Developing a Cloud Scheme With Prognostic Cloud Fraction and Two Moment Microphysics for ECHAM-HAM

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Abstract We present a new cloud scheme for the ECHAM-HAM global climate model (GCM) that includes prognostic cloud fraction and allows for subsaturation and supersaturation with respect to ice separately in the cloud-free and cloudy air. Stratiform clouds form by convective detrainment, turbulent vertical diffusion, and large-scale ascent. For each process, the corresponding cloud fraction is calculated, and the individual updraft velocities are used to determine cloud droplet/ice crystal number concentrations. Further, convective condensate is always detrained as supercooled cloud droplets at mixed-phase temperatures (between 235 and 273 K), and convectively detrained ice crystal number concentrations are calculated based on the updraft velocity. Finally, the new scheme explicitly calculates condensation/evaporation and deposition/sublimation rates for phase-change calculations. The new cloud scheme simulates a reasonable present-day climate, reduces the previously overestimated cirrus cloud fraction, and in general improves the simulation of ice clouds. The model simulates the observed in-cloud supersaturation for cirrus clouds, and it allows for a better representation of the tropical to extra-tropical ratio of the longwave cloud radiative effect. Further, the ice water path, the ice crystal number concentrations, and the supercooled liquid fractions in mixed-phase clouds agree better with observations in the new model than in the reference model. Ice crystal formation is dominated by the liquid-origin processes of convective detrainment and homogeneous freezing of cloud droplets. The simulated ice clouds strongly depend on model tuning choices, in particular, the enhancement of the aggregation rate of ice crystals.

Plain Language Summary This paper describes a new cloud scheme for the global climate model ECHAM-HAM that better represents the ice cloud formation processes. It calculates the formation of clouds by convection, turbulent vertical diffusion, and large-scale ascent. For each cloud formation process, the scheme calculates the cloud volume and the number concentration and size of cloud droplets and ice crystals. Further, it calculates how cloud droplets and ice crystals grow with time until they are large enough to form precipitation and are removed from the cloud. We show how the introduction of new formulations of the cloud processes affects the simulated clouds. The new ice cloud fraction compares better to satellite observations. In-cloud properties including ice crystal number concentrations, the fraction of supercooled liquid clouds, and the radiative effects of clouds are also compared to observations. We conclude that the new cloud scheme better captures the observed properties of ice clouds and improves our capability to simulate and understand ice clouds.

1. Introduction

Clouds have a major impact on the climate system and are a dominating source of uncertainty for future climate projections. Clouds often form in rising air that cools due to adiabatic expansion when exceeding saturation with respect to liquid water. Slow and large-scale upward motions occur along a warm front of a cyclone in the storm track regions. Fast and small-scale upward motions occur in convective events like deep convection in the tropics or shallow convection over the oceans. While large-scale motions are resolved by a global climate model (GCM), the detailed small-scale motions must be parameterized for the coarse model grid (with resolutions on the order of 100 km). Clouds also occur at different scales and usually are not homogeneously distributed over a GCM grid box. Reasons for this are small-scale motions, such as convection, boundary layer turbulence, orography, and gravity waves that transport moisture and can lead to subgrid scale cloud formation. Further, water vapor can be inhomogeneously distributed in a grid box,
which upon cooling will initiate cloud formation only in the parts of the grid box that are supersaturated with respect to liquid water/ice.

In state-of-the-art GCMs, subgrid-scale motions and clouds are simulated by a combination of parameterizations. For example, in ECHAM6.3 (Mauritsen et al., 2019; Stevens et al., 2013), subgrid-scale motions are parameterized by a turbulence and a convection scheme, while long-lived clouds are simulated by a stratiform cloud scheme that includes a formulation for cloud fraction and parameterizes the cloud microphysical processes. The turbulence scheme mixes air vertically mostly within the boundary layer. The convection scheme is responsible for the larger vertical subgrid moisture, tracer, and heat transport, which includes cloud and precipitation formation during convective events. However, the convective events are assumed to be subgrid scale and short-lived; that is, convective clouds form and decay within one model time step. The stratiform cloud scheme, on the other hand, simulates the formation, dissipation, and precipitation formation of long-lived, large-scale clouds. These clouds are not only important for precipitation formation but also for the radiation balance. The convection scheme interacts with the stratiform cloud scheme through condensate detrainment that can lead to a long-lived cloud (e.g., convective anvil) if it does not evaporate/sublimate. Finally, the cloud fraction formulation describes the fraction of the GCM grid box that is filled by stratiform clouds, which affects both the radiative transfer calculations as well as the in-cloud processes in the stratiform cloud microphysics scheme.

Stratiform cloud schemes include increasingly sophisticated ice microphysics formulations as the interest in mixed phase and ice clouds has grown in the scientific community (e.g., Barahona et al., 2014; Barahona & Nenes, 2009; Dietlicher et al., 2018; Kärcher et al., 2006; Kuebbeler et al., 2014; Lohmann, 2002; Morrison & Gettelman, 2008; Morrison & Milbrandt, 2015). In earlier GCMs, cloud parameterization schemes applied a saturation adjustment approach to calculate the cloud liquid water and ice mass mixing ratios (e.g., Kogan & Martin, 1994; Lohmann & Roeckner, 1996; McDonald, 1963; Sundqvist, 1978); however, cloud-free and in-cloud ice supersaturations have frequently been observed in high altitude ice clouds (e.g., Gao et al., 2004; Heymsfield et al., 1998; Jensen et al., 2005; Krämer et al., 2009; Ovarlez et al., 2002; Peter et al., 2006; Spichtinger et al., 2004). As ice crystal growth time scales at cold temperatures are much longer than cloud droplet growth time scales at warmer temperatures, ice crystals can exist in supersaturation and subsaturation for multiple GCM time steps. Therefore, applying a saturation adjustment approach is too simplified for ice clouds (Wilson & Ballard, 1999) and has been abandoned from microphysics schemes (e.g., Lohmann & Kärcher, 2002). Further, as ice nucleation requires tens of percent of supersaturation (Koop et al., 2000), and ice nucleating particles can influence the ice formation process (e.g., Kanji et al., 2017), detailed in situ cirrus nucleation schemes have been developed that simulate the competition between homogeneous and heterogeneous ice nucleation and the deposition of water vapor on preexisting ice crystals (e.g., Barahona & Nenes, 2009; Kärcher et al., 2006; Liu & Penner, 2005). Gasparini and Lohmann (2016) and Gasparini et al. (2018) identified convective detrainment and in situ nucleation of ice crystals as the main formation pathways of cirrus cloud ice crystals in the ECHAM-HAM GCM (Neubauer et al., 2019). Therefore, to investigate mixed phase and ice clouds, a cloud scheme should include a cloud fraction formulation that considers convective detrainment, ice supersaturation and subsaturation and subgrid-scale cloud formation. However, while there are cloud microphysics schemes that are able to integrate these processes (including the ECHAM-HAM microphysics scheme; Kuebbeler et al., 2014; Lohmann et al., 2007; Lohmann & Kärcher, 2002; Neubauer et al., 2019), the implementation of the ice microphysics is often inconsistent with the underlying assumption of the cloud fraction formulation, as most cloud fraction formulations were designed specifically for liquid clouds and implicitly assume saturation adjustment. In this study, we first discuss which cloud fraction concept is capable of consistently simulating ice clouds without the assumption of saturation adjustment and which includes convective detrainment, ice supersaturation and subsaturation and in situ cirrus cloud formation. We then describe our formulation of such a cloud scheme. Finally, we present simulations with the new cloud scheme with the ECHAM-HAM GCM.

2. Concepts for Cloud Schemes

The simplest formulation of a cloud scheme assumes a homogeneous distribution of the total water content (water vapor plus liquid and frozen water) in a grid box, with either 0% or 100% cloud fraction. This “binary” scheme does not allow for a conservation of in-cloud properties like the cloud droplet number concentration...
(CDNC) or ice crystal number concentration (ICNC) during advection. If, for example, half of a cloud with a CDNC of 100 cm$^{-3}$ is advected into a previously cloud-free adjacent grid box, the new CDNC in both grid boxes will be 50 cm$^{-3}$ after advection, because the properties of the cloudy and the cloud-free air masses are mixed. This simple example of numerical diffusion at cloud edges shows the limits of numerical grid-based models. Processes are only properly simulated if they occur at larger scales than the model resolution. This numerical diffusion at cloud edges may be negligible in cloud resolving models where one cloud spans over many grid boxes. However, in GCMs most clouds are subgrid scale with cloud edges in many grid boxes. Therefore, more information must be added to the cloud scheme to more realistically simulate in-cloud processes. Information about the subgrid distribution of cloud liquid water, cloud ice, and water vapor in the grid box can be added by either adding prognostic variables that trace specific details of the distribution or by assuming a distribution that is based on theory, measurements, or high-resolution models. Both the assumption about the distribution and the number of prognostic variables define the flexibility of the scheme and determine which processes can be simulated. We will now discuss the basic advantages and disadvantages of well-known cloud scheme designs, with a focus on the ability to simulate mixed phase and cirrus clouds. More discussion of the general advantages and disadvantages of the different approaches can be found in Wilson et al. (2008) and Tompkins (2002).

The family of statistical cloud schemes assumes that the total water distribution follows a certain shape (e.g., triangular in Smith, 1990; Gamma distributions in Tompkins, 2002; double Gaussian in Golaz et al., 2002), which can be described by a probability density function (PDF). Some schemes trace properties of the PDF, such as width and skewness, as prognostic variables so that the distribution can be altered by atmospheric processes (e.g., Tompkins, 2002). These statistical schemes have the advantage of self-consistency as the cloud fraction and the condensed water can always be diagnosed from the supersaturated part of the PDF. In addition, if the PDF has sufficient degrees of freedom (e.g., Golaz et al., 2002; Tompkins, 2002), it can be adjusted to convective detrainment events. However, for cirrus clouds, the statistical approach is challenging: The existence of ice supersaturation does not necessitate the existence of an ice cloud, and therefore, cloud fraction cannot be diagnosed from the PDF. Thus, for ice clouds, it is not clear how to divide the PDF into cloudy and cloud-free air. Depending on the presence of ice nucleating particles (INPs), in situ cirrus clouds can start to form at a few percent supersaturation, with the exact supersaturation depending on the INP type and temperature (Kanji et al., 2017). In the absence of suitable INPs, homogeneous freezing of solution droplets occurs at higher supersaturations of 40% to 60%, depending on the temperature (Koop et al., 2000). Further, the cloud ice mass cannot be diagnosed from the PDF as saturation adjustment cannot be assumed for cirrus clouds (e.g., Lohmann & Kärcher, 2002; Wilson & Ballard, 1999), leaving the ice mass dependent on the amount of water vapor deposited on the ice crystals in previous time steps. Therefore, current statistical cloud schemes are implemented for liquid clouds only (Bogenschutz et al., 2012; Tompkins, 2002), and the implementation of ice clouds in statistical schemes seems difficult.

Relative humidity (RH)-based cloud schemes diagnose cloud fraction from the grid-mean RH when it exceeds a tunable critical value ($RH_{crit}$). The cloud fraction reaches one at grid-mean saturation (Sundqvist et al., 1989). RH-based schemes are simplifications of statistical schemes that can be inferred from a PDF with a constant width and skewness (Smith, 1990). Therefore, RH-based cloud schemes face the same problems in simulating cirrus clouds with different RH thresholds based on the formation process as statistical schemes (Dietlicher et al., 2019; Gettelman et al., 2010). Further, for ice clouds, an RH-based cloud scheme has to switch from RH with respect to liquid water as long as cloud droplets are present to RH with respect to ice after the cloud has been glaciated, the implementation of which is ambiguous (see, e.g., the discussion in Dietlicher et al., 2019). This can lead to an unphysical increase of cloud fraction during glaciation of a subgrid cloud.

This discussion of concepts of cloud schemes shows that a proper simulation of cirrus and mixed-phase clouds with partial cloudiness is difficult to realize with a statistical or RH-based approach. However, it is possible with a Tiedtke (1993)-like prognostic cloud scheme that traces all in-cloud variables, including cloudy air, in-cloud water vapor, and cloud water and ice as prognostic variables, and allows for an independent evolution of the in-cloud and cloud-free properties as suggested by Kärcher and Burkhardt (2008). In such a scheme, cloud ice is a prognostic variable, which allows ice crystals to exist in a subsaturated or supersaturated environment. Further, as cloudy air is simulated as a prognostic variable (from which cloud fraction is directly diagnosed), the in-cloud water vapor can be reduced, for example, by depositional growth on...
ice crystals, and large ice crystals can sediment without changing the cloud fraction. In addition, a convective detrainment event can be implemented as a source term for cloudy air, condensate, and in-cloud water vapor independent of the relative humidity in the grid box (e.g., Wilson et al., 2008). Further, a prognostic scheme allows for a straightforward calculation of in-cloud processes as all variables are directly available and do not have to be derived from a PDF. Disadvantages of prognostic schemes are that inconsistent model states (cloud fraction with no condensate) are in theory possible as a result of numerical inaccuracy. In addition, there is no information available about the distribution of water vapor and condensate within the in-cloud and cloud-free parts of the grid box, which requires further assumptions as discussed in Tiedtke (1993) or Wilson et al. (2008).

Tiedtke (1993) first implemented a prognostic one-moment (mass mixing ratio) cloud scheme that predicted the cloudy air mass, with no separation between in-cloud and cloud-free water vapor. Cloud water and ice were traced together and distinguished based on temperature. Kärcher and Burkhardt (2008) developed a concept for a cirrus scheme with separate prognostic variables for in-cloud and cloud-free water vapor and ICNC based on homogeneous nucleation of solution droplets but did not include heterogeneous nucleation on INPs. This cirrus scheme has been implemented in a GCM and extended to heterogeneous nucleation by Wang and Penner (2010) and Wang et al. (2014); however, they did not advect cloud fraction and the in-cloud water vapor. Wilson et al. (2008) developed a prognostic scheme for liquid and ice clouds. However, their scheme was only designed for mass mixing ratios and does not include a separation between in-cloud and cloud-free water vapor, or specific formulations for cirrus clouds. Yukimoto et al. (2012) included prognostic variables for CDNC and ICNC in the Tiedtke (1993) scheme; however, it is not clear if and how the model simulates clear-sky supersaturation and the associated cirrus cloud formation.

