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Empirical values and assumptions in the convection of numerical models

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Abstract. Convection influences climate and weather events over a wide range of spatial and temporal scales. Therefore, accurate predictions of the time and location of convection and its development into severe weather are of great importance. Convection has to be parameterized in Numerical Weather Prediction models, Global Climate Models, and Earth System Models (NWPs, GCMs, and ESMs) as the key physical processes occur at scales much lower than the model grid size. The convection schemes described in the literature represent the physics by simplified models that require assumptions about the processes and the use of a number of parameters based on empirical values. The present paper examines these choices and their impacts on model outputs and emphasizes the importance of observations to improve our current understanding of the physics of convection.

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Table 1. List of acronyms.

Acronym	Meaning
AM4.0	Atmospheric Model version 4
AOT	Aerosol Optical Thickness
ARM	Atmospheric Radiation Measurement
ARW	Advanced Research WRF
AS	Arakawa-Schubert scheme
ATEX	Atlantic Trade-Wind Experiment
BCL	Buoyant Condensation Level
BMJ	Betts-Miller-Janjić
BOMEX	Barbados Oceanographic and Meteorological EXperiment
CAM	Community Atmosphere Model
CAPE	Convective Available Potential Energy
CCM3	Community Climate Model version 3
CCN	Cloud Condensation Nuclei
CCSM	Community Climate System Model
CDNC	Cloud Droplet Number Concentration
CESM	Community Earth System Model
CFSv2	Climate Forecast System version 2
CIN	Convective Inhibition
CISK	Conditional Instability of the Second Kind
COARE	Coupled Ocean-Atmosphere Response Experiment
CP	Cumulus Parameterization
CRCP	Cloud Resolving Convective Parameterization
CRM	Cloud Resolving Model
CSRM	Cloud System Resolving Model
CWF	Cloud Work Function
DBL	Downdraft Base Layer
dCAPE	Dynamic Convective Available Potential Energy
DDL	Downdraft Detrainment Level
ECHAM	General circulation model developed by the Max Planck Institute for Meteorology
ECMWF	European Centre for Medium-Range Forecasts
EL	Equilibrium Level
ENSO	El Niño-Southern Oscillation
ESM	Earth System Model
GARP	Global Atmospheric Research Program
GATE	GARP Atlantic Tropical Experiment
GCM	Global Circulation/Climate Model
GEOS-5	Goddard Earth Observing System, Version 5 model
GFDL	Geophysical Fluid Dynamics Laboratory
GFS	Global Forecast System
GISS GCM	Goddard Institute for Space Studies Global Climate Model

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Acronym	Meaning
GOAmazon	Green Ocean Amazon field campaign
HadGEM3 GA2.0	Hadley Centre Global Environmental model Global Atmosphere version 2
HCF	Heated Condensation Framework
HWRF	Hurricane Weather Research and Forecasting model
IFS	Integrated Forecasting System
IN	Ice Nuclei
IOP	Intensive Observation Period
ITCZ	Intertropical Convergence Zone
KF	Kain-Fritsch scheme
LBN	Level of Neutral Buoyancy
LCL	Lifting Condensation Level
LES	Large Eddy Simulation
LFC	Level of Free Convection
LFS	Level of Free Sinking
LWC	Liquid Water Content
MJO	Madden-Julian Oscillation
MM5	Mesoscale Model version 5
MMF	Multiscale Model Framework
MP	Microphysics Parameterization
NAM	North American Mesoscale model
NCAR	National Center for Atmospheric Research
NCEP	National Centers for Environmental Prediction
NWP	Numerical Weather Prediction
PBL	Planetary Boundary Layer
PCAPE	Integral over pressure of the buoyancy of an entraining ascending parcel with density scaling
PDF	Probability Density Function
PML	Potential Mixed Layer
QE	Quasi-Equilibrium
RACORO	Routine AAF (ARM Aerial Facility) CLOWD (Clouds with Low Optical Water Depths) Optical Radiative Observations
RAS	Relaxed Arakawa-Schubert scheme
RCM	Regional Climate Model
RH	Relative Humidity
RICO	Rain In Cumulus over the Ocean field campaign
SAS	Simplified Arakawa-Schubert scheme
SCM	Single Cloud Model
SGP97	Southern Great Plains 97
SNU	Seoul National University
SP	Super-Parameterization
SPCZ	South Pacific Convergence Zone
SST	Sea Surface Temperature

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Acronym	Meaning
TKE	Turbulent Kinetic Energy
TWP-ICE	Tropical Warm Pool – International Cloud Experiment
UIUC	University of Illinois, Urban–Champaign
UM	Unified Model
USL	Updraft Source Layer
WRF	Weather Research and Forecasting model

1 Introduction

Numerical Weather Prediction models, Global Climate Models, and Earth System Models (NWP, GCMs, and ESMs) generate precipitation through two parameterizations: microphysics of precipitation (MP hereafter) and cumulus parameterization (CP) schemes. They produce what is known as large-scale precipitation and convective precipitation, respectively. While other schemes, such as planetary boundary layer (PBL) parameterization, also affect precipitation occurrence, the especially intricate processes by which water vapor becomes cloud droplets or ice crystals and then liquid or solid precipitation are intended to be modeled by the two former modules.

The empirical values and assumptions embedded in the MP were explored in Tapiador et al. (2019a). In the present paper, attention is focused on the CP, aiming to provide a comprehensive account of the empirical choices and assumptions behind the representation of convective precipitation in models. Indeed, precipitation is the most important component of the water cycle. Extreme hydrological events in the form of floods are responsible for the loss of thousands of lives every year and great damage to property, while droughts affect water resources, livestock, and crop production. Both extremes represent important threats for human life and developing economies (Trenberth, 2011; Pham-Duc et al., 2020). Changes in the hydrological cycle also affect human activities such as the production of electricity in hydropower plants, where a better optimization of electricity production depends on water input (García-Morales and Dubus, 2007; Tapiador et al., 2011).

Precipitation is also a key environmental parameter for biota. The types of vegetation and animal life that exist in a certain area are conditioned by temperature but even more by precipitation. Changes in the precipitation regime alter plant growth and survival and consequently impact the food chain (McLaughlin et al., 2002; Choat et al., 2012; Barros et al., 2014; Deguines et al., 2017). Prolonged droughts may increase the risk of wildfires, with the associated loss of local species (Holden et al., 2018). Therefore, it is not surprising that providing an accurate representation of precipitation in models is an active research topic. Specifically, in the climate realm it is already known that the effects of climate change will strongly modify the distribution and variability of precipitation around the world (Easterling et al., 2000; Dore, 2005; Giorgi and Lionello, 2008; Trenberth, 2011), posing many risks to life and human activities (Patz et al., 2005; McGranahan et al., 2007; IPCC, 2014; Woetzel et al., 2020). Thus, it is important to provide an explicit account of how models produce rain and snow in order to fully understand the outputs of the simulations.

The paper is organized as follows. A brief note on model parameterization, tuning, and the importance of convection follows (Sect. 1.1 and 1.2). Then, the main strategies to model cumulus convection are briefly presented to provide the framework to the rest of the paper (Sect. 2). The core of the review is in the following three sections, which present the assumptions and empirical values in the trigger (Sect. 3), the cloud model (Sect. 4) and the closure of the scheme (Sect. 5). The paper concludes with notes and considerations on the topic, bringing together the most important results. The acronyms used through the paper may be found in Table 1.

1.1 Model parameterizations

Parameterizations in numerical models address the fact that some significant physical processes in nature occur at scales much lower than the grid size used in models (Arakawa and Schubert, 1974; Stensrud, 2007; McFarlane, 2011). That is the case of convection, where spatial resolutions of at least 100 m are required to realistically solve its dynamics (Bryan et al., 2003). However, typical horizontal grid resolutions in current models range from a kilometer scale for high resolution NWP applied to a particular area, to dozens of kilometers in global NWPs, GCMs, and ESMs. With these model grids, convection is a subgrid-scale process not explicitly resolved. The physics is represented by a simplified model that requires assumptions about the processes and the use of several parameters based on empirical values. These are used as thresholds, constraints, or mean values of a number of processes, whereas the former simplification requires a compromise between reducing complexity and a fair representation of the atmosphere.

While sometimes neglected and seldom explicit, tuning is an integral procedure of modeling (Hourdin et al., 2017; Schmidt et al., 2017; Tapiador et al., 2019a, b). It consists of estimating sensible values for the empirical parameters to reduce the discrepancies between model outputs and observations. An example of these discrepancies is shown in Fig. 1. Hence, tuning may have a significant influence on model results and can help identify the parts of the model that need further attention. However, blind tuning can mask fundamental problems within the parameterization, leading to non-realistic physical states of the system, compensating for errors that translate into an inappropriate budget equilibrium, or affect other metrics (Tapiador et al., 2019a). This is particularly important for climate models, since projections and simulations of future climates always include the *ceteris paribus* assumption (Smith, 2002). Indeed, parameters that work well for the present climate may not do so for the future. Understanding the range of validity of the choices and the logical steps for the selections can help produce stronger and more robust simulations.

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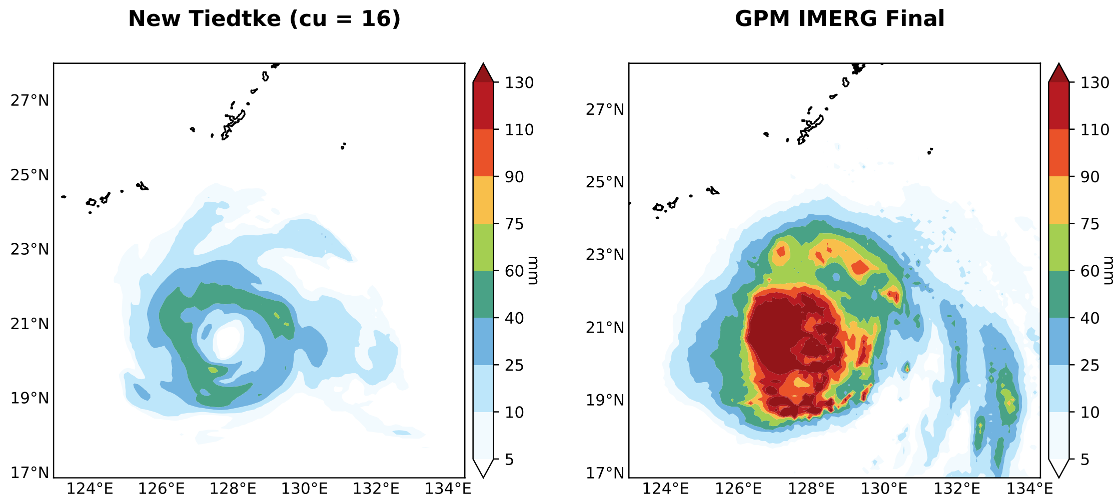

Figure 1. Comparison between simulated 6-hour accumulated surface liquid precipitation with the New Tiedtke convection parameterization in the WRF model (left) and GPM IMERG Final run (right) for Typhoon Chaba on 2016/09/25 from 18.00 UTC. The accumulated precipitation includes cumulus, shallow cumulus and grid scale rain. The domain is located over the Philippine Sea with a horizontal grid size of 10 kilometers.

1.2 Convection: a key process in models

There is a wide range of recent research topics in convection. These topics include machine learning to parameterize moist convection (Gentine et al., 2018; O'Gorman and Dwyer, 2018; Rasp et al., 2018); stochastic parameterizations of deep convection (Buizza et al., 1999; Majda et al., 1999, 2001; Majda and Khouider, 2002; Khouider et al., 2003; Majda et al., 2003; Shutts, 2005; Plant and Craig, 2008; Dorrestijn et al., 2013; Khouider, 2014; Wang et al., 2016); the use of convective parameterization on "gray scales" (Kuell et al., 2007; Gerard et al., 2009; Mironov, 2009); aerosols and their influence on convection (Heever and Cotton, 2007; Storer et al., 2010; Heever et al., 2011; Morrison and Grabowski, 2013; Grell and Freitas, 2014; Kawecki et al., 2016; Peng et al., 2016; Han et al., 2017; Grabowski, 2018); microphysics impacts (Grabowski, 2015); impact of new cumulus entrainment (Chikira and Sugiyama, 2010; Lu and Ren, 2016); orographic effects on convection (Panosetti et al., 2016); new mass flux formulations (Han et al., 2017); large eddy simulations (LES) (Siebesma and Cuijpers, 1995; Brown et al., 2002; De Rooy and Siebesma, 2008; Heus and Jonker, 2008; Neggers et al., 2009; Dawe and Austin, 2013) and scale-aware cumulus parameterization (Wagner et al., 2018).

Such a wealth of papers illustrates the strength of this research topic in a vast number of fields. Of these, developing parameterization schemes for models is a thriving subfield, with several teams advancing the field (see Sect. 2 below). Difficulties persist, however. Convective processes have been identified as a major source of uncertainty in the latest decadal survey (National Academies of Sciences, Engineering and Medicine, 2018), and dedicated efforts are needed to fill the gaps in our present knowledge of the processes involved.

