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1. Introduction

The goal of physical parameterization in numerical models of the atmosphere is to adequately describe the physical processes in the atmosphere. These processes occur to a large extent on a sub-grid scale which the model cannot otherwise compute due to a lack of model resolution.

The present report is concerned with an important and complex part of physical parameterization, namely the description of subgrid scale condensation of water vapor with associated phase changes, and the formulation of precipitation plus evaporation processes, hereafter referred to as CCPE processes = (Convection -Condensation - Precipitation - Evaporation). A specific scheme for describing these processes will be documented in detail. This scheme is being developed in collaboration with the HIRLAM community (Lynch et al., 2000), and a version of the scheme is used operationally at the Danish Meteorological Institute (DMI). The scheme has novel features related to cloud parameterization and the treatment of precipitation release.

When decribing cloud physics in atmospheric models an important point is that the condensation and precipitation processes are strongly dependent on both resolved scale motions described by the model dynamics and subgrid scale motions on all scales from the convective cloud motions over the full depth of the troposphere down to the smallest turbulent scales. In the present context this means that the computations related to the CCPE processes are dependent on the formulation of the model dynamics and turbulence of the HIRLAM forecasting system (Sass et al., 2002; Undén et al., 2002) The vertical transport processes are by far the most important to parameterize due to the presence of the earth's surface as a boundary for vertical fluxes and the large vertical gradients of most meteorological parameters. However, when approaching the scale of about 1 km , the socalled cloud resolving scale, the importance of the lateral subgrid scale transports increases.

When developing a parameterization of CCPE processes several problems or questions emerge:

It is a question how the whole range of subgrid scale transports can be adequately described. Practically all schemes developed for atmospheric models in the past distinguish between 'turbulence parametrization' and 'convection parameterization'. The former describes the small scale transports up to a dimension of a few hundred metres while the latter describes vertical transports extending up to the full depth of the troposphere. It is often questioned whether the turbulence scheme and the convection scheme each describe separate scales of the subgrid transports leading to worries about possible 'double counting' the heat and moisture transports.

It is normally assumed that parameterization can be made from model parameters defined for a given grid square or an air column vertically above a surface grid square. However, this assumption becomes increasingly problematic as the grid size becomes small. This problem shows up when treating deep convection since a convective cloud sometimes move horizontally by more than one grid distance during its evolution cycle. A similar problem occurs in relation to precipitation release from high elevations where the precipitation particles may travel several grid distances 'downstream' before reaching the ground. Also the assumption in many large scale models that an ensemble of convective clouds exist in balance with the large scale forcing breaks down as the grid size is reduced towards the cloud scale. Hence it appears that the parameterization of the CCPE processes in connection with convection are particularly difficult.

The condensation processes in a statically stable atmosphere inside a grid box desribed by a ‘stratiform’ condensation scheme is also influenced by subgrid scale motions giving rise to subgrid scale condensation (partial cloud cover). As the grid size becomes smaller, however, the amplitude of the subgrid moisture variations becomes smaller and must decrease towards zero as the grid size goes towards zero. One may claim that the subgrid parameterization should automatically include assumptions about the scale dependency of the process description even if it is difficult to prove the validity of a specific scale dependency. It is often difficult to validate which formulations should be preferred by means of solid observational evidence. The results of very high resolution large-eddy simulations might be a tool to validate proposed scale dependent formulations. The alternative is to make specific model tuning of various parameters when changing the model resolulion. In recent years international comparisons have been useful as a guidance, e.g., the GEWEX (Global Energy and Water Cycle Experiment) cloud system study (GCSS).

There have been many approaches during the past decades to describe the processes connected to clouds and condensation. This is partly because of the large amount of subjects involved in the CCPE process description. In the HIRLAM community a revview has been written on the use of convection schemes in mesoscale models (Bister, 1998). One trend has been to apply mass flux concepts to parameterize convection. It may be argued that a mass flux approach is more physically based than are formulations relying on alternative approaches. However, the details of the processes governing the evolution of mass fluxes are not well known.

For the socalled stratiform condensation process in a statically stable atmosphere the assumtions on the character of the subgrid scale condensation influences the microphysics including precipitation release.

Section 2 to section 4 contain a documentation of a research version of the operational cloud- and condensation scheme used at DMI. The scheme is named STRACO which stands for ‘Soft TRAnsition COndensation’ (gradual transitions between convective and stratiform regimes).

The prognostic model variables used by the scheme comprise specific humdidity q, cloud condensate q_c, temperature T, the horizontal wind components u and v and the surface pressure p_s. In addition, a vertical velocity is used.

Section 2 is devoted to a description of the convection scheme. Section 3 is concerned with the subgrid scale cloud parameterization and the ‘stratiform’ condensation process. Section 4 contains a desciption of the microphysics involved, including parameterization of phase changes and physics related to precipitation release. A brief discussion and concluding remarks are finally provided in section 5.

2. Parameterization of convection

Currently the operational convection scheme is based on a moisture budget closure. This implies that the moisture source from the model's dynamics and turbulence scheme during a physics time step may be redistributed in a convective air column leading to heating and moistening. For the basic redistribution to become active it is assumed that the total vertically integrated supply of humidity should be positive. In this respect the scheme may be viewed as a further development of ideas expressed by Kuo (1974). The reasoning behind this type of closure is that a substantial convective activity needs a moisture source (moisture 'convergence') to be sustained. In the present scheme the treatment of the moisture convergence includes the effect of surface evaporation flux.

The vertical transports include a formulation of the vertical redistribution of cloud condensate since 'prognostic cloud water' is a feature of the scheme.

Convection can start from any level in the atmosphere provided that the onset of convection is supported by the model's cloud ascent formulation. Some convection schemes treat only deep convection originating from the lowest model layer.

Also fluxes of moisture and heat across the interface to the stable atmosphere above the convective entity are taken into account. Moreover, the precipitation release formulation differs radically from the parameterization in schemes which do not have 'cloud condensate' as a prognostic variable.

Before describing the equations connected to the convective closure the method to assess the moist convective part(s) of the atmosphere will be described.

2.1. Determination of convective entities

The convective part of the STRACO scheme defines vertical sections of the atmosphere which form convective 'entities'. The vertical extent of a convective entity is determined by adiabatic cloud 'parcel' lifting including latent heat release. The cloud parcel starts at the bottom of a new part of the atmosphere to be investigated for convective instability, using a trigger perturbation temperature and specific humidity, respectively. A weak resolution dependence of the perturbations is suggested. The perturbations are defined in (1) and (2). It is seen that the perturbations goes to zero as the grid size goes towards zero which is the governing constraint.

$$\Delta T_{per} = \frac{1}{a_1 + a_2 \cdot \sqrt{\frac{D_T}{D}}} \tag{1}$$

$$\Delta q_{per} = a_3 \cdot q_k \cdot \sqrt{\frac{D}{D_T}} \tag{2}$$

$a_1 = 0.6\mathrm{K}^{-1}$ and $a_2 = 0.5\mathrm{K}^{-1}$, $a_3 = 0.02$ $D_T = 1.0 \cdot 10^4\mathrm{m}$. D is the model grid size (m). q_k is the specific humidity at the bottom of the convective entity considered.

