-323.

Ogura, Y., 1963: The evolution of a moist convective element in a shallow, conditionally unstable atmosphere: A numerical calculation. *J. Atmos. Sci.*, **20**, 407-424.

Orville, J. D., 1965: A numerical study of the initiation of cumulus clouds over mountainous terrain. *J. Atmos. Sci.*, **22**, 684-699.

Sommeria, G. and J. W. Deardorff, 1977: Subgrid-scale condensation in models of nonprecipitating clouds. *J. Atmos. Sci.*, **34**, 344-355.

Stephans, G. L., 1978: Radiation profiles in extended water clouds. II: Parameterization schemes. *J. Atmos. Sci.*, **35**, 2123-2132.