In this paper, we describe a prognostic Tiedtke (1993)-like cloud scheme with prognostic cloud fraction, in-cloud water vapor, and cloud droplet/ice crystal number concentrations, with two-moment (mass mixing ratios and number concentrations) cloud microphysics. The scheme is designed for liquid and ice clouds, allows for supersaturation, and includes a cirrus nucleation scheme that simulates ice nucleation at supersaturation via homogeneous and heterogeneous nucleation. Finally, it is implemented and validated in the ECHAM-HAM GCM.

3. Model Description

The host model, ECHAM6.3-HAM2.3 (Neubauer et al., 2019; Tegen et al., 2019), is the general circulation model ECHAM6.3 (Crueger et al., 2018; Mauritsen et al., 2019; Stevens et al., 2013) that is coupled to the aerosol model HAM2 (Tegen et al., 2019; Zhang et al., 2012). ECHAM solves prognostic equations for temperature, pressure, wind, and water and tracer advection and parameterizes radiative transfer, turbulence, convection, and the interaction with the ocean surface and land surface (Stevens et al., 2013). HAM simulates emission, nucleation, transport, and removal of solution droplets, dust, and other aerosol particles (Stier et al., 2005; Zhang et al., 2012). ECHAM-HAM also includes a two-moment stratiform cloud microphysics scheme (Lohmann et al., 2007) that uses the RH-based diagnostic cloud cover formulation by Sundqvist et al. (1989). This cloud scheme is the predecessor of our new scheme from which we adopt many concepts. Therefore, we compare our new cloud scheme with ECHAM6.3-HAM2.3, which we refer to as ECHAM-HAM in the following.

3.1. Prognostic Variables and Equations

The new cloud scheme has a Tiedtke (1993)-like structure with prognostic variables for cloudy air $q_{ac}$, grid-mean water vapor $q_v$, in-cloud water vapor $q_{vc}$, cloud liquid water $q_l$, cloud ice $q_i$, cloud droplet number concentration $N_c$, and ice crystal number concentration $N_i$. The cloud fraction $CF$ is the volume of cloudy air per total air volume in the grid box, which is equivalent the cloudy air mass mixing ratio $q_{ac}$. Following Tiedtke (1993), the variables $q_{vc}$, $q_l$, $q_i$, $N_c$, and $N_i$ are defined as in-cloud but are implemented as grid-mean variables to assure mass and number conservation during advection. We refer to the grid-mean values as $q_{vc}$, $q_l$, $q_i$, $N_c$, and $N_i$ and calculate the in-cloud values by dividing by the cloud fraction, for example, $q_{ac} = \frac{q_{vc}}{CF}$. In this paper we describe all changes of water vapor in terms of changes of the in-cloud water vapor $q_{ac}$ and the cloud-free water vapor $q_{cf}$ instead of $q_{ac}$, as this allows for a clearer description of the processes. $q_{cf}$ is always clearly defined by $q_{cf} = \frac{q_{ac}}{(1 - CF)}$ as long as $CF < 1$ and with $q_{ac} = q_v - q_{cf}$. 

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To formulate the prognostic equations, we define air as cloudy when it contains a minimum in-cloud water/ice mass mixing ratio of $10^{-18}$ kg/kg and CDNC/ICNC of 1 m$^{-3}$, which corresponds to approximately 1 cloud droplet/ice crystal with a 1 $\mu$m radius in 1 m$^3$. Further, we group the processes in the cloud scheme by cloud formation processes in the cloud-free, in-cloud processes, precipitation related processes, ice crystal sedimentation, and cloud dissipation. Microphysical in-cloud processes either change the phase of water or the number concentrations of hydrometeors but do not change the total water in the grid box. Precipitation reduces the cloud liquid water or cloud ice and can increase the water vapor in lower levels or the number concentrations of hydrometeors but do not change the total water in the grid box. Unlike precipitation, it acts as a source of cloud ice in the levels below. The details of the in-cloud processes, precipitation related processes, and ice crystal sedimentation are discussed in sections 3.5 and 3.6.

For simplicity in the new cloud scheme, we follow Tiedtke (1993) by assuming a homogeneous distribution of water vapor, cloud droplets, and ice crystals within the cloudy part of the grid box. Thus, cloud fraction is not influenced by any of the in-cloud or precipitation processes except if the entire cloud dissipates (Tiedtke, 1993). Even though dissipation is a result of an in-cloud or precipitation process, we state it as an individual process to emphasize that in-cloud processes usually do not change the cloud fraction. This leads to the prognostic equation of cloud fraction $CF$:

$$\frac{\partial CF}{\partial t} + \mathbf{u} \cdot \nabla CF = \left[ \frac{\partial CF}{\partial t} \right]_{\text{Form}} + \left[ \frac{\partial CF}{\partial t} \right]_{\text{Diss}} + \left[ \frac{\partial CF}{\partial t} \right]_{\text{Sedi}}$$

(1)

with the term on the left-hand side representing advection in the 3-D wind field $\mathbf{u}$ and turbulent vertical diffusion and on the right-hand side the terms for the cloud formation processes (Form), cloud dissipation (Diss), and ice crystal sedimentation (Sedi) into lower levels. The equation for the grid-mean cloud-free water vapor is

$$\frac{\partial q_{vf}}{\partial t} + \mathbf{u} \cdot \nabla q_{vf} = \left[ \frac{\partial q_{vf}}{\partial t} \right]_{\text{Form}} + \left[ \frac{\partial q_{vf}}{\partial t} \right]_{\text{Diss}} + \left[ \frac{\partial q_{vf}}{\partial t} \right]_{\text{Sedi}}$$

(2)

with a decrease of the cloud-free water vapor when the cloud volume increases by cloud formation and sedimentation of ice crystals into a level, when cloud-free air turns into cloudy air. Vice versa, the cloud-free water vapor increases when the cloud volume decreases during cloud dissipation and sedimentation out of a level. In addition, the water vapor can increase by evaporation/sublimation of raindrops/snow. The prognostic equation of the grid-mean in-cloud water vapor is similar:

$$\frac{\partial q_{ic}}{\partial t} + \mathbf{u} \cdot \nabla q_{ic} = \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Form}} + \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Incl}} + \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Precip}} + \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Sedi}}$$

(3)

but the terms have the opposite signs for cloud formation, dissipation, and sedimentation, and the in-cloud water vapor can be changed by in-cloud condensation/deposition and evaporation/sublimation (Incl).

Accordingly, the prognostic equations for the cloud water mass mixing ratio and CDNC are given as

$$\frac{\partial q_{c}}{\partial t} + \mathbf{u} \cdot \nabla q_{c} = \left[ \frac{\partial q_{c}}{\partial t} \right]_{\text{Form}} + \left[ \frac{\partial q_{c}}{\partial t} \right]_{\text{Incl}} + \left[ \frac{\partial q_{c}}{\partial t} \right]_{\text{Precip}}$$

(4)

$$\frac{\partial N_{c}}{\partial t} + \mathbf{u} \cdot \nabla N_{c} = \left[ \frac{\partial N_{c}}{\partial t} \right]_{\text{Form}} + \left[ \frac{\partial N_{c}}{\partial t} \right]_{\text{Incl}} + \left[ \frac{\partial N_{c}}{\partial t} \right]_{\text{Precip}}$$

(5)

and for the cloud ice mass mixing ratio and ICNC:

$$\frac{\partial q_{ic}}{\partial t} + \mathbf{u} \cdot \nabla q_{ic} = \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Form}} + \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Incl}} + \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Precip}} + \left[ \frac{\partial q_{ic}}{\partial t} \right]_{\text{Sedi}}$$

(6)

$$\frac{\partial N_{ic}}{\partial t} + \mathbf{u} \cdot \nabla N_{ic} = \left[ \frac{\partial N_{ic}}{\partial t} \right]_{\text{Form}} + \left[ \frac{\partial N_{ic}}{\partial t} \right]_{\text{Incl}} + \left[ \frac{\partial N_{ic}}{\partial t} \right]_{\text{Precip}} + \left[ \frac{\partial N_{ic}}{\partial t} \right]_{\text{Sedi}}$$

(7)
3.2. Advection, Vertical Mixing, Surface Processes, and Convective Transport of the Prognostic Variables

As the prognostic variables for $q_v$, $q_c$, $q_i$, $N_c$, and $N_i$ are already implemented in ECHAM-HAM, we only have implemented variables for and $q_{ac}$ and and $q_{vc}$ for the new cloud scheme. As for and $N_c$, and $N_i$, the prognostic variables for $q_{ac}$ and $q_{vc}$ are implemented as tracers, which allows for consistent advection of all cloud variables by the advection scheme. In the physics routine, before the cloud scheme is called, tendencies are calculated for vertical mixing by the turbulence scheme, for surface interactions by the surface layer parameterization, and for shallow, deep, and mid-level convection by the convection scheme.

Vertical mixing is calculated for tracers and the new prognostic cloud variables in the same way as for water vapor, cloud water, and ice. However, evaporation of water from the ocean or the land surface is only calculated for the grid-mean water vapor. The parameterization of the surface fluxes effectively leads to transporting a fraction $f_s$ of air with the properties of the surface layer (e.g., water vapor at saturation over the ocean) to the lowest model level. $f_s$ is calculated by the turbulent vertical diffusion scheme and depends on, for example, the turbulent kinetic energy. We use the same approach and transport air with the properties of the surface layer to the fraction $f_s$ of the cloudy air in the lowest model level. While the exchange of air of the surface layer with the lowest model level is assumed to be homogeneously distributed, different water vapor amounts in the cloudy and cloud-free air lead to an inhomogeneous distribution of the water vapor surface fluxes. Thereby, evaporation of water over the ocean will increase the cloud-free water vapor, but condensation of in-cloud water vapor in the case of fog is also possible.

The convection scheme (Nordeng, 1994; Tiedtke, 1989) parameterizes convective events with a mass-flux approach and calculates tendencies for temperature, moisture, momentum, and tracers. It is used, for example, for aerosol transport and wet scavenging in ECHAM-HAM. In the implementation in ECHAM-HAM no cloud variables are transported by the convection scheme, and only the grid-mean water vapor is changed. Consistently, we assume that the convection events occur in the cloud-free part of the grid box and accordingly the in-cloud water vapor should not change. The implementation of all convective processes in the cloud-free portions of the grid box is pragmatic for our study; however, ideally convective processes should include the transport of all cloud variables, including cloudy air, cloud water, and in-cloud water vapor. This ideal implementation will be addressed in future model development.

3.3. Cloud Formation

In ECHAM, the cloud scheme is called after the advection, vertical mixing, and convection schemes. It calculates the tendencies of the cloud variables due to cloud processes. Cloud formation can occur in the cloud-free part of the grid box and turns cloud-free air into cloudy air by activation of cloud droplets or nucleation of ice crystals. Cloud formation could be described by a single process that calculates cloud formation for the supersaturated part of an assumed water vapor distribution. However, at temperatures below 235 K, ice crystals can form below water saturation (see discussion in section 2). Therefore, a second cloud formation process for ice cloud formation needs to be included. Further, as a fully prognostic cloud scheme has the flexibility to simulate two separate air masses (cloudy and cloud-free), convective detrainment can be directly added as an additional cloud formation process by using the information about the detrained air and condensate from the convection scheme (e.g., Tiedtke, 1993; Wilson et al., 2008). Further, the number concentrations of activated cloud droplets and nucleated ice crystals strongly depend on the updraft velocity. Therefore, in addition to convective detrainment we add cloud formation due to vertical transport by subgrid scale turbulent motion as an additional cloud formation process to separate fast turbulent updrafts from slow and large-scale updrafts. Similar to Tiedtke (1993), this leads to three cloud formation types (Figure 1):

1. Convective detrainment of cloud droplets and ice crystals
2. Liquid and ice cloud formation due to turbulent vertical diffusion
3. Large-scale liquid and ice cloud formation

With the equation,
In the case of a cloud formation event, for example, large-scale liquid cloud formation (LSL) during the time step $\Delta t$, a fraction of the grid box transfers from cloud-free to cloudy: $\Delta CF_{\text{LSL}} = \frac{\partial CF}{\partial t}|_{\text{LargeScaleLiquid}} \cdot \Delta t$, with the water vapor $q_{\text{vc,LSL}}$, the cloud droplet number concentration $N_{c,\text{LSL}}$, and liquid water mass $q_{c,\text{LSL}}$ in the newly formed cloud. These properties are added to a preexisting cloud by weighting with the air volume. As previously cloud-free air turns cloudy, the grid-mean values of the in-cloud and cloud-free water vapor change according to $\Delta q_{\text{vf}} = -\Delta CF_{\text{LSL}} \cdot q_{\text{vc,LSL}}$ and $\Delta q_{\text{vc}} = \Delta CF_{\text{LSL}} \cdot \left(q_{\text{vc,LSL}} - q_{c,\text{LSL}}\right)$. Further, the newly formed cloud droplets are added to the grid-mean CDNC and mass: $\Delta N_{c} = \Delta CF_{\text{LSL}} \cdot N_{c,\text{LSL}}$ and $\Delta q_{c} = \Delta CF_{\text{LSL}} \cdot q_{c,\text{LSL}}$, respectively. After all grid-mean properties and the cloud fraction have been updated, the updated in-cloud (cloud-free) properties can be calculated by dividing by the new cloud fraction (cloud-free fraction). Finally, the tendencies for the prognostic variables $\frac{\partial q_{\text{vc}}}{\partial t}$, $\frac{\partial q_{\text{vf}}}{\partial t}$, $\frac{\partial q_{c}}{\partial t}$, and $\frac{\partial N_{c}}{\partial t}$ can be calculated by dividing by the time step length. The case of ice cloud formation is handled accordingly.

In the following, we describe how CF, the in-cloud water vapor, CDNC, ICNC, and the mass mixing ratios are calculated for each cloud formation process.

### 3.3.1. Convective Detrainment of Cloud Droplets and Ice Crystals

The convection scheme is based on the mass flux scheme of Tiedtke (1989), with updates by Nordeng (1994), and parameterizes temperature and moisture tendencies for moist convective events (Möbius & Stevens, 2012; Nordeng, 1994; Stevens et al., 2013; Tiedtke, 1989). The scheme calculates an air parcel ascent with condensation of water vapor, precipitation formation, entrainment of environmental air, and detrainment of cloudy air from the convective plume into the environment. The latter represents the link to the cloud scheme as the detrained cloudy air can form a long-lived cloud (e.g., anvil cirrus). In ECHAM-HAM, consistent with the RH-based cloud fraction scheme, the detrained condensate does not evaporate/sublimate if the grid-mean water vapor is above the critical RH for cloud fraction before the convective event. On the contrary, in Tiedtke (1993) and Wilson et al. (2008), the detrained air is always added to the cloud variables, and the condensate is never evaporated/sublimated immediately, but processes for mixing of cloudy and cloud-free air exist that can lead to evaporation/sublimation of the condensate with time.