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Owing to the influence of convection on climate and weather events over a large range of spatial and temporal scales, one of the most important objectives of the latest decadal survey is to improve the predictions of the timing and location of convective storms, and their evolution into severe weather. Besides the drawbacks associated with the spatial resolution, the multiscale interactions leading to the organization and evolution of convective systems are difficult to observe and represent.

Improving the observed and modeled representation of natural, low-frequency modes of weather/climate variability was identified in the survey as one of the most important challenges of the coming decade. Including interactions between large-scale circulation and organization of convection such as Madden–Julian Oscillation (MJO) or El Niño–Southern Oscillation (ENSO) aims to improve predictions by 50 % at lead times of 1 week to 2 months, which will have a high societal impact.

It is essential to further understand the physics and dynamics of the underlying processes, currently crudely parameterized in the majority of models. Advanced observations of atmospheric convection and high-resolution models are also needed. While models will likely increase their nominal resolution in the next decade, it is also likely that global, century-long simulations from multi-ensembles under different assumptions will need to resort to parameterizing the most computing-intensive tasks.

2 Overview of the main schemes in cumulus convection modeling

Soon after Charney and Eliassen (1964), and Ooyama (1964) introduced the idea of cumulus parameterization, two approaches emerged: the convergence and the adjustment schemes (Arakawa, 2004). Later, a new scheme was introduced by Ooyama (1971): mass-flux parameterization. Despite all these schemes attempting to explain the interaction between cumulus clouds and the large-scale environment, the choice of empirical values for certain parameters and the simplifications in the physics yield different convective parameterizations and strategies. Indeed, as shown in Fig. 2 for the total accumulated precipitation, even today model outputs look different depending on the cumulus parameterization used.

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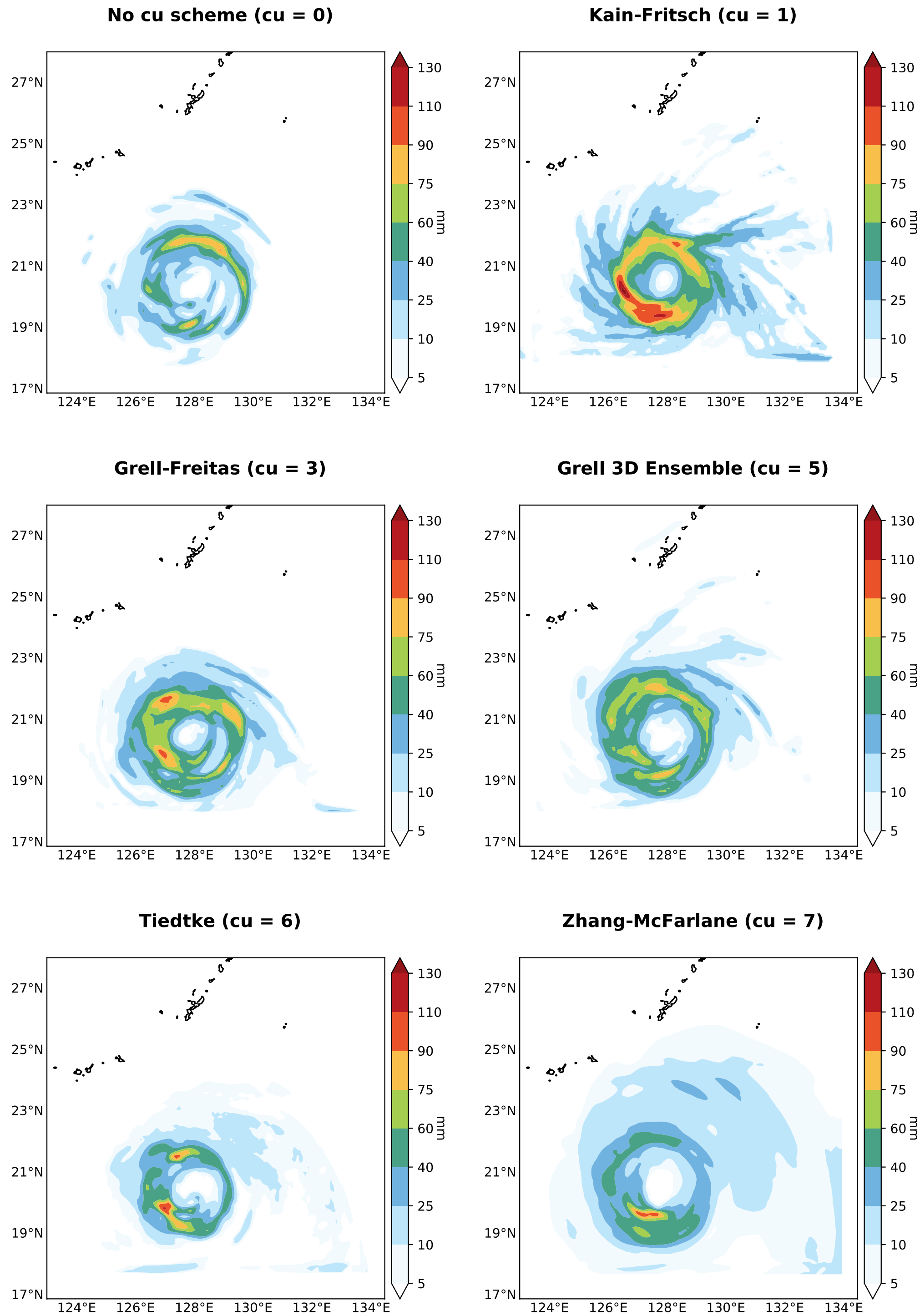

Figure 2. Simulated 6-hour accumulated surface liquid precipitation for Typhoon Chaba without using a CP (upper left) and using five different CPs in the WRF model. The accumulated precipitation includes cumulus, shallow cumulus, and grid scale rain. The simulations start on 2016/09/25 at 18.00 UTC. The domain is located over the Philippine Sea with a horizontal grid size of 10 kilometers.

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The main assumptions in convective parameterizations concern the trigger model, the representation of the mutual interaction between cumulus clouds and the large-scale environment (cloud model), and the closure of the scheme.

As of 2020, the main cumulus convection schemes publicly available for NWPs are convergence schemes, adjustment schemes, mass flux schemes, cloud system resolving models (CSRM), and super-parameterization (SP). The purpose of this paper is not to compare the performances of the schemes but to investigate their empirical values and assumptions so the focus on the following section is on these.

2.1 Convergence schemes: the key role of the total moisture convergence parameter

Convergence schemes consider that synoptic scale convergence destabilizes the atmosphere, while the heat released through condensation in cumulus clouds stabilizes it. Typical examples of this approach are Charney and Eliassen (1964), Ooyama (1964) and Kuo (1974). Charney and Eliassen (1964) did not use cloud models to explain these interactions. Instead, the concept of conditional instability of the second kind (CISK) was introduced. Ooyama (1964) used a similar formulation, but represented the heating released through condensation in cumulus clouds in terms of a mass flux and considered the entrainment of ambient air. Kuo (1965, 1974) used a simple cloud model scheme to describe the interaction between a large-scale environment and cumulus clouds. One of the key assumptions in this scheme is that the total moisture convergence can be divided into a fraction b, which is stored in the atmosphere, and the remaining fraction $(1 - b)$, which precipitates and heats the atmosphere. This parameter was further modified by Anthes (1977), who proposed a relationship between b and the mean relative humidity (RH) in the troposphere, with $b \leq 1$. In the evaluation of rainfall rates using the Global Atmospheric Research Program Atlantic Tropical Experiment (GATE) scale phase III, Krishnamurti et al. (1980) obtained the most realistic precipitation rates for $b \approx 0$. In a later paper, Krishnamurti et al. (1983) introduced an additional subgrid-scale moisture supply to account for the observed vertical distributions of heat and moisture. The total moisture supply was expressed as $I = (1 + \eta)I_L$, with I_L the large-scale moisture supply. The authors used a multiple regression approach to find the values of b and η. Another approach consists of using the wet-bulb characteristics to locally determine the partition between precipitation and moistening (Geleyn, 1985).

Due to its formulation, the Kuo scheme cannot produce a realistic moistening of the atmosphere and cannot represent shallow convection. Moreover, it assumes that convection consumes water and not energy, which violates causality (Raymond and Emanuel, 1993; Emanuel, 1994). Despite these drawbacks, it can produce acceptable results in various applications (Kuo and Anthes, 1984; Molinari, 1985; Pezzi et al., 2008), such as in GCMs and NWP models (Rocha and Caetano, 2010; Mbienda et al., 2017). The convective parameterization scheme demands the least computational power and is thus sometimes used for large, centennial simulations.

2.2 Adjustment schemes: two strategies to remove instability

In adjustment schemes, the atmospheric instability is removed through an adjustment towards a reference state. Therefore, the physical properties of clouds are implicit and no cloud models are needed. The first proposed adjustment scheme was the moist

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convective adjustment by Manabe et al. (1965), also known as the hard adjustment. In this parameterization, moist convection occurs if the air is supersaturated and conditionally unstable. The instability is removed through an instantaneous adjustment of the temperature to a moist-adiabatic lapse rate, and of water vapor mixing ratio to saturation. Moreover, all the condensed water in this process precipitates immediately. The main problems of this scheme are the production of very large precipitation rates, and its saturated final state after convection, which is rarely observed in nature.

The so-called soft or relaxed adjustment schemes attempt to alleviate these problems by assuming that the hard adjustment occurs only over a fraction α of the grid area, or by specifying the final mean RH (Cotton and Anthes, 1992). For example, Miyakoda et al. (1969) defined saturation as 80 % RH, while Kurihara (1973) performed the adjustment based on the buoyancy condition of a hypothetical cloud element instead of the saturation criterion.

Further improvements to the adjustment schemes were introduced by Betts and Miller (1986), whose scheme is also known as a penetrative adjustment scheme. The authors proposed an adjustment of large-scale atmospheric temperature and humidity to reference profiles over a specified time scale τ (adjustment timescale). The reference profiles, different for shallow and deep convection, are quasi-equilibrium states based on observational data from GATE, Barbados Oceanographic and Meteorological Experiment (BOMEX), and Atlantic Trade-Wind EXperiment (ATEX). For the construction of the temperature reference profile, Betts (1986) used a mixing line model (Betts, 1982, 1985). Then, the moisture reference profile was calculated from the temperature profile by specifying the pressure difference between air parcel saturation level and pressure level at cloud base, freezing level, and cloud top. Therefore, the three adjustment parameters used in this scheme are the adjustment timescale τ, the stability weight W_s, and the saturation pressure departure, S_p.

The sensitivity of the scheme to the adjustment parameters has been evaluated by numerous authors. For instance, Baik et al. (1990) analyzed the influence of different values of each adjustment parameter on the simulation of a tropical cyclone, while Vaidya and Singh (1997) did the same for the simulation of a monsoon depression using four sets of values, including those from Betts and Miller (1986) and Slingo et al. (1994). In all cases, the adjustment parameters had to be modified depending on the different climate regimes. While Baik et al. (1990) set $W_s = 0.95$ and $S_p = (-30, -37.5, -38)$ hPa as the optimal parameters to simulate a tropical cyclone, Vaidya and Singh (1997) obtained the best forecast for a monsoon depression with $W_s = 1.0$ and $S_p = (-60, -70, -50)$ hPa. Despite the improvements achieved through adjusting the parameters for different climate conditions, the original Betts-Miller scheme occasionally produced heavy spurious rainfall over warm water and light precipitation over oceanic regions (Janjić, 1994). To overcome this problem, Janjić (1994) proposed considering a range of reference equilibrium states, and characterizing the convective regimes by a parameter called "cloud efficiency", which is related to precipitation production and depends on cloud entropy. This parameter is the sort of empirical value that requires attention when future climates are to be simulated. The modified scheme, known as the Betts-Miller-Janjić (BMJ) scheme, is one of the most widely used adjustment schemes in NWP models (Vaidya and Singh, 2000; Evans et al., 2012; Fiori et al., 2014; Fonseca et al., 2015; García-Ortega et al., 2017), despite its large bias for light rainfall (Gallus and Segal, 2001; Jankov and Gallus, 2004; Jankov et al., 2005). Convective adjustment schemes are computationally efficient, which makes them suitable for large-scale simulations.

