An Ensemble Cumulus Convection Parameterization with Explicit Cloud Treatment

TILL M. WAGNER AND HANS-F. GRAF

Centre for Atmospheric Science, University of Cambridge, Cambridge, United Kingdom

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ABSTRACT

The paper describes a convection parameterization employing a new formulation of the quasi-equilibrium closure hypothesis of Arakawa and Schubert. The scheme models a full spectrum of different cumulus clouds and its evolution within one time step of the host global climate model. Each cloud is simulated using a one-dimensional Lagrangian entraining parcel model, which includes mixed phase microphysics and vertical velocity. Hence, the model delivers explicit information on distribution of vertical velocities, precipitation intensities, cloud heights, and cloud coverage. The parameterization is evaluated in the ECHAM single-column model for midlatitude summer and tropical convection. Results show an improved temporal distribution, including the diurnal cycle, of convective heating and moistening in comparison to the Tiedtke–Nordeng scheme, which is the standard convection parameterization within ECHAM. The amount and temporal distribution of precipitation are clearly improved compared with the original parameterization. The convective cloud field model (CCFM) does not produce spurious convection events occurring with the standard parameterization.

1. Introduction

The parameterization of clouds in global climate models has been recognized for decades as a source of substantial uncertainty concerning the estimates of climate variability and sensitivity (Charney et al. 1979; Senior and Mitchell 1993; Randall et al. 2007). However, because of its complexity and discouraging difficulty, progress in convective cloud parameterization has been very slow. Randall et al. (2003) even speaks of a cloud parameterization deadlock. The problem of convection parameterization is far from being solved in a satisfactory way. The Intergovernmental Panel on Climate Change Fourth Assessment Report (IPCC AR4) lists cumulus convection as one of the major problems in global climate modeling. Precipitation remains one of the least reliable variables in climate models. It is not at all well simulated by current climate models, which have “too many days with weak precipitation” (Randall et al. 2007). A reasonable treatment of the physical processes associated with convective clouds is of great importance for many other physical processes in an atmospheric general circulation model (AGCM). Convection controls to a large degree the vertical transport of moisture, chemical tracers, energy, and momentum. When precipitation forms in convective clouds, the net latent heat release directly couples convection to the large-scale dynamics. Cumulus convection not only takes part directly in the global energy and water cycle (transport), but also indirectly by the outflow of cloud water at the top of convective clouds.

Traditionally, within an AGCM cloud processes are separated into convective and stratiform clouds. Water detrainment from convective clouds is used as a source for stratiform clouds including cirrus. Therefore, convection not only leads to a redistribution of moisture and energy within cumulus towers themselves, but also to a decoupling of these quantities from their primary sources. Stratiform clouds in turn have great importance for the radiation budget of the earth. Effects of aerosols on clouds have so far been studied mainly for stratiform clouds [for a review see Lohmann and Feichter (2005)], while their effects on convective clouds, although observed with increasing confidence (e.g., Rosenfeld and Lensky 1998; Andreae et al. 2004), have only started to be addressed (Nober et al. 2003; Lohmann 2008). Since aerosol effects on convective clouds may be highly nonlinear (Graf 2004; Stevens and Feingold 2009), cumulus parameterizations should provide parameters that are
important for the interaction of aerosols and cloud microphysics, such as vertical velocity, which is important for condensation nucleus activation.

Open issues concerning convective clouds include the following:

(i) the effect of aerosols on convective clouds, including large-scale dynamic effects;
(ii) convective transport of chemical tracers, water vapor, and condensate;
(iii) maximum precipitation intensities from convective and large-scale clouds and the diurnal cycle of convection;
(iv) mixed phase microphysics and icing; and
(v) cloud coverage by convective clouds of different heights and its influence on radiation balance.

To solve these problems, the spectrum or probability density function (PDF) in terms of cloud height, radius, and vertical velocities of convective clouds and their microphysics (including interaction with aerosols and different chemical species, soluble and nonsoluble) needs to be explicitly determined at a much higher vertical resolution (~100 m; Graf 2004) than in current climate models. To this end we propose to utilize a newly developed convective cloud field model (CCFM) in an AGCM framework.

Several different attempts have been made to parameterize convection. These approaches are different in complexity and their underlying physical assumptions. While Kuo (1965, 1974) assumed a statistical equilibrium for water substance, in the convective adjustment approach (Manabe and Strickler 1964; Betts 1986; Betts and Miller 1986) an unstable temperature profile is adjusted back toward a profile that is neutral or nearly neutral to convection. Other convection schemes are based on cloud models. Arakawa and Schubert (1974) introduced the idea of an explicit cloud spectrum parameterization. Their scheme describes a spectrum of mass fluxes. Most of the current cumulus convection parameterizations are formulated as bulk mass flux schemes (Emanuel and Raymond 1993; Emanuel 1994). They determine the overall mass flux of all cumulus clouds in one AGCM grid column (Tiedtke 1989; Gregory and Rowntree 1990; Kain and Fritsch 1990; Donner 1993).

One problem with current mass flux schemes is that they describe a single mean convective cloud representing the properties of the cloud spectrum whereas in the real world convective activity always produces many different individual clouds. In general there are many small clouds and just a few deep clouds. Hence, standard bulk mass flux schemes have great difficulties in correctly representing convective-scale transport (Yano et al. 2004; Lawrence and Rasch 2005). The mass flux approach causes a lack of information about cloud dynamics and microphysics, although Zhang et al. (2005) report on the successful introduction of explicit microphysics in the convective mass flux scheme of ECHAM5 (still simulating one “mean” cloud). In contrast to Arakawa and Schubert (1974), Donner (1993), Donner et al. (2001), and Naveau and Moncrieff (2003) describe a spectrum of simplified clouds and not mass fluxes. Cloud dynamical and microphysical structures are represented in a more precise way, but both schemes are based on either observations or high-resolution simulations and therefore are to a certain degree case dependent. Using a cloud-resolving model instead of convection parameterization (also called superparameterization; Grabowski and Smolarkiewicz 1999; Grabowski 2001) is another approach to solve the cumulus problem. However, these schemes are computationally far too expensive to be applied in long climate integrations.