In our scheme, we want to allow for detrainment of cloudy air but also consider evaporation/sublimation in case of detrainment into dry environmental air. Therefore, we assume a mixing that is faster than one time step of the detrained air with the environmental cloud-free air. Unfortunately, the mixing ratio of the detrained air with the environmental air is unknown, and we have to introduce the tuning parameter

$$
\frac{\partial CF}{\partial t}_{\text{Form}} = \frac{\partial CF}{\partial t}_{\text{ConvDetr}} + \frac{\partial CF}{\partial t}_{\text{Turb}} + \frac{\partial CF}{\partial t}_{\text{LargeScaleLiquid}} + \frac{\partial CF}{\partial t}_{\text{LargeScaleIce}}
$$

In the following, we describe how CF, the in-cloud water vapor, CDNC, ICNC, and the mass mixing ratios are calculated for each cloud formation process.
\( \gamma_{\text{detMix}} \). If \( \gamma_{\text{detMix}} \) is equal to unity, it corresponds to mixing of detrained and environmental air in equal parts, as assumed by Tiedtke (1989) for the calculation of the convective downdraft. If not enough cloud-free air is available due to a too high cloud fraction in the grid box, the mixing ratio is reduced accordingly. Knowing the temperature and the water vapor of the detrained air, we calculate the saturation ratio after mixing. In the case of detrainment as cloud droplets \( (T > 235 \text{ K}, \text{ see below}) \) and a subsaturation with respect to liquid water that suffices to evaporate all detrained condensate, no cloud formation occurs. Otherwise, the detrained cloud fraction is calculated based on the detrained air mass flux \( \dot{M}_\text{conv,detr} \) from the convection scheme as its fraction of all air in the grid box with the level height \( \Delta z = \Delta p/(g \rho_{\text{air}}) \) and consider that mixing with environmental air leads to an increase in cloud fraction:

\[
\frac{\partial CF}{\partial t}_{\text{ConvDetr}} = \frac{M_{\text{conv,detr}}}{\Delta p/g} \cdot (1 + \gamma_{\text{detMix}})
\]

Note that below the level of zero buoyancy the detrained air is warmer and moister than the environmental air. If the mixing with environmental air did not reduce the average in-cloud water vapor to or below saturation, adding all detrained water vapor as in-cloud water vapor can lead to additional condensation.

### 3.3.1.1. Number Concentrations of Convective Detrainment

As the convection scheme is only a one-moment scheme (mass), the number concentrations of the detrained condensate must be calculated. Parameterizations of the detrained number concentrations strongly simplify the processes in the convective plume. In the convection scheme all condensate is assumed to be frozen at \( T < 273.15 \text{ K} \) (Tiedtke, 1989), and in ECHAM the detrained condensate is added to the cloud variables accordingly (Lohmann & Roeckner, 1996). This assumption is unrealistic as homogeneous nucleation of cloud droplets only occurs below 235 K (Koop et al., 2000), while heterogeneous nucleation occurs at temperatures below 273.15 K in case of suitable INPs (Kanji et al., 2017). Therefore, in ECHAM-HAM the detrained condensate is assumed to be liquid if \( T > 235 \text{ K} \) in the absence of sufficient ice crystals. If there are sufficient ice crystals present at mixed-phase temperatures \( (235 \text{ K} < T < 273 \text{ K}) \), a Wegener-Bergeron-Findeisen process (Korolev & Isaac, 2003) occurs, and the detrained condensate is assumed to be frozen. In ECHAM-HAM, the detrained ICNC is calculated from the detrained ice mass based on a temperature-dependent ice crystal size climatology, and in Morrison and Gettelman (2008) ICNC is calculated from an assumed temperature-independent ice crystal size. We argue that calculating the detrained ICNC based on a constant or temperature-dependent size does not represent the physical process and that the assumed size represents a (hidden) tuning parameter. Instead, we suggest that when a convective event detrains at \( T < 235 \text{ K} \), the detrained ICNC should be calculated based on the updraft velocity when the ascending air parcel cools to \( T < 235 \text{ K} \), where cloud droplets freeze homogeneously. Further, all condensate should be detrained as liquid droplets at \( T > 235 \text{ K} \) as they can freeze at a later time step in the cloud scheme via heterogeneous freezing, allowing for a more accurate simulation of the Wegener-Bergeron-Findeisen process.

In the new cloud scheme, we calculate the CDNC at the convective cloud base similar to Lohmann et al. (2007), with a cloud condensation nuclei (CCN) activation scheme that depends on the updraft velocity (section 3.4.1). In ECHAM-HAM a turbulent enhanced updraft velocity was used for CCN activation at the convective cloud base. In the new cloud scheme, the convective updraft \( w_{\text{conv}} \) is calculated from the convective available potential energy (CAPE) with \( w_{\text{conv}} = 0.5 \cdot \sqrt{\text{CAPE}} \); the factor 0.5 allows for better agreement with observations (Lohmann et al., 2016). We assume that the CDNC remains constant during the ascent from the convective cloud base to the convective cloud top or the level of \( T = 235 \text{ K} \). Therefore, we neglect a reduction of the CDNC by precipitation and entrainment of environmental air, but we also neglect an increase of the CDNC by activation of additional cloud droplets. At the \( T = 235 \text{ K} \) level, we calculate the ICNC with our scheme for homogeneous freezing of cloud droplets, which parameterizes the ICNC based on the updraft velocity \( w_{\text{conv}} \) (discussion and equation in section 3.4.2). Above that level, we assume a constant ICNC up to the convective cloud top. Thus, the detrained CDNC depends on the CCN concentration and the updraft velocity at the convective cloud base, and the detrained ICNC depends on the CDNC and the updraft velocity at the \( T = 235 \text{ K} \) level. In the case of a convective cloud base at \( T < 235 \text{ K} \), the cirrus nucleation scheme (section 3.4.3) is used to calculate the ICNC at the cloud base.
3.3.1.2. Accounting for Immediate Aggregation of the Convectively Detrained Ice Crystals

In ECHAM-HAM calculating the detrained ICNC based on a temperature-dependent size, or based on the updraft velocity as suggested above, leads to a too high ICNC compared to observations (see section 4.2 and 4.5) which leads to an overestimation of the cloud radiative effect. Reasons might be that important processes in the convective plume like collision-coalescence or aggregation or during detrainment that strongly reduce the ICNC and increase their size to >100 µm as observed in convective anvils (e.g., Jensen et al., 2009, 2018) are missing. The ice crystals may also be distributed into too large areas in the GCM grid boxes when the detrained ice crystals are added to a preexisting large-scale cirrus cloud. In ECHAM-HAM the problem of too many ice crystals is addressed by a very high enhancement factor of the aggregation rate of ice crystals to snowflakes of 900 (Neubauer et al., 2019). However, this approach does not only affect the convectively detrained ice crystals but also all other ice crystals that form via in situ cirrus cloud formation, homogeneous freezing of cloud droplets during large-scale ascents, and heterogeneous freezing of cloud droplets at mixed-phase temperatures. We therefore propose to account for immediate aggregation by reducing the detrained ICNC by dividing it by an updraft-dependent factor that is 1 at 0 m/s and increases linearly with the slope as tuning parameter \( f_{\text{detIC}} \). We suggest an updraft dependency as we expect aggregation to be enhanced by turbulent motions in the convective plume, in line with observations and other studies (e.g., Benmoshe & Khain, 2014; Franklin et al., 2007; Pinsky & Khain, 1998).

3.3.2. Cloud Formation Due to Turbulent Vertical Diffusion

The vertical diffusion scheme vertically mixes air mostly in the boundary layer based on the turbulent kinetic energy (TKE) by upward and downward transport. Thereby, it simulates parts of the convective motions which can lead to cloud formation. However, as the turbulence scheme in ECHAM6 (Brinkop & Roeckner, 1995; Stevens et al., 2013) does not consider cloud formation, we implement our own calculations for which we use the information from the turbulence scheme, with no changes to the scheme itself.

In the turbulence scheme, upward mixing is simulated from the lowest to the highest level by replacing a certain fraction of air (depending on TKE) by air from the level below. Before continuing the upward mixing to the next level, the air is mixed in the grid box. Therefore, there is no ascending plume as in the convection scheme, and we can only simulate cloud formation for rising air to the adjacent level. From the turbulence scheme, we know the fraction of air in a grid box that was transported upwards \( F_{\text{turb, fromBelow}} \) the temperature, cloud fraction \( CF_{\text{below}} \) and cloud-free water vapor of the upward moving air. We calculate if the cloud-free air is supersaturated after a dry-adiabatic air parcel ascent from the level below. If so, a cloud is forming from the supersaturated water vapor accounting for latent heat release, increasing the cloud fraction

\[
\frac{\partial CF}{\partial t}\bigg|_{\text{Turb}} = \frac{F_{\text{turb, fromBelow}} \cdot (1 - CF_{\text{below}})}{\Delta t}
\]

which is assumed to be saturated. Following Lohmann et al. (2007), we calculate the turbulent updraft velocity from the turbulent kinetic energy (TKE) as \( w_{\text{turb}} = 0.7 \cdot \sqrt{\text{TKE}} \). CDNC is calculated using the CCN activation scheme if \( T > 235 \text{ K} \) (section 3.4.1) and the ICNC using the cirrus nucleation scheme if \( T < 235 \text{ K} \) (section 3.4.3).

3.3.3. Large-Scale Liquid Cloud Formation

Relative humidity increase and cloud formation can occur independent of convective and turbulent motions due to large-scale ascent of air masses, for example, along a warm conveyor belt in the storm tracks, and radiative cooling of air. As this cloud formation does not depend on parameterized subgrid motions, there is no information available to calculate cloud fraction, and an assumption of the water vapor distribution in the cloud-free area is required to calculate the newly forming cloud. Tiedtke (1993) assumed a top-hat distribution, with a critical RH that has to be exceeded for cloud formation to occur in a cloud-free grid box. High-resolution cloud-resolving simulations of Tompkins (2002) show a Gaussian-like distribution of water vapor that is often positively or negatively skewed due to convective events. As we implemented convective detrainment as separate cloud formation process in this study, we assume a Gaussian distribution of water vapor for large-scale cloud formation. Further, a homogeneous temperature distribution is assumed which leads to a Gaussian distribution of saturation ratio over water \( G(S) \) in the cloud-free part of the grid box that is assumed to be centered around the cloud-free saturation ratio \( S_f \). For an easier handling of the unbounded...
Gaussian distribution, $G(S)$ is only integrated in the $\pm 3\sigma$ interval. In the case of a completely cloud-free grid box, we assume $G(S)$ to have a standard deviation $\sigma_{CF} = \frac{1}{3}(1 - RH_{crit})$, with the tuning parameter $RH_{crit}$. $RH_{crit}$ is the minimum RH that needs to be exceeded for large-scale cloud formation to occur and thereby $RH_{crit}$ keeps a similar meaning as in the RH-based cloud fraction formulation and as in Tiedtke (1993).

The change in cloud fraction is calculated from the supersaturated area:

$$\frac{\partial CF}{\partial t} = \left. \frac{1 - CF}{\Delta t} \right|_{LargeScaleLiquid} \int_{S_f}^{1 + 3\sigma} G(S)dS, \tag{11}$$

and the mass of the newly formed cloud droplets is the supersaturated water vapor.

Figures 2a and 2b schematically show the calculation of the cloud fraction from the moisture distribution in the case of a cloud-free grid box. Figures 2c, 2d, and 2e show the calculation of additional cloud formation in the case of a partly cloudy grid box. Here we assume that if the cloud originated from large-scale liquid cloud formation in the previous time step, and the cloud-free saturation ratio and cloud fraction have not changed, there will be no additional cloud formation. This is achieved by a stepwise approach, first by assuming a PDF with $\sigma = \frac{1}{3}(1 - RH_{crit})$ and deriving its mean saturation ratio that corresponds to the cloud fraction in the grid box. Next, the corresponding cloud-free mean saturation ratio $S_f$ can be inferred from the subsaturated part of the PDF. Finally, the standard deviation of the PDF in the cloud-free area is chosen according to $\sigma_{CF > 0} = \frac{1}{3}(1 - S_f)$, so that additional cloud formation according to Equation 11 only occurs when the cloud-free saturation ratio $S_f$ exceeds $S_f$. With this implementation, the width of the moisture distribution will be largest in a cloud-free grid box and decrease towards zero in a completely cloudy grid box. Note, however, that the distribution in the cloud-free area is different in Figures 2b and 2d. This is a consequence of the Gaussian distribution and can only be avoided by assuming a top-hat distribution for $G(S)$. However, we decide to keep the Gaussian distribution as it is in better agreement with the high-resolution cloud-resolving modeling of Tompkins (2002).

To calculate the CDNC in the newly formed large-scale cloud, we use the grid-mean large-scale updraft velocity and assume a minimum updraft velocity of 1 cm/s. The CDNC is calculated using the CCN activation scheme (section 3.4.1).
3.3.4. Large-Scale Ice Cloud formation (Cirrus Clouds)

For large-scale ice cloud formation, we make the same assumptions as for large-scale liquid cloud formation; that is, we assume a Gaussian distribution of water vapor centered around the cloud-free saturation ratio, with the same standard deviation as for liquid cloud formation, but with respect to ice (Figure 3a). Contrary to liquid cloud formation, there are multiple ice crystal nucleation mechanisms, with heterogeneous ice nucleation on INPs requiring lower supersaturations than homogeneous freezing of solution droplets. Therefore, we split the saturation ratio PDF into three parts: no cloud formation below saturation with respect to ice, only heterogeneous ice nucleation between saturation and the critical relative humidity for homogeneous freezing of solution droplets $RH_{crit, hom}$ (after Koop et al., 2000), and homogeneous ice nucleation for all saturation ratios above $RH_{crit, hom}$ (Figure 3b). We calculate the cloud fractions for the heterogeneously and the homogeneously nucleated ice crystals from the saturation ratio PDF and calculate ice crystal nucleation for both parts individually using the cirrus nucleation scheme (section 3.4.3) at the median saturation ratios and the large-scale updraft velocity. Finally, the number concentrations and mass mixing ratios of homogeneously and heterogeneously nucleated ice crystals are weighted with their respective cloud fractions and combined to a single ICNC and cloud fraction.