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2.3 Mass flux schemes: assuming the rates of mass detrainment and entrainment

Because of the nature of both convergence and adjustment schemes, a cloud model is not needed to describe the interaction between cumulus clouds and the large-scale environment. This is not the case for the mass-flux schemes, where convective instability is removed through the vertical transport of heat, moisture, and momentum. The first formulation of this type was introduced by Ooyama (1971). The author assumed that cumulus clouds of different sizes coexist, and that they could be represented by an ensemble of independent non-interacting buoyant elements. The definition of the so-called dispatcher function would close the parameterization. However, the author left this question open. Yanai et al. (1973) and Arakawa and Schubert (1974), hereafter AS, considered that an ensemble of cumulus clouds in a large-scale system is confined to an area σ that is large enough to contain the ensemble, and small enough compared to the large-scale system. The equations of mass continuity, heat, and moisture continuity are

$$\begin{aligned} &\frac{\partial \bar{s}}{\partial t} + \nabla \cdot \overline{sv} + \frac{\partial \overline{s\omega}}{\partial p} = Q_R + L(\bar{c} - \bar{e}), \\ &\frac{\partial \bar{q}}{\partial t} + \nabla \cdot \overline{qv} + \frac{\partial \overline{q\omega}}{\partial p} = \bar{e} - \bar{c}, \\ &\nabla \cdot \bar{v} + \frac{\partial \bar{\omega}}{\partial p} = 0 \,, \end{aligned} \tag{1}$$

where s is the dry static energy, v is the horizontal velocity, ω is the vertical velocity, Q_R the heating rate due to radiation, L the latent heat of vaporization, c the rate of condensation per unit mass of air, e the rate of evaporation of cloud water, q the water vapor mixing ratio, and the bar denotes horizontal averages over the hypothesized area. Using several assumptions, such as that mass exchange between cumulus cloud and the large-scale environment takes place through detrainment of cloud air D and entrainment of environmental air E, and following the analysis performed by Gregory and Miller (1989) (the reader is referred to Bechtold (2009) for a detailed explanation), the budget equations for a single entraining plume are

$$\begin{aligned} &\frac{\partial \sigma_i}{\partial t} + \delta_i - \varepsilon_i - \frac{\partial M_i}{\partial p} = 0, \\ &\frac{\partial \sigma_i s_i}{\partial t} + \delta_i s_{Di} - \varepsilon_i \bar{s} - \frac{\partial M_i s_i}{\partial p} = Lc_i \\ &\frac{\partial \sigma_i q_i}{\partial t} + \delta_i q_{Di} - \varepsilon_i \bar{q} - \frac{\partial M_i q_i}{\partial p} = -c_i \\ &\frac{\partial \sigma_i l_i}{\partial t} + \delta_i l_{Di} - \frac{\partial M_i l_i}{\partial p} = c_i - r_i \,, \end{aligned} \tag{2}$$

where δ and ε are the rates of mass detrainment and entrainment per unit pressure interval, M the cumulus mass flux, l the mixing ratio of liquid water, and r the rate of rainwater generation. Subscript i denotes the *ith* cumulus cloud, and subscript D the value in the detraining air.

Mass flux convective parameterization schemes still are the most common convective parameterizations used in ESMs, Regional Climate Models (RCMs), and NWP models.

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2.4 Cloud System Resolving Models (CSRM)

The performances of the previous schemes prompted the search for new strategies to model convection. Krueger (1988) put forward the CSRM idea (also known as the explicit convection, convection-permitting or cloud ensemble models) to explicitly simulate convective processes over a kilometer scale, instead of using parameterizations. Convective parameterizations tend to produce too little heavy rain and too much light rain (Kooperman et al., 2018), and have problems representing diurnal precipitation cycles over land (Pritchard et al., 2011). The use of convection-permitting models can solve errors associated with other convective parameterizations (Kendon et al., 2012; Prein et al., 2013; Brisson et al., 2016), but entails an extremely high computational cost, which limits its application in climate modeling (Wagner et al., 2018; Randall et al., 2019). However, it is also widely used in NWP (Kain et al., 2006; Gebhardt et al., 2011).

2.5 Super-Parameterization (SP)

Hybrid approaches also exist. SP (also known as cloud-resolving convective parameterization (CRCP) or multiscale model framework (MMF)) is an approach between parameterized and explicit convection, which consists of replacing the convective parameterizations by 2D cloud resolving models (CRMs), or even a 3D LES model, at each grid cell of a GCM (Grabowski and Smolarkiewicz, 1999; Grabowski, 2016). SP is mostly applied in GCMs (Grabowski, 2001; Khairoutdinov and Randall, 2003; Khairoutdinov et al., 2005; Zhu et al., 2009; Jung and Arakawa, 2014; Sun and Pritchard, 2016). Several studies have compared the performance of SP with convective parameterizations, in particular, using the Community Atmosphere Model (CAM).

Among the most notable improvements achieved by SP in CAM are simulations of heavy rainfall events that are much more similar to observations, a better diurnal precipitation cycle over land (Khairoutdinov et al., 2005; DeMott et al., 2007; Zhu et al., 2009; Holloway et al., 2012; Rosa and Collins, 2013), and the production of a realistic MJO (Thayer-Calder and Randall, 2009; Holloway et al., 2013). However, simulations with SP also have problems that need solving, such as the failure to simulate light rainfall rates reported by Zhu et al., (2009). The computational cost of this approach is also higher than the one for convective parameterizations (Krishnamurthy and Stan, 2015) but smaller than the computational cost for global CSRMs to perform climate simulations (Randall et al., 2003).

This paper considers all the aforementioned convective parameterizations with emphasis on the mass-flux schemes.

3 Trigger function: assumptions and empiricisms

In a CP, the accurate simulation of convection greatly depends on the trigger function. The trigger function has to determine whether convectively unstable air at the boundary layer leads to the onset of convection and if so, activate the CP.

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There are as many strategies to initiate convection as there are convection schemes. This section focuses on the assumptions and empirical values of the most important trigger functions, the starting levels, and the impacts of the trigger formulations on the simulation of convective processes. Table 2 lists the most common choices used in the main trigger function types.

Table 2: A sample of empirical values and assumptions used in the main trigger function types.

Empirical value or assumption	**Choices in the literature**	**Reference**
Large-scale moisture convergence	Yes	Kuo (1974); Anthes (1977); Tiedtke (1989)
CWF	Positive	Arakawa and Schubert (1974); Pan and Wu (1995); Han et al. (2019)
	Fixed value	Moorthi and Suarez (1992)
Large-scale vertical velocity ω	Controls δT to trigger convection	Fritsch and Chappell (1980); Kain and Fritsch (1990); Bechtold et al. (2001); Kain (2004); Ma and Tan (2009); Berg et al. (2013)
CAPE	At least some CAPE	Betts (1986); Betts and Miller (1986); Janjić (1994)
	Must be positive	Zhang and McFarlane (1995); Xie and Zhang (2000); Bechtold et al. (2004); Zhang and Mu (2005a); Wu (2012)
dCAPE	$dCAPE > 100\ \mathrm{J\ kg^{-1}}$	Xie and Zhang (2000); Zhang (2002); Song and Zhang (2009); Zhang and Song (2010)
	$dCAPE > 45\ \mathrm{J\ kg^{-1}\ h^{-1}}$	Song and Zhang (2018)
Stochastic	Stochastic perturbation in the large-scale vertical velocity ω in KF trigger	Bright and Mullen (2002)
	Markov process	Majda and Khouider (2002); Khouider et al. (2003); Stechmann and Neelin (2011)
	Bayesian Monte Carlo	Song et al. (2007)
	Adds a stochastic feature to the SAS trigger	Zhang et al. (2014)
	Adds a stochastic trigger to Emanuel (1991)	Rochetin et al. (2014a)
Dilute dCAPE	$dilute\ dCAPE > 70\ \mathrm{J\ kg^{-1}}$	Neale et al. (2008)
	$dilute\ dCAPE > 55\ \mathrm{J\ kg^{-1}\ h^{-1}}$	Song and Zhang (2017)
HCF	Yes	Tawfik and Dirmeyer (2014); Bombardi et al. (2015); Tawfik et al. (2017)

3.1 Trigger function types

According to the physical variable used as the main trigger condition, the most commonly used trigger functions in CPs may be classified into (1) moisture convergence, (2) cloud work function (CWF), (3) convective available potential energy (CAPE), and (4) large-scale vertical velocity. Other triggers used are (5) stochastic and (6) heated condensation framework (HCF) triggers. Table 3 lists the assumptions and empirical values used in the main trigger function types, which are discussed below.

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3.1.1 Moisture convergence trigger

The main condition to activate convection, together with the existence of a deep layer of conditional instability, is exceeding a minimum threshold value of the vertically integrated moisture convergence. This is the case in the Anthes-Kuo scheme (Kuo, 1965; Anthes, 1977) and in the original Tiedtke scheme (Tiedtke, 1989). The latter has undergone several modifications since its publication. For instance, Gregory et al. (2000) substituted the condition of positive buoyancy to activate deep convection by a minimum cloud depth threshold in the ECMWF convective parameterization. Zhang et al. (2011) proposed a modified version of the Tiedtke scheme with the aim of improving the representation of marine boundary layer clouds over the southeast Pacific. Among these modifications, deep convection is allowed to occur only when the vertically averaged relative humidity (RH) exceeds 80 %. A new modified Tiedtke scheme used in the Integrated Forecasting System (IFS) and in the Weather Research and Forecasting model (WRF) model uses the trigger criteria from Jakob and Siebesma (2003), and Bechtold et al. (2004), which include the search for unstable parcels within the lowest 300 hPa above the ground. The simulation of the diurnal cycle of precipitation using this new trigger and new entrainment rates improved in comparison to previous versions of IFS (Bechtold et al., 2004).

3.1.2 CWF trigger

The first CWF trigger was introduced by AS, who proposed that convection activation depends on a threshold value of the CWF, which is defined as the integral buoyancy force of each entraining cloud between cloud base and cloud top. Several variations of the original CWF trigger function have been suggested. In the relaxed Arakawa-Schubert scheme (RAS) (Moorthi and Suarez, 1992), the activation of convection depends on a critical value of the CWF, while the simplified Arakawa-Schubert scheme (SAS) (Pan and Wu, 1995) triggers convection if the CWF is positive, as shown in Table 2. Another condition to activate convection in SAS is based on the pressure difference between the starting point and the level of free convection (LFC), which defines a threshold value for the convection inhibition (CIN) factor. With the aim of decreasing convection in large-scale subsidence regions and increasing it in large-scale convergent regions, Han and Pan (2011) modified the limit to reach the LFC, which is now proportional to large-scale vertical velocity ω. Further improvements to the SAS activation criteria include a grid-spacing dependency in the convective trigger function (Lim et al., 2014), considering the spatial resolution dependency, and a new definition of the CIN threshold value applying a scale-aware factor (Kwon and Hong, 2017). Different versions of the AS scheme are currently used in the Global Forecast System (GFS) of the National Centers for Environmental Prediction (NCEP), the Mesoscale Model 5 (MM5), the Goddard Earth Observing System model version 5 (GEOS-5), the Geophysical Fluid Dynamics Laboratory (GFDL) model, and in the WRF model.

3.1.3 CAPE trigger

Many CPs have been proposed to simplify the formulation and implementation of the AS scheme. Among other assumptions, some CPs substitute the convection trigger based on CWF by CAPE, defined in a similar way as CWF but without including dilution of ascending parcel by entrainment. For instance, BMJ developed a new parameterization based on empirical results, in which the activation of convection requires the existence of CAPE. In this scheme, cloud base is the lifting condensation level (LCL) of a lifted parcel with the largest CAPE in the lowest 130 hPa of the model. From there, the parcel is lifted moist adiabatically until the equilibrium level (EL) is reached. In general, the cloud top is at the level immediately beneath EL. Moreover, deep convection continues if the cloud depth is greater than a certain value and covers at least two model layers (Baldwin et al., 2002). Finally, deep convection activates if the adjustment using reference profiles of temperature (based on a moist adiabat) and moisture (based on imposed sub-saturation at the cloud base) results in the column drying. The BMJ scheme is currently used in NCEP North American Mesoscale model (NAM), MM5, and WRF models. Another important convective parameterization also using a CAPE trigger is the Zhang-McFarlane scheme (Zhang and McFarlane, 1995). To improve climate simulations in the Canadian Climate Center GCM, the authors proposed a simplified version of the AS scheme that includes a positive CAPE trigger. However, it initiates convection too often during the day, which led Xie and Zhang (2000) to modify the scheme. They kept the positive CAPE condition and added a second condition based on the change of CAPE due to large-scale forcing (dCAPE). This new trigger improved the simulations of the Intertropical Convergence Zone (ITCZ) and MJO (Zhang, 2002; Song and Zhang, 2009; Zhang and Song, 2010). Alternative formulations of convection trigger include the addition of an RH threshold of 80 % in the convection trigger (Zhang and Mu 2005a, b) to suppress convection if the boundary layer air is too dry. Another modification is the inclusion of dilution in CAPE calculation due to entrainment (dilute CAPE) by Neale et al. (2008) to reduce excessive precipitation over land in the simulations of ENSO.