The convective parcel ascent is carried out by assuming that the convective air parcel remains saturated at the saturation specific humidity of the convective air parcel temperature which evolves by the adiabatic lifting process and a 'dilution' process. This is a volumetric mixing fraction per unit length of vertical ascent. This environmental mixing process normally tends to cool the convective air parcel by evaporation of cloud parcel cloud condensate into dry environmental air. Such process is often quite powerful to reduce the buoyancy of the convective cloud and may strongly reduce the vertical extent of convection which stops as the cloud buoyancy becomes negative.

A volume fractional entrainment ϵ_e per unit length of vertical parcel ascent is tentatively described according to the equation below.

$$\epsilon_e = \Big(K_{\epsilon 0} + \frac{K_{\epsilon 1}}{Ri_*}\Big) \cdot \Big(\frac{z}{(K_{\epsilon 3} + z)}\Big) \cdot \frac{D_0}{D} \tag{3}$$

In (3) Ri_* is a Richardson number which enables that effects of wind shear is incorporated. It is argued that increasing wind shear gives rise to more mixing of the convective cloud with the environments.

$$Ri_* = \Big(\frac{\theta}{g}|\frac{\partial V}{\partial z}|^2\Big)^{-1} \cdot \Big(K_{\epsilon 2} + |\frac{\partial \theta}{\partial z}|\Big)$$

In (3) $K_{\epsilon 0} = 1.3 \cdot 10^{-4}\mathrm{m}^{-1}$, $K_{\epsilon 1} = 7.5 \cdot 10^{-4}\mathrm{m}^{-1}$.

The second brackets of (3) expresses a height dependency of the entrainment process being dimensionless and increases from zero at the surface towards 1 at great heights. $K_{\epsilon 2} = 1.0 \cdot 10^{-4}\mathrm{K} \cdot \mathrm{m}^{-1}$ and $K_{\epsilon 3} = 500\mathrm{m}$.

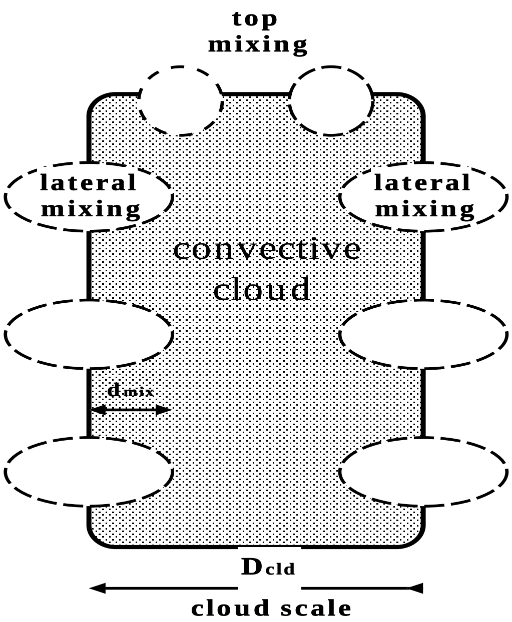

Figure 1: Schematic picture of the mixing processes at the edges (top and sides) of a convective cloud. The turbulent mixing scale is d_{mix}, and the cloud scale is D_{cld}. For details see text

The horizontal resolution dependence is described by the last term. Currently, the fraction between a constant value $D_0 \approx 10\mathrm{km}$) and the model grid size D is constrained to be no less than 1.

If it is maintained that the parameterized convection should describe effects of subgrid scale features being 1-2 orders of magnitude smaller in areal extent than the grid square, it is reasonable to assume that the dimension of the parameterized convective clouds (parcels) decreases roughly proportional to the grid size. It seems reasonable to assume that the ratio between the surface and the volume of a convective cloud increases as the horizontal dimension of the cloud decreases. For a ball-shaped cloud the ratio goes to infinity and is inversely proportional to the size of the convective cloud. As a consequence one might expect that the dilution process is becoming more efficient at high model resolution since the ratio D_{mix}/D_{cld} between the characteristic turbulent scale D_{mix} and the size D_{cld} of the cloud increases (see figure 1). As a first approximation it is therefore assumed that the volumetric dilution becomes inversely proportional to grid size which is expressed by the last term of eq. (3). An important consequence of the resolution dependent dilution is that the vertical extent of parameterized convection will automatically be reduced as the model grid size is reduced. As a consequence the parameterized 'deep convection' will automatically tend to be 'switched off'.

Model experimentation confirms that there can be a pronounced and sometimes subtle interaction between a model's turbulence scheme and convection scheme. For the present model it has been found that it is beneficial to impose a regulating criterion which controls the depth of a convective entity for a rather weak integrated moisture convergence. A maximum depth D_{cv} in Pa of a convective entity is determined by

$$D_{cv} = C_1 + C_2 \cdot Q_* \tag{4}$$

In (4) Q_* is the moisture accession, if positive ($\mathrm{kg} \cdot \mathrm{kg}^{-1} \cdot \mathrm{s}^{-1}$ Pa), otherwise it is zero. More precisely , it is the average specific humidity rate of change over the convective entity times the pressure thickness of the convective entity in Pa. $C_1 = 1.25 \cdot 10^4\mathrm{Pa}$ $C_2 = 4.0 \cdot 10^7 \mathrm{s} \cdot \mathrm{kg} \cdot \mathrm{kg}^{-1}$ The significance of the above formula is for cases where the vertically integrated moisture accession is small.

The level of non-buoyancy determines a transition to the stable atmosphere above. The effect of overshooting eddies penetrating into the stable layer above the moist unstable atmosphere is parameterized and a depth of this 'extension zone' is estimated (Sass, 2001). It is at least one model layer and is limited to be at most 25 hPa thick. The convective transports of heat and moisture, including cloud condensate, across the interface between the moist unstable and the stable atmosphere, is often termed 'shallow convection'. This effect is also parameterized when a deep moist convective atmosphere is involved (see below).

2.2. Convective equations

The relevant equations describing the processes in connection with convection are described below:

$$\left(\frac{\partial q}{\partial t}\right)_{ADC} = \left(\frac{\partial q}{\partial t}\right)_{AD}(1-\delta_*) + \widehat{Q}_a \beta \frac{F_q}{\widehat{F}_q}\delta_* + K_c q_c (q_s - q_e) + S_q + E_{pc} \tag{5}$$

$$\left(\frac{\partial q_c}{\partial t}\right)_{ADC} = \left(\frac{\partial q_c}{\partial t}\right)_{AD} + \widehat{Q}_a (1-\beta) \frac{F_c}{\widehat{F}_c}\delta_* - K_c q_c (q_s - q_e) + S_c - G_{pc} \tag{6}$$

$$\left(\frac{\partial T}{\partial t}\right)_{ADC} = \left(\frac{\partial T}{\partial t}\right)_{AD} + \frac{L'}{c_p}\left(\widehat{Q}_a (1-\beta) \frac{F_h}{\widehat{F}_h}\delta_* - K_c q_c (q_s - q_e)\right) + S_T - \frac{L'}{c_p} E_{pc} \tag{7}$$

The left hand sides of these equations express the combined effect of both dynamical advection, turbulence and convection. $\frac{\partial}{\partial t}()_{AD}$ signifies a tendency excluding convection. $\widehat{Q}_a$ is the total moisture accession per unit mass and time in the convective cloud. L' and c_p represent the specific latent heat of fusion or sublimation, depending on the microphysical conditions, and the specific heat capacity at constant pressure, respectively (see section 4).