Addressing the open questions mentioned above we use a parameterization in which convective cloud properties of a spectrum of clouds are determined using an entraining parcel model of very high adaptive vertical resolution (~100 m). In principle any other type of cloud model could be used and the modular structure of our scheme allows an easy exchange. Calculating vertical velocities enables the introduction of more sophisticated microphysics including aerosol–cloud interaction. Because of the explicit treatment of vertical velocity and the high resolution, mixed phase processes and precipitation can be represented as realistically as the microphysics scheme allows. Moreover, entrainment and detrainment processes as a function of vertical acceleration and deceleration as well as turbulent cloud–environment mixing dependent on vertical wind shear are potentially resolvable.

The simulation of an explicit cloud spectrum implies spectra of vertical velocities, entrainment and detrainment profiles, precipitation intensities, and cloud heights. This allows for improved vertical transport of water vapor, energy, and chemical tracers as well as cloud coverage by convective clouds of different top heights and, in the future, the consideration of outgoing longwave radiation fluxes from convective cloud fields.

We will first describe the theoretical framework of the CCFM followed by results from single-column model simulations of intensive operational periods (IOPs) of the Atmospheric Radiation Measurement Program (ARM) Southern Great Plains (SGP) site and the Tropical Warm Pool International Cloud Experiment (TWP-ICE).

2. Model description

The CCFM is a spectral convection parameterization for use in general circulation models. It is based on
several key concepts of the Arakawa and Schubert (1974) convection scheme and parameterizes the statistical effect of convection by taking an ensemble of characteristic different types of cumulus clouds into account. It determines their influence on the environment and their mutual interaction via this environmental influence. A first conceptual version of the CCFM was described in Nober and Graf (2005) and a simplified form was successfully applied in a limited area model by Graf and Yang (2007) and Graf et al. (2009). The major differences to the Arakawa and Schubert (1974) parameterization are the use of an explicit cloud model and a different closure formulation. The use of contemporary numerics makes the CCFM computationally affordable and its modular construction leads to uncomplicated implementation of alterations and expansions of different parts of the model. CCFM is only active at time steps and grid cells that allow for convective activity defined by positive convective available potential energy (CAPE).

The parameterization is implemented and tested in the general circulation model ECHAM5, which is the fifth-generation and most recent version of the atmospheric general circulation model developed at the Max Planck Institute for Meteorology (Roeknner et al. 2003). Moreover, it is embedded into the aerosol–climate model ECHAM-HAM, which is an extension of the ECHAM5 model and incorporates aerosol processes and inventories (Stier et al. 2005). ECHAM originated from the numerical weather prediction model of the European Centre for Medium-Range Weather Forecasts (ECMWF). The basic prognostic variables in ECHAM are vorticity, divergence, temperature, surface pressure, and mass mixing ratios of water vapor and cloud water/ice. The governing dynamical equations are formulated in spherical harmonics, whereas all parameterizations and the advective transport are defined on a Gaussian grid. The vertical axis is discretized using a hybrid coordinate system. The standard convection scheme in ECHAM5 is the Nordeng (1994) instability closure modification.

a. Cloud model

A Lagrangian one-dimensional steady-state entraining plume model is used to simulate cumulus clouds that can potentially develop in the environment of the grid cell. It resolves only the vertical dimension and therefore assumes the cloud properties are constant over the slice plane of the cloud. Moreover, it assumes that the cloud is in a steady state, so the cloud does not change over the time step of the host atmospheric model. Finally, the cloudy air in the updraft is assumed to be diluted by entrainment of environmental air by turbulent mixing through the boundary of the cloud and detrainment at the cloud top.

The first model of cumulus clouds using an entraining mass flux approach was formulated by Stommel (1947, 1951). Further developments and the introduction of the notion of a plume for a buoyancy-driven fluid motion originated from the works of Batchelor (1954), Morton et al. (1956), Morton (1957), and Turner (1963), who modeled convection in laboratory experiments and formulated the first set of equations describing the rising plume.

This set of equations was employed in a modified form in the 1960s by Simpson et al. (1965), Simpson and Wiggert (1969), and Weinstein and MacCready (1969) to model cumulus clouds in order to verify cloud seeding experiments. This type of cumulus cloud model was never free from critique (Warner 1970, 1972; Simpson 1971, 1972), mostly because it lacked the ability to simultaneously predict the cloud height and precipitation rate correctly. Nevertheless, because of its computational affordability it is perceived as a valuable compromise between accuracy and cost and in statistical cloud parameterizations, where individual clouds are less important than the spectrum of clouds, it may lead to quite realistic results, as we will show in the current work.

Entraining plume models consist of basically three equations: a vertical momentum equation, a thermodynamic energy equation, and a continuity equation to determine the three unknowns: vertical velocity, temperature, and cloud radius. Additionally, they usually comprise some microphysical parameterization to capture the different phases of water in the cloud, mainly to calculate the precipitation rate.

In our study we use the standard formulation for the dynamical and thermodynamical equations as given in Simpson and Wiggert (1969) or Kreitzberg and Perkey (1976). The cloud base is defined as condensation level of an adiabatically lifted parcel originating from the surface. The cloud top is defined as where the vertical velocity drops below 0.2 m s⁻¹. The entrainment rate μ for the plume model is assumed to be inversely proportional to the cloud’s radius. Therefore, different initial radii lead to different entrainment rates, thus altering the dilution of the cloud by environmental air. This in turn usually produces very different cumulus clouds in the sense of different cloud properties such as height, precipitation rate, amount of detrained cloud water, and vertical transport. This approach is rather simplistic and we are currently working to improve entrainment since this seems to be the parameter leading to most of the CCFM deficiencies (see below).

The cloud microphysics largely follows the mixed phase scheme for stratiform clouds as employed by the
host model ECHAM5 for stratiform clouds (Lohmann and Roeckner 1996).

Besides the number concentration of cloud liquid water and cloud ice, respectively, the mass mixing ratios of the following five water species are considered: water vapor, cloud liquid water, cloud ice, rain, and snow.