3.1.4 Large-scale vertical velocity trigger

Drawing on the observations in Fritsch and Chappell (1980) suggesting a positive impact of background vertical motion on convective development, Kain and Fritsch (1990) (KF) proposed a trigger based on large-scale vertical velocity. In this scheme, the first potential source layer for convection, also known as the updraft source layer (USL), is a layer of at least 60 hPa thickness that is constructed by mixing vertically adjacent layers, beginning at the surface. The temperature and pressure of the parcel at its LCL is calculated, as well as a temperature perturbation δT, which is proportional to w (see Table 3). If the sum of the parcel temperature and the temperature perturbation is higher than the environmental temperature, the parcel is released from its LCL. Above the LCL, the parcel is lifted upwards with entrainment, detrainment, water loading, and a vertical velocity determined by the Lagrangian parcel method (Bechtold et al., 2001). Convection is activated if the vertical velocity remains positive for a minimum depth of 3–4 km. Otherwise, the USL is moved up one model level and the procedure starts again. This process continues until a suitable USL is found or the search has moved up above the lowest 300 hPa of the atmosphere, where the search is terminated. To extend the application of the KF scheme to a broad range of scales, Bechtold et al. (2001) related the temperature perturbation to the grid-scale vertical velocity through a slightly different mathematical expression (see Table 3). It is widely used at ECMWF. Other authors, such as Ma and Tan (2009), included moisture advection in the temperature perturbation to improve the KF scheme for the case of weak synoptic forcing. Berg et al. (2013) defined a probability density function (PDF) that generates a range of virtual potential temperature and water vapor mixing ratio to

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substitute δT in the trigger function. With this new trigger, the scheme more realistically accounts for subgrid variability within the convective boundary layer in a way. Both the modified version of the KF scheme, and the KF itself, are used in the WRF mode.

Table 3: A sample of empirical values and assumptions used in the trigger.

Components	Empirical value or assumption	Choices in the literature	Reference
Buoyancy threshold	Includes a temperature perturbation δT linked to the large-scale vertical velocity ω	$T_{LCL} + \delta T > T_{env}$, $\delta T = k\ \omega^{1/3}$, where k is a unit number with dimensions K s$^{1/3}$ cm$^{-1/3}$	Fritsch and Chappell (1980)
		$\delta T = \pm k \cdot \overline{\omega_n}^{1/3}$, where $k = 6$ K m$^{-1/3}$ s$^{1/3}$. $\overline{\omega_n}$ is the normalized ω using a reference grid space of 25 km	Bechtold et al. (2001)
		$\delta T = k[\omega_{LCL} - c(z)]^{1/3}$, with k a unit number with dimensions K s$^{1/3}$ cm$^{-1/3}$and $c(z) = \begin{cases} \omega_0(z_{LCL}/2000), & z_{LCL} \leq 2000 \\ \omega_0 & z_{LCL} > 2000 \end{cases}$, where $\omega_0 =$ 2 cm s^{-1}, and z_{LCL} is the height (m) of the LCL above the ground	Kain (2004)
	Includes a constant δT	$\delta T = 0.65$ K	Emanuel and Živković-Rothman (1999)
		$\delta T = 0.90$ K	Bony and Emanuel (2001)
	Includes δT composed of horizontal δT_h and vertical δT_v components with associated normalized moisture advections (R_h and R_v)	$\delta T = R_h\ \delta T_h + R_v\ \delta T_v$	Ma and Tan (2009)
	Uses probability density function (PDF)	Substitute δT in the trigger function by a generated range of virtual potential temperature and water vapor mixing ratio q_v	Berg et al. (2013)
CIN	Must be smaller than a certain threshold	$CIN < 10$ J kg^{-1}	Donner (1993); Donner et al. (2001)
		$CIN < 100$ J kg^{-1}	Wilcox and Donner (2007)
Cloud base	At LCL		Betts (1986); Betts and Miller (1986); Janjić (1994)
	Height at which air parcel is moistly saturated and $T_{parcel} - T_{env} > -0.5$ K		Tiedtke (1989); Baba (2019)
	Determined from sounding	Cloud base is lower than LNB	Emanuel (1991)
	Can be anywhere in the troposphere		Grell (1993)
	Below PBL top		Zhang and McFarlane (1995)
	Might be above PBL top		Zhang and Mu (2005a)
	Lowest level where an adiabatic parcel is supersaturated		Wu (2012)
Cloud depth	Should be higher than a certain threshold value	$CD > 300$ hPa	Kuo (1965); Anthes (1977)
		$CD > 3 - 4$ km	Kain and Fritsch (1990)

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Components	Empirical value or assumption	Choices in the literature	Reference
		$CD > 150$ hPa	Hong and Pan (1998); Han and Pan (2011); Stratton and Stirling (2012)
		$CD > 3$ km	Bechtold et al. (2001)
		$CD > 200$ hPa	Gregory (2001); Jakob and Siebesma (2003); Bechtold et al. (2004)
	Minimum cloud depth is a function of the parcel temperature at LCL T_{LCL}	$CD_{min} = \begin{cases} 4000, & T_{LCL} > 20\,°\mathrm{C} \\ 2000, & T_{LCL} < 0\,°\mathrm{C} \\ 2000 + 100\,T_{LCL}, & 0\,°\mathrm{C} \le T_{LCL} \le 20\,°\mathrm{C} \end{cases}$	Kain (2004)
Cloud radius	Constant		Arakawa and Schubert (1974)
		$R = 1500$ m	Fritsch and Chappell (1980); Bechtold et al. (2001)
	Depends on the large-scale vertical velocity at LCL ω_{LCL}	$R = \begin{cases} 1000, & W_{KL} < 0 \\ 2000, & W_{KL} > 10 \\ 1000 + W_{KL}/10, & 0 \le W_{KL} \le 10 \end{cases}$ where $W_{KL} = \omega_{LCL} - c(z)$ (see buoyancy threshold for Kain (2004)	Kain (2004)
Cloud top	Determined by a temperature condition	Level where $T_{cloud} = T_{env}$	Kuo (1974); Fritsch and Chappell (1980); Wu (2012)
	Determined by LNB		Arakawa and Schubert (1974); Tiedtke (1989)
	Immediately beneath EL		Betts (1986); Betts and Miller (1986); Janjić (1994)
	Determined by the vertical velocity of the parcel w	Level where $w^2 < 0$	Bechtold et al. (2001)
		$w = 0\ \mathrm{m\,s^{-1}}$	Jakob and Siebesma (2003); Bechtold et al. (2004)
		$w < 0.2\ \mathrm{m\,s^{-1}}$	Wagner and Graf (2010)
Entrainment rate	Convection is suppressed if the entrainment in the updraft ε^u, is smaller than a certain threshold value ε_c^u	$\varepsilon_c^u = 1 \cdot 10^{-4}\ \mathrm{m^{-1}}$	Anderson et al. (2004); Kim et al. (2011)
		$\varepsilon_c^u = 0.5\,\delta M_e$, where δM_e is the mixing rate in $\mathrm{kg\,s^{-1}}$	Anderson et al. (2004); Kim et al. (2011)
RH	Set to a constant value	$RH = 100$ %	Manabe et al. (1965)
		$RH = 85$ %	Hamilton et al. (1995)
	Must be greater than a certain threshold value	$RH > 80$ %	Zhang and Mu (2005a, b); Zhang et al. (2011)
		$RH > 75$ % at lifting level	Wu (2012)
		$RH > 40$ %	Zhao et al. (2018)
Vertical velocity of the parcel		$w > 0$	Kain and Fritsch (1990); Jakob and Siebesma (2003); Bechtold et al. (2004); Kain (2004)

3.1.5 Stochastic trigger

The traditional convective triggers lead to deficiencies in the simulation of different atmospheric events, as stated in Sect. 2. A promising strategy to reduce these deficiencies is the use of stochastic triggering (Rochetin et al. 2014a, b). Instead of using a deterministic parameterization in which the subgrid-scale response is fixed to a certain resolved-scale state, the response is sampled from a suitable probability distribution (Dorrestijn et al., 2013). For example, Majda and Khouider (2002), and Khouider et al. (2003) used a stochastic model based on CIN using a Markov process. Stechmann and Neelin (2011) used a two-state Markov jump process as their stochastic trigger. Bright and Mullen (2002) modified the KF trigger function by applying stochastic perturbation to w, while Song et al. (2007) included several random parameters in the trigger criteria using a Bayesian learning procedure. Zhang et al. (2014) added a stochastic term to the SAS trigger function in the Hurricane Weather Research and Forecasting model (HWRF), and Rochetin et al. (2014a, b) used LES to introduce a stochastic trigger in the Emanuel parameterization (Emanuel, 1991).

3.1.6 HCF trigger

Unlike some of the trigger criteria already discussed, a more recent trigger function by Tawfik and Dirmeyer (2014), the HCF, is not based on the lifting parcel method, but uses vertical profiles of temperature and humidity. First, it finds the buoyant condensation level (BCL), which is the level at which saturation would occur through buoyant mixing as a result of sensible heating from the surface. To find the BCL, it increases the near-surface potential temperature through small increments and mixes the specific humidity from the surface to the level of neutral buoyancy, i.e., the top of the potential mixed layer (PML). If saturation does not occur at this level, the procedure to find the BCL is repeated until saturation is reached, while if saturation occurs, several variables are determined. The first variable is the buoyant mixing potential temperature, θ_{BM}, also known as the convective threshold. This is the temperature that the 2 m potential temperature needs to reach the BCL. The second variable, the potential temperature deficit, θ_{def}, is defined as the difference between the θ_{BM} and the 2 m potential temperature, or the sum of all the temperature increments needed to attain the BCL. Hence, it is a measure of convective inhibition similar to CIN in the parcel-based approach. In HCF, convection will activate when $\theta_{def} \leq 0$. The HCF trigger reduces the number of false positives compared to the parcel-based trigger. When the HCF trigger is implemented in the NCEP Climate Forecast System version 2 (CFSv2), the representation of the Indian monsoon and tropical cyclone intensity improves (Bombardi et al., 2016). In the Community Earth System Model (CESM), the strategy improves the frequency of heavy precipitation events and reduces the overactivation of convection in the model (Tawfik et al., 2017).

3.2 Starting levels

The LFC, USL or starting level for updraft is located at, or near, the cloud base or at the top of the planetary boundary layer. Different methods are applied for calculating the LFC in the literature, such as those used by KF and BMJ already described

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in Sect. 3.1, or the one used by Grell (1993), who determined the USL as the maximum value of the moist static energy, h. Table 4 lists a sample of the main assumptions and empirical values used to determine the starting levels.

While the starting level for the ascending currents (updrafts) is reasonably evident, the starting level for the descending currents (downdrafts), usually called the level of free sinking (LFS), may start at any vertical level no lower than the cloud base. Several convective parameterizations, such as those proposed by Tiedtke (1989) or Bechtold et al. (2001), follow the definition suggested by Fritsch and Chappell (1980), who assumed that LFS is the level at which the temperature of a saturated mixture of equal amounts of updraft and environmental air becomes smaller than the environmental temperature. In contrast, Grell (1993) determined LFS as the minimum value of h, and Zhang and McFarlane (1995) matched LFS with the lowest updraft detrainment level. However, if the minimum value of h is lower than the bottom level of updraft detrainment, LFS is determined as in Grell (1993).

Table 4: A sample of empirical values and assumptions used in the starting levels.

Components	Empirical value or assumption	Choices in the literature	Reference
USL/LFC	Level of maximum moist static energy h	[many choices]	Arakawa and Schubert (1974); Grell (1993); Zhang and McFarlane, (1995); Wu (2012)
	Near-surface air…		Tiedtke (1989); Donner (1993); Bechtold et al. (2001); Tawfik and Dirmeyer (2014)
	…must be reached within a certain pressure level	$p_{max} = 300$ hPa	Kain and Fritsch (1990); Jakob and Siebesma (2003); Bechtold et al. (2004)
	…must be reached within a certain upper limit of CIN	$CIN_{max}^{old} = 150$ hPa	Pan and Wu (1995)
	…must be reached within an upper limit in a certain range of CIN and in proportion to the large-scale vertical velocity ω	$CIN_{max} = 180 - 30\, f_\omega$, with $f_\omega = 1 - min\left[max\left(\frac{\omega-\omega_{max}}{\omega_{min}-\omega_{max}}\right)\right]$, where $\omega_{min} = -5 \cdot 10^{-3}\ (-1 \cdot 10^{-3})$ hPa s^{-1} and $\omega_{max} = -5 \cdot 10^{-4}\ (-2 \cdot 10^{-5})$ hPa s^{-1}, respectively, over land (ocean). CIN varies within the range 120−180 hPa	Han and Pan (2011)
		ω_{min} and ω_{max} are computed assuming that ω depends on the horizontal resolution of the model	Lim et al. (2014); Han et al. (2019)
		$\mathrm{CIN}_{\mathrm{max}}^{\mathrm{new}} = (1 - \sigma)\,\mathrm{CIN}_{\mathrm{max}}$, where σ is a scale-aware factor	Kwon and Hong (2017)
LFS	Level at which the temperature of a saturated mixture of equal amounts of updraft and environmental air becomes less than T_{env}		Fritsch and Chappell (1980); Tiedtke (1989); Bechtold et al. (2001)
	Level of minimum environmental saturated equivalent potential temperature between LCL and cloud top		Kain and Fritsch (1990); Wu (2012)
	Coincides with the level of minimum moist static energy h if lower than the base of the detrainment layer. If not, it matches the detrainment level		Grell et al. (1991); Zhang and McFarlane (1995)