Another key difference between large-scale liquid and ice cloud formation is that not all supersaturated vapor is assumed to deposit on the newly nucleated ice crystals immediately. If the ice crystals are too small or not numerous enough, the new cloudy air can still be supersaturated, and this supersaturation has to be deposited on the ice crystals in the subsequent time steps.

In previous cirrus cloud studies (Barahona et al., 2014; Kuebbeler et al., 2014; Shi et al., 2015), the deposition of water vapor on preexisting ice crystals was identified as an important process that reduces ICNC (by an order of magnitude in Shi et al., 2015), as it depletes the supersaturation during an air parcel ascent. Therefore, it competes with and reduces ice crystal nucleation, leading to a higher relative importance of heterogeneous nucleation (Shi et al., 2015). However, we argue that the effect of preexisting ice crystals is overestimated in cloud schemes that do not clearly separate in-cloud and cloud-free processes or overestimate the occurrence of fully cloudy grid boxes. In these implementations preexisting ice crystals can reduce ice nucleation everywhere in the grid box, while they should only be able to do so in the cloudy part of the grid box. Thus, (homogeneous) ice nucleation is suppressed too often in too large an area, and the formation pathway is shifted too often towards heterogeneous nucleation. In our implementation of ice nucleation in the cloud-free part of the grid box, there are no preexisting ice crystals.

3.4. Cloud Droplet Activation and Ice Crystal Nucleation Schemes

3.4.1. Cloud Droplet formation Via CCN Activation

There are two cloud droplet activation schemes implemented in ECHAM-HAM. The standard CCN activation scheme in ECHAM-HAM is the Abdul-Razzak and Ghan (2000) scheme (Neubauer et al., 2019). It calculates the activated CCN based on temperature, updraft velocity, and aerosol concentration by parameterizing air parcel ascent and considering that larger aerosol particles activate first and consume the available supersaturated water vapor (Abdul-Razzak & Ghan, 2000). All soluble/internally mixed aerosols of HAM except those in the nucleation mode are used as potential CCN (Lohmann & Neubauer, 2018).
Alternatively, the empirical Lin and Leaitch (1997) scheme parameterizes CCN activation based on the updraft velocity and the aerosol concentration of all soluble/externally mixed aerosol with a radius larger than 30 nm. CCN activation in convective clouds uses a smaller aerosol activation threshold of 20 nm.

The updraft velocity $w$ is dependent on the CCN activation regime:

$$w = \begin{cases} 
  w_{\text{conv}} = 0.5 \cdot \sqrt{CAPE}, & \text{convection} \\
  w_{\text{turb}} = 0.7 \cdot \sqrt{TKE}, & \text{turbulence} \\
  w_{\text{LS}} = \bar{w}, & \text{large scale}
\end{cases}$$

(12)

with the convective updraft $w_{\text{conv}}$ being used for convective detrainment (section 3.3.1.1), the turbulent updraft $w_{\text{turb}}$ for turbulent vertical diffusion (section 3.3.2), and the large-scale updraft $w_{\text{LS}}$ for large-scale liquid cloud formation (section 3.3.3). A minimum velocity of 1 cm/s is assumed for all updrafts.

### 3.4.2. Homogeneous Freezing of Cloud Droplets

While cloud droplets can remain in the liquid phase below 273 K, they will freeze homogeneously if they are transported to regions where the temperature is below their homogeneous freezing temperature. Larger cloud droplets start to freeze spontaneously at 239 K, but at 235 K the homogeneous freezing rates are large enough that all cloud droplets freeze very rapidly (Kärcher & Seifert, 2016). While the cloudy air is cooled from 239 K to 235 K deposition on earlier frozen cloud droplets reduces the water vapor. This decrease of RH by deposition dominates over the increase of RH by cooling when the updraft/cooling rate is small. In this case, RH can be reduced to below saturation over liquid water, which leads to evaporation of the remaining cloud droplets and therefore not all cloud droplets freeze. Thus, the ICNC is proportional to the cooling rate and strength of the updraft. Kärcher and Seifert (2016) report a dependence of ICNC by homogeneous freezing $N_{i,\text{HomFrz}}$ on the updraft velocity $w$ with $N_{i,\text{HomFrz}} \propto w^{3/2}$ in the fast-growth regime ($w < 10$ m/s) and $N_{i,\text{HomFrz}} \propto w$ for larger updrafts in the slow-growth regime (Kärcher & Lohmann, 2002). We approximate their results with

$$N_{i,\text{HomFrz}} = 3 \cdot 10^6 \cdot w^{1/2} \text{ m}^{-3} \text{ for } w < 10 \text{ m/s}$$

(13)

and extrapolate this result to larger updrafts:

$$N_{i,\text{HomFrz}} = 9.49 \cdot 10^6 \cdot w \text{ m}^{-3} \text{ for } w \geq 10 \text{ m/s}$$

(14)

and use this relation (limited by the available CDNC) for the process of homogeneous freezing of cloud droplets. For convective updrafts $w_{\text{conv}}$ is used (section 3.3.1.1) while for large-scale freezing the sum of the large-scale updraft $w_{\text{LS}}$ and the turbulent updraft $w_{\text{turb}}$ is used (section 3.5.3). For future model development, we will consider choosing $\max(w_{\text{conv}}, w_{\text{turb}})$ for convective updrafts, as the turbulent vertical diffusion scheme simulates parts of the boundary layer convective motions.

### 3.4.3. Cirrus Nucleation Scheme

In addition to cloud droplet freezing, ice crystals in the cloud scheme can form by homogeneous and heterogeneous nucleation directly from the vapor phase at temperatures colder than 235 K (cirrus regime). The cirrus nucleation scheme by Kärcher et al. (2006) parameterizes the competition between homogeneous and heterogeneous ice nucleation during an adiabatic air parcel ascent and includes water vapor deposition on preexisting ice crystals. In our cloud scheme, we implement an updated version of Kärcher et al. (2006) that was rewritten for optimized performance and improved code readability in which the water vapor deposition formulation has been updated to the formula given in Lohmann et al. (2016), and the initial ice crystal size at nucleation has been set to the aerosol size. Finally, to calculate ice nucleation on dust particles, ICNC is derived from the activated fraction of the dust particles (Kuebbeler et al., 2014) which depends on the temperature and supersaturation following Ullrich et al. (2017). Thus, the activated fraction will increase during an air parcel ascent, which is not considered in the Kärcher et al. (2006) scheme. As a result, heterogeneous ice crystal nucleation in our updated cirrus scheme is now conceptually comparable to the nucleation spectra in Barahona and Nenes (2009).

The cirrus nucleation scheme is coupled with the aerosol module HAM by calculating heterogeneous ice nucleation for uncoated and coated dust particles in the accumulation and coarse modes. Coating of dust
by sulfuric acid reduces the ice nucleation ability. We simulate this effect by calculating the activated fraction after Ullrich et al. (2017), but at a 20% lower supersaturation, following Kuebbeler et al. (2014) who also assumed a 20% shifted ice nucleation onset supersaturation for ice nucleation on coated dust compared to the uncoated dust. For homogeneous freezing, all soluble aerosols from HAM are used except those in the nucleation mode. The updraft velocities are as described for CCN activation in Equation 12.

3.5. In-Cloud Processes That Do Not Change Cloud Fraction

For all in-cloud processes we assume that the water vapor, cloud liquid water, and ice are distributed homogeneously within the cloud volume. Therefore, all in-cloud processes do not change cloud fraction, except if the cloud completely dissipates. Further, to avoid overly complex microphysics in this study, we assume all cloud droplets and ice crystals to have the same size within the grid box. The integration of a more complex description of the ice crystal properties, as in Morrison and Gettelman (2008) and Dietlicher et al. (2018), is left for further studies.

3.5.1. Condensation/Evaporation and Deposition/Sublimation Rates

While we assume large-scale liquid clouds to be saturated at formation, we do not apply saturation adjustment for any liquid or ice clouds after formation. Instead, we explicitly calculate the condensation (evaporation) rate for cloud droplets and deposition (sublimation) rate for ice crystals (equations in Appendix A.1), using the in-cloud saturation ratios over liquid water and ice. Condensation and deposition are limited by the in-cloud supersaturated water vapor with respect to liquid water and ice, respectively, while evaporation and sublimation are limited by the liquid water mass and ice mass. In mixed-phase clouds, in case of sub-saturation with respect to liquid water but supersaturation with respect to ice, as in Dietlicher et al. (2018), the Wegener-Bergeron-Findeisen process (Korolev & Isaac, 2003) is accounted for by limiting deposition by the supersaturated water vapor with respect to ice plus the liquid cloud water. Additional liquid water is evaporated when the deposition rate over one time step exceeds the supersaturated water vapor plus the evaporating liquid water. Following the assumption that all cloud droplets/ice crystals have the same size and are distributed homogeneously, evaporation/sublimation shrinks all cloud droplets/ice crystals equally but does not reduce the number concentration or the cloud fraction unless the entire cloud dissipates.

3.5.2. Formation of Additional Cloud Droplets and Ice Crystals in the Cloud

If the cloudy air is still supersaturated after condensational and depositional growth (e.g., when condensational/depositional growth on the available cloud droplets/ice crystals is not sufficient to deplete all supersaturation), we allow for activation and nucleation of additional cloud droplets and ice crystals, respectively. The number concentrations and mass mixing ratios of the additional cloud droplets and ice crystals are calculated using the CCN activation and cirrus nucleation schemes (sections 3.4.1 and 3.4.3). For the calculations, the large-scale updraft velocity plus the turbulent updraft $w_{turb}$ is used. To consider the effect of preexisting cloud droplets and ice crystals, the updraft velocity is reduced by the equivalent downdraft velocities that correspond to the condensation and deposition rates, respectively (as described in Appendix A.2).

3.5.3. Freezing of Cloud Droplets and Melting of Ice Crystals

When cloud droplets are advected to areas in which temperatures are below 235 K, the freezing formulation described in section 3.4.2 is used to calculate ICNC, using the sum of the large-scale updraft velocity $w_{LS}$ and the turbulent updraft $w_{turb}$. Note, that compared to ECHAM-HAM, this process will lead to fewer ice crystals as not all cloud droplets freeze, but some evaporate (section 3.4.2).

At mixed-phase temperatures between 235 K and 273.15 K, ice crystals can form by immersion freezing of cloud droplets on mineral dust particles. We implemented the freezing parameterizations by DeMott et al. (2015) and Ickes et al. (2017). DeMott et al. (2015) parameterized the concentrations of INPs based on measurements of particles larger 0.5 μm for which we use all HAM coarse mode particles as proxy. The Ickes et al. (2017) parameterization calculates immersion freezing of internally mixed accumulation and coarse mode dust particles based on classical nucleation theory within a freezing time of 10 s. For both schemes, if the predicted frozen CDNC exceed the existing in-cloud ICNC, we freeze additional cloud droplets.

Ice crystals that are advected to areas with temperatures above 273.15 K melt as long as the latent heat consumption is sufficient to keep the temperature above 273.15 K. We transfer melting ice crystals to the rain
drop category as these ice crystals are usually much larger than cloud droplets, and can therefore be assumed to rain out quickly.

### 3.5.4. Aggregation and Riming

In ECHAM-HAM, aggregation is only calculated to transfer ice crystals to the (diagnostic) snowflake category, and only snowflakes can collide with cloud droplets to form rime. In the new scheme, ice crystals can grow by aggregation with other ice crystals and by riming following collisions with cloud droplets and remain in the grid box. Aggregation and riming are calculated based on collision kernels of the colliding hydrometeors that depend on the ice crystals’ projected area and fall velocity (Lohmann et al., 2016). However, not every collision will lead to aggregation and riming, as we use the collection efficiencies included in ECHAM-HAM. The equations for the ice crystal sedimentation velocity $v_{\text{sed}}$ and the aggregation and riming rates can be found in Appendix A.3.

### 3.6. Precipitation and Ice Crystal Sedimentation

We keep the diagnostic rain approach from ECHAM-HAM that assumes that cloud droplets remain within the cloud, and raindrops must either reach the surface within one time step or evaporate below cloud. As in ECHAM-HAM, autoconversion of cloud droplets to raindrops is calculated after Khairoutdinov and Kogan (2000), and CDNC is reduced proportionally to the change in mass mixing ratio. The autoconversion rate is multiplied by a constant enhancement factor to account for the underestimation of process rates due to the assumption of homogeneously distributed cloud water in the cloudy area (Pincus & Klein, 2000); Boutle et al. (2014) suggested a resolution-dependent enhancement factor that we will consider for future model development. Further, as the Khairoutdinov and Kogan (2000) scheme was developed for large eddy simulations, we will also consider the implementation of the prognostic precipitation approach by Sant et al. (2015) in future model development. As in ECHAM-HAM, we keep the combination of slow sedimentation of all ice crystals and fast sedimentation of snowflakes that can reach the surface within one time step or sublimate below cloud base. However, we have changed the snowflake formation as we want to allow ice crystals to remain in the grid box after aggregation. Therefore, we only transfer ice crystals in the grid box into the snow category once they exceed a tunable volume radius of $r_{\text{s,min}} = 200 \mu m$. This minimum size for snowflakes was increase from 100 $\mu m$ in ECHAM-HAM in the tuning process of the model (section 4.2) to allow for a higher ice water path. As in ECHAM6, but differently from ECHAM6-HAM2, we do not allow for accretion of newly formed rain with cloud droplets in the same grid box or aggregation of newly formed snow with ice crystals in the same grid box. The assumption of a monodisperse cloud droplet and ice crystal size distribution for the calculation of precipitation and sedimentation, however, underestimates the size-sorting effect (e.g., Milbrandt & McTaggart-Cowan, 2010).

#### 3.6.1. Interaction of Raindrops and Snowflakes While Falling Through a Level

Raindrops and snowflakes by definition reach the ground within one time step and are simulated in terms of their mass fluxes $F_{\text{rain, mass}}$ and $F_{\text{snow, mass}}$ in mass per m$^2$ and number fluxes $F_{\text{rain, conc}}$ and $F_{\text{snow, conc}}$ in raindrops/snowflakes per m$^2$. We assume all newly formed raindrops to have a radius of 100 $\mu m$, which is a typical size for a drizzle drop (Lohmann et al., 2016) and all newly formed snowflakes to have the radius $r_{\text{s,min}}$.