The microphysical processes simulated in the scheme are condensational growth of cloud droplets, depositional growth of ice crystals, heterogeneous and homogeneous freezing of cloud droplets, autoconversion of cloud droplets to form rain, aggregation of ice crystals to form snow, accretion of cloud droplets by snow, and the melting of ice crystals and snow. Different from Lohmann and Roeckner (1996), we utilize the autoconversion rate following Berry (1967).

If coupled with the aerosol–climate model ECHAM-HAM, the cloud droplet number concentration (CDNC) is connected to aerosol by an empirical relation to the HAM, the cloud droplet number concentration (CDNC) following Berry (1967). And Roeckner (1996), we utilize the autoconversion rate varying in their cloud-base radius. The maximum cloud-base vertical velocity. The minimal vertical velocity was introduced because the single-column mode does not calculate TKE. Including roughness via TKE improves results (Graf and Yang 2007). These cloud types \( i \) can potentially develop under the given environmental conditions.

The cloud work function \( \text{(CWF)} \) as introduced by Arakawa and Schubert (1974) describes the production of kinetic energy (mechanical work against gravity) of an individual convective cloud characterized by its vertical velocity \( w \), cloud radius \( r \), density \( \rho \), and buoyancy. The cloud work function is usually normalized by its cloud-base mass flux and can be defined as

\[
A(T^c_t, T^v_t) = \frac{1}{w_b r_b \rho_b} \int_{\text{cloud base}}^{\text{cloud top}} \frac{T^v_t - T^v_{\text{env}}}{T^v_t} g w r^2 \rho \, dz, \tag{3}
\]

where \( w_b \) is velocity, \( r_b \) is radius, and \( \rho_b \) is density at cloud base; also, \( T^c_t \) and \( T^v_{\text{env}} \) are the virtual temperature in the cloud and the environment, respectively. The vertical profile of \( T^c_t \) is provided by the cloud model described in the previous section.

According to Lord and Arakawa (1980), the kinetic energy budget for the ensemble of clouds of type \( i \) can be written as

\[
d \frac{d K_i}{d t} = A_i M_{b, i} - K_i \tau_{\text{dis} i}, \tag{4}
\]

where \( \tau_{\text{dis} i} \) is a characteristic time scale for the dissipation of cumulus clouds and the cloud-base mass flux of the ensemble of clouds of type \( i \) is given by

\[
M_{b, i} = x_i w_b r_b \frac{\pi r_b^2}{a}, \tag{5}
\]

where \( x_i \) is the number of clouds of type \( i \), and \( a \) is the area size of the model grid cell. The subscript \( i \) denotes the cloud type and the subscript \( b \) denotes that the value is given at cloud base.

Analogous to the Arakawa and Schubert (1974) quasi-equilibrium hypothesis, we assume a rapid response of convective processes to the slow large-scale processes. This means that the characteristic time scale of the large-scale disturbances \( \tau_{ls} \), which also determines the temporal evolution of the kinetic energy budget [Eq. (4)] and of the cloud work function \( A_i \), is much larger than the characteristic time of cloud dissipation \( \tau_{\text{dis} i} \) (i.e., \( \tau_{ls} \gg \tau_{\text{dis} i} \)). Therefore the left-hand side of Eq. (4) can be neglected (Lord and Arakawa 1980) and the kinetic energy budget [Eq. (4)] reduces to

\[
A_i M_{b, i} = \frac{K_i}{\tau_{\text{dis} i}}, \tag{6}
\]
Equation (6) is referred to as kinetic energy quasi-equilibrium equation (Lord and Arakawa 1980), since it states that the production of kinetic energy by convective clouds is in equilibrium with the dissipation of cumulus cloud kinetic energy (for time scales much smaller than \( \tau_{\text{dis}} \)).

Following both Arakawa and Xu (1992) and Randall and Pan (1993) the cloud-base mass flux \( M_{b,i} \) is related to the kinetic energy by the following relationship:

\[
K_i = \alpha_i M_{b,i}^2, \tag{7}
\]

where \( \alpha_i \) is a constant parameter associated with the cloud type \( i \). Substituting Eq. (7) into the kinetic energy budget [Eq. (6)] gives

\[
\frac{A_i}{M_{b,i}} = \frac{\alpha_i}{\tau_{\text{dis}}}, \tag{8}
\]

Equation (8) provides a diagnostic relationship between the cumulus mass flux and the cloud work function.

Several convection parameterizations utilize a similar diagnostic relationship between cumulus mass flux and cloud work function or CAPE as closure (Moorthi and Pan (1993); Zhang and McFarlane (1995); Gregory et al. (2000); Neale et al. (2008). Evaluation of formulations employing CAPE using atmospheric observations suggest that convective quasi-equilibrium (CQE) holds for longer (i.e., diurnal) time scales, but must be considered problematic on shorter time scales especially for midcontinent North America (Donner and Phillips 2003).

Instead of directly using Eq. (8) by estimating \( \alpha_i \) via the updraft vertical velocity and making an assumption on \( \tau_{\text{dis}} \), for which we currently lack a theory (Pan and Randall 1998), we will follow the derivation of Arakawa and Schubert (1974) and link the mass flux to the change of cloud work function due to large-scale processes. An alternative approach to derive Eq. (8) is to insert relationship (7) into the kinetic energy budget [Eq. (4)] and assume convective quasi-equilibrium.

Assuming that the parameter \( \alpha_i \) as well as the cloud dissipation time scale \( \tau_{\text{dis}} \) is independent of time (Lord and Arakawa 1980; Arakawa and Xu 1992; Randall and Pan 1993) results in the following equation describing the evolution of the cloud-base mass flux:

\[
\frac{d}{dt} M_{b,i} = M_{b,i} \frac{d}{dt} A_i. \tag{9}
\]

The time change of the cloud-base mass flux through stabilizing and destabilizing processes, given by \( dA_i/dt \), can be attributed either to a change in the vertical velocities \( w_i \), which will lead to more vigorous convection, or to a change in the cloud coverage by cloud types \( i \) reflected by the cloud numbers \( x_i \), leading to more or less cumulus clouds. Robe and Emanuel (1996) and Emanuel and Bister (1996) showed that the vertical velocities of individual clouds do not differ much with variations of radiative cooling in radiative equilibrium calculations. Therefore, cumulus convection reacts to changes in atmospheric instability mainly through changes in the number of clouds \( x_i \), rather than through changes in the intensity (e.g., vertical velocity) of convection. Hence, we assume invariant cloud types \( i \) within one time step of the host model. Making use of the explicit spectrum of individual clouds \( x_i \) according to Eq. (5), the time derivative of the cloud-base mass flux can be written as

\[
\frac{d}{dt} M_{b,i} = w_{b,i} \rho_{b,i} \frac{\pi r_{i,j}^2}{a} \frac{d}{dt} x_i. \tag{10}
\]

Within the next time step a new initial cloud spectrum is being computed according to the then-changed vertical profiles.