F_h is a function describing the vertical variation of convective heating.

$$F_h = T_{vc} - T_{ve} + \epsilon_T \tag{8}$$

F_q is a function describing the vertical variation of convective moistening.

$$F_q = q_{sc} - q_e + \epsilon_q \tag{9}$$

F_c is a function describing the vertical variation of convective condensate supply.

$$F_c = q_{cc} + \epsilon_c \tag{10}$$

In the above equations for F_h, F_q and F_c index v stands for 'virtual'. Index e means 'environmental' (outside clouds). $\widehat{F}$ stands for a vertical average value for the convective cloud. Finally, c-index means a value applicable to cloud and s signifies a saturation value. The constants $\epsilon_T, \epsilon_q, \epsilon_c$ are currently set to zero.

The parameter β is a moistening parameter (Kuo, 1974). It represents moistening due to convective transports, without condensation. In the present scheme, contrary to models without prognostic cloud condensate, moistening can take place also from evaporation of cloud condensate. As a consequence, this β-term is considered of reduced importance in the present scheme.

$$\beta = \left(1 - \frac{\sum_{j=jbot}^{jtop} \frac{q}{q_s}\Delta p}{p_{jbot} - p_{jtop}}\right)^{n_1} \tag{11}$$

In (11) p represents 'pressure', and $jbot$ and $jtop$ are the model level numbers for the bottom and top of convection, respectively. Currently n_1 is set to a value of 2.

The parameter δ_* is an important one, because it determines a link between convective moisture transports on one hand and turbulence plus dynamics effects on the other.

$$\delta_* = \left(\frac{\Delta p_c}{p_{00}}\right)^{n_2} \tag{12}$$

In (12) Δp_c is the total depth of the convective entity considered, and p_{00} is a constant cloud depth used for scaling. Currently $p_{00} = 2 \cdot 10^4\,\mathrm{Pa}$, $n_2 = 1$. δ_* is constrained to be no larger than 1. The effect of this formulation is that δ_* goes to zero for extremely shallow phenomena. A very small δ_* means that the convection scheme is decoupled as is reasonable in the limit of very shallow phenomena and high resolution where dynamics and turbulence should suffice.

The third term in the main equations involving $(q_s - q_e)$ is an evaporation /sublimation term of cloud condensate. (K_c is a constant). A similar formulation, involving cloud cover as the leading term in place of q_c, has been used by others, e.g., (Tiedtke, 1993) in the ECMWF cloud scheme. The parameterization of this term is difficult, partly because of the problem to describe the surface area of clouds and the degree of mixing at the edges of the clouds which are in a subsaturated environment.

The fourth term in the same equations, involving S_q, S_c and S_T respectively, describes the effect of fluxes of heat and moisture across the interface between a moist convective atmosphere and the stable atmosphere above. For example, this parameterization is needed at stratocumulus cloud tops unless the turbulence scheme is specifically designed for computing such fluxes. We let the convection scheme describe the effect of larger eddies in an environment where condensation takes place. The computations make use of the cloud parcel ascent computation of the convection scheme. Physically we may think of the heat- and moisture transports as accomplished by mainly the larger eddies penetrating through the stable layer on the top of a cloud layer. The penetration of these eddies into the stable layer can be estimated from the cloud parcel ascent. The observational and modelling evidence that stratocumulus are associated with a substantial entrainment of (dry) air from the stable layer into the cloud (Nicholls and Leighton, 1986; Duynkerke et al., 1995) makes it reasonable to assume that this hypothesis of 'overshooting' eddies as the mechanism for entrainment of dry air is a reasonable concept. We denote by w_b a characteristic vertical velocity of convective motions in the cloud right below the cloud top and will estimate a distance D_e of penetration into the stable layer.

This depth is estimated as follows: From dimensional analysis it has been argued that the vertical velocity w_r of an idealized thermal depends on its size r, the dimensionless buoyancy B of the 'bubble' and the acceleration of gravity g (m s^{-2}) according the following combination (Rogers and Yau, 1989).

$$w_r = c_b\sqrt{gBr} \tag{13}$$

In (13) B is the virtual temperature difference between the cloud parcel and environment, divided by the environmental temperature. c_b =1.2. Choosing $r = 50$ m as representing the dimension of convective eddies near cloud top we get

$$w_b = w_0 \cdot \sqrt{B}$$

$w_0 =\approx 27$ m s^{-1}.

B is computed in the cloud ascent of the convection scheme (see section 2.1).

We estimate the maximum penetration depth D_e from the deceleration in the stable layer. Utilizing the start velocity of w_b for the deceleration we get

$$D_e = w_0 \sqrt{\frac{B \cdot T}{g|\gamma_c - \gamma|}} \tag{14}$$

In (14) D_e is in metres, T is temperature (K), γ_c in K m^{-1} is the lapse rate associated with moist adiabatic ascent ($\gamma_c > 0$) and γ in K m^{-1} is the ambient lapse rate in the stable layer. It is demanded that $\gamma < \gamma_c$. A numerical security computation has been implemented to avoid extreme behaviour if the two lapse rates become almost equal, and the penetration is not allowed to exceed a depth corresponding to 25 hPa.

A maximum fluctuation s' of the variable s possible at the interface between cloud and the stable layer is estimated to be approximately equal to the increase above the value s_- at the interface on the cloudy side up to the value s_e at depth D_e into the stable layer. The value s_e is estimated from the mean gradient of the variable at cloud top. More specificly, it is assumed that the flux at the cloud top of the scalar s is a factor $\tilde{w}$ times s' where $\tilde{w}$ is a velocity scale. It is reasonable to assume that the velocity scale $\tilde{w}$ is closely linked to a typical convective cloud velocity at the top of convective clouds. It is therefore argued that a reasonable estimate is

$$\tilde{w} = w_1 \cdot \sqrt{B} \tag{15}$$

In (15) $w_1 = 0.15 \mathrm{m} \cdot \mathrm{s}^{-1}$ has been determined on the basis of numerical experimentation.

The sensible heat flux F_H (J m^{-2} s^{-1}) at the level of transition between cloud and the stable layer is computed according to

$$F_H = \rho c_p \cdot \tilde{w} \cdot D_e \cdot \frac{\partial \theta}{\partial z} \tag{16}$$

In (16) c_p is the specific heat capacity at constant pressure. ρ is air density, θ is potential temperature.

Similarly we get for the moisture flux of total specific humidity q_t (kg $\cdot$ m^{-2} s^{-1}) at the transition level

$$F_{q_t} = \rho \cdot \tilde{w} \cdot D_e \frac{\partial q_t}{\partial z} \tag{17}$$

We assume that the fluxes of heat and moisture determined from the above formulas are distributed linearly with height in the convective cloud of depth D_- and in a stable layer D_+. The latter should approximately be equal to D_e apart from the constraints set by vertical resolution. If D_e is larger than the depth of one model layer above cloud a sufficient number of levels are included to exceed D_e. Currently, the specific humidity q and cloud condensate q_c are processed independently according to the method described above, but the flux of the moist conserved variable of 'total specific humidity' is then also linear, which preserves moisture structures in a well mixed cloud.