Raindrops and snowflakes can interact with water vapor molecules and hydrometeors on their way to the surface via condensation, evaporation, deposition, sublimation, accretion, and riming. In the new scheme, raindrops and snowflakes can grow by condensation and deposition, while in ECHAM-HAM they could only evaporate and sublimate. Accretion and riming only depend on the hydrometeors within the precipitation column, whereas condensation, evaporation, deposition, and sublimation depend on the relative humidity and also on the time that the raindrops and snowflakes need to fall through one level. In ECHAM-HAM evaporation and sublimation is calculated in every grid box for a full time step length. However, a better estimate of the residence time is the time needed for raindrops and snowflakes to fall through each grid box, that is, $dt = \Delta z / v_{\text{rain, snow}}$. We use raindrop fall velocities $v_{\text{rain}}$ (Equation A10) reported by Lohmann et al. (2016) and the ice crystal sedimentation velocity $v_{\text{sed}}$ as snowflake fall velocity $v_{\text{snow}}$ (Equation A4), as it is size-dependent and also valid for snowflake sized ice crystals. We assume maximum overlap of precipitation within clouds. Further, growth by condensation/deposition leads to a size increase of all raindrops and snowflakes, whereas evaporation/sublimation leads to a reduction in the number concentration, mimicking that the first raindrops/snowflakes of a precipitation event will evaporate/sublimate completely.
Accretion of cloud droplets by raindrops, riming of cloud droplets on snowflakes, and accretion of ice crystals by snowflakes are based on the hydrometeors’ projected area (equations in Appendix A.5), with the collection efficiencies from ECHAM-HAM. CDNC and ICNC are reduced proportionally to the reductions in cloud water and ice masses. These processes do not change the rain/snow number fluxes but result in larger raindrops and snowflakes.

3.6.2. Ice Crystal Sedimentation

Ice crystals can sediment into the next level below. Assuming a homogeneous distribution of monodisperse ice crystals, the fraction of ice crystals falling into the next level is given as

$$F_{\text{FallOut}} = \frac{v_{\text{sed}} \cdot \Delta I}{\Delta Z}$$

(15)

If sedimentation is treated like vertical advection, the cloud fraction has to be reduced by the same factor and added to the next level. A different assumption would be to not reduce the cloud fraction in the upper level (assuming the cloud to spread the full vertical depth) and increase the cloud fraction in the lower level only if the horizontal extent of the sedimenting ice crystals is larger than preexisting cloud fraction in this level (assuming that ice crystals sediment through the already existing cloud before forming virga). Here we use a mixture of both approaches to cover both cases. Half of $F_{\text{FallOut}}$ is reduced and added in the level below, assuming maximum overlap for the changes in the in-cloud water vapor. For the other half, the difference to the cloud fraction in the lower level is added to the lower level. When the cloud fraction is increased, and cloud-free air is turned into cloudy air, ice crystals sediment into the cloud-free conditions where they might sublimate in the next time step. Finally, for ice crystal sedimentation in mixed-phase clouds, all cloud fraction changes are weighted by the fraction of the grid-mean ICNC to the sum of CDNC and ICNC, as a few sedimenting ice crystals from a cloud with many cloud droplets do not significantly influence the cloud optical thickness (due to the hydrometeor size and phase dependence of the extinction coefficient) and hence the cloud fraction.

3.7. Variable Updates, Minimum Values, Radiation, and Subgrid Scale Marine Stratocumulus and Orographic Cirrus Clouds

We have rewritten most of the cloud routine for better readability and to assure consistency for all processes. In particular, all changes affecting both the number concentrations and mass mixing ratios are calculated and applied together. To always assure consistency of all cloud variables, the cloud variables are updated in between the cloud processes in the following order. First, convective cloud formation and cloud formation due to turbulent vertical diffusion are calculated and updated, as the turbulence and convection schemes are called before the cloud scheme. Next, large-scale cloud formation in the cloud-free air and condensation and additional cloud droplet activation/ice crystal nucleation in the cloudy air are calculated in parallel and applied afterwards. Thereafter, the ice microphysical processes of heterogeneous ice crystal nucleation, aggregation, and riming are calculated in parallel and the changes applied afterwards. Finally, the precipitation and sedimentation processes are calculated in a top-to-bottom loop and applied.

In the standard version of ECHAM-HAM, a minimum CDNC of 10 or 40 cm$^{-3}$ is applied to compensate for unrealistically low aerosol concentrations (Neubauer et al., 2019). In the new scheme, a minimum CDNC of 40 per cm$^{3}$ is prescribed only during cloud droplet activation; however, CDNC can evolve freely below this minimum concentration afterwards. This implementation of a minimum CDNC is more physical as it does not influence the CDNC during subsequent processes like freezing in mixed-phase clouds and precipitation formation. Further, CDNC is not set to the activated CCN at cloud base (as done in ECHAM-HAM) as we want to preserve the prognostic nature of the scheme to allow vertical cloud droplet transport in the cloud. In addition, in ECHAM-HAM the CDNC and ICNC tracers were implemented as in-cloud values instead of grid-mean values, which is inconsistent with the cloud water and ice prognostic variables and therefore was changed to grid-mean values in the new model. Finally, the homogeneous freezing temperature was reduced from 238 K in ECHAM-HAM to 235 K which is closer to the latest literature values (Koop & Murray, 2016).

In the radiation scheme, cloud fraction is interpreted as a horizontal cover, and the cloud is assumed to occupy the entire vertical extent of the grid box. This missing vertical cloud fraction is a limitation for thin clouds and should be overcome in future studies by introducing a vertical cloud fraction or by increasing the vertical resolution. The interaction of the cloud scheme with the radiation scheme is kept the same as
described in Lohmann et al. (2007): The in-cloud liquid water and ice mass mixing ratios and number concentrations and the grid-mean water vapor are used for the radiation calculations. The use of the cloudy and cloud-free water vapor could be added to the radiation scheme in future studies.

The default cloud fraction scheme in ECHAM includes a parameterization for vertical subgrid cloudiness for marine stratocumulus (Mauritsen et al., 2019). It allows for a cloud fraction of 100% for grid boxes with grid-mean RH below 100% in case of a low-level temperature inversion over the ocean to “correct for the fact that the grid-box mean RH in vicinity of inversions is not a good predictor of the cloud fraction in the vertical” (Mauritsen et al., 2019). However, we did not include a similar formulation in our new scheme as a consistent model state is not possible. We leave improvements to this approach for future research.

Different to Kuebbeler et al. (2014) and Gasparini and Lohmann (2016), we did not include a parameterization for subgrid scale orographic cirrus clouds (Joos et al., 2008) even though in Gasparini and Lohmann (2016) these orographic updrafts are a dominating source for homogeneous nucleation of solution droplets and observations show increased ICNC over mountains (Sourdeval et al., 2018a, 2018b). The reason is that subgrid scale orographic cirrus clouds are mostly stationary clouds; that is, the cloud constantly forms in the updraft and dissolves in the downdraft. Therefore, including orographic cirrus clouds only as a source for ice clouds strongly overestimates the global effect of orographic cirrus formation, as the cloud fraction sink of the subgrid scale downdraft is missing and the ice crystals are transported too far away from mountainous regions. In future studies, orographic cirrus clouds could be added to the model by including an additional diagnostic ice mode that is not advected and stationary over the mountains. However, a detailed formulation for this is beyond the scope of this study.

3.8. Changes in the Aerosol Model HAM

In the aerosol model HAM, we included nucleation scavenging for insoluble dust if heterogeneous ice nucleation on dust particles is simulated by the cirrus nucleation scheme and included bugfixes for the in-cloud nucleation scavenging for cloud droplets and ice crystals. Further, in HAM, dust can be transferred from the externally mixed to the internally mixed dust mode by coating with sulfuric acid (Zhang et al., 2012). However, as there are many sulfuric acid water droplets at cirrus cloud altitudes, and the number concentrations of the internally mixed species are not traced individually, internal mixing of a few large dust particles with many small sulfuric acid-water droplets effectively leads to a decrease of the dust size after it is coated, and after it is transferred to the internally mixed mode. This introduces two problems: first, coated dust sediments are slower from high altitudes than uncoated dust, which is unphysical, and, second, coated dust concentrations that are calculated for the ice nucleation processes in the cloud scheme are too high. We adjust to the second problem by introducing a minimum size of 200 nm for the accumulation mode and 500 nm for the coarse mode coated dust particles when their number concentrations are calculated for the cloud scheme. However, we cannot address the first problem in this study and note that in our simulations the size of the coated dust particles is too small, which reduces their sedimentation rate, therefore increasing their atmospheric lifetime and leading to an overestimation of heterogeneous cirrus nucleation.

4. Results

4.1. Model Setup and Tuning

Simulations are performed at T63 spectral resolution (1.875° × 1.875°) with 31 vertical levels in the troposphere up to 10 hPa resulting in a vertical resolution of <100 m near the surface and 1,000 m at the highest cirrus altitudes and a time step of 450 s. All simulations were performed for the years 2003–2012, the ECHAM-HAM reference period for present day simulations (Neubauer et al., 2019), with October to December 2002 as spin-up, and with prescribed SST and sea ice cover (monthly climatology derived from AMIP data for 2000–2015, Taylor et al., 2000), and greenhouse gas concentrations from 2008. The tuning factors for the deep, shallow, and mid-level convective turbulent entrainment rates (0.2/km, 3/km, and 0.1/km), for the fraction of the convective mass flux that overshoots the level of neutral buoyancy (0.2), and the inhomogeneity factors that are introduced to compensate for the plane-parallel bias (Cahalan et al., 1994) of liquid clouds (0.8) and ice clouds (0.7) have the same values as in ECHAM-HAM (Neubauer et al., 2019). In all simulations except in our ECHAM-HAM reference simulation, we remove the upper limit of mid-level convection of 400 hPa (Gasparini et al., 2018; Giorgetta et al., 2018) and additionally allow
Table 1
Description of the Simulations

| Name                      | Description                                                                                                                                                                                                 |
|---------------------------|-------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|
| REF                       | Reference simulation of ECHAM-HAM similar to Neubauer et al. (2019) but with nucleation scavenging bugfixes                                                                                               |
| REF-ConvCirrus            | As REF, but allowing convective precipitation independently of the convective cloud thickness and with different convective precipitation tuning (see section 4.1 for details). Further, the cirrus scheme of section 3.4.3 with heterogeneous ice nucleation and considering preexisting ice crystals is used. |
| REF-ConvCirrus-DetrLiquid| As REF-ConvCirrus, but detraining all convectively detrained condensate in the liquid phase at $T > 235$ K.                                                                                                  |
| NEW-OldIce                | New cloud scheme with old ice microphysics (only aggregation of ice crystals for snow formation) and calculating the detrained ICNC based on the ice crystal size climatology. Further, every cloud droplet that is advected to cirrus temperatures freezes.       |
| NEW-NewIceMicro           | As NEW-OldIce, but ice crystals can grow by aggregation and need to grow to 200 $\mu$m radius to form snow (minimum snow radius was 100 $\mu$m before), using the DeMott et al. (2015) mixed-phase ice nucleation scheme, removing the ice crystal sedimentation velocity enhancement factor of 3 and introducing that not every cloud droplet that is advected to cirrus temperatures freezes (see section 3.4.2). |
| NEW-NewDetrICNC           | As NEW-NewIceMicro, but the convectively detrained ICNC is calculated based on the convective updraft velocity (see section 3.3.1.1).                                                                          |
| NEW-ReducedDetrICNC       | As NEW-NewDetrICNC, but the convectively detrained ICNCs are reduced as described in section 3.3.1.2.                                                                                                      |

initiation of mid-level convection at levels higher than 300 hPa as both of these limits introduce an unphysical split between the mid and upper troposphere. Further, in all simulations, as in Giorgetta et al. (2018), we reduce the time scale for the relaxation of CAPE from 2 hr to 1 hr. This is used as closure to calculate the deep convective mass flux. Finally, we allow precipitation to form in convective clouds independent of the thickness of the convective cloud. In ECHAM, a minimum cloud thickness of 150 hPa (300 hPa) over oceans (land) was required for the initiation of convective precipitation. This thickness criterion was removed during the porting of the convection scheme from ECHAM to ICON (ICoahedral Nonhydrostatic GCM, the successor of ECHAM; Giorgetta et al., 2018; Crueger et al., 2018), and we also remove this criterion in our simulations with ECHAM-HAM as it reduces the overestimation of the cloud radiative effect in open-cell shallow convective areas.

Note that removing the thickness criterion for convective precipitation formation leads to more washout of sea salt aerosol, which decreases the aerosol optical depth (AOD) in the open-cell convection areas over the ocean. This reduces a high bias of AOD in ECHAM-HAM (Tegen et al., 2019), but in combination with the changes in nucleation scavenging, the sea salt and dust aerosol burdens decrease significantly, and the simulated AOD is lower than observed. The aerosol emissions in ECHAM-HAM are below the averages of the Aerosol Comparisons between Observations and Models project AeroCom (Tegen et al., 2019; Textor et al., 2006).

### 4.2. Step-by-Step Moving to the New Cloud Scheme

In this section, we show, step-by-step, the effects of moving from the standard ECHAM-HAM model setup to the new cloud scheme with all features as discussed in section 3. Table 1 describes all simulations, Table 2 shows the tuning parameters, and Table 3 summarizes top-of-the-atmosphere fluxes, cloud radiative effects (CRE), and key cloud properties for all simulations. Figure 4 shows the corresponding annual zonal-mean cloud fraction, Figure 5 the annual zonal-mean values of key cloud properties, and Figure 6 the annual zonal-mean in-cloud ICNC. The simulations are approximately tuned to radiative balance (Table 3), but for the intermediate simulations we favor the same tuning parameters over an exact radiative balance for better comparability. In Figure 4, the Cloud Feedback Model Intercomparison Project Observational Simulator Package (COSP, Boda-Salcedo et al., 2011, Chepfer et al., 2010) is used to allow for a direct comparison of the modeled cloud fraction to the satellite observations of the GCM-Oriented CALIPSO Cloud Product (GOCCP, Chepfer et al., 2010). In Figure 6, the ICNC from the DARDAR-Nice product are shown for the years 2006–2016 for comparison (Sourdeval et al., 2018a, 2018b).