The equation for the cloud-base mass flux evolution [Eq. (9)] then simplifies to

\[
\frac{dx_i}{dt} = \frac{x_i}{A_i} \frac{dA_i}{dt}, \quad i = 1, \ldots, n. \tag{11}
\]

Equation (11) is a coupled system of first-order ordinary differential equations describing the evolution of the cumulus cloud population.

Analogous to Arakawa and Schubert (1974), the time derivative of the cloud work function \( A_i \) can be split into two terms:

\[
\frac{d}{dt} A_i = \left( \frac{dA_i}{dt} \right)_{\text{cu}} + \left( \frac{dA_i}{dt} \right)_{\text{ls}}, \tag{12}
\]

with the first term on the right-hand side describing the stabilizing effect of convective clouds of any type \( j \) and the second term describing the destabilizing effect of the large-scale environment. More precisely, the effect of cloud type \( j \) on clouds of type \( i \) via stabilizing influence on the environment is defined through the so-called kernel \( K_{ij} \) as \( \left( dA_i/dt \right)_{\text{cu}} = \sum_{j=1}^{n} K_{ij} x_j \). The second term on the right-hand side of Eq. (12) representing solely the large-scale destabilizing processes that force the occurrence of cloud \( i \) is denoted \( \left( dA_i/dt \right)_{\text{ls}} = F_i \). In the case of atmospheric convection, large-scale processes force atmospheric convection and \( F_i \) will be positive, whereas the kernel \( K_{ij} \) has to be negative since cumulus clouds reduce atmospheric instability and therefore inhibit further production of convection. In practice the condition of \( F_i > 0 \) is employed to test for the occurrence of convective activity.
Using the above notation, the evolution for the cloud population yields
\[
\frac{dx_i}{dt} = \frac{F}{A_i} x_i + \sum_{j=1}^{n} K_{ij} x_j, \quad i = 1, \ldots, n. \tag{13}
\]

With the definition of \( r_i := F/A_i \) and \( a_{ij} := -K_{ij}/F_i \), the cloud evolution equation can be further simplified and written as
\[
\frac{dx_i}{dt} = r_i x_i \left(1 - \sum_{j=1}^{n} a_{ij} x_j \right), \quad i = 1, \ldots, n. \tag{14}
\]

The cloud evolution in the form of Eq. (14) is a multivariate Lotka–Volterra system that describes a competitive system of \( n \) interacting species. An important property of the Lotka–Volterra system is the so-called carrying capacity \( \xi_i \), defined as the equilibrium amount of one cloud type \( i \) in the absence of other clouds—that is, the solution of
\[
0 = r_i x_i \left(1 - \sum_{j=1}^{n} a_{ij} x_j \right), \quad x_j = 0 \text{ for } j \neq i. \tag{15}
\]

Thus, the carrying capacity \( \xi_i \) of cloud type \( i \) is given by
\[
\xi_i := \frac{1}{a_{ii}} = -\frac{F_i}{K_{ii}}. \tag{16}
\]

The cloud evolution equation [Eq. (14)] is the closure formulation for the convective cloud field model. It defines an evolution of a cloud spectrum within a region, depending on destabilizing effects of large-scale processes and stabilizing effects of the cumulus cloud spectrum itself.

To numerically solve the CCFM closure equation the cloud forcing coefficients \( r_i \) and the cloud interaction matrix \( a_{ij} \) have to be defined. Following the concept of Lord (1982), the coefficients will be determined by directly calculating the time change of the CWF, utilizing the host model operator splitting. This means we calculate the cloud work function at three different stages during the time step calculation of the host model.

The first stage is at the end of the previous large-scale model time step (i.e., after the large-scale flow is calculated and all subgrid-scale parameterized processes are accounted for). Variables at this stage are denoted by the superscript \( m1 \). The second stage is after the computation of the large-scale processes, such as radiation and vertical diffusion by turbulence. A physical quantity \( \varphi \) at this state is defined as
\[
\varphi^b = \varphi^{m1} + \left( \frac{\Delta \varphi}{\Delta t} \right)_{ls}, \tag{17}
\]

where \( \Delta t \) is the time step of the host atmospheric model and \( (\cdot)_{ls} \) refers to the contribution of large-scale processes.

At the third stage large-scale processes have been computed and the effect of one single cumulus cloud of type \( i \) has been accounted for. A physical variable \( \varphi \) at this state is defined as
\[
\varphi^k = \varphi^{m1} + \left( \frac{\Delta \varphi}{\Delta t} \right)_{ls} + \left( \frac{\Delta \varphi}{\Delta t} \right)_{i}, \tag{18}
\]

where \( (\cdot)_{i} \), refers to the contribution of cumulus cloud \( i \).

Then, the forcing of cloud \( i \) due to large-scale processes (potentially leading to atmospheric instability) is defined as the rate of change in time of the CWF of cloud type \( i \) through large-scale processes and can be written as
\[
F_i = \frac{A(T_{cu^{i1}}, T_{v}^{i1}) - A(T_{cu^{i0}}, T_{v}^{m1})}{\Delta t}, \tag{19}
\]

where \( T_{cu^{i0}} \) denotes the virtual temperature within the cumulus cloud of type \( i \).

Accordingly, the suppression of cumulus clouds of type \( i \) due to the occurrence of clouds of type \( j \) is defined as the time rate change of the CWF of cloud type \( i \) through stabilizing effects of clouds of type \( j \), that is,
\[
K_{ij} = \frac{A(T_{cu^{j1}}, T_{v}^{i1}) - A(T_{cu^{j0}}, T_{v}^{m1})}{x_j \Delta t}. \tag{20}
\]

The influence of clouds on the environment is normalized to the effect of one single cloud; therefore, \( x_j = 1 \).