A semi-implicit treatment of the scheme has been introduced to reduce the risk of noise or instability when computing updates of the prognostic variables. It involves partial derivatives of the fluxes described above, as well as the associated layer depths.

Finally, the terms involving G_{pc} and E_{pc} concern generation and evaporation of convective precipitation, respectively (see section 4).

The equations above are applied in the layers of the convective entities while the stratiform condensation applies to the remaining parts of the atmosphere.

3. Cloud cover and subgrid scale condensation

In the previous section the equations governing the subgrid scale vertical convective transports of heat, humidity and cloud condensate have been described. While these equations describe the evolution of temperature (T), specific humidity (q) and cloud condensate (q_c) in a grid box for the convective part of the atmosphere the distribution of humidity within the grid box has not yet been treated. It is natural to describe the humidity variation by a statistical probability distribution function (PDF) which will define both saturated and unsaturated portions of the grid box. By definition, the fractional cloud cover is the saturated fraction of the grid box with cloud condensate in some concentration. Hence cloud cover will be defined by the PDF which needs to be defined not only for the convective parts of the atmosphere, but also for the stratiform parts. There are substantial differences between the convectively unstable and the stable stratiform regimes as will be described below:

We first consider the problem of defining convective cloud cover. For clarity an 'overline' symbol is used in this section for grid box average values, $\overline{q}$ for specific humidity, $\overline{q}_c$ for specific cloud condensate and $\overline{q}_t = \overline{q} + \overline{q}_c$ for total specific humidity. The prognostic moisture variables $\overline{q}$ and $\overline{q}_c$ have known values at a given time step of a model run. Furthermore, the relevant saturation specific humidity to describe supersaturation is $q_s(T_c)$, which is the saturation specific humidity valid for the convective cloud temperature T_c. This temperature is available from the convective cloud ascent model. One may argue that a probability function describing the variation of total specific humidity around the grid box average value defines the supersaturation in the grid box. In the convective situation the challenge is that the distribution of total specific humidity can vary a lot across the grid box, and the moisture distribution may be quite asymmetric. This is because moisture is exchanged over large depths in the atmosphere. Also temperature varies to some extent. A true description of supersaturation taking into account both temperature and moisture variations is therefore very complex. The extreme situation where cloud cover becomes 100 % may then be a combination of saturated fractions of the grid box with different temperatures. In the present description we have already introduced the convective cloud temperature T_c which is generally higher than the grid box mean value $\overline{T}$. However, in order to simplify cloud cover computations near grid box saturation, and in order to avoid a too high level of complexity, it is demanded that the convective cloud temperature goes towards the grid box mean value $\overline{T}$ when $\overline{q}$ goes towards $\overline{q}_s(\overline{T})$. This means that the preliminary convective cloud temperature T_c is corrected close to grid box saturation conditions. This is done when the relative humidity exceeds $1 - A_{st}$ where A_{st} is defined later in this section in the context of stratiform condensation.

$$T'_c = T_c + \Big(\frac{\frac{\overline{q}}{q_s(\overline{T})} + A_{st} - 1}{A_{st}}\Big)^2 \cdot (\overline{T} - T_c) \tag{18}$$

In (18) T_c' is the corrected convective cloud temperature.

After this simplification we try to describe the probability function defining the variation of total specific humidity by means of a piecewise rectangular probability density function. The most simple formulation involves 2 rectangular boxes which allows for a simple asymmetric PDF. This type of formulation has been used in the past. The formulation below, however, represents an enhancement consisting of 3 boxes when $\overline{q}_t < q_s(T_c)$. The application of 3 boxes in a rather dry atmosphere is consistent with a simple conceptual picture of convective clouds with a large specific humidity embedded in an environment with a fairly homogeneous humidity. Under these conditions a double-peaked PDF may be expected which fits with a 3-box structure (see figure 2).

The 3 boxes have amplitudes ψ_1, ψ_* and ψ_2, respectively (figure 2). The amplitudes are yet unknown, but may be determined by solving the equations (21), (22) and (23) for the situation that $\overline{q_t} < q_s(T_c)$. These equations have a solid basis. Eq.(21) defines that the total integral of the PDF equals 1 when integration is done over the entire humidity spectrum. Eq.(22) expresses that the average value of q_t as determined from the PDF should be equal to the grid box total specific humidity. Eq.(23) is a computation of the grid box mean cloud condensate from the PDF. The integration limits q_{min} and q_{max} have so far not been defined.

The strategy is to parameterize q_{min} which appears as a free parameter. Based on experimentation q_{min} is currently defined as follows:

$$q_{min} = \begin{cases} q_{min_1} & \text{if } \overline{q}_t < q_s \cdot (1 - A_{st}) \\ q_{min_1} \cdot (1-y) + q_{min_2} \cdot y & \text{if } q_s \cdot (1 - A_{st}) \leq \overline{q}_t \leq q_s \end{cases}$$

$$y = \Big(\frac{\frac{\overline{q}_t}{q_s(T_c)} + A_{st} - 1}{A_{st}}\Big)^2$$

$$q_{min_1} = \overline{q}_t(1 - C_{w1}\frac{\overline{q}_c}{q_s(T_c)} - C_{w2}) \tag{19}$$

$$q_{min_2} = q_s(T_c) - C_{w1} \cdot \overline{q}_c \tag{20}$$

In (19) and (20) $C_{w1} = 4$ and $C_{w2} = 0.02$

The interpolation formulas for q_{min} makes it possible to obtain a symmetric PDF at $\overline{q}_t = q_s(T_c)$. The associated range of variability at this point is $\overline{q}_t - q_{min} = 4\overline{q}_c$. This result may be shown by integration of the equation for cloud condensate (23) below.

Then the equations (21),(22) and (23) constitue a system of 3 equations with 4 unknowns namely the amplitudes ψ_1,ψ_*, ψ_2 and the integration limit q_{max}. By formally solving the system of 3 equations it is possible to express q_{max} by means of ψ_* and the other known parameters. This solution is specified in (25). At this stage it is possible to specify the amplitude ψ_* with some freedom as a tuning parameter under the restriction that q_{max} is larger than q_s. This leads to a limit ψ_{*l} on ψ_* according to eq.(26). It is noted that the convective cloud cover f_{cv} is obtained by integrating ψ_2 which operates over the saturated part of the grid box.