In REF-ConvCirrus, the convective precipitation formation thickness criterion is removed, and the increase in precipitation and reduction of convectively detrained condensate is compensated by a reduced critical RH for low-level cloud formation, a lower enhancement of the autoconversion rate of cloud water to rain, and a
decreased conversion rate of cloud water to rain in convective clouds, which is in the range of the ECHAM6 setups (Crueger et al., 2018; Mauritsen et al., 2012). Less convective precipitation leads to more detrainment of cloud ice at cirrus altitudes, which is compensated by a larger enhancement factor increasing the aggregation rate of ice crystals to snow in the stratiform microphysics scheme. Finally, to make the REF-ConvCirrus simulation more comparable to the simulations with the new cloud scheme, it uses the new cirrus nucleation scheme with homogeneous and heterogeneous cirrus nucleation (section 3.4.3). The reduced convective precipitation leads to an increased tropical high-level cloud fraction (Figure 4) and a better representation of the longwave cloud radiative effect (LW CRE, Figure 5). Here, more convective detrainment increases the LW CRE in the tropics, while faster snow formation decreases the LW CRE globally, resulting in a better representation of the ratio between the tropical and extra-tropical LW CRE. The changes in convective precipitation and the increase of the critical RH lead to a reduction of the tropical cloud fraction at 2 km and a slight increase of the cloud fraction in the levels below (Figure 4), leading to an improved simulation of the shortwave cloud radiative effect (SW CRE, Figure 5). However, while these changes lead to a better representation of the CRE, the REF-ConvCirrus simulation

### Table 2

| Tuning Parameters of All Simulations With the Parameters That Have Been Changed for the Specific Simulation When Going From Left to Right in Bold |
|-------------------------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|
|                              | REF OldIce     | REF-ConvCirrus DetrLiquid | REF-ConvCirrus NewIceMicro | REF-ConvCirrus NewDetrICNC | REF-ReducedDetrICNC | NEW             |                |
| -----------------------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|
| Convective conversion of water to rain | 0.0009         | **0.0002**     | 0.0002         | 0.0002         | 0.0002         | 0.0002         | 0.0002         | 0.0002         |
| Autoconversion of cloud water to rain | 9              | 4              | 4              | **2.5**        | 2.5            | 2.5            | 2.5            | 1.8            |
| Aggregation of ice crystals | 900            | **3,000**      | 3,000          | 700            | 700            | 700            | 2              | 2              |
| Ice crystal sedimentation velocity | 3             | 3              | 3              | 3              | 1              | 1              | 1              | 1              |
| Minimum snowflake radius | 100 μm         | 100 μm         | 100 μm         | 100 μm         | **200 μm**     | 200 μm         | 200 μm         | 200 μm         |
| RH_{crit} for cloud fraction/large-scale cloud formation | 75–97.5%       | 75–90%         | 75–90%         | **90%**        | 90%            | 90%            | 90%            | 90%            |
| Conv. detr. air mixing with environmental air | —              | —              | —              | —              | —              | —              | **80%**        | 80%            |
| Implicit aggregation of the detrained ICNC (reduction factor per m/s) | —              | —              | —              | —              | —              | —              | —              | **5.0**        | 5.0            |

### Table 3

| Overview of Global-Mean Key Values of Radiation and Cloud Properties of All Simulations |
|-----------------------------------------|----------------|----------------|----------------|----------------|----------------|----------------|----------------|
| Observations | REF SW (W/m²) | REF LW (W/m²) | REF SW CRE (W/m²) | REF LW CRE (W/m²) | New IceMicro | New DetrICNC | New ReducedDetrICNC | New |
| SW (W/m²) | 241.2 | 238.3 | 241.8 | 241.9 | 242.2 | 238.0 | 236.9 | 242.7 | 240.9 |
| LW (W/m²) | −240.2 | −238.1 | −240.8 | −240.9 | −241.1 | −236.9 | −235.2 | −240.1 | −240.0 |
| Net (W/m²) | 1.0 | 0.3 | 1.0 | 1.0 | 1.1 | 1.1 | 1.7 | 2.6 | 0.9 |
| SW CRE (W/m²) | −45.8 | −50.4 | −47.6 | −47.5 | −47.6 | −51.9 | −53.1 | −47.2 | −49.0 |
| LW CRE (W/m²) | 27.9 | 23.8 | 20.9 | 20.4 | 21.2 | 26.0 | 27.5 | 22.5 | 22.6 |
| Net CRE (W/m²) | −17.9 | −26.6 | −26.7 | −27.1 | −26.4 | −25.9 | −25.6 | −24.7 | −26.4 |
| CC (%) | 68 ± 5 | 68.4 | 67.1 | 66.6 | 60.9 | 66.2 | 67.4 | 64.2 | 64.5 |
| LWP (g/m²) | 68.0 | 75.9 | 77.9 | 65.4 | 76.7 | 76.9 | 61.7 | 69.5 |
| LWP ocean (g/m²) | 81.4 (30–90) | 73.9 | 79.6 | 81.6 | 71.7 | 83.3 | 83.6 | 67.7 | 75.9 |
| IW (g/m²) | 25 ± 7 | 14.5 | 5.9 | 4.7 | 2.1 | 2.7 | 2.8 | 26.5 | 26.4 |
| CDNC (x10^10 m⁻³) | 3.1 | 3.2 | 3.4 | 3.0 | 2.0 | 2.2 | 2.2 | 1.8 | 1.9 |
| ICNC (x10^10 m⁻³) | 0.075 | 0.054 | 0.040 | 0.066 | 0.020 | 0.050 | 0.017 | 0.017 |
| Precip (mm/day) | 2.7 ± 0.2 | 3.0 | 3.0 | 3.0 | 3.1 | 3.0 | 3.0 | 3.1 | 3.1 |
| AOD | 0.12–0.18 | 0.11 | 0.07 | 0.07 | 0.06 | 0.06 | 0.06 | 0.06 | 0.06 |

**Note.** The table shows the values of shortwave (SW), longwave (LW), and net radiative fluxes at the top of the atmosphere (TOA), the according cloud radiative effects (CRE), the total cloud cover (CC), the global and oceanic liquid water path (LWP, LWP_ocean), the ice water path (IWP), the vertically integrated cloud droplet and ice crystal number concentrations (CDNC and ICNC), precipitation, and the aerosol optical depth (AOD). The sources of the observations are CERES observations for the TOA radiative fluxes (climatology 2005–2015, version CERES-EEAF-Ed4.0; Kato et al., 2013; Loeb et al., 2009; Loeb et al., 2018), CC from Stubenrauch et al. (2013) and Matus and L’Ecuyer (2017), LWP_ocean from Platnick et al. (2015, 2017), Stengel et al. (2017), and Elsaesser et al. (2017), IWP from Li et al. (2012), precipitation central value from Adler et al. (2018) and the range from Adler et al. (2012). AOD from AeroCom (Kinne et al., 2006), MODIS (Wei et al., 2019, 2019), and AATSR (Popp et al., 2016).
has a very small ice water path (IWP, Table 3, Figure 5) due to the high snow formation rate. Together with the too large high-level cloud fraction, this indicates a structural problem in the ice cloud fraction and microphysics formulation, which is in accordance with Gasparini et al. (2018) who observed too high extinction of cirrus clouds and too small ice crystals in ECHAM-HAM when comparing them to the ice crystal climatology by Heymsfield et al. (2014). The pattern of higher in-cloud ICNC below 8 km in REF is absent in REF-ConvCirrus (Figure 6) due to the removal of the mid-level convection limits. Further, ICNC is decreased in extratropical cirrus clouds due to faster aggregation and heterogeneous cirrus nucleation (Figure 6).

In REF-ConvCirrus-DetrLiquid all convectively detrained condensate is detrained as liquid cloud water at $T > 235$ K, which is more reasonable as discussed in section 3.3.1.1. This leads to a strong reduction of the ICNC in mixed-phase clouds (Figure 6) that is in better agreement with the ICNC satellite product, in which the highest ICNC occur in cirrus clouds. However, REF-ConvCirrus-DetrLiquid still has a too low IWP and a too large high-level cloud fraction.

NEW-OldIce uses the new scheme described in section 3 with the prognostic cloud fraction and prognostic in-cloud water vapor. However, for this simulation we keep key ice microphysical concepts from the reference version of ECHAM-HAM; that is, aggregating ice crystals are immediately considered as snow, every cloud droplet that is advected to cirrus temperatures freezes and forms an ice crystal, and the convectively detrained ICNC are calculated based on the ice crystal size climatology. Therefore, this simulation with the new cloud scheme is as close to REF-ConvCirrus-DetrLiquid as possible. The new cloud scheme results in a decrease of high-level cloud fraction (Figure 4), as in situ cirrus cloud formation can occur in parts of the grid box. The low-level cloud fraction increases with the new cloud scheme (Figure 4), as a critical RH is not required for cloud formation by convective detrainment and turbulence. The new cloud scheme requires a weaker enhancement of the autoconversion rate of cloud water to rain and of the aggregation rate of ice

**Figure 4.** Annual zonal mean cloud fraction for all simulations of Table 1 as seen by the COSP simulator for the years 2003–2012 and from CALIPSO-GOCCP for comparison (Chepfer et al., 2010).
crystals to snow (Table 2), as shown by cloud ice residing in smaller cloud volumes, with higher ICNC (Figure 6), leading to higher aggregation rates. Additionally, the reduced autoconversion enhancement can be explained by a lower CDNC (Table 3), which results from applying the minimum CDNC only for CCN activation and thereby only to the new cloud volume. Pincus and Klein (2000) discuss that the assumption of homogeneously distributed cloud water in the cloudy area can lead to an underestimation of process rates by up to a factor of 2, while Boutle et al. (2014) discuss that the Khairoutdinov and Kogan (2000) scheme underestimates the autoconversion rates by up to a factor of 5 at GCM resolutions of 100 km. Thus, the autoconversion rate enhancement factor of 2.5 in NEW–OldIce (Table 2) can be justified as compensating for the missing subgrid-scale variability, while the enhancement factor of 9 in REF (Table 2) and the ice crystal aggregation enhancement factor of 900 in REF and NEW–OldIce (Table 2) are too high. Thus, with all these changes, the new cloud scheme helps to reduce model biases. However, it cannot improve the too low IWP (Table 3, Figure 5). The simulated ICNCs exceeding 1 cm$^{-3}$ are also too high using the new cloud scheme (Figure 6), as the observed ICNCs do not exceed 300 L$^{-1}$ in the annual average.

In NEW–NewIceMicro, the ice microphysics scheme described in section 3 is used; that is, ice crystals can grow by aggregation and riming and remain in the grid box, and not every cloud droplet that is advected to cirrus temperatures freezes and forms an ice crystal (section 3.4.2). Further, an ice crystal sedimentation velocity enhancement factor of 3, that is applied in ECHAM-HAM, was removed in this simulation, and the minimum snowflake radius was increased from 100 to 200 μm (Table 2). These changes lead to a decrease of the ICNC everywhere in the model (Figure 6). On the other hand, cloud fraction is increased in the cirrus levels (Figure 4), as ice clouds have a longer lifetime due to the removal of the ice crystal sedimentation velocity enhancement factor and a delayed conversion from ice crystals to snowflakes. Corresponding to the increased cirrus cloud fraction, the cloud radiative effects increase significantly (Table 2 and Figure 5).

In NEW–NewDetrICNC, the convectively detrained ICNCs are calculated based on the new formulation of section 3.3.1.1 and not according to the ice crystal size climatology. This leads to an increase of the tropical ICNC (Figure 6) and LW CRE (Figure 5). However, IWP remains very low (Table 3, Figure 5) as a high aggregation rate is still needed to achieve radiative balance (Table 2). Aggregation is more efficient at warmer

Figure 5. Annual zonal mean values of key cloud properties for all simulations of Table 1 for the years 2003–2012. Shown are total cloud cover as seen by the COSP simulator (a), shortwave and longwave cloud radiative effect (SW and LW CRE) (b and c), precipitation (d), and liquid and ice water path (LWP and IWP) (e and f) of all simulations and satellite observations. Observation sources: Cloud cover from CALIPSO GCCP (Chepfer et al., 2010), AVHRR-PM (Stengel et al., 2017), and MODIS collection 6.1 (Platnick et al., 2015, 2017); SW and LW CREs from CERES (Loeb et al., 2018), ERBE (Barkstrom, 1984), and TOVS (Susskind et al., 1997); precipitation from GPCP (Adler et al., 2018); LWP from Elsaesser et al. (2017), MODIS collection 6.1 (Platnick et al., 2015, 2017), and ATSR-2-AATSR (Stengel et al., 2017); IWP from Li et al. (2012).
subzero temperatures due to the quasi-liquid layer and thus leads to a stronger reduction of large ice crystals and a low IWP. However, the LW CRE is dominated by small ice crystals at cold temperatures. Therefore, reducing the global, annual mean LW CRE by enhancing the ice crystal aggregation rate addresses the wrong ice crystals and leads to a too fast removal of ice crystals from mixed-phase clouds. Consequently, the ICNC at mixed-phase temperatures are too low compared to the observed ICNC.

In NEW-ReducedDetriICNC, we directly reduce the convectively detrained ICNCs as described in section 3.3.1.2, and the ice crystal aggregation rate of all ice crystals no longer requires an enhancement factor larger than 2 to account for the subgrid-scale variability. This leads to a strong reduction of the tropical high-level ICNC, an increase of ice crystals in mixed-phase clouds (Figure 6), and an increase of the IWP (Figure 5), in better agreement with observations. Next to the ICNC and IWP, the cirrus cloud fraction agrees better with observations (Figure 4), as non-convective ice crystals can sediment further before they are converted to snowflakes due to the reduced aggregation rate of non-convective ice crystals.

To obtain the final setup of the new cloud scheme, NEW, the autoconversion rate enhancement factor is reduced. Thereby, the low-level clouds and LWP slightly increase, the SW CRE intensifies, and the TOA imbalance is in the observed range (Table 3).