Thus, the cloud forcing coefficient \( r_i \) defined as
\[
r_i = \frac{F_i}{A_i} = \frac{A(T_{cu^{i1}}, T_{v}^{i1}) - A(T_{cu^{i0}}, T_{v}^{m1})}{A(T_{cu^{i0}}, T_{v}^{ls}) \Delta t} \tag{21}
\]
describes the relative time change of the CWF of cloud type \( i \) as a result of large-scale processes. Finally, the interaction coefficient \( a_{ij} \) is given by
\[
a_{ij} = -\frac{K_{ij}}{F_i} = \frac{A(T_{cu^{j1}}, T_{v}^{i1}) - A(T_{cu^{j0}}, T_{v}^{m1})}{A(T_{cu^{j0}}, T_{v}^{ls}) - A(T_{cu^{j0}}, T_{v}^{i1})} \tag{22}
\]

and describes the proportionality of the suppression of cloud \( i \) by competing clouds of type \( j \) and the large-scale forcing of clouds of type \( i \).

As a realistic first guess for the spectrum of convective clouds, we define the initial spectrum \( x_i^0 \) through the carrying capacity of cloud types \( x_i \), divided by the number of different cloud types \( n \), that is,
\[
x_i^0 = \frac{\xi_i}{n} = -\frac{F_i}{K_{ii}}. \tag{23}
\]
This can be interpreted as dividing the total production of convective instability by large-scale processes into $n$ shares, which are equally distributed over all cloud types.

In theory, the cloud evolution equation [Eq. (4)] needs to be integrated until the cloud population reaches equilibrium. Such a cloud population would also satisfy the classical quasi-equilibrium equation of Arakawa and Schubert [1974; cf. their Eq. (150)]. In practice we find that the cloud numbers show a reduced variability in the majority of situations after around 1 h. This seems plausible since this time scale is on the order of the cumulus cloud lifetime. Thus, over this time span all clouds felt the influence of other members of the cloud population and weak members, in terms of low large-scale forcing and high suppression by other clouds, are already vaporized, of convective instability by large-scale processes into fundamental assumptions of the quasi-equilibrium hypothesis, will be the subject of a separate paper.

c. Effect on the large-scale variables

1) TEMPERATURE AND MOISTURE

The in-cloud variables calculated by the cloud model on a high vertical resolution ($\sim 100$ m) are restricted to and stored at the vertical levels defined by the host model. The calculation of the convective heating (cooling) and drying (moistening) follows the formulation of Tiedtke (1989). Convective clouds are considered to modify the large-scale environment by vertical transport and processes of condensation (evaporation), resulting in removal of atmospheric water vapor through precipitation. These effects can be expressed in geometric coordinates by

$$
\left( \frac{\partial s}{\partial t} \right)_\text{cu} = \bar{L}_\nu (c - e) - \frac{1}{\rho} \frac{\partial}{\partial z} \left( \rho w^s s \right),
$$

(24)

$$
\left( \frac{\partial q_v}{\partial t} \right)_\text{cu} = -(c - e) - \frac{1}{\rho} \frac{\partial}{\partial z} \left( \rho w^q q_v \right),
$$

(25)

where $s$ is the dry static energy, $L_{\nu}$ is the latent heat of vaporization, $z$ is the geometric height, $q_v$ is the water vapor mixing ratio, and $c$ and $e$ are the rate of condensation and evaporation, respectively [thus, $(c - e)$ is the rate of net condensation]. The overbar denotes averages over the horizontal area of the grid cell and the prime denotes the deviation from the horizontal averages due to convective processes.

The convective heating by subgrid-scale vertical transport of dry static energy can be written as

$$
\frac{-1}{\rho} \frac{\partial}{\partial z} (\rho w^s s) = \frac{-1}{\rho} \frac{\partial}{\partial z} [((M)_{\text{cu}} - M_{\text{cu}})],
$$

(26)

where $(M)_{\text{cu}} = \sum_i^n \sigma_i \omega_i \rho_i s_i$ is the total static energy flux of the cumulus cloud population, with $\sigma_i = \pi r^2 v \lambda / \alpha$ denoting the cloud cover fraction and $\omega_i$, $\rho_i$, and $s_i$ denoting the vertical velocity, density, and static energy within clouds of type $i$, respectively. Furthermore, $M_{\text{cu}}$ is the static energy flux in the cloud-free part of the grid cell induced by compensating subsidence, where $M_{\text{cu}} = \sum_i^n \sigma_i \omega_i \rho_i$ denotes the total convective mass flux.

The rate of net condensation within the cumulus cloud can be expressed as the change of condensate mass flux

$$
(c - e) = \frac{1}{\rho} \frac{\partial}{\partial z} \left( [M(q_i + q_r + q_s + q_s)]_{\text{cu}} \right),
$$

(27)

where $q_i$, $q_r$, $q_s$, and $q_s$ denotes the mass mixing ratios of cloud liquid water, rain, cloud ice, and snow, respectively.

Combining the equations for convective heating through subgrid-scale transport of static energy [Eq. (26)] and the release of latent heat by net condensation (freezing) [Eq. (27)] and furthermore assuming hydrostatic balance

$$
\frac{1}{\rho} = -g \frac{\partial z}{\partial p},
$$

(28)

the convective heating (cooling) [Eq. (24)] can be written as

$$
\left( \frac{\partial s}{\partial t} \right)_\text{cu} = g \frac{\partial}{\partial p} \left( (M)_{\text{cu}} - M_{\text{cu}} \right) - L_v \left[ M(q_i + q_r) \right]_{\text{cu}} - L_s \left[ M(q_i + q_s) \right]_{\text{cu}}. \tag{29}
$$

Here $q_i$, $q_r$, and $q_s$ denote mixing ratios of ice and snow, respectively, which originate from freezing of liquid water. Similarly, $q_s$ denotes the mixing ratio of cloud ice originating from deposition (sublimation).