$$\int_{q_{min}}^{\overline{q}_t} \psi_1 dq_t + \int_{\overline{q}_t}^{q_s} \psi_* dq_t + \int_{q_s}^{q_{max}} \psi_2 dq_t = 1 \tag{21}$$

$$\int_{q_{min}}^{\overline{q}_t} \psi_1 \cdot q_t \cdot dq_t + \int_{\overline{q}_t}^{q_s} \psi_* \cdot q_t \cdot dq_t + \int_{q_s}^{q_{max}} \psi_2 \cdot q_t \cdot dq_t = \overline{q}_t \tag{22}$$

$$\int_{q_s}^{q_{max}} \psi_2 \cdot (q_t - q_s) dq_t = \overline{q_c} \tag{23}$$

$$f_{cv} = \frac{2q_c}{q_{max} - q_s} \tag{24}$$

$$q_{max} = (b_1 + b_2\psi_*)/(b_3 + b_4\psi_*) \tag{25}$$

$$b_1 = (\overline{q}_t - q_{min}) \cdot (q_s + 2q_c) - 2q_s \cdot (\overline{q}_t + \overline{q}_c)$$

$$b_2 = q_s \cdot (q_s - \overline{q}_t) \cdot (q_s + q_{min} + 2\overline{q}_t)$$

$$b_3 = \overline{q}_t + q_{min} - 2\overline{q}$$

$$b_4 = (q_s - \overline{q}_t) \cdot (q_s - q_{min})$$

The limit ψ_{*l} of ψ_* is

$$\psi_{*l} = (b_3 q_s - b_1)/(b_2 - b_4 q_s) \tag{26}$$

Hence the applicable values of ψ_* may be written as

$$\psi_* = \delta_\psi \cdot \psi_{*l}$$

It may be determined whether δ_ψ must be chosen larger than or smaller than 1 by differentiating with respect to ψ_*

$$\frac{\partial(q_{max} - \overline{q}_t)}{\partial\psi_*}$$

in the point ψ_{*l}. In this way it may be concluded that δ_ψ should be smaller than 1 (first case) if

$$b_2(b_3 + b_4\psi_{*l}) - b_4(b_1 + b_2\psi_{*l}) < 0 \tag{27}$$

On the other hand, δ_ψ should be larger than 1 if the sign of the expression in (27) is positive (second case). Tentatively the values $\delta_\psi = 0.07$ (first case) and 1.07 (second case) have been set which appear to give reasonable results. The selection of optimal values, which may be determined from a more involved computation, requires experimentation with a given model.

Hence it may be concluded that the above solution allows for some freedom regarding the choice of q_{min} and ψ_*. Having made an appropriate choice the unknowns are then ψ_1, ψ_2 and q_{max}, and the cloud cover f_{cv} may be computed from (24) and (25).

For humid conditions $\overline{q}_t > q_s(T_c)$ it is assumed that the PDF is symmetric and that the humidity variation covers at least the range from $q_s(T_c)$ to $\overline{q}_t$. Integrating the equation for cloud condensate (23) then gives for the range A_c defined from $q_{min} = \overline{q}_t \cdot (1 - A_c)$

$$A_c = \frac{-b_6 + \sqrt{b_6^2 - 4b_5 b_7}}{2b_5} \tag{28}$$

In (28)

$$b_5 = \frac{1}{2}\overline{q}_t^2$$

$$b_6 = \overline{q}_t \cdot (\overline{q}_t - q_s - 2\overline{q}_c)$$

$$b_7 = \frac{1}{2}(\overline{q}_t - q_s)^2$$

The cloud cover f_{cv} becomes

$$f_{cv} = Min\Big(\frac{(\overline{q}_t - q_s)}{2A_c\overline{q}_t}, 1\Big) \tag{29}$$

In the stratiform regime we need to define both cloud cover and the stratiform condensation process. The generally smaller vertical scales and larger horizontal scales associated with stratiform condensation makes it reasonable to assume a smaller spatial variation and a symmetric PDF of q_t as an approximation to the true PDF. The PDF, which is rectangular, is shown schematically in figure 2 (to the right). q_s is now equal to the grid box saturation value $q_s(\overline{T})$. Again the temperature variation in the grid box is neglected. The assumption is also made that clouds extend vertically from the bottom to the top of a model layer. These assumptions are quite common among modellers (Sundqvist et al., 1989; P.J.Rasch and J.E.Kristjansson, 1997). The amplitude ψ is evidently given by (30)

$$\psi = \frac{1}{2 \cdot (q_{max} - \overline{q}_t)} \tag{30}$$

The maximum value q_{max} of total specific humidity in the stratiform case is formally written

$$q_{max} = \overline{q}_t \cdot (1 + A_{st}) \tag{31}$$

The dimensionless amplitude A_{st} is determined from a time dependent equation to be described below (see equation 35).

At first it is noted that, for a given value of A_{st}, the equilibrium cloud condensate value q_{ceq} as determined from the PDF, may change from the current value $\overline{q}_c$ as a result of changing temperature and specific humidity during the run. The system of equations (32) and (33) defines the condensation process to establish an equilibrium among the model variables by a first order adjustment. Higher accuracy can be obtained if (32) and (33) are applied several times in an iterative fashion. In practice, it has been chosen to apply a relaxation towards the equilibrium using a small relaxation time period. At least half of the supersaturation is removed per time step. However, full adjustment in a time step is done in the case of grid box supersaturation.

$$\Delta q_{st} = \frac{(\overline{q}_c - q_{ceq})}{1 + f \cdot \frac{L'}{c_p} \left(\frac{\partial q_s}{\partial T}\right)} \tag{32}$$

$$q_{ceq} = 0.5 \cdot \widetilde{\psi} \cdot \widetilde{q}_s^2 + \widetilde{\psi} \cdot \widetilde{q}_{max} \cdot (0.5\widetilde{q}_{max} - \widetilde{q}_s) \tag{33}$$

In (32) Δq_{st} refers to the change of specific humidity during the adjustment process. f is the current value of the fractional cloud cover. L' is the specific latent heat as a function of temperature (to be defined in the next section). In (33) the ˜ symbol applies to preliminary model variables to be adjusted during the condensation process.

In this case the cloud cover f_{st} associated with the equilibrium density function is:

$$f_{st} = \frac{1 + A_{st} - \frac{q_s(T)}{q_t}}{2A_{st}} \tag{34}$$

In general, the dimensionless amplitude A_{st} describing the variation of total specific humidity is not constant, but is a function of model resolution, space and time. A tentative time dependent formulation of A_{st} is expressed in (35) which is intended to

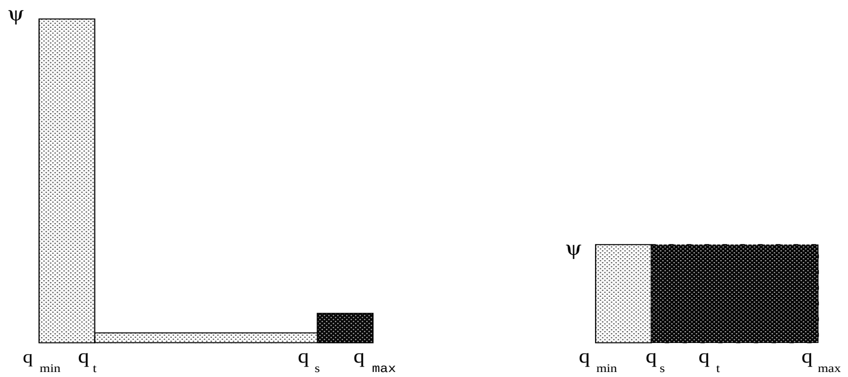

Figure 2: Two examples of piecewise rectangular probability density functions of total specific humidity. The dark dotted regions describe supersaturated cloudy parts. The left figure applies to convective conditions and the figure to the right to stratiform conditions. q_{min} is the minimum value of total specific humidity with non-zero probability and q_{max} is the maximum value. q_t is the grid box value of total specific humidity and q_s is the saturation value for cloud computations. See text for details

describe at least qualitatively the effect of model resolution, elevation above the surface, stationary forcing effects, unsteady flow effects and the effect of precipitation release. The reasoning behind each term of the equation is described below:

$$\frac{\partial A_{st}}{\partial t} = -K_1 \cdot (A_{st} - A_{cli}) + \frac{K_2}{q_t} \cdot \frac{|\partial q_t|}{\partial t} \cdot (A_{max} - A_{st}) + \frac{K_3}{q_t} \cdot Min(\frac{\partial q_c}{\partial t}_{prc}, 0.) \cdot (A_{st} - A_{min}) \tag{35}$$