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Components	Empirical value or assumption	Choices in the literature	Reference
	Level above the minimum moist static energy *h*		Grell (1993); Pan and Wu (1995)
	The highest level where equal parts of evaporatively cooled environmental air and cloudy air become unstable with respect to the environment		Nordeng (1994)
		Located within the range 120−150 hPa above USL	Kain (2004)
	Level where the saturated updraft terminates	150 hPa above the ground	Stratton and Stirling (2012)
	Level of minimum moist static energy *h*		Baba (2019)

3.3 Impact of trigger functions on convective models

Differences between trigger functions depend on the identification of the source layer of convective air and on how this layer of unstable air can give rise to convection. While near-surface air is selected as the source layer in some CPs (Tiedtke, 1989; Donner, 1993; Bechtold et al., 2001; Tawfik and Dirmeyer, 2014), in others, the choice is the layer of maximum moist static energy, *h* (Arakawa and Schubert, 1974; Grell, 1993; Zhang and McFarlane, 1995; Wu, 2012). On the other hand, different convection triggers are used to determine whether unstable air turns into convection, as mentioned in the previous section. However, the best way to construct a trigger function is still unknown and, in many cases, an ad hoc formulation leads to poor performance in the activation of convection at the right location and time (Suhas and Zhang, 2014; Song and Zhang, 2017). Comparison between the performance of different trigger functions and observations from different climates leads to improvements in the formulation of the activation criteria for convection. Suhas and Zhang (2014) used three intensive observation period (IOP) datasets from the Atmospheric Radiation Measurement (ARM) program, and long-term single-column models (SCMs) to evaluate the performance of different trigger functions (Arakawa-Schubert scheme, Bechtold scheme, Donner scheme, Kain-Fritsch scheme, Tiedtke scheme, and four variants of the Zhang-McFarlane scheme). The dilute dCAPE trigger function showed the best performance in both the tropics and midlatitudes, while the undilute dCAPE was as good as the dilute dCAPE only for the tropics. Furthermore, the Bechtold and the dilute CAPE trigger functions were among the best performing schemes. As a follow-up, Song and Zhang (2017) used observations from the Green Ocean Amazon (GOAmazon) field campaign to evaluate and improve the trigger functions selected in Suhas and Zhang (2014), with the addition of the HCF. In their study, the dCAPE-type triggers also ranked first, followed by the Bechtold and HCF triggers. The dCAPE trigger improved with an optimization of the entrainment rate and dCAPE threshold, while the undilute dCAPE trigger performed better with the inclusion of a 700-hPa upward motion.

The convection trigger criterion plays a crucial role in the simulation of a wide number of atmospheric events. The impact of the trigger function on the correct simulation of the diurnal cycle of convection and precipitation in atmospheric models has been widely studied, especially over land (Bechtold et al., 2004; Knievel et al., 2004; Lee et al., 2007a, b, 2008; Hara et al., 2009; Evans and Westra, 2012). The common problem in the simulation of the diurnal cycle is that it peaks too early and its amplitude is too high (Yang and Slingo, 2001; Collier and Bowman, 2004). Moreover, the diurnal cycle of precipitation peaks too early over land (in general, 2 to 4 hours before the observed maxima) (Dai, 2006), which is related to the formulation of the trigger function (Betts and Jakob, 2002; Bechtold et al., 2004). Lee et al. (2008) performed a sensitivity analysis with four different trigger functions implemented in the relaxed Arakawa-Schubert scheme (RAS) and found significant differences in the diurnal cycle of precipitation over the Great Plains in the United States. Several studies have performed sensitivity analyses and found possible ways to improve the simulation of the diurnal cycle. Models with finer resolution provided a better simulation in the amplitude, variability, and timing of the diurnal cycle (Wang et al., 2007; Sato et al., 2009). The inclusion of the effect of moisture advection in the trigger function improved the distribution and intensity of convective precipitation in the MM5 (Ma and Tan, 2009). The use of different initiation and termination conditions in the SAS scheme led to a better diurnal variation of precipitation (Han et al., 2019) although it increased the excessive precipitation and did not alleviate the bias in the phase of precipitation intensity. The modification of both the trigger and closure criteria by considering cold pools could minimize the bias in the diurnal cycle of convection (Rio et al., 2009, 2013). Another important case are the deficiencies in the simulation of the MJO (Lin et al., 2006), which are often improved by the modification of the trigger function. For example, Wang and Schlesinger (1999) found that a better representation of the MJO was possible by adding a moisture trigger to the convective parameterization used in the atmospheric general circulation model at the University of Illinois, Urban–Champaign (UIUC). Zhang and Mu (2005b) used the same approach in the National Center for Atmospheric Research (NCAR) Community Climate Model version 3 (CCM3) as well as Lin et al. (2008) in the Seoul National University (SNU) atmospheric general circulation model. Another example is a better representation of the Indian summer monsoon rainfall by the addition of HCF to the trigger function in the Climate Forecast System version 2 (CFSv2) (Bombardi et al., 2015).

The lack of "convective memory" effects in the models based on the quasi-equilibrium (QE) assumption causes a convective parameterization to be triggered, regardless of the convection stage, as long as the convection criteria are met. Different ways to include the memory effect have been proposed, such as using prognostic cumulus kinetic energy (Pan and Randall, 1998), or an ensemble of cold pools (Grandpeix and Lafore, 2010; Del Genio et al., 2015).

4 Cloud model: types and choices

The cloud model represents the interaction between cumulus clouds and the large-scale environment. Thus, it determines the vertical distribution of convective heat and moisture through the parameterization of the mass flux profile, the entrainment/detrainment, and the microphysics. This section discusses the main types of mass flux and entrainment/detrainment schemes adopted in the literature, as well as the main assumptions and empirical values employed in the formulation of the cloud model.

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4.1 Mass flux scheme types

According to the approach used to estimate the unknown quantities in Eq. (2), mass flux schemes are classified into bulk, spectral and episodic mixing models.

4.1.1 Bulk models

The ensemble of clouds within a grid box is represented by a single cloud model. Yanai et al. (1973) are the main representatives of this type of scheme. In their diagnostic study, clouds are classified according to their cloud tops, and the steady plume hypothesis (Morton et al., 1956) is applied. It is assumed that all clouds have a common cloud base height, and that the values on detrainment are identical to the values inside the plume. In mesoscale models, Fritsch and Chappell (1980) and Kain and Fritsch (1992) also applied the steady hypothesis, as did Singh et al. (2019) in their study of the relationship between humidity, instability, and precipitation in the tropics. Tiedtke (1989), and Gregory and Rowntree (1990) applied the same approach as Yanai et al. (1973) in their schemes at the ECMWF, and at the U.K. Meteorological Office. The scheme used at ECMWF has undergone several modifications since then (Nordeng, 1994; Gregory et al., 2000; Li et al., 2007; Zhang et al., 2011; Kim and Kang, 2012; Stevens et al., 2013). Many mass flux parameterizations use the bulk-cloud approach (Siebesma and Holtslag, 1996; Bechtold et al., 2001; Neggers et al., 2009; Yano and Baizig, 2012; Loriaux et al., 2013) with different formulations of their cloud models (i.e., formulation of the mass flux at cloud base, entrainment, detrainment, microphysics).

4.1.2 Spectral models

In contrast to bulk models, spectral models select a certain parameter to group the plumes into different types, each of them with a cloud model. The majority of spectral approaches use a constant entrainment rate, while other authors choose the pressure depth (Hack et al., 1984), or the radius and vertical velocity at cloud base (Nober and Graf, 2005). In contrast to Yanai et al. (1973), AS applied the quasi-equilibrium hypothesis (QE), which assumes that convection is in a quasi-equilibrium with the large-scale environment. Since the publication of the original version, the AS scheme has undergone several modifications. Moorthi and Suarez (1992) proposed a simplified version called the relaxed Arakawa-Schubert (RAS) parameterization with a simpler closure formulation. Grell (1993) changed the spectrum of cloud sizes in AS for a single cloud top at a particular location and time. Pan and Wu (1995) developed the so-called simplified Arakawa-Schubert model (SAS), which is a modified version of the model proposed by Grell (1993). Han and Pan (2011) further modified SAS to overcome unrealistic grid-scale precipitation and develop a mass flux parameterization for shallow convection.

4.1.3 Episodic mixing models

Drawing on the continuous entrainment and average buoyancy used in entraining/detraining plume models in both bulk and spectral formulations, Emanuel (1991, 1994) proposed the so-called episodic mixing model, which is based on the stochastic mixing model of Raymond and Blyth (1986), and the observations of Taylor and Baker (1991), among others. Thus, Emanuel assumed that mixing is highly inhomogeneous and episodic, and applied the buoyancy sorting hypothesis, which is the basis of a number of cumulus parameterizations (James and Markowski, 2010; Park, 2014), especially those focused on shallow convection (Bretherton et al., 2004; De Rooy and Siebesma, 2008; Neggers et al., 2009; Pergaud et al., 2009). The Emanuel scheme and its modified versions (Emanuel and Živković-Rothman, 1999; Grandpeix et al., 2004; Peng et al., 2004) are widely used in RCMs (Zou et al., 2014; Raju et al., 2015; Bhatla et al., 2016; Gao et al., 2016; Kumar and Dimri, 2020).

The aforementioned mass flux scheme types are explained from the point of view of the ascending currents. However, convective downdrafts, i.e., descendent currents caused by evaporation of condensate and rainwater loading, should be taken into account. Simply put, they may be considered as bottom-up updrafts. Downdrafts are of great importance in atmospheric convection. As Plant and Yano (2015) highlighted, they have opposite effects on the organization and evolution of convective systems. The transport of cooler and drier air into the subcloud layer may stabilize it and therefore inhibit convection or may lead to the development of new convective elements if downdrafts cause an increase in low-level convergence. The majority of convective parameterizations include downdrafts with assumptions about their starting level, entrained and detrained air, or the amount of condensate available for evaporation. However, many schemes, such as Grell (1993), the Zhang-McFarlane scheme used in CESM, or the Tiedtke scheme in the ECHAM model, have described downdrafts as simple saturated plumes, i.e., "inverse plume", with a mass flux proportional to the updraft mass flux (Thayer-Calder, 2012). Other authors have proposed a more complex parameterization including unsaturated downdrafts in their formulations and a downdraft mass flux based on Eq. (2) (Emanuel, 1991; Xu et al., 2002).

4.2 Entrainment and detrainment

The mixing of air masses due to entrainment of environmental air into clouds and detrainment of cloudy air into the environment are key processes in convective parameterizations (Blyth, 1993; Luo et al., 2010; Donner et al., 2016) as they modify the vertical profiles of heat and moisture within cloudy air. Sanderson et al. (2008) identified the entrainment rate as one of the dominant parameters affecting climate sensitivity after evaluating thousands of GCM simulations. Other authors, such as Rougier et al. (2009), Klocke et al. (2011) and Zhao (2014) have obtained similar conclusions in their analyses. In addition, the influence of convective detrainment of water vapor and hydrometeors from cumulus clouds is an important source of water that strongly impacts climate simulations (Ramanathan and Collins, 1991; Lindzen et al., 2001).

In this section, attention is drawn to the most important model types of entrainment and detrainment, the main assumptions and empirical values used in the literature, and the impact that the different formulations have in convective models. The main assumptions and empirical values used in the formulation of entrainment and detrainment are listed in Tables 5 and 6 and in Tables 7 and 8, respectively.

4.2.1 The choice of lateral vs cloud-top entrainment

Since Stommel (1947) provided the first description of cumulus cloud dilution by entrainment of environmental air, two conceptual models are still competing: the lateral entrainment model and the cloud-top entrainment model.

In the lateral entrainment model, Stommel (1947) considered that environmental air enters the cloud through the lateral cloud edges and continuously dilutes cloudy air during its ascent, regardless of whether it is considered a plume or a bubble. Several aircraft observations and experiments in water tanks (Turner, 1962; Morton, 1965) contributed to the formulation of the lateral entrainment theory. However, authors such as Warner (1970) pointed out the deficiencies of this theory in predicting the right profile of liquid water content (LWC).

In order to address these deficiencies, Squires (1958) proposed another entrainment model, the cloud-top entrainment. This author suggested that environmental air enters the cloud predominantly at or near the cloud top, descends through penetrative downdrafts created by evaporative cooling, and dilutes the cloud by turbulent mixing. Paluch (1979) provided more evidence for cloud-top entrainment in her study on cumulus clouds over Colorado. The author found that the cloud water-mixing ratio and the wet equivalent potential temperature follow a line at a single level, the so-called "mixing line", which connects cloud base and cloud top. Paluch interpreted it as an evidence for a two-point mixing scenario. Further studies (Boatman and Auer, 1983; Lamontagne and Telford, 1983; Jensen et al., 1985; Reuter and Yau, 1987) confirmed Paluch's results. However, several authors have criticized the mixing line source levels (Blyth et al., 1988; Malinowski and Pawlowska-Mankiewicz, 1989; Raga et al., 1990; Grabowski and Pawlowska, 1993; Neggers et al., 2002; Zhao and Austin, 2005), and the interpretation of the mixing line (Betts and Albrecht, 1987; Taylor and Baker, 1991; Grabowski and Pawlowska, 1993; Siebesma, 1998; Böing et al., 2014).