Equivalently, the convective moistening [Eq. (25)] can be written as

$$
\left( \frac{\partial q_v}{\partial t} \right)_\text{cu} = g \frac{\partial}{\partial p} \left( (Mq_v)_{\text{cu}} - M_{\text{cu}} \right) q_v + \left[ M(q_i + q_r + q_s) \right]_{\text{cu}}. \tag{30}
$$

The effect of the cumulus cloud population on a physical quantity $\varphi$, such as tracers or horizontal momentum, is given by
\[
\left( \frac{\partial \varphi}{\partial t} \right)_{\text{cu}} = g \frac{\partial}{\partial p} \left[ (M\varphi)_{\text{cu}} - M_{\text{cu}} \varphi + (M\varphi_S)_{\text{cu}} \right],
\]

where \( \varphi_S \) is the physical quantity \( \varphi \) originating from a source/sink term \( S \) within the convective updraft.

2) PRECIPITATION AND DETRAINMENT

The precipitation rate is defined as the vertically integrated generation of rain and snow within the cumulus cloud:

\[
P = \int_{\text{base}}^{\text{top}} \frac{\partial}{\partial z} [M(q_t + q_s)]_{\text{cu}} \mathrm{d}z.
\]

Evaporation of precipitation falling through unsaturated air below the cloud base is not taken into account in the current model version.

Cloud water, cloud ice, moisture, and other transported tracers are detrained from the cloud at the cloud top. Moisture, cloud water, and cloud ice are of significant importance for the formation of stratiform clouds at the cumulus cloud-top level.

The rate of detraining is defined as the total updraft flux of quantity \( \varphi \) at the cloud top and is therefore given by

\[
\left( \frac{\partial \varphi}{\partial t} \right)_{\text{det}} = \frac{1}{\rho \Delta z} (M\varphi)_{\text{cu}}^{\text{top}}
\]

\[
= -\frac{g}{\Delta p} (M\varphi)_{\text{cu}}^{\text{top}},
\]

where \( \Delta z \) and \( -\Delta p \) denote the height of the detrainment layer in geometric and pressure coordinates, respectively. As usual, \( \varphi \) denotes any physical quantity such as water vapor, cloud water, cloud ice, chemical tracers, or aerosols.

d. Single-column model

The single-column model (SCM) of ECHAM5 used in this evaluation study is set up as follows. One vertical column of the horizontal area size of approximately 2.8° × 2.8° (corresponding to a spectral triangular truncation at T42) is extracted. The vertical coordinate is discretized into 19 levels in a hybrid grid and the temporal resolution is 30 min.

The boundary conditions or forcing data to drive the SCM replacing the atmospheric flow computation are the large-scale horizontal and vertical advective tendencies of mass, temperature, moisture, and additionally surface pressure. The initial values defining the starting conditions of the SCM are the values of temperature, moisture, horizontal and vertical wind, and surface pressure. To fit the vertical and temporal resolution of the SCM, the provided forcing data are linearly interpolated.

e. Evaluation and boundary/initial-value data

Three datasets are used for the evaluation of the CCFM in the single-column model. The first two are produced by ARM, which aims to improve the scientific understanding of the fundamental physical processes in the atmosphere and provides data products that promote the advancement of climate models. These datasets are produced by observations from IOPs at the Southern Great Plains ARM site in Oklahoma. These observations include 3-hourly radiosonde data at five locations, hourly wind data from 17 stations, microwave radiometer data for water vapor and cloud liquid water, and precipitation measurements at over 50 surface meteorological stations, as well as satellite data to estimate the earth radiation budget. The forcing data are then constructed from these observations using the variational method as described by Zhang and Lin (1997) and Zhang et al. (2001). A full description of used observations, production of forcing data, and different SCM designs is given in Ghan et al. (2000).

The two particular ARM datasets used in this study are the summer 1995 IOP, which was conducted between 18 July and 4 August 1995, and the summer 1997 IOP, which took place from 18 June to 17 July 1997. The summer 1995 IOP was employed for the first SCM intercomparison study (Ghan et al. 2000). The meteorological situations during the summer 1995 IOP were those of a typical continental summertime regime. The weather conditions ranged from varying cloudiness with precipitation every other day in the first half of the IOP, with several clear, hot days in the beginning of the second half, to large-scale synoptic forcing with increased cloudiness and occasional strong precipitation. The summer 1997 IOP was already utilized for cloud-resolving model (Xie et al. 2002) and convection parameterization intercomparisons (Xu et al. 2002). The meteorological situation during the summer 1997 IOP was characterized by several strong precipitation events with a few intermittent dry and clear days.

The third dataset aims to test the parameterization for tropical marine convection. Therefore, a dataset from the Stratospheric–Climate Links with Emphasis on the Upper Troposphere and Lower Stratosphere/Aerosol and Chemical Transport in Deep Convection (SCOUT-O3/ACTIVE) campaign, which is associated with the Tropical Warm Pool-International Cloud Experiment (TWP-ICE) in Darwin, Australia, is used. The dataset is based on the Diagnostique des Domaines Horizontaux (DDH) analysis of the 24-h high-resolution (40 km) forecast of the ECMWF integrated forecast system. Hence, it is directly based not on observations but on a high-resolution weather prediction model reinitialized from analysis.
every 24 h. Within the TWP-ICE campaign, observations at the ARM site in Darwin, Australia, have also been made in order to produce the first SCM forcing dataset for tropical convection. However, the forcing dataset was not available in time for this study. The dataset employed here covers the whole month of November 2005 and the meteorological situation during this time was characterized by the premonsoon period with pronounced diurnal evolution of strong deep convection.

Data for the liquid water path (LWP) or precipitable water (PW) are not available from the abovementioned dataset. Therefore, the ECMWF Re-Analysis (ERA)-Interim reanalysis data are shown for qualitative comparison. We selected the grid box located at 12.8°S, 130.5°E, which is the closest location to the Darwin central point at 12.25°S, 130.53°E. Whereas the PW data are 6-hourly data based on analysis, the LWP data are 3-hourly forecast data.