In (35)

$$A_{max} = \frac{\alpha_1 + \alpha_2 \cdot (1 - (\frac{p}{p_s})^3)}{(\alpha_1 + \alpha_2) \cdot (1 + \sqrt{\frac{\alpha_3}{D}})} \tag{36}$$

Currently $\alpha_1 = 2$, $\alpha_2 = 9$, $A_{min} = 0.005$ $\alpha_3 = 3.6 \cdot 10^5$m K_1=$4.63 \cdot 10^{-5} \cdot \mathrm{s}^{-1}$ $K_2 = 3, K_3 = 9$. The maximum allowed value A_{max} of A_{st} in (35) describes effects of elevation above surface and of model grid size. In (36) p and p_s are model level pressure and surface pressure, respectively. The term involving p/p_s describes a significant reduction of A_{st} towards the surface. A similar effect have been included by others (Sundqvist et al., 1989). The term involving grid size D in the denominator of (36) implies that A_{st} goes towards zero for the grid size going to zero. Obviously this is reasonable in a continuous formulation. Results from the litterature (Redelsperger and Sommeria, 1986) emphasize the virtues of having a subgrid scale parameterization of condensation even at horizontal resolutions of a few kilometres grid size. This is consistent with results from large eddy simulations in recent years, indicating that subgrid scale variability is well pronounced at a grid size of few kilometres. The square root in the denominator is used in order to incorporate the effect that subgrid scale variations are significant already at a grid size of a few kilometres.

The second term of (35) is a crude parameterization of nonstationary flow effects. The term expresses that the change of A_{st} towards the maximum allowed value is proportional (dimensionless factor K_2) to the relative rate of change of total specific humidity. Such a formulation is reasonable to the extent that resolved scale advections are reflected also in a subgrid scale variation, that is, augmented and reduced advections compared to grid box mean value exist inside the grid box.

The first term describes a relaxation towards the value A_{cli} which stands for a 'climatic' type of stationary forcing. Currently the value of A_{cli} is a fixed fraction of A_{max} ($A_{cli} = 0.75 \cdot A_{max}$). A more refined treatment could take into account local stationary forcing effects, e.g., due to varying topography. The relaxation factor corresponds to an e-folding time og 6 hours. The formulation inplies that, in the absence of non-stationarity and precipitation release the amplitude A_{st} approaches A_{cli} exponentially with the given e-folding time.

Finally the last term which is qualitatively similar in appearance to the second term, expreses always a reduction of A_{st} towards a minimum value A_{min}. $(\partial q_c/\partial t)_{prc}$ expresses the reduction rate of cloud condensate due to precipitation release in the grid box. The reasoning behind this term is the following: Consider a saturated grid box with cloud condensate in spatially varying amount consistent with a given positive value of A_{st}. It is then possible that the cloud condensate will fall out if the precipitation release parameterization is active, e.g. through a collission and coallescence process associated with precipitation particles entering the grid box from higher altitudes. This 'sweepout' process increases with increasing precipitation flux. The third term describes that it is possible to reduce cloud condensate substantially during precipitation release ('cloud thinning') without substantial compensating condensation. Physically the cloud thinning process due to precipitation 'sweepout' seems realistic.

Finally, the non-stationary conditions as regards $\overline{q}$, $\overline{q}_c$ and $\overline{T}$ will sometimes lead to the onset of stratiform condensation after convective conditions or vice-versa. After convection is no more supported, the convective cloud may exist in a subsaturated environment which is sufficiently dry such that subgrid scale stratiform condensation is not supported. In this situation the evaporation terms, that is, the 3rd terms of (5), (6) and (7) describe also the evaporation of cloud condensate after convection.

To describe the actual transitions the cloud cover f is made time dependent by relaxing towards the equilibrium cloud cover f_{eq} which may be either a stratiform (f_{st}) or a convective equilibrium (f_{cv}).

$$\frac{\partial f}{\partial t} = -K_f(f - f_{eq}) \tag{37}$$

Currently $K_f^{-1} = 900$ s.

4. Microphysics

The model's micro-physics concerns a parameterization of processes related to the formation/decay and fallout of precipitation particles. The microphysics represent the smallest scales down to molecular processes in the atmosphere. An overview of central topics in cloud physics can be found in the litterature, see for example Rogers and Yau (1989).

Some of the equations described so far, e.g., (5), (6) and (7) already contain the effects of microphysics. The terms G_{pc} and E_{pc} describing precipitation release and evaporation of precipitation in the convective case need to be specified for these equations. The similar terms should be formulated in stratiform conditions. Also the effect of melting/freezing of precipitation needs to be specified. These effects are described below. The treatment follows closely the formulations by Sundqvist (1989) and Sundqvist (1993) except for some extensions and few exceptions.

The potential advantage of having cloud condensate as a prognostic variable is that condensation and latent heating may occur without automatically giving rise to precipitation release. For the present scheme this statement applies also to convective condensation. As described in section 2 the present treatment of convection allows for several convective layers or entities in a vertical air column. Stratiform condensation may also occur in parts of an air column, and the cloudiness can be partial. This makes the precipitation release parameterization and an associated description of evaporation of precipitation very complicated.

Technically, this scheme separates between convective precipitation and stratiform precipitation. Both may be present at the same time in a vertical column. A possible mutual interaction between stratiform and convective precipitation fluxes is strongly restricted in the current formulation. This is a common feature of most precipitation schemes used in atmospheric models. The precipitation release formulas described below are basically the same for stratiform and convective precipitation release. The differences involve mainly different constants in the formulas. The precipitation flux from individual layers of the atmosphere are added down to the surface to produce surface precipitation rate. However, evaporation of precipitation is accounted for in this process. Also a treatment of the transitions between water and ice phases makes it possible to distinguish between rain and snow.