### 4.3. Cloud Cover

The performance of the new cloud scheme is evaluated against the GCM-Oriented CALIPSO Cloud Product (GOCCP, Chepfer et al., 2010) using the Cloud Feedback Model Intercomparison Project Observational Simulator Package (COSP, Bodas-Salcedo et al., 2011; Chepfer et al., 2010) that simulates the satellite retrieval algorithms using GCM variables. In Figure 7, we compare annual means of low-, mid-, and high-level cloud cover following the ISCCP cloud definition and the annual zonal mean cloud fraction of the NEW simulation and include the difference between REF and the satellite observations for comparison.
The structure of low-level clouds is captured well by the new model. The marine stratocumulus cloud cover is improved but still lower than in the observations. The high bias in the Western tropical Pacific is reduced; however, the cloud cover is slightly overestimated over oceans. On the other hand, the overestimation of the low-level cloud cover over the Eurasian continent has increased. Mid-level cloud cover is underestimated in the new scheme as well as in REF; however, the low bias is reduced in the intertropical convergence zone (ITCZ) over the oceans due to the reduced convective precipitation rate and the increased cloud formation by convective detrainment in NEW. High-level cloud cover has significantly changed from REF as the new cloud scheme allows for subgrid cirrus cloud formation. Therefore, the overestimation of high-level cloud cover has been reduced everywhere in the extra-tropics, but by too much in the storm tracks. The total cloud cover has changed mainly following the changes in subtropical low-level cloud cover. The zonal mean cloud fraction structure is improved in the new scheme as the cirrus cloud overestimation was reduced. However, cirrus cloud cover is still overestimated in the tropics, and polar cirrus clouds are located at too high altitudes.

### 4.4. Cloud Fraction Sources

Figure 8 shows the cloud fraction source terms for the different cloud formation processes as discussed in section 3.3. Cloud formation by detrainment from deep convection is a dominant process in the tropics in
the ITCZ at mid and high levels and can reach up to 16 km in altitude. Shallow convection, on the other hand, is limited to the lower troposphere due to its large entrainment/detrainment rates but leads to more cloud formation than deep convection. Mid-level convection mostly occurs in the storm tracks and in the tropics over the continents. Cloud formation due to turbulent vertical diffusion dominates cloud formation in the marine stratocumulus areas and in general contributes strongly to low-level cloud formation in the boundary layer. Additionally, there is also some ice cloud formation caused by turbulent vertical diffusion at higher altitudes; however, its contribution to ice cloud formation is very small. Large-scale liquid cloud formation occurs in the storm tracks together with cloud formation due to turbulent vertical diffusion but also at higher altitudes. Interestingly, the design of the new cloud scheme leads to almost no large-scale liquid cloud formation in the tropics and subtropics over the ocean, as almost all clouds there form via convective detrainment or turbulence and there is no large-scale ascent in the subtropical downward branch of the Hadley circulation. These results agree with our understanding of the atmospheric circulation with deep convection occurring in the ITCZ, subtropical low cloud formation occurring by shallow convection and turbulent vertical motions, and large-scale ascent of air masses in the storm tracks (e.g., Hartmann, 2015; Holton, 1973; Vallis, 2017). In situ large-scale ice cloud formation is the dominating source of high-level cloud fraction and is mostly located in the storm tracks and in the tropics over the continents. However, the cloud fraction source terms are not fully representative for cirrus cloud formation pathways as liquid-origin cirrus clouds from freezing of supercooled cloud droplets is not a cloud fraction source and therefore not included in Figure 8.
4.5. Ice Crystal Sources

Figure 9 shows the ICNC source terms for the different ice crystal formation pathways. Convective detrainment and homogeneous freezing of cloud droplets are the dominant ice crystal sources. In the tropics, convective detrainment from deep convection dominates ice crystal formation in the ITCZ, and mid-level convection strongly contributes to ice crystal formation in the extra-tropics (Figures 8 and 9). Homogeneous nucleation of solution droplets occurs in the upper tropical troposphere and in the storm tracks, with more homogeneous nucleation in the southern hemisphere storm tracks. Homogeneous nucleation is partly suppressed by heterogeneous nucleation in the northern hemisphere storm tracks. In situ ice nucleation on mineral dust particles is strongest in the northern hemisphere due to larger mineral dust sources, while its contribution in the tropics and the southern hemisphere is small, which is reasonable and consistent with previous GCM studies (Barahona et al., 2014; Shi et al., 2015). However, compared to Gasparini and Lohmann (2016), the heterogeneous ice nucleation contribution has decreased which is likely due to our changes in the treatment of preexisting ice crystals (section 3.3.4) and the changes in the aerosol scheme HAM (see section 3.8) like the inclusion of nucleation scavenging of insoluble dust. Convective detrainment contributes more to ICNC (even though ICNC is reduced to account for immediate aggregation, section 3.3.1.2) and dominates over homogeneous nucleation of solution droplets in more areas than in Gasparini and Lohmann (2016) due to the different convection tuning (less convective precipitation leads to more detrainment) and no parameterization for orographic cirrus clouds in the new scheme.

Homogeneous freezing of cloud droplets mostly occurs in the storm tracks where large-scale ascent lifts supercooled cloud droplets to temperatures below the homogeneous freezing temperature and leads to almost as many ice crystals as convective detrainment. Thus, in agreement with Krämer et al. (2016) and Wernli et al. (2016), a significant fraction of cirrus clouds originates from liquid clouds and in our simulations liquid-origin ICNC outnumber in situ nucleated ICNC. Thus, while recent research has focused on improving the in situ ice crystal formation processes, these results show the need for a careful parameterization of also the liquid-origin ice crystal sources in GCMs. Finally, immersion freezing on dust particles is strongest in the northern hemisphere due to higher atmospheric dust concentrations.

Ice crystals from homogeneous and heterogeneous in situ cirrus nucleation are larger at lower altitudes and warmer temperatures (Figure 9) due to a higher saturation vapor pressure, allowing ice crystals to consume more water vapor and grow larger before depleting the supersaturation. The ice crystals from homogeneous nucleation are on average larger than the heterogeneously nucleated ice crystals. This is contrast to the general assumption that homogeneous nucleation leads to many small ice crystals while heterogeneous nucleation leads to fewer large ice crystals (e.g., Kärcher & Lohmann, 2003). However, as homogeneous nucleation requires higher supersaturations, there is more water vapor available for ice crystals to grow. Further, the ICNC from homogeneous nucleation depends on the updraft velocity (Jensen et al., 2016; Kärcher & Lohmann, 2002), and thus, homogeneous nucleation in slow updrafts results in cirrus clouds with low ICNC (Krämer, Rolf, et al., 2016). Thus, with higher water vapor availability and a low ICNC, the homogeneously formed ice crystals can grow larger than the heterogeneously formed ice crystals. Note, however, that our scheme currently does not include mesoscale gravity waves and might therefore underestimate updraft velocities and ICNC from homogeneous nucleation (Haag & Kärcher, 2004; Jensen et al., 2016; Kärcher et al., 2019; Kärcher & Podgajen, 2019; Schoeberl et al., 2015). Convectively detrained ice crystals are the largest as their ICNC has been reduced to account for aggregation in convective plumes (section 3.3.1.2). Ice crystal sizes above 100 μm have been observed in convective anvils (e.g., Jensen et al., 2009, 2018); however, the detrained ice crystals are very large in the extra-tropics, which could partly explain the too low extra-tropical high-level cloud cover (Figure 7). When cloud droplets freeze, the resulting ice crystals initially have approximately the size of the droplets (Figure 9), but they will grow afterwards by depositional growth.

Figure 10 compares the spatial patterns of the simulated ICNCs with the observations of Sourdeval et al. (2018a, 2018b); the ICNCs of the other simulations are shown in Figure A1. In general, the new scheme strongly reduces the ICNC overestimation evident in REF, and the patterns of the simulated ICNCs agree well with observations. However, ICNC is underestimated at the poles, over mountains, and in most of the extra-tropics at temperatures below −50°C, likely because of missing meso-scale gravity waves (as discussed above). In the tropics ICNC is overestimated over the maritime continent between −50°C and −30°C and underestimated over South America and Africa at temperatures below −50°C. Finally, the underestimation of ICNC over the tropical continents at cold temperatures results from a too strong
Figure 9. Annual grid-mean ice crystal number concentration source terms for homogeneous nucleation of solution droplets (a) and heterogeneous ice nucleation on mineral dust particles at cirrus temperatures (b) both from large-scale and turbulent vertical diffusion cirrus cloud formation, ice crystals from convective detrainment after immediate aggregation (c), ice crystals from homogeneous freezing of supercooled cloud droplets that are advected to temperatures below 235 K (d), and ice crystals from heterogeneous freezing of cloud droplets on dust particles above 235 K (e) for the NEW simulation. All values are shown as vertical integrals including the global mean value in units of ice crystals $m^{-2} s^{-1}$ (first column) and as zonal mean in units of ice crystals $l^{-1} s^{-1}$ (middle column). The third column shows the according ice crystal size in μm after formation and, in the case of cirrus nucleation, growth by deposition for one time step.
reduction of convectively detrained ICNC by immediate aggregation, as the ICNCs are much higher in these areas in NEW-NewDetrICNC (Figure A1). The underestimation of tropical ICNC is consistent with an underestimation of the tropical LW CRE (Figure 5). Due to the priority of achieving radiative balance during the tuning process, the LW CRE and the tropical ICNC were reduced, leading to disagreement between TOA outgoing LW and LW CRE and the observed values.

4.6. In-Cloud and Cloud-Free Saturation Ratio

Figure 11 shows the simulated cloud fraction (without using the COSP simulator) and the cloud-free and in-cloud saturation ratios with respect to ice and liquid water. The cloud-free and grid-mean saturation ratios are similar, except at the lowest altitudes where more water vapor is contained in the cloudy than in the cloud-free areas. In the areas of large-scale cloud formation (Figure 8), the patterns of the grid-mean saturation ratio with respect to ice are similar to those of cloud fraction due to the cloud fraction depending on the RH for the large-scale cloud formation processes. However, the tropical and subtropical low-level and tropical high-level cloud fraction do not follow the grid-mean RH pattern, as in these areas clouds mostly form by convection and turbulence (Figure 8), for which the cloud fraction does not depend on the grid box relative humidity, but on the mass fluxes from below. The extra-tropical cirrus cloud fraction differs between the hemispheres in NEW while the saturation ratios are similar. Here, heterogeneous in situ

![Figure 11. Annual mean in-cloud ICNCs from the DARDAR-Nice satellite product for the years 2006-2016 (a) and the NEW simulation (b) for different temperature ranges.](image)

![Figure 11. Annual zonal mean cloud fraction (a) and cloud-free, and in-cloud saturation ratios with respect to water (b, c) and grid-mean, cloud-free, and in-cloud saturation ratios with respect to ice (d, e, f) for the NEW simulation. The saturation ratio of 99% is shown by the gray line.](image)
cirrus formation (Figure 9) leads to thin cirrus clouds that are not seen by the COSP simulator (Figure 7). Thus, an RH-based cloud fraction scheme cannot capture these details of the small-scale convective and turbulent events and the formation pathways of cirrus clouds.

Figure 11 shows that the too low mid-level cloud cover (Figure 7) is consistent with a low cloud-free saturation ratio. The tropical mid troposphere and the extra-tropical mid-levels are not moist enough for sufficient liquid cloud formation at $T > 235$ K. Cloud-free ice supersaturation only occurs in the annual mean in the Antarctic upper troposphere/lower stratosphere (UTLS). In a sensitivity simulation without ice nucleation on mineral dust particles (not shown), the cloud-free water vapor is also supersaturated in the Arctic UTLS. Thus, heterogeneous cirrus cloud formation on mineral dust particles appears to keep the annual mean cloud-free saturation ratio below saturation with respect to ice as cirrus clouds form early enough to transform any supersaturated cloud-free air into cloudy air. However, the in-cloud saturation ratios with respect to ice are supersaturated in the annual average in extra-tropical cirrus clouds in both hemispheres but are most pronounced in the southern hemisphere (Figure 11). Thus, the ice crystals that form are not numerous enough or sediment too quickly to deplete all supersaturation. Tropical cirrus clouds on the other hand are subsaturated in the annual global average, likely due to convectively detrained ice crystals mixing with and sedimenting into subsaturated air. However, there is also a saturated layer at high tropical altitudes, that is, in the tropical tropopause layer (TTL). In-cloud supersaturations in the TTL and the extra-tropics have been observed in aircraft measurements (e.g., Gao et al., 2004; Jensen et al., 2005; Krämer et al., 2009; Ovarlez et al., 2002; Peter et al., 2006; Spichtinger et al., 2004) and are consistent with previous modeling studies (e.g., Jensen & Pfister, 2005; Kärcher et al., 2014; Spichtinger & Gierens, 2009a, 2009b). Finally, the in-cloud cirrus relative humidities as a function of temperature agree well with the measurements by Krämer et al. (2009) (Figure A2; updated values according to Krämer et al., 2016, and personal communication with Martina Krämer), but as in Kuebbeler et al. (2014) the variability of in-cloud relative humidity is somewhat underestimated.

Liquid clouds at $T > 273$ K are saturated with respect to liquid water (Figure 11). In contrast, clouds at mixed-phase temperatures are supersaturated with respect to ice, but subsaturated with respect to liquid water in the annual average (Figure 11). At these temperatures, clouds form in the liquid phase at supersaturation with respect to water, but transition to the ice phase with saturation with respect to ice once the Wegener-Bergeron-Findeisen process dominates.

### 4.7. Supercooled Liquid Fraction

Climate models struggle at representing mixed-phase clouds, and glaciate supercooled liquid clouds too early, corresponding with too low supercooled liquid fractions (e.g., Cesana et al., 2015; Dietllicher et al., 2019; Komurcu et al., 2014; Lohmann & Neubauer, 2018). Figure 12 compares the supercooled liquid fractions (SLF) of clouds at mixed-phase temperatures from all simulations described in Table 1. We calculate the SLF as the fraction of cloud liquid water to cloud liquid water and ice for all grid boxes in the temperature range $T \pm 2.5$ K and weight each SLF by the model level thicknesses to account for cases of more than one grid box per column falling into this temperature range as is done for the satellite data (Komurcu et al., 2014). Every simulation in Figure 12 has a higher SLF than REF. REF-ConvCirrus has a higher SLF likely due to more detrained cloud liquid water as a result of less convective precipitation. Further, REF-ConvCirrus has a faster aggregation rate of ice crystals to snow, which decreases cloud ice and increases SLF. The SLF increases in REF-ConvCirrus-DetrLiquid as detrained condensate is always in the liquid phase at mixed phase temperatures. Switching to the new cloud scheme increases the SLF (Figure 12), which might be explained by the absence of a threshold parameterization of the Wegener-Bergeron-Findeisen process. Further, all three simulations with a fast aggregation rate have a very high SLF that is consistent with very low in-cloud ICNCs at mixed-phase temperatures (Figure 6). In NEW-ReducedDetrICNC, the SLF decreases as the aggregation rate of all ice crystals is no longer enhanced, but the convectively detrained ICNCs are reduced to account for immediate aggregation. Thus, the aggregation rate in mixed-phase clouds is much smaller, resulting in higher ICNC (Figure 6). The SLFs of NEW and REF-ConvCirrus are comparable and both agree well with the observations. Thus, the SLF depends on the aggregation rate enhancement factor, the phase of convective detrainment, the parameterization of the Wegener-Bergeron-Findeisen process, and the cloud fraction formulation.
4.8. Cloud Type Contributions to the Cloud Radiative Effects

Figure 13 shows the contributions of liquid clouds with $T > 273$ K (warm clouds), supercooled liquid clouds at $T < 273$ K, ice in mixed-phase clouds at $T > 235$ K, and cirrus clouds (ice at $T < 235$ K) to the cloud radiative effects (CRE) in REF and NEW. The CRE contributions were diagnosed by two calls to the radiation

![Figure 13](image-url)

**Figure 12.** Annual global mean supercooled liquid fractions for all simulations from Table 1 for the years 2003–2012 and from CALIPO observations (Komurcu et al., 2014).