Which of the two models predominates in cumulus convection remained unclear for many years. The increase in computational power in recent decades has promoted the use of LES to study entrainment and detrainment mainly in shallow cumulus clouds. Several authors, such as Heus et al. (2008) and Böing et al. (2014), have applied LES to identify the dominant process in mixing in cumulus clouds, concluding that cloud-top entrainment is insignificant compared to lateral entrainment.

4.2.2 Main empirical values in entrainment and detrainment formulations

Aircraft observations and experiments in water tanks (Turner, 1962; Morton, 1965) led to the formulation of the lateral entrainment theory, which anticipates that the fractional entrainment rate (hereafter entrainment rate) changes with the cloud radius (Malkus, 1959; Squires and Turner, 1962)

$$\frac{1}{M}\frac{\partial M}{\partial z} = \varepsilon \simeq \frac{C}{R}, \quad (3)$$

where M is the mass flux, z is the height, ε denotes the entrainment rate, C is a constant, and R is the radius of the rising plume. As De Rooy et al. (2013) pointed out in their review article on entrainment and detrainment in cumulus convection, many cloud models still use this formulation (Arakawa and Schubert, 1974; Kain and Fritsch, 1990; Donner, 1993), sometimes assuming a constant entrainment rate.

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Houghton and Cramer (1951) improved this theory by taking into account the increase of vertical velocity due to buoyancy. Thus, the authors distinguish between dynamical entrainment due to larger-scale organized inflow, $\varepsilon_{\mathrm{dyn}}$, and turbulent entrainment caused by turbulent mixing, $\varepsilon_{\mathrm{turb}}$ (often described with an eddy diffusivity approach). Hence, the change of mass flux with height, including the detrainment, δ, of negative buoyant mixtures, is given by

$$\frac{1}{M}\frac{\partial M}{\partial z} = \varepsilon_{\mathrm{dyn}} + \varepsilon_{\mathrm{turb}} - \delta_{\mathrm{dyn}} - \delta_{\mathrm{turb}}. \qquad (4)$$

Tiedtke (1989) and Nordeng (1994) assumed that turbulent entrainment is inversely proportional to cloud radii, as in Simpson and Wiggert (1969) and Simpson (1971). They used typical cloud sizes for different types of convection to fix the values of entrainment rates. For penetrative and midlevel convection, the entrainment rate was fixed to $\varepsilon_{\mathrm{turb}} = 1 \cdot 10^{-4}\ \mathrm{m}^{-1}$, which is a typical value for tropical clouds in (Simpson, 1971). For shallow convection, the entrainment rate was based on typical values for large trade cumuli, $\varepsilon_{\mathrm{turb}} = 3 \cdot 10^{-4}\ \mathrm{m}^{-1}$ (Nitta, 1975). Gregory and Rowntree (1990) also assumed a turbulent entrainment rate, but inversely proportional to the height, while in Bechtold et al. (2008), $\varepsilon_{\mathrm{turb}}$ depends on the saturation specific humidity (Table 5). Dynamical entrainment $\varepsilon_{\mathrm{dyn}}$ is proportional to moisture convergence and occurs only in the lower part of the cloud layer up to the level of strongest vertical ascent in Tiedtke (1989). In Nordeng (1994), it is based on momentum convergence. Gregory and Rowntree (1990) did not include it in their parameterization, whereas in Bechtold et al. (2008), it depends on RH. For downdraft, Bechtold et al. (2014) set $\varepsilon_{\mathrm{turb}} = 3 \cdot 10^{-4}\ \mathrm{m}^{-1}$ and $\varepsilon_{\mathrm{dyn}}$ as a function of B. A common practice in the definition of entrainment rates for downdraft consists in assuming a similar parameterization as for updrafts (Table 6).

Kain and Fritsch (1990) introduced another type of parameterization based on the buoyancy sorting. In their parameterization, homogeneous mixing of cloudy and environmental air was assumed, leading to mixtures with different buoyancy properties that have the same probability of occurrence. Moreover, the authors modified Eq. (3) to make it pressure-dependent. The fraction of environmental air that makes the mixture neutrally buoyant is the so-called critical mixing fraction χ_c, which determines whether a mixture entrains or detrains after mixing. Thus, entrainment of positive buoyant mixtures occurs if $\chi < \chi_c$, while $\chi > \chi_c$ leads to immediate detrainment of negative buoyant mixtures. Therefore, detrainment can occur at any level where $\chi > \chi_c$, unlike in the Arakawa-Schubert scheme, where only the cloud top detrainment is considered. Moreover, the maximum entrainment rate is proportional to pressure and inversely proportional to updraft radius. However, the Kain-Fritsch scheme had deficiencies, such as excessive detrainment or the production of unrealistic deep saturated layers. To handle the excessive detrainment, Bretherton et al. (2004) modified χ_c by defining a critical eddy-mixing distance d_c based on observations and LES results that revealed fractions of negative buoyant air in the updrafts (Taylor and Baker, 1991; Siebesma and Cuijpers, 1995). Thus, d_c is the distance that negative buoyant mixtures in absence of entrainment can continue upwards before their velocity drops to zero, i.e., before detraining. Mixtures of this kind are included in the definition of χ_c together with positive buoyant mixtures, which leads to new definitions of entrainment/detrainment rates. In newer versions of the KF scheme, a mitigation of unrealistic deep saturated layers is achieved by assuming that the entrainment of environmental air cannot be lower than 50 % of the total environmental air involved in the mixing process in the updraft, and that cloud radius

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depends on the convergence of the subcloud layer (Kain, 2004). Recently, Zheng et al. (2016) modified the minimum entrainment equation in Kain (2004) to include both organized and turbulent entrainment. The authors made the equation scale-dependent and expressed it in terms of subcloud layer depth instead of cloud radius. Another scheme based on the buoyancy-sorting hypothesis, but assuming episodic mixing, is the Emanuel scheme (Emanuel, 1991), where, in contrast to the KF scheme, the resulting mixtures just ascend or descend to their level of neutral buoyancy to detrain.

Apart from buoyancy, another environmental quantity that might influence entrainment, and therefore convection, is RH. A number of studies have analyzed the effect of RH in parameterization of entrainment/detrainment rates, drawing different conclusions. For instance, Jensen and Del Genio (2006) found a positive correlation between entrainment rate and RH in their analysis of remote sensing observations and soundings at Nauru Island, while Bechtold et al. (2008) and Zhao et al. (2018) found a negative correlation using the Atmospheric Model version 4 (AM4.0). The same conclusion was achieved by Stirling and Stratton (2012) using a CRM formulation and the Met Office Unified Model (Met Office UM). Recently, Lu et al. (2018) identified deficiencies in the previous studies that could lead to erroneous conclusions regarding the effects of RH on entrainment, such as the use of conserved quantities related to RH to estimate entrainment rates, or that no observations had thus far been used to determine the relationship between RH and entrainment. To address these deficiencies, the authors analyzed aircraft observations from the Routine AAF (ARM Aerial Facility) CLOWD (Clouds with Low Optical Water Depths) Optical Radiative Observations (RACORO) (Vogelmann et al., 2012) and RICO field campaigns (Rauber et al., 2007) for shallow cumulus and concluded that ε and RH are positively correlated. Nonetheless, there is no general consensus on the effects of environmental RH on entrainment rates (Lu et al., 2018).

Other approaches use in-cloud quantities instead of only the environmental quantities to estimate the entrainment rate. For instance, Neggers et al. (2002) used LES results in their multi-parcel model to formulate the entrainment rate as inversely proportional to the product of an eddy turnover time scale and the updraft speed w. Based on the assumption that entrainment reduces B, Gregory (2001) proposed an entrainment rate that depends on B and inversely on the square of the updraft speed w and used different parameters for shallow and deep convection. This parameterization achieved satisfactory results in various analyses (Chikira and Sugiyama, 2010; Del Genio and Wu, 2010) but proved to be cloud- and altitude-dependent. Recently, Baba (2019) modified Gregory's parameterization of the entrainment rate by relating it to the detrainment rate and B. This new parameterization led to improvements in the simulation of MJO, equatorial waves, and precipitation over the western Pacific region.

Mapes and Neale (2011) addressed the so-called "entrainment dilemma", in which the excessive entrainment values tend to excessively restrain convection, while insufficient entrainment values abundantly ease its activation. To overcome this, they proposed a new formulation of the entrainment rate dependent on a prognostic variable called *organization*, which expresses the interaction between the environment and convection. In their formulation, the rain evaporation rate controls the *organization* and produces more deep convection for lower values of the entrainment rate.

Instead of the lateral entrainment hypothesis, other authors have proposed a stochastic parameterization of entrainment. Romps and Kuang (2010) suggested two probability density functions to specify their stochastic entrainment parameterization and

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assumed that the entrainment rate follows a stochastic Poisson process. Above the condensation level, Sušelj et al. (2013) used a stochastic approach similar to Romps and Kuang (2010), but adjusted for steady-state updrafts, while below the condensation level, the entrainment rate is constant. Using this formulation, the authors achieved good results for several shallow cumulus convection events.

Table 5: A sample of empirical values and assumptions used in the parameterization of entrainment in the updraft.

Type	Empirical value or assumption	Choices in the literature	Reference
Turbulent	Constant	$\varepsilon^u_{turb} = 1 \cdot 10^{-4}\ \mathrm{m}^{-1}$ for penetrative (only occurs in the lower part of the cloud layer) and midlevel convection	Tiedtke (1989); Nordeng (1994); Zhang et al. (2011); Möbis and Stevens (2012)
		$\varepsilon^u_{turb} = 2 \cdot 10^{-4}\ \mathrm{m}^{-1}$	Wang et al. (2007)
	Inversely proportional to height z	$\varepsilon^u_{turb} = C^u_t/z$, with $C^t_t = 3\,A_e\,f(p)$, where $A_e = 1.5$ for all levels above LCL, and $f(p) = p/p_s^2$, with p_s *the* surface pressure	Gregory and Rowntree (1990)
		$C^u_t = 0.55 + 8.0\left(1.2 - \frac{z_{LCL}}{100}\right)^2$, with $0.55 \le C^u_t \le 3.5$	Stratton and Stirling (2012)
	Proportional to the environmental humidity $\bar{q}$	$\varepsilon^u_{turb} = c_0 F_{\varepsilon,0}$, where $F_{\varepsilon,0} = \left(\frac{\overline{q_s}}{\overline{q_{s,b}}}\right)^2$ and $\overline{q_s}$ and $\overline{q_{s,b}}$ are the saturation specific humidity at the parcel level and cloud base, respectively	Bechtold et al. (2008); Han and Pan (2011); Zhang and Song (2016)
Dynamical	Proportional to moisture convergence		Tiedtke (1989); Möbis and Stevens (2012)
	Depends on momentum convergence	$\varepsilon^u_{dyn} = \frac{1}{2}\frac{B}{w^2_{d,LFS} - \int_z^{LFS} B\,dz} + \frac{1}{\rho}\frac{d\rho}{dz}$, where $w_{d,LFS} = 1\ \mathrm{m\,s^{-1}}$ is the downdraft velocity at LFS	Nordeng (1994); Möbis and Stevens (2012)
	Proportional to the environmental humidity $\bar{q}$	$\varepsilon^u_{dyn} = c_1 \frac{\overline{q_s} - \bar{q}}{\bar{q}} F_{\varepsilon,1}$, where $F_{\varepsilon,1} = \left(\frac{\overline{q_s}}{\overline{q_{s,b}}}\right)^3$, c_1 is a tunable parameter, and $\overline{q_s}$ and $\overline{q_{s,b}}$ are the saturation specific humidity at the parcel level and cloud base, respectively	Bechtold et al. (2008)
		$\varepsilon^u_{dyn} = d_1(1 - RH)F_{\varepsilon,1}$where d_1 is a tunable parameter	Han and Pan (2011)
	Occurs when cloud parcels accelerate upward and the buoyancy B is positive		Zhang et al. (2011)
No distinction	Inversely proportional to cloud radius R	$\varepsilon^u = C^u_e/R$, with $C^u_e = 1$	Malkus (1959)
		$C^u_e = 0.2$	Squires and Turner (1962); Simpson and Wiggert (1969); Arakawa and Schubert (1974); Lin and Arakawa (1997); Wagner and Graf (2010)
	Function of a critical mixing fraction χ_c	$\chi < \chi_c$	Kain and Fritsch (1990); Bechtold et al. (2001)
		$\varepsilon^u \ge M_u \frac{C^u_e \delta p}{R} \chi_c$, where M_u is the updraft mass flux at cloud base, $C^u_e = 0.03\ \mathrm{m\,Pa^{-1}}$, and $\chi_c = 0.5$	Kain (2004)
	Does not exist around cloud edges		Grell et al. (1994)
	Defined by the requirement that the temperature of the plume that	Reaches its maximum value at the height of minimum h for a saturated state	Zhang and McFarlane (1995)