**FIG. 1.** Observed and simulated 3-hourly mean total precipitation: (a) summer 1995 SGP IOP, (b) summer 1997 SGP IOP, (c) November 2005 Darwin TWP ICE.
f. Configurations

Two different configurations of the ECHAM5 single-column model are evaluated. The first setup is the standard ECHAM5 model configuration, which employs the Tiedtke (1989) scheme including the Nordeng (1994) CAPE closure. This configuration is referred to as the Tiedtke setup. The second setup is the ECHAM5 model incorporating the convective cloud field model instead of the standard Tiedtke (1989) convection parameterization. This configuration is referred to as the CCFM setup.

3. Results

a. Precipitation

The precipitation produced by convection is one of the most important climate model outputs and the rate of precipitation is directly connected to the convective heating and drying. Figure 1 shows the comparison of the observed and simulated 3-hourly surface precipitation rates for all three SCM evaluation cases.

For all three test cases the CCFM parameterization captures all precipitation events, solely missing the occurrence of convection at the 26 July 1995 during the summer 1995 SGP IOP. Moreover, hardly any and only very light spurious occasions of precipitation are produced.

In contrast, the simulation utilizing the Tiedtke parameterization creates several spurious precipitation events, while missing other strong convection occurrences, such as in the beginning of the first phase (19–23 July 1995) of the summer 1995 SGP IOP, as shown in Fig. 1a, or later in the convection-free phase in the same test case. Generally, during the summer 1997 SGP IOP (Fig. 1b) the Tiedtke scheme has difficulties predicting the time and amount of precipitation correctly. Because of the strong diurnal cycle of tropical convection the events during the November 2005 Darwin TWP-ICE case (Fig. 1c), the Tiedtke scheme performs better than for the less regular occurrences of convection of the midlatitude cases, although the total amount during individual events is not very accurately estimated.

The overall average precipitation of the CCFM and the Tiedtke parameterization estimates given in Table 1 for the SGP test cases are almost equal and only slightly underestimate the observed precipitation. In contrast, for the 2005 Darwin case the CCFM overestimates the overall precipitation, whereas the Tiedtke scheme underestimates the precipitation by almost the same amount.

The corresponding correlation coefficients are shown in Table 2. As a visual inspection of the precipitation distribution in Fig. 1 already suggests, the correlation of the precipitation distribution between the CCFM and the observations is much stronger than for the case of the Tiedtke parameterization.

One advantage of the CCFM is the explicit information about the cloud spectrum. Figure 2 shows the distribution of precipitation intensity calculated for the summer 1997 SGP IOP. Precipitation rates produced by different clouds are classified into five categories of intensity. The intensity is given as precipitation rate for the cloud-covered area and the large-scale model time step only.

Most convection events that produce only small amounts of total convective precipitation (<0.5 kg m$^{-2}$ h$^{-1}$) are dominated by low-intensity precipitation events. For cases of convection producing medium amounts of total convective precipitation (<1 kg m$^{-2}$ h$^{-1}$) (24 and 25 June 1997 as well as 9, 11, 12, 17, and 18 July 1997) no prevalent regime is found. Each case is dominated by different precipitation intensities. Even cases of convection

\begin{table}[h]
\centering
\begin{tabular}{lccc}
\hline
\hline
Precip (kg m$^{-2}$ h$^{-1}$) & Observation & 0.318 & 0.178 & 0.274 \\
 & CCFM & 0.280 & 0.171 & 0.328 \\
 & Tiedtke & 0.281 & 0.168 & 0.229 \\
LWP (kg m$^{-2}$) & Observation & 0.112 & 0.030 & 0.033* \\
 & CCFM & 0.135 & 0.113 & 0.041 \\
 & Tiedtke & 0.077 & 0.029 & 0.078 \\
IWP (kg m$^{-2}$) & CCFM & 0.042 & 0.032 & 0.012 \\
 & Tiedtke & 0.036 & 0.028 & 0.005 \\
PW (kg m$^{-2}$) & Observation & 39.1 & 36.5 & 48.1 \\
 & CCFM & 57.3 & 42.1 & 65.1 \\
 & Tiedtke & 48.8 & 37.0 & 39.5 \\
TCC & Observation & 0.60 & 0.43 & 0.739 \\
 & CCFM & 0.72 & 0.857 & 0.902 \\
 & Tiedtke & 0.64 & 0.749 & 0.479 \\
\hline
\end{tabular}
\caption{Average precipitation, liquid water path, ice water path, and precipitable water. Data from reanalysis instead of observations are marked by an asterisk.}
\end{table}

\begin{table}[h]
\centering
\begin{tabular}{lccc}
\hline
\hline
Precip & CCFM & 0.74 & 0.81 & 0.85 \\
 & Tiedtke & 0.41 & 0.34 & 0.66 \\
LWP & CCFM & 0.69 & 0.47 & 0.48 \\
 & Tiedtke & 0.48 & 0.26 & 0.78 \\
PW & CCFM & 0.76 & 0.82 & 0.78 \\
 & Tiedtke & 0.78 & 0.62 & 0.78 \\
TCC & CCFM & 0.21 & 0.33 & 0.23 \\
 & Tiedtke & 0.23 & 0.38 & 0.23 \\
\hline
\end{tabular}
\caption{Correlation coefficients between the observed and simulated distributions of precipitation, and liquid water path.}
\end{table}

1 At this sample size the correlation coefficients are significant at confidence levels of 0.99 (using the Student’s $t$ test).
producing large amounts of precipitation are not necessarily dominated by high-intensity events (cf. 4 July 1997).

The occurrence of convection of 26 July 1997 is displayed in more detail. In the beginning of the event clouds with medium precipitation intensity dominate. As the convection becomes more vigorous, an increasing fraction of precipitation is produced by intensively precipitating clouds. Toward the end of the event the precipitation intensity reduces again. Overall, the simulated distribution seems to give a realistic picture, although observations of this detail are not available and hence a direct evaluation of the results is not possible.

The information of the precipitation intensity distribution may be used to estimate the frequency of extreme precipitation events. Obviously, distributions of other cloud properties such as vertical velocity and cloud height are important as well. These can be utilized to estimate processes and events that are connected to extreme values of cloud properties (e.g., lightning or penetrative convection into the lower stratosphere).

b. Liquid water path and precipitable water

The liquid water path is displayed in Fig. 3. Obviously, the CCFM overestimates the overall LWP by detraining too much cloud water and ice at the cloud top, whereas the Tiedtke scheme produces amounts that fit the observations better. This can be also seen from Table 1, which shows a strong overestimation of CCFM LWP for the summer 1997 SGP IOP. Despite this overestimation, the CCFM shows a better correlation between the observed and the simulated distribution of the LWP than the Tiedtke scheme (cf. Table 2). This points to a superior closure formulation of the CCFM but also to possible deficits in the entrainment assumptions of the cloud model, as mentioned above.