**Figure 13.** Annual zonal-mean shortwave and longwave cloud radiative effects (CRE) for liquid clouds with $T > 273$ K (warm clouds), cloud water at $T < 273$ K (supercooled liquid clouds), cloud ice at $T > 235$ K (mixed-phase ice clouds), and ice clouds at $T < 235$ K (cirrus clouds) in REF (a) and NEW (b), and regional distribution of the net cloud radiative effect contributions in NEW (c). The gray area represents the difference between adding up the individual contributions and the total CRE.
subroutine: One call is the normal radiative transfer calculation including all clouds, and the other is a radiative transfer calculation where one of the above-mentioned cloud water or cloud ice contributions was removed. As there can be shielding effects between the cloud types, adding up the individual components leads to a lower CRE than the total CRE (gray area in Figure 13). When comparing the CRE contributions between REF and NEW, we observe that warm clouds contribute less to the ITCZ peak in simulation NEW. This could be explained by the reduced Eastern Pacific ITCZ low-level cloud cover in simulation NEW (Figure 7). Further, in simulation NEW the supercooled liquid clouds have stronger CREs especially in the extra-tropical storm tracks, and the mixed-phase ice clouds have weaker CREs. This is consistent with the increased supercooled liquid fraction in simulation NEW (Figure 12). Finally, the cirrus cloud CRE has decreased in the extra-tropics likely due to the lower cirrus cloud cover in simulation NEW (Figure 7).

Figure 13 further shows the regional distribution of the net CRE of the different cloud types for the NEW simulation. The cirrus net CRE correlates well with the high-level cloud cover (Figure 7) and is strongest in the ITCZ. The net CRE of mixed-phase ice clouds is very low. The net CRE of supercooled liquid clouds is strongest in the southern hemispheric storm tracks. Finally, warm clouds have the strongest cooling effect in the marine stratocumulus areas.

5. Summary

In this study, we present a new cloud scheme for ECHAM-HAM that includes a prognostic cloud fraction and two-moment microphysics. The design of the cloud scheme follows the concept by Tiedtke (1993) and includes a prognostic variable for cloudy air from which the cloud fraction is directly diagnosed. The model is designed to allow for subsaturation and supersaturation with respect to ice in the cloud-free and the cloudy part of the grid box by including another prognostic variable for in-cloud water vapor following Kärcher and Burkhardt (2008). Clouds can form as a result of convective activity and turbulent vertical diffusion, for which the mass fluxes and updraft velocities of the different parameterizations are used to calculate the respective cloud fraction and cloud droplet activation/ice crystal nucleation. Further, large-scale cloud formation is described by assuming a Gaussian water vapor distribution in the cloud-free part of the grid box. Next, a new parameterization to calculate the convectively detrained ICNC was implemented that depends on the convective updraft velocity instead of an ice crystal size climatology. In the cloud microphysics scheme, there is no saturation adjustment as all condensation/deposition rates can be calculated explicitly due to the additional prognostic variable for the in-cloud water vapor. Further, ice crystals can grow by aggregation and riming with cloud droplets and remain in the ice crystal category. They are only converted to the snow category when they grow to radii beyond 200 μm or when they exceed this size due to aggregation.

The new cloud scheme leads to a reduction and better representation of cirrus cloud fraction. In-cloud supersaturation is simulated frequently for cirrus clouds. Further, the new cloud scheme leads to a larger IWP and a smaller ICNC, both leading to better agreement with observations. Ice crystal formation is dominated by the liquid-origin pathways of convective detrainment and homogeneous freezing of cloud droplets. However, in situ ice crystal formation might be underestimated due to missing parameterizations of mesoscale gravity waves and subgrid scale orographic cirrus formation. In situ heterogeneous ice nucleation dominates over homogeneous nucleation only in the northern hemisphere where dust aerosols are more abundant. The new cloud scheme allows for larger supercooled liquid fractions in mixed-phase clouds that are closer to observations. Accordingly, the cloud radiative effect of supercooled liquid clouds increases, and the cloud radiative effect of ice in mixed-phase clouds decreases.

We also showed that the simulated ice crystals are strongly affected by tuning choices. Namely, the enhancement of the aggregation rate of all ice crystals and the aggregation of ice crystals detrained from convective clouds have a strong effect on the simulated ice clouds. As these tuning choices can only be validated indirectly, the simulation of ice clouds still requires further investigation in future research. In conclusion, the new cloud scheme with prognostic cloud fraction allows for a more physical simulation of ice clouds.
Appendix A.

A.1 Condensation and Deposition Rates

All condensation and deposition rates in this study are calculated according to Lohmann et al. (2016)

\[
\frac{dm_w}{dt} = 4\pi r L_v \left( \frac{L_v}{R_v T} - 1 \right) + \frac{1}{D_v \rho_v s} \frac{S_w - 1}{K T L_v R_v T - 1} + \frac{1}{C_{18}/C_{19}} \frac{D_v \rho_v}{C_{18}/C_{19}} \frac{dm_i}{dt} = \alpha 4\pi C_{Si} \frac{S_i - 1}{L_s \rho_v s} \left( \frac{L_s}{R_v T} - 1 \right) + \frac{R_v T \rho_v s}{C_{18}/C_{19}} + \frac{R_v T \rho_v s}{C_{18}/C_{19}} \frac{dm_i}{dt} \]

(A1)

with the mass of one cloud droplet \(m_w\), the radius of the cloud droplet \(r\), the saturation ratio with respect to water \(S_w\), the latent heat of vaporization \(L_v\), the thermal conductivity coefficient \(K\), the temperature \(T\), the specific gas constant of water vapor \(R_v\), the water vapor diffusion coefficient in air \(D_v\), the saturation water vapor density \(\rho_v\), the mass of one ice crystal \(m_i\), the deposition coefficient \(\alpha = 0.5\), the capacitance of the ice crystal for which we assume plate-type crystals with \(C = \frac{2r}{\pi}\), the saturation ratio w.r.t. ice \(S_i\), the latent heat of sublimation \(L_s\), and the saturation vapor pressure w.r.t. ice \(e_{i,s}\).

A.2 Formulating Condensation and Deposition Rates as Equivalent Downdraft Velocities

The change of the saturation ratio over water/ice during an adiabatic air parcel ascent can be described based on the updraft velocity \(w\) and the condensation/deposition rate on cloud droplets/ice crystals in the air parcel \(\frac{dm_w}{dt}\) (Pruppacher & Klett, 2010)

\[
\frac{dS_w}{dt} = \left( \frac{\varepsilon L_v}{R_a T^2 c_p} - \frac{g}{R_a T} \right) w = \left( \frac{1}{q_{v,sat,w/i}} + \frac{\varepsilon L_s^2}{R_a T^2 c_p} \right) \frac{dm_w}{dt} \]

(A2)

with the latent heat of vaporization/sublimation \(L_{v/s}\), the heat capacity of air \(c_p\), \(\varepsilon = M_w/M_a = 0.62\), and the specific gas constant of air \(R_a\). Introducing the temperature and pressure-dependent variables \(\alpha\) and \(\beta\) allows us to reformulate the condensation/deposition rate to an equivalent downdraft velocity \(w_{cond/depo}\) (Kärcher et al., 2006)

\[
\frac{dS_w}{dt} = \alpha w - \beta \frac{dm_w}{dt} = \alpha \left( w - \frac{\beta \frac{dm_w}{dt}}{\alpha} \right) = \alpha(w - w_{cond/depo}) \]

(A3)

A.3 Ice Crystal Sedimentation Velocity, Sticking Coefficient, Aggregation, and Riming

The ice crystal sedimentation velocity \(v_{sedi}\) is calculated from the mass of one ice crystal \(m\) after Spichtinger and Gierens (2009a)

\[
v_{sedi} = \alpha m^\beta \cdot \left( \frac{p}{30000} \right)^{-0.178} \left( \frac{T}{233} \right)^{-0.394}
\]

\[
\alpha = 63292.4 \quad \beta = 0.573 \quad \text{for } m < m_1 \\
\alpha = 329.7 \quad \beta = 0.309 \quad \text{for } m_1 < m < m_2 \quad \text{and} \quad m_1 = 2.166 \cdot 10^{-9} \text{ kg} \\
\alpha = 8.78 \quad \beta = 0.0954 \quad \text{for } m_2 < m \quad \text{and} \quad m_2 = 4.26 \cdot 10^{-8} \text{ kg}
\]

For the calculation of self-aggregation of ice crystals, we follow Levkov et al. (1992) for the formulation of the sticking coefficient \(E_{ii}\)
\[ E_{ii} = 0.025 \cdot (T - 273.15) \tag{A5} \]

Following Murakami (1990), the aggregation rate of ice crystals is
\[ \frac{\partial N_i}{\partial t}_{\text{agg}} = 0.5 \gamma_{\text{agg}} \frac{\pi}{6} D_i^{2} \text{eff} \nu_{\text{sedi}} E_{iw} X N_i^2 \tag{A6} \]

with assuming a fall velocity dispersion of \( X = 10\% \), the effective ice crystal diameter \( D_{\text{eff}} \) and the aggregation enhancement factor \( \gamma_{\text{agg}} \). The riming rate is calculated from the cloud water mass mixing ratio in the ice crystal fall volume
\[ \frac{\partial q_i}{\partial t}_{\text{rim}} = E_{iw} \frac{\pi}{6} r_i^{2} \text{eff} \nu_{\text{sedi}} q_w N_i \tag{A7} \]

with the collection efficiency of ice crystals and cloud droplets \( E_{iw} \) after Mitchell (1990).

**A.4 Rain and Snow Formation**

The autoconversion rate of cloud droplets to rain is calculated after Khairoutdinov and Kogan (2000)
\[ \frac{\partial q_w}{\partial t}_{\text{rain}} = -\gamma_{\text{auto}} 1350 \left( 10^{-6} N_w \right)^{-1.79} q_w^{2.47} \tag{A8} \]

with the autoconversion enhancement factor \( \gamma_{\text{auto}} \). The rate of ice crystals aggregating to snowflakes is calculated after Murakami (1990) based on how fast ice crystals aggregate to snowflake size:
\[ \frac{\partial q_i}{\partial t}_{\text{snow}} = -\gamma_{\text{agg}} q_i N_i \frac{\pi}{6} D_i^{2} \nu_{\text{sedi}} E_{iw} \tag{A9} \]

with the same values for \( X \) and \( E_{ii} \) as in Equation A6.

**A.5 Interaction of Rain and Snow While Falling Through a Level**

We can calculate the rain drops size \( r_{\text{rain}} \) from the mass and concentration fluxes assuming spherical rain drops. We then calculate the fall velocity after Lohmann et al. (2016)
\[ v_{\text{rain}} = \min \left( \begin{array}{c}
1.2 \cdot 10^6 \frac{1}{\text{m/s}} r_{\text{rain}}^2 \text{for } r_{\text{rain}} < 30 \mu m \\
8000 \frac{1}{\text{s}} \text{for } 30 \mu m < r_{\text{rain}} < 600 \mu m \\
201 \frac{\text{m}}{\text{s}} \sqrt{r_{\text{rain}}/600} \mu m < r_{\text{rain}}, 9 \text{ m/s}
\end{array} \right) \tag{A10} \]

For snowflakes we keep the formulation of ECHAM-HAM for the effective diameter of the snowflakes:
\[ D_{\text{snow}} = \max \left( 20 \mu m, 0.01 \sqrt{\frac{1000 F_{\text{snow, max}}}{3.8 \cdot 10^{-4} F_{\text{snow, conc}}}} \right) \tag{A11} \]

and calculate the fall velocity of snow \( v_{\text{snow}} \) with the same equation as for ice crystal sedimentation \( v_{\text{sedi}} \). Accretion of cloud droplets by rain drops is based on the volume the rain drops fall through
\[ \frac{\partial q_w}{\partial t}_{\text{acc, rain}} = -\pi r_{\text{rain}}^{2} F_{\text{rain, conc}} q_w \tag{A12} \]

assuming maximum overlap. Similarly riming of cloud droplets and aggregation of ice crystals by snowflakes is calculated as
\[ \frac{\partial q_{w/i}}{\partial t}_{\text{rim/agg - snow}} = -\gamma_{\text{rim/agg}} \pi_{\text{snow}}^{2} F_{\text{snow, conc}} q_{w/i} E_{w/i}^{1/2} \] 

(A13)

with the same collection efficiencies \( E_{w/i} \) as in Appendix A.3.

**A.6 Figures**

Figure A1 compares the simulated in-cloud ICNC to observations, and Figure A2 compares the simulated in-cloud relative humidities to observations.
Data Availability Statement
The ECHAM-HAMMOZ model is made freely available to the scientific community under the HAMMOZ Software License Agreement, which defines the conditions under which the model can be used. More information can be found at the HAMMOZ website (https://redmine.hammoz.ethz.ch/projects/hammoz). All model data for the figures can be found at https://doi.org/10.5281/zenodo.3587985.

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Figure A2. In-cloud relative humidity from aircraft observations (Krämer et al., 2009; updated values according to Krämer, Afchine, et al., 2016) between 25°S and 75°N from 1999–2014 (a) and in-cloud relative humidity from the NEW simulation sorted in temperature bins online during the simulation of 2008 for all model grid boxes (b).

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