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Type	Empirical value or assumption	Choices in the literature	Reference
	detrains at a certain level *z* equals T_{env}		
	Function of the buoyancy of the parcel *B* and the in-cloud updraft velocity, *w*	$\varepsilon^u = C_e^u \frac{aB}{w^2}$, where $C_e^u = 0.5$ (G01)	Gregory (2001), Kim et al. (2013)
		$C_e^u = 0.6$	Chikira and Sugiyama (2010)
		$C_e^u = 0.3$	Del Genio et al. (2012)
		$C_e^u = (\frac{1}{RH} - 1)$	Kim and Kang (2012)
		$C_e^u = 0.52$	Hirota et al. (2014)
	Function of the in-cloud vertical velocity *w* and a turnover timescale τ_t	$\varepsilon^u = \frac{\eta}{\tau_t}\frac{1}{w}$, with $\tau_t = 300$ s and $\eta = 0.9$ for BOMEX and 1.2 for SCMs	Neggers et al. (2002)
		$\eta/\tau_t = 2.4 \cdot 10^{-3}$ s^{-1}	Chikira and Sugiyama (2010)
	Inversely proportional to height *z*	$\varepsilon^u = C_e^u/z$, where $C_e^u = 0.55$	Jakob and Siebesma (2003)
		$C_e^u = 1$	Han and Pan (2011) (only in subcloud layers)
	Depends on a critical eddy-mixing distance d_c and a critical mixing fraction χ_c	$\varepsilon^u \propto \frac{C_e^u}{d_c}\chi_c^2$, where $C_e^u = 1.5$	Bretherton et al. (2004)
	Function of the buoyancy *B* and the in-cloud vertical velocity *w*	$\varepsilon^u = max\left[0, \frac{1}{1+\beta_1}\left(\frac{a_1\beta_1 B}{w^2} - b\right)\right]$, where $a_1\beta_1(1+\beta_1)^{-1} = 0.315$, and $b = 0.002$	Rio et al. (2010)
	Stochastic parameterization	ε^u follows a Poisson process	Romps and Kuang (2010)
	Depends on a prognostic variable		Mapes and Neale (2011)
	Depends on RH and the height of the LCL z_{LCL} for the early stages of developing convection over land		Stirling and Stratton (2012)
	Depends on the PBL depth and the height *z*. Sets a maximum value for ε^u	$\varepsilon^u = \mu/\min(z, z_{PBL})$ with $\mu = 0.185$ as default value and $\varepsilon^u_{max} = 1 \cdot 10^{-4}$ m^{-1}. The value of μ is modified within the paper (μ × 2, μ × 5, μ/2)	Oueslati and Bellon (2013)
	Function of the pressure *p*	$\varepsilon^u = 4.5\, F \frac{p(z)\rho)g(z)}{p_s^2}$ with $F = 0.9$ as a default value and p_s *the* surface pressure	Klingaman and Woolnough (2014)
	Uses PDFs	Lognormal, gamma and Weibull distributions	Guo et al. (2015)
	The entrained mass depends on the pressure depth of a model layer Δp, horizontal grid spacing Dx, and the height of LCL above the ground z_{LCL}	$\Delta M_e = M_b \frac{\alpha\beta}{z_{LCL}} \Delta p$, where M_b is the updraft mass flux at cloud base, $\alpha = 0.03$, and $\beta = [1 + ln(25/Dx)]$	Zheng et al. (2016)
	Set to a constant value	$\varepsilon^u = 2.5 \cdot 10^{-4}$ m^{-1}	Song and Zhang (2017)
	Function of buoyancy *B* and detrainment rate δ^u	$\varepsilon^u w^2 = C_1 B - C_2 \delta^u w^2$ with $C_1 = C_2 \approx 0.2$	Baba (2019)

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Table 6: A sample of empirical values and assumptions used in the parameterization of entrainment in the downdraft.

Type	Empirical value or assumption	Choices in the literature	Reference
Turbulent	Set to a constant value	$\varepsilon_{turb}^{d} = 2 \cdot 10^{-4}\ \mathrm{m^{-1}}$	Tiedtke (1989); Nordeng (1994); Möbis and Stevens (2012); Baba (2019)
		$\varepsilon_{turb}^{d} = 3 \cdot 10^{-4}\ \mathrm{m^{-1}}$	Bechtold et al. (2014)
Dynamical	Function of in-cloud buoyancy B and downdraft velocity at the LFS $w_{\ d,LFS}$	$\varepsilon_{dyn}^{d} = \frac{-B}{w_{d,LFS}^{2} - \int_{z}^{LFS} B\, dz} + \frac{1}{\rho}\frac{d\rho}{dz}$, where $w_{\ d,LFS} = 1\ \mathrm{m\ s^{-1}}$ is the downdraft velocity at the LFS	Baba (2019)
	Function of in-cloud buoyancy B		Bechtold et al. (2014)
No distinction	Set to a constant value	$\varepsilon^{d} = 2 \cdot 10^{-4}\ \mathrm{m^{-1}}$ (K13)	Gerard and Geleyn (2005); Gerard (2007); Kim et al. (2013)
	Proportional to ε^{u}. Its maximum value ε_{max}^{d} is constrained	$\varepsilon^{d} = 2\ \varepsilon^{u}$ and $\varepsilon_{max}^{d} = 2/(z_D - z_b)$ where z_D is height of the detrainment level, and z_b is the cloud base height	Zhang and McFarlane (1995)

Less attention has been paid to the parameterizations of the detrainment process. Many convection schemes set it as a constant value (see Tables 7 and 8), while others consider detrainment to be negligible (Lu et al., 2012). Tiedtke (1989) and Nordeng (1994) assumed a turbulent detrainment inversely proportional to cloud radii and fixed its value to $\delta_{\mathrm{turb}} = 1 \cdot 10^{-4}\ \mathrm{m^{-1}}$ for penetrative and midlevel convection (see Table 7). On the other hand, Gregory and Rowntree (1990) assumed a turbulent detrainment rate inversely proportional to the height and smaller than $\varepsilon_{\mathrm{turb}}$, while Bechtold et al. (2008) set δ_{turb} to a constant value. Dynamical detrainment δ_{dyn} occurs above the cloud top in Tiedtke (1989), while in Nordeng (1994), it is computed for a spectrum of clouds detraining at different heights. In Gregory and Rowntree (1990), it is activated when B is less than 0.2 K and in Bechtold et al. (2008), it is proportional to the decrease in updraft vertical kinetic energy at the top of the cloud. For downdraft, Bechtold et al. (2014) set $\delta_{\mathrm{turb}} = \varepsilon_{\mathrm{turb}}$, and enforced δ_{dyn} over the lowest 50 hPa. As in the case of entrainment rates in downdrafts, a common practice in the definition of detrainment rates for downdraft consists in assuming a similar parameterization as for updrafts (Table 8).

Table 7: A sample of empirical values and assumptions used in the parameterization of detrainment in the updraft.

Type	Empirical value or assumption	Choices in the literature	Reference
Turbulent	Constant	$\delta_{turb}^{u} = 1 \cdot 10^{-4}\ \mathrm{m^{-1}}$	Tiedtke (1989); Nordeng (1994); Bechtold et al. (2008); Zhang et al. (2011)
		$\delta_{turb}^{u} = 0.75 \cdot 10^{-4}\ \mathrm{m^{-1}}$	Bechtold et al. (2014)
	Proportional to the entrainment rate ε_{turb}^{u}	$\delta_{turb}^{u} = C_{dt}^{u} \cdot \varepsilon_{turb}^{u}$ where $C_{dt}^{u} = 2/3$	Gregory and Rowntree (1990)
		$C_{dt}^{u} = (1 - RH)$	Derbyshire et al. (2011); Walters et al. (2019)
		$C_{dt}^{u} = 1$	Stratton and Stirling (2012)
Dynamical	Initiated if the buoyancy of the parcel is less than a minimum value, B_{min}	$B_{min} = 2 - 3\ \mathrm{K}$	Yanai et al. (1973)
		$B_{min} = 0.2\ \mathrm{K}$	Gregory and Rowntree (1990)

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Type	Empirical value or assumption	Choices in the literature	Reference
	Only at levels of neutral buoyancy		Tiedtke (1989)
	Non-zero above the lowest possible organized detrainment level z_{low}	$\delta^u_{dyn} = \frac{1}{\sigma}\frac{d\sigma}{dz}$, where $\sigma = \sigma_0 \cos\left(\frac{\pi}{2}\frac{z-z_{low}}{z_{ct}-z_{low}}\right)$ with z_{ct} the cloud top height, and σ the horizontal area covered by the updraft. z_{low} is the level of neutral buoyancy with entrainment rate $\varepsilon = \frac{1}{2(\zeta+z-z_{cb})}$, where the subscript cb means cloud base, and $\zeta = 25$ m corresponds to an excess buoyancy of 1 K at cloud base and a vertical velocity of 1 m s^{-1} at that level.	Nordeng (1994),
	Proportional to the decrease in updraft vertical kinetic energy at the top of the cloud		Bechtold et al. (2008); Zhang and Song (2016)
	Proportional to the loss of buoyancy		Derbyshire et al. (2011)
No distinction	Occurs only in a thin layer at cloud top		Arakawa and Schubert (1974)
	Only at levels of neutral buoyancy		Emanuel (1991); Moorthi and Suarez (1992)
	Does not exist around cloud edges		Grell et al. (1994)
	Depends on a critical eddy-mixing distance dc and a critical mixing fraction χ_c	$\delta^u \propto \frac{C^u_d}{d_c}\chi_c^2$, where $C^u_d = 1.5$	Bretherton et al. (2004); Zhao et al. (2018)
	Function of a non-dimensionalized mass flux $\hat{m}$ and a non-dimensionalized height $\hat{z}$	$\delta^u = \frac{0.1\, ln\left(1+\hat{z}\frac{h}{z_b}\right) - \ln \hat{m}}{h\,\hat{z}}$	De Rooy and Siebesma (2008)
	Depends on in-cloud vertical velocity w, buoyancy B and the difference in the water mixing ratio (Δq) between the mean plume (q_l) and the environment (q)	$\delta^u = max\left[0, -\frac{a_1\beta_1}{1+\beta_1}\frac{B}{w^2} + c\left(\frac{\frac{\Delta q}{q}}{w^2}\right)^d\right]$, where $a_1 = 2/3$, $\beta_1 = 0.9$, $c = 0.012\ \mathrm{s}^{-1}$ and $d = 0.5$	Rio et al. (2010)
	Constant at all levels	$\delta^u = \varepsilon_0$, with ε_0 the entrainment at cloud base	Han and Pan (2011)
	Function of buoyancy B and in-cloud vertical velocity w	$\delta^u = -C^u_d \frac{aB}{w^2}$where C^u_d takes different values	Kim et al. (2013)
	Function of buoyancy B	$\delta^u = B/2$	Baba (2019)

Table 8: A sample of empirical values and assumptions used in the parameterization of detrainment in the downdraft.

Type	Empirical value or assumption	Choices in the literature	Reference
Turbulent	Set to a constant value	$\delta^d_{turb} = 2 \cdot 10^{-4}\ \mathrm{m}^{-1}$	Tiedtke (1989): Nordeng (1994); Baba (2019) neglects it when the downdraft is thermodynamically positive buoyant or reaches below the cloud base
		$\delta^d_{turb} = 3 \cdot 10^{-4}\ \mathrm{m}^{-1}$	Bechtold et al. (2014)
Dynamical	Enforced over the lowest 50 hPa		Bechtold et al. (2014)
	When the downdraft is thermodynamically positive buoyant or reaches below the cloud base	δ^d_{dyn} inversely proportional to layer thickness (if in-cloud) or to height (if below cloud base)	Baba (2019)
No distinction	Set to a constant value that is replaced when vertical velocity	$\delta^d = 2 \cdot 10^{-4}\ \mathrm{m}^{-1}$	Gregory (2001)

Type	Empirical value or assumption	Choices in the literature	Reference
	decreases with height, usually near cloud top		
	Only at levels of neutral buoyancy		Emanuel (1991)
	Only over a fixed layer of 60 hPa that extends from DDL to DBL	$\delta^d = 0\ \mathrm{m}^{-1}$ apart from the detrainment layer	Bechtold et al. (2001)
	Linear function of pressure between the top of USL and the base of the downdraft		Kain (2004)
	Proportional to the updraft convergence of the updraft mass flux		Gerard and Geleyn (2005)
	When downdraft becomes positively buoyant, with 75% of its mass detraining at each subsequent		Kim et al. (2013)
	Only in the lowest 1000 m above the ground or starting at LFC, whichever is located higher above the ground		Grell and Freitas (2014)

4.2.3 Impact of entrainment and detrainment on convective models

The discussion above illustrates the many nuances in the modeling of convection, the importance of empirical values in the final results and the need to further research to disentangle the many details involved. It is accepted that the parameterizations of entrainment and detrainment still have great uncertainties (Romps, 2010; Becker and Hohenegger, 2018) and problems in producing a realistic representation of convection (Mapes and Neale, 2011). For example, Siebesma and Holtslag (1996) evaluated a mass flux shallow cumulus based on BOMEX results and found that lateral entrainment and detrainment rates were one order of magnitude larger than those used in Tiedtke scheme (Tiedtke, 1989). Using an RCM over the Maritime Continent region, Wang et al. (2007) demonstrated that changes in the values of the fractional entrainment/detrainment rates in Tiedtke scheme affect the simulation of the tropical precipitation diurnal cycle. Over land, Del Genio and Wu (2010) used a CRM to study the transition from shallow to deep convection in diurnal cycles and inferred entrainment rates. Subsequently, the authors compared results from three different entrainment parameterizations to the results obtained with CRM and concluded that the best results were achieved by the entrainment parameterization of Gregory (2001). The entrainment rate depends on the parcel buoyancy, the convective updraft speed, and a free parameter representing the fraction of the buoyant turbulent kinetic energy generation used for entrainment. On the other hand, Stratton and Stirling (2012) improved the timing and amplitude of the diurnal cycle of tropical convection in the Met Office climate model by setting the entrainment as a function of the height of LCL.