We first describe the stratiform precipitation release. The rate of precipitation release $\mathrm{kg} \cdot \mathrm{kg}^{-1}\mathrm{s}^{-1}$ is given by (38)

$$G_p = \Phi \cdot q_c \cdot \left(1 - exp(-\frac{q_c}{f \cdot \mu})^2\right) \tag{38}$$

In (38) f is the fractional cloud cover and q_c is specific cloud condensate as in previous sections. Φ is a rather complex function with unit s^{-1} and describes an inverse time scale associated with the precipitation release formulation.

$$\Phi = \Phi_1 \cdot \Phi_2 \cdot \Phi_3$$

The first term Φ_1 represents an inverse time scale depending on the dynamical model state represented by the vertical velocity ω in the pressure system. Φ_1 decreases in proportion to $-\omega$. The values of Φ decreases for non-positive ω and is constant (ϕ_{st}) for subsidence conditions. It is limited to be no less than ϕ_{00}. The term simulates the effect of precipitation particles being carried with the updraft velocity in the precipitating clouds while it is advected also horizontally. The model resolved vertical velocity adds (subtracts) to the fall velocity, and therefore a longer time scale is associated with the precipitation fallout under conditions of a sufficiently high vertical velocity. The term is novel and simulates the same effect as a prognostic precipitation field which may be advected with the air flow while falling towards the surface with a characteristic velocity. This effect will get increasingly important at a high model resolution since, on average, larger vertical velocities will then be simulated.

$$\Phi_1 = \phi_{st} + K_{\phi_1}\omega_*(\phi_{st} - \phi_{00}) \tag{39}$$

$$\omega_* = \begin{cases} 0 & \text{if } \omega > 0 \\ \omega & \text{if } \omega \leq 0 \end{cases}$$

$\omega = \frac{dp}{dt}$ is the vertical velocity in pressure coordinates. $K_{\phi_1} = 0.02\mathrm{Pa}^{-1} \cdot \mathrm{s}$, $\phi_{00} = 1.0 \cdot 10^{-5}\mathrm{s}^{-1}$, $\phi_{st} = 1.0 \cdot 10^{-4}\mathrm{s}^{-1}$.

$$\Phi_2 = 1 + \sqrt{\frac{P_{co}}{K_{B_1}}} + K_{B_2}\delta_{BF} \tag{40}$$

In (40) $K_{B_1} = 1.0 \cdot 10^{-4}\mathrm{kg} \cdot \mathrm{m}^{-2} \cdot \mathrm{s}^{-1}$, $K_{B_2} = 4$. The term involving K_{B_1} describes the effect of the collection process (collision and coalescence) on the precipitation release (Rogers and Yau, 1989). P_{co} is the precipitation intensity ($\mathrm{kg} \cdot \mathrm{m}^{-2} \cdot \mathrm{s}^{-1}$) per grid square divided by the maximum of the fractional cloud covers present above the vertical level considered. Hence P_{co} is an estimation of the precipitation flux per unit area entering the cloud at the local level considered.

The second term, proportional to K_{B_2}, describes the Bergeron-Findeisen effect (Bergeron, 1935; Findeisen, 1938). This process enhances the precipitation release as ice crystals become present in a water cloud. The coefficient function δ_{BF} has been parameterized according to Sundqvist (1993) and is given by (41) and (42).

$$\delta_{BF} = \widetilde{\delta_{ice}} \cdot (1 - \delta_{ice})\Delta \widetilde{E}_{wi} \tag{41}$$

$$\widetilde{\delta_{ice}} = \delta_{ice} + (1 - \delta_{ice}) \cdot \frac{P_{ice}}{P_{tot}} \tag{42}$$

In (41) and (42) δ_{ice} is a basic function for the probability of ice crystals in clouds (Sundqvist, 1993). This formulation describing an increasing probability of ice crystals in the interval between 273 K and 232 K is based on extensive statistics on the occurrence of ice crystals in clouds (L.T.Matveev, 1984). The functional form is given in the appendix. This function is also used to describe specific latent $L'(T)$ as a function of temperature, appearing in previous equations.

$$L'(T) = L_v + \delta_{ice} \cdot L_i \tag{43}$$

In (43) L_v is the specific latent heat of evaporation and L_i is the specific latent heat of frezing/melting.

The $\Delta\widetilde{E_{wi}}$ function is the difference in satuation vapor pressure over water and ice, divided by its own maximum value. Hence this function is dimensionless and has its largest values between -10° C and -20° C. In is seen that the Bergeron-Findeisen term describes an enhanced precipitation release only if the ice fraction is less than 100 % and if precipitation as snow (ice) entering the layer is positive.

The equation (42) describes a modified ice fraction when computing precipitation release. The fraction of precipitation release from the layer being snow or ice will then be modified from δ_{ice} to $\widetilde{\delta}_{ice}$. A latent heating due to freezing associated with the modified ice fraction is taken into account in the temperature equation. It is noted that this modelling of ice fraction increase as a result of the total precipitation flux from above (stratiform +convective) represents a weak coupling between the two precipitation streams, stratiform and convective precipitation, respectively. In all other aspects the two precipitation streams are currently described independently.

Finally, the term Φ is modelled as a dimensionless function. It has been introduced in order to make precipitation release sufficiently efficient at very low temperatures below 238 K (Källén, 1996).

$$\Phi_3 = \begin{cases} 1 & \text{if } T > 238 \\ 1 + \frac{238-T}{2} & \text{if } T \in [230, 238] \\ 5 & \text{if } T < 230 \end{cases} \tag{44}$$

Looking back on on the precipitation release formula (38) the function in the brackets implies that the precipitation release becomes efficient as the specific cloud condensate inside cloud (q_c/f) reaches a magnitude of μ which is the second important function in the description of precipitation release. This function is specified in (45), (49) and (47):

$$\mu = \frac{\mu_1 \cdot \mu_2}{\Phi_2} \tag{45}$$

$$\mu_1 = \mu_{st} + K_{\mu_1} \cdot \omega_*(\mu_{st} - \mu_{cv}) \tag{46}$$

$$\mu_2 = (1 - \delta_{ice})^2 + \delta_{ice} \cdot \xi(T) \tag{47}$$

In (46) $K_{\mu_1} = K_{\phi_1}$ and the functional form is similar to Φ_1 except for the constants, that is, μ_{st} is a threshold value of cloud condensate in the case of $\omega_* = 0$. The maximum value allowed (μ_{cv}) equals the value normally used for the convection scheme for the case of ω_*=0 (see 49). Currently $\mu_{st} = 5.0 \cdot 10^{-4}$, $\mu_{cv} = 3.0 \cdot 10^{-3}$.

The functions μ_2 and Φ_2 have been adopted from a previous formulation (Sundqvist et al., 1989; Källén, 1996). The function of temperature μ_2 has been found necessary to describe a realistical amount of cloud condensate at low temperatures. The function $\xi(T)$ is given in the appendix.

For convective precipitation release similar formulas are used, except for some difference described in (48) and (49).

$$\Phi_{1_{cv}} = \phi_{cv} + K_{\phi_1} \cdot \omega_*(\phi_{cv} - \phi_{00}) \tag{48}$$

$$\mu_{1_{cv}} = \mu_{st} + K_{\mu_3} \cdot (\mu_{cv} - \mu_{st}) \tag{49}$$

$$K_{\mu_3} = Min(-K_{\mu_1}\omega_* + K_{\mu_2}B, 1)$$

In these equtions index ‘cv’ indicates use for convective conditions. In (48) the currently used value of ϕ_{cv} is $2.5 \cdot 10^{-4}\mathrm{s}^{-1}$. The value of K_{μ_3} is constrained to be no larger than 1. K_{μ_2} =250. The term involving buoyancy B as defined in section 2 is included to provide a continuous formulation between a stratiform and a convective regime. In most situations $\mu_{1_{cv}} = \mu_{cv}$.