Figure 4 displays the comparison of the total precipitable water distribution for the test cases at the ARM SGP site. Both the Tiedtke and the CCFM parameterizations overestimate the PW amount for the SGP cases and slowly accumulate moisture, which corresponds to the underestimation of precipitation rate for both models. The enhanced precipitable water is the result of reduced precipitation in few cases (20 and 26 July 1995 for the 1995 ARM case and 24 June 1997 for the 1997 ARM case). The PW differences last until the model produces a higher precipitation than observed (e.g., Tiedtke for 31 July 1995 or Tiedtke for 27 June 1997). For the SGP cases CCFM barely overestimates the precipitation; therefore, once atmospheric moisture is enhanced it remains so in the single-column model. The situation is different for the Darwin case; here the Tiedtke scheme underestimates precipitation as well as PW, whereas the CCFM overestimates precipitation as well as PW. Compared to the Tiedtke scheme the CCFM produces a warmer surface temperature, which leads to an enhanced latent heat flux and in turn more clouds and precipitation. One cause is the missing incorporation of evaporating precipitation below the cloud base another reason is the missing adjustment and tuning of the CCFM toward radiational balance of the ECHAM CCFM setup. The correlation (cf. Table 2) is strong for both schemes, with the CCFM being almost equal for the 1995 case and better for the 1997 case.

The time average of precipitation, liquid and ice water path, precipitable water, and total cloud cover (TCC), which is defined only for stratiform clouds, is given in
The overestimation of liquid and ice water path by the CCFM shows an enhanced formation of stratiform clouds, which is also shown in the larger total cloud cover.

The correlation coefficients between the distribution of observed and simulated precipitation, LWP, PW, and total cloud cover are given in Table 2. In addition to the total cloud cover, the CCFM produces a better time distribution of the considered variables. The correlation for variables taken from reanalysis is not calculated since they do not stem from a dataset that is consistent with the forcing data and therefore are not to the same extent expected to be reproduced by the simulation.

Overall, the data confirm that the CCFM produces good estimates of the precipitation rate that are consistent with the convective heating and moistening rates, but it overestimates the cloud-top detrainment, thus producing an increased liquid and ice water path, which in turn results in an overestimation of precipitable water and cloud cover. Nonetheless, the strong correlation of the CCFM with the observations shows the good ability of the CCFM to correctly diagnose the occurrence of moist convection.
4. Summary and conclusions

In this work we have presented a comprehensive description of the convective cloud field model, a spectral convection parameterization with a new formulation of the quasi-equilibrium closure hypothesis of Arakawa and Schubert (1974). The discussed parameterization models the evolution of a convective cloud field within one time step of the host model. In contrast to the original Arakawa and Schubert (1974) closure, the presented formulation guarantees a nonnegative solution, which is of crucial importance when using the closure in a numerical model. Each cumulus cloud type is modeled by a one-dimensional Lagrangian entraining parcel model, which includes a description for temperature, water vapor, and four different water species (cloud liquid water, cloud ice, rain, and snow) as well as vertical velocity. The cloud microphysics includes mixed phase processes and is consistent with the microphysical model used in the stratiform cloud model of the host model ECHAM. Since the full spectrum of different clouds is modeled, the parameterization provides valuable...
information about the distribution of cloud characteristics within one grid cell of an AGCM. This potentially enables a more realistic introduction of nonlinear cloud processes such as CCN activation and aerosol effects on convective clouds.

The new convection parameterization was evaluated in the ECHAM SCM under summertime midlatitude and tropical conditions using observations and operational analysis data. The results show that the new parameterization has very good ability to simulate the amount and temporal occurrence of the observed precipitation events. The vertical extent of the convective events is on average slightly overestimated, which is caused by the crude but commonly employed entrainment parameterization. However, in comparison to the Tiedtke (1989) scheme, which is the standard convection parameterization of ECHAM, the spatiotemporal patterns of the CCFM simulated convection events, as shown by the convective heating and moistening rates, are clearly improved.

The CCFM simulation tends to overestimate the observed values of LWP and PW, and for these variables the Tiedtke scheme produces better estimates than the CCFM. However, LWP and PW are also influenced by other parameterizations, such as the stratiform cloud scheme, which is to a certain extent adjusted toward working with the original (Tiedtke) convection scheme, whereas the CCFM has not been tuned. These overestimates of the CCFM are again connected to the entrainment and detrainment parameterization, which produces a moist bias at the detrainment level at the cloud top. Currently work has been undertaken to evaluate the Lagrangian entraining parcel model against a cloud-resolving model to test new entrainment descriptions in order to improve the modeling of entrainment and detrainment processes and the single cumulus cloud model in general. Another reason for CCFM errors is the poor vertical resolution of the host model, which leads to highly smoothed vertical profiles, excluding, for example, inversion layers in the free troposphere and their effects on deep convection Graf (2004).

In conclusion, the CCFM clearly improves the spatiotemporal distribution of precipitation and convective activity in comparison to the Tiedtke scheme within ECHAM. The currently observed weaknesses compared to the Tiedtke parameterization may likely be overcome by adjusting and tuning the CCFM to the ECHAM model. Since the cloud spectrum is explicitly modeled, a large amount of additional information is available and the information of vertical velocity allows for the incorporation of important microphysical processes such as cloud condensation nuclei (CCN) activation in a physically consistent way. ECHAM-CCFM was also run in climate model setup for 20 years forced by observed SST [the second Atmospheric Model Intercomparison Project (AMIP2)]. The results show that our model is within the range of standard ECHAM and reanalysis data [National Centers for Environmental Prediction (NCEP) and the 40-yr ECMWF Re-Analysis (ERA-40)]. The analysis and results of these global integrations of the model will be shown in a follow-up paper. Overall, we believe that the CCFM has the potential to be a very useful tool.

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