The formulations of melting and of evaporation of precipitation are rather uncertain. Several formulations have been used in numerical models. Since less energy is involved with melting compared with evaporation or sublimation a very simple formulation is used for the melting process. As precipitation falls through a layer with temperature above 273 K while the precipitation as snow (ice) is present, the rate of melting M is described according to the following equation (50)

$$M = K_{ml} \cdot \frac{c_p}{L_i}(T - T_{ml}) \tag{50}$$

$K_{ml} = 4.0 \cdot 10^{-4} \cdot \mathrm{s}^{-1}$ The formulation defined by (50) describes in most situations complete melting in a layer of a thickness no more than a few hundred metres.

Finally, evaporation of precipitation takes place in subsaturated model layers. As noted by Tiedtke (1993) parameterizations of this process in operational numerical models may be considered rather uncertain at present. This is reflected in various approaches giving different results. Most formulations apply a formulation E_p ($\mathrm{kg} \cdot \mathrm{kg}^{-1} \cdot \mathrm{s}^{-1}$) proportional to the square root of precipitation intensity, e.g. Sundqvist (1989). Further, the evaporation should in some way depend on the atmospheric subsaturation. Some models use subsaturation $q_s - q$ as a basic parameter (Tiedtke, 1993), others relative humidity (Sundqvist et al., 1989). Precise computations, using the basic physics, require integrations over droplet spectra, utilizing fall velocities and detailed computations of the diffusion process at the surface of the droplets, taking into account features such as ventilation effects (Rogers and Yau, 1989). Currently (51) is used which depends on the subsaturation $q_s - q$ and obeys a square root dependency on precipitation intensity at moderate to high precipitation fluxes, but allows for some increase of evaporation rate at low precipitation intensities. Curently $K_{e_1} = 1.0 \cdot 10^{-3}$, $K_{e_2} = 1.0 \cdot 10^{3}$ and $K_{e_3} = 6.0 \cdot 10^{9}$. Optimal values of these coefficients are not well known at present.

$$E_p = K_{e_1} \cdot \frac{(q_s - q)}{(1 + \frac{L'}{c_p}\frac{\partial q_s}{\partial T})} \cdot \left(\sqrt{P} + \frac{K_{e_2} P}{(1 + K_{e_3} P^2)}\right) \tag{51}$$

5. Discussion and conclusions

It is clear from the presentation so far that that parameterization of CCPE processes in numerical models used for weather forecasting is complicated, for the following reasons:

There are still some general uncertainties on how to construct atmospthetic models in an optimal way for a given set of applications. Apart from the numerical challenge it is still not known how to describe most accurately the subgrid scale fluxes of heat, moisture and momentum. A key problem is to describe all scales in a realistic way for models of a coarser resolution than the cloud resolving models. The turbulence formulation remains a problem even in the cloud resolving models.

The complexity of the CCPE processes and the need to apply parameterizations which are sufficently efficient from a computational point of view sets rather strong limitations on the methods which can be used. As a consequence, this often leads to the development of efficient parameterizations with rather many tuning constants as is the case for the present scheme. Optimal values of tuning constants will depend on other components of the meteorological model including the model dynamics.

For the present forecast model tuning of some parameters is likely to give more optimal results, in particular as regards the formulations of cloud cover and precipitation release which include new features. In addition, the turbulence scheme does currently not diagnose moist unstable conditions in cloudy regions. This implies that the turbulence intensity inside clouds is underestimated and is compensated for by the shallow convection parameterization of the convection scheme. In case that the turbulence scheme is upgraded to an adequate scheme for cloudy conditions some associated tuning must be expected with regard to the shallow convection parameterization. At a very high resolution also the amplitude of the moisture fluctuations described by the subgrid scale condensation scheme should be made consistent with the moisture variation described by the turbulence scheme.

In view of the difficulties involving tuning it is important to decide on a strategy which facilitates real progress in terms of more accurate parameterizations. It seems natural to choose a strategy which splits up the CCPE processes in parts which can be studied more theoretically in order to check, tune and possibly modify existing parameterizations. As an example, the significant spread of the results with different parameterizations for evaporation of precipitation calls for comparisons with detailed theoretical computations based on size distributions of precipitation particles. This treatment should include varying fall velocity of cloud particles and a theoretically advanced treatment of the diffusion proceses at the surface of precipitation particles, including ventilation effects.

Another challenging problem is the forecasting of cirrus clouds. A correct determination of these clouds seems to require that the water phase and the ice phase are stored separately in the model. This means that ‘prognostic cloud condensate’ should be split up into ‘prognostic cloud water’ and ‘prognostic cloud ice’. Ideally, a knowledge of the presence of freezing nuclei is also necessary. Progress in this area may require a refined aerosole treatment. The present scheme does not have such features and can be expected to overestimate the occurrence of cirrus in some situations.

It is also relevant to carry out experiments where the performance of a turbulence scheme combined with the description of the CCPE processes can be compared with detailed measurements, e.g. through the design of 1-dimensional column experiments with specified forcing determined from large field experiments. Related results obtained with very high resolution large eddy simulation models are often very important as a part of this experimental framework.

Experiments along these lines have already been undertaken with the present parameterization package, e.g. using international data sets prepared from field experiments such as BOMEX, ASTEX and experiments designed in projects such as the European Project on Cloud Systems (EUROCS). In order to penetrate into all aspects of the parameterizations many different studies should be carried out. Results obtained so far are promising as regards the realism of the cloud parameterization described in section 3. However, these results are outside the scope of the present report and will be presented elsewhere.

Appendix A. Functions in microphysics

The probability δ_{ice} is according to Matveev (1984):

$$\delta_{ice*} = 1 - A_{ice} \cdot (1 - exp(-\chi^2))$$

$$\delta_{ice} = \begin{cases} 0 & \text{if } T \geq 273 \\ \delta_{ice*} & T \in [232, 273[\\ 1 & \text{if } T \leq 232 \end{cases}$$

$$A_{ice} = \frac{1}{1 - exp\Big[- \Big(\frac{(T_1 - T_{ci})}{(T_2 - T_{ci})\sqrt{2}} \Big)^2 \Big]}$$

$T_1 = 273\text{K}$, $T_2 = 299\text{K}$, $T_{ci} = 232K$

$$\chi = \frac{T - T_{ci}}{(T_2 - T_{ci})\sqrt{2}}$$

The function $\xi(T)$ used to compute a reduced cloud condensate at subfreezing temperatures follows from (Sundqvist et al., 1989; Källén, 1996)

$$\xi(T) = \begin{cases} \frac{4}{3} \cdot exp\Big(- [(T - 273)\frac{2}{30}]^2 \Big) & \text{if } T \geq 250 \\ 0.075 \cdot \Big(1.07 + \frac{y}{1+y} \Big) & T \in [232, 250[\\ 0.075 \cdot \Big(1.07 - \frac{y}{1+y} \Big) & \text{if } T < 232 \end{cases}$$

$$y = x + x^2 + \frac{4}{3}x^3$$

$$x = \frac{|T - 232|}{18}$$

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