Perhaps not surprisingly, MJO simulations are also sensitive to entrainment (Hannah and Maloney, 2011; Del Genio et al., 2012; Kim et al., 2012; Klingaman and Woolnough, 2014). Hannah and Maloney (2011) applied the RAS scheme in a GCM and analyzed the influence of minimum entrainment rate and rain evaporation fraction in the simulation of MJO. Larger values of any of the two parameters led to a better representation of the MJO and interseasonal variability, although higher values of

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minimum entrainment produced a drier and cooler atmosphere in contrast to the effect of higher values of rain precipitation fraction. Through a version of the Goddard Institute for Space Studies Global Climate Model (GISS GCM) with the entrainment rate proposed by Gregory (2001), Del Genio et al. (2012) efficiently reproduced the MJO transition from shallow to deep convection. Klingaman and Woolnough (2014) evaluated the effects of 22 model configurations and subgrid parameterizations on the simulation of MJO in the Hadley Centre Global Environmental model Global Atmosphere version 2 (HadGEM3 GA2.0) and tested the changes in 14 hindcast cases. A better representation of the MJO for both hindcast and climate simulations was achieved by increasing entrainment and detrainment rates for mid-level and deep convection. A better representation of MJO was also achieved by Kim et al. (2012) using a GCM to evaluate the tropical subseasonal variability. However, this improvement was at the expense of an increased bias in the mean state, typical for other GCMs with stronger MJO (Kim et al., 2011).

Other studies have evaluated the impact of entrainment/detrainment formulation on large-scale features, such as the double Intertropical Convergence Zone (ITCZ) (Chikira, 2010; Chikira and Sugiyama, 2010; Möbis and Stevens, 2012; Oueslati and Bellon, 2013). Möbis and Stevens (2012) used both the Tiedtke and Nordeng schemes in an aquaplanet GCM to evaluate the sensitivity of ITCZ to the choice of the convective parameterization. The Tiedkte scheme produced a double ITCZ, while the Nordeng scheme, with a higher lateral entrainment rate, led to a single ITCZ. In the works by Chikira (2010) and Chikira and Sugiyama (2010), the entrainment rate from AS was replaced by a formulation that depends on the surrounding environment following Gregory (2001) and Neggers et al. (2002). With this new formulation, variability and climatology improved, including the double ITCZ and the South Pacific Convergence Zone (SPCZ). Oueslati and Bellon (2013) obtained similar improvements in their study of the effects of entrainment on ITCZ by increasing entrainment in a hierarchy of models (coupled ocean–atmosphere GCM, atmospheric GCM, and aquaplanet GCM), at the cost of an overestimation of precipitation in the center of convergence zones. The role of entrainment on large-scale features was also underlined by Hirota et al. (2014) in their comparison of four atmospheric models with different entrainment formulations over tropical oceans.

Based on Zhang (2002) and using sounding data from the Coupled Ocean-Atmosphere Response Experiment (COARE), the South Pacific Convergence Zone (SGP97) and the Tropical Warm Pool – International Cloud Experiment (TWP-ICE), Zhang (2009) concluded that the entrainment of environmental air also affects CAPE and closure assumptions in CPs. The drier the entrained air, the stronger is the dilution effect that acts to reduce CAPE. Moreover, dilute CAPE shows a better correlation with the consumption of CAPE than undilute CAPE.

As mentioned in Sect. 4.2.2, less attention has been paid to the parameterizations of the detrainment process. Based on LES results for shallow convection, De Rooy and Siebesma (2008) proposed a new detrainment parameterization that led to improvements for ARM, BOMEX, and RICO shallow convection cases. Moreover, the authors revealed a greater variation in the detrainment rates from hour to hour and case to case than the variation in the entrainment rates. Derbyshire et al. (2011) confirmed this finding using a CRM and an adaptive detrainment model. Later, De Rooy and Siebesma (2010) showed that detrainment strongly influences the vertical structure of the mass flux.

4.3 Microphysics in convective clouds

The representation of microphysical processes in cumulus parameterizations is key to simulations of climate change (Ramanathan and Collins, 1991; Rennó et al., 1994; Lindzen et al., 2001). Convective microphysics greatly affects the representation of convective clouds due to its influence on detrainment of water vapor and hydrometeors, and the interaction between clouds and aerosols (Khain et al., 2005; Koren et al., 2005; Rosenfeld et al., 2008; Song and Zhang, 2011; Song et al., 2012; Tao et al., 2012). However, many convective parameterization schemes treat microphysical processes crudely, specifying an empirically determined conversion rate from cloud water to rainwater (Arakawa and Schubert, 1974; Tiedtke, 1989; Zhang and McFarlane, 1995; Han and Pan, 2011) or a certain precipitation efficiency like in Emanuel (1991) (see Table 9). A brief description of the main assumptions and empirical values used in the representation of microphysics in CPs is presented here for the same of completeness. For a detailed review of microphysics parameterizations, the reader is referred to Zhang and Song (2016) for convection and Tapiador et al. (2019a) for a full account.

Table 9: A sample of empirical values and assumptions used in precipitation efficiency.

Empirical value or assumption	Choices in the literature	Reference
Varies linearly between 150 mb and 500 mb	$PE = \begin{cases} 0, & p_b - p_i < 150 \text{ hPa} \\ \frac{p_b - p_i - 150}{350} & 150 \text{ hPa} < p_b - p_i < 500 \text{ hPa} \\ 1, & p_b - p_i > 500 \text{ hPa} \end{cases}$, where p_b is the pressure at cloud base	Emanuel (1991)
Function of the detrainment pressure	$PE = \begin{cases} 0.975, & p < 500 \text{ hPa} \\ 0.500 + 0.475\frac{800 - p}{300} & 500 \text{ hPa} < p < 800 \text{ hPa} \\ 0.500, & p > 800 \text{ hPa} \end{cases}$	Moorthi and Suarez (1992); Anderson et al. (2004); Li et al. (2018)
Function of wind shear and subcloud RH		Grell (1993); Grell and Dévényi (2002)
Varies with lower and middle troposphere RH		Emanuel (1994)
Proportional to a maximum precipitation efficiency PE_{max}	$PE^i = \left[1 - \frac{l_c(T^i)}{CLW^i}\right] PE_{max}$, where T^i is the in-cloud temperature, CLW^i is the in-cloud condensed water mixing ratio, and $l_c(T^i)$ is the temperature-dependent threshold condensed water value above which precipitation occurs. $PE_{max} = 1$ (EZ99) and 0.999 (BE01)	Emanuel and Živković-Rothman (1999); Bony and Emanuel (2001)
Function of wind shear and cloud base height		Bechtold et al. (2001)
Proportional to CCN	$PE \approx M_V^{\alpha_3 - 1} N_b^{\beta_3}$, where M_V is the total volume of condensed water accumulated over the cloud lifetime, N_d Nd is the droplet concentration, $\alpha_3 = 1.9$, and $\beta_3 = 1.13$	Jiang et al. (2010); Grell and Freitas (2014)

4.3.1 Conversion of cloud water to rainwater

Despite the importance of microphysical processes in the simulation of surface precipitation, radiation or cloud cover, only a few convection schemes attempt to realistically represent these processes. A common approach is to assume that a specified fraction of the condensate is instantaneously removed as rain. In Yanai et al. (1973) and Tiedtke (1989), the conversion rate

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from cloud water to rainwater is assumed to be proportional to cloud water mixing ratio q_l with an empirical function $K(z)$ conversion coefficient that depends on height, as shown in Table 10. Other assumptions include a constant conversion coefficient C_c (Arakawa and Schubert, 1974; Grell, 1993; Zhang and McFarlane, 1995) or define a temperature-dependent threshold water content l_{wc}, above which all cloud water is converted to precipitation (Emanuel and Živković-Rothman, 1999). Few schemes with a more realistic treatment of the conversion of cloud water to rainwater can be found in the literature on convection. Autoconversion of cloud water in the convection scheme is considered in Sud and Walker (1999), following Sundqvist (1978), as well as in Zhang et al. (2005). The latter included the autoconversion of cloud water and other microphysical processes for both cloud water and ice in the Tiedtke scheme. However, neither the size nor the number concentration of both hydrometeors is considered explicitly. This makes it impossible to account for aerosol-convection interaction, which is of great importance in climate simulations. To overcome this shortcoming, Song and Zhang (2011) and Song et al. (2012) added mass mixing ratio and number concentration of each hydrometeor in their parameterization. Another more realistic treatment of condensation is that proposed by Bony and Emanuel (2001). In this scheme, the condensed water produced at the subgrid scale is predicted by the convection scheme, while its spatial distribution is predicted by a statistical cloud scheme through a probability distribution function of the total water. Indeed, the parameterization of the microphysics is more comprehensively devoted to this specific problem.

Table 10: A sample of empirical values and assumptions used in the conversion of cloud water to precipitation.

Empirical value or assumption	Choices in the literature	Reference
Proportional to the liquid water content l_w and an empirical function $K(z)$ that depends on height z	$r = K(z) l_w$, where $k(z) = \begin{cases} 0, & z \le z_b + 1500\ \mathrm{m} \\ 2 \cdot 10^{-3}\ \mathrm{m}^{-1}, & z > z_b + 1500\ \mathrm{m} \end{cases}$	Yanai et al. (1973); Tiedtke (1989)
Constant conversion rate C_c		Arakawa and Schubert (1974); Grell et al. (1991)
	$r = C_c \frac{M_u l_w}{\rho}$, where $C_c = 6 \cdot 10^{-3}\ \mathrm{m}^{-1}$, M_u is the updraft mass flux, l_w is the liquid water content and ρ is the air density	Lord et al. (1982); Wu (2012)
	$r = C_c M_u l_w$, where $C_c = 2 \cdot 10^{-3}\ \mathrm{m}^{-1}$	Zhang and McFarlane (1995)
	$C_c = 2 \cdot 10^{-3}\ \mathrm{m}^{-1}$	Han and Pan (2011); Han et al. (2016) (use an exponential decaying rate of C_c below the freezing temperature)
No distinction between liquid and solid forms	The amount of condensate removal from the updraft depends on the mean vertical velocity in a layer of depth δz, and the concentration of condensate at the bottom of the layer	Kain and Fritsch (1990)
All water content in excess of a threshold of the cloud water content l_{wc} is converted to precipitation	$l_{wc} = \begin{cases} l_0, & T \ge 0\ ^\circ\mathrm{C} \\ l_0 (1 - T/T_c), & T_c < T < 0\ ^\circ\mathrm{C} \\ 0, & T \le T_c \end{cases}$, where $l_0 = 1.1\ \mathrm{g\ kg^{-1}}$ is a warm cloud autoconversion threshold, and $T_c = -55\ ^\circ\mathrm{C}$ is a critical temperature below which all cloud water is converted to precipitation	Emanuel and Živković-Rothman (1999)

Empirical value or assumption	Choices in the literature	Reference
Both liquid and solid precipitation depend on a condensate to precipitation conversion factor c_r and the in-cloud vertical velocity w	$\Delta r + \Delta i \propto 1 - exp(-c_r \Delta z / w)$, where Δi is the generation of solid precipitation, and $c_r = 0.02\ \mathrm{s}^{-1}$	Bechtold et al. (2001)
Convective precipitation depends linearly on cloud water content l_w and a function of temperature T and the cloud droplet number concentration (CDNC). It forms if the convective layer is at least 150 hPa deep		Nober et al. (2003)
Function of temperature T	$r = \begin{cases} a \cdot \exp[b\,T(z)], & T \leq 0\ °\mathrm{C} \\ a, & T > 0\ °\mathrm{C} \end{cases}$, where $a = 2.0 \cdot 10^{-3}\ \mathrm{m}^{-1}$ and $b = 0.07\ °\mathrm{C}^{-1}$	Han et al. (2016)

4.3.2 Evaporation in downdrafts

Downdrafts are greatly affected by evaporation of hydrometeors and detrained cloud droplets due to latent cooling. Therefore, a realistic representation of this microphysical process is needed. However, only a limited number of convective parameterizations, such as Emanuel (1991), include an explicit calculation of this process, as shown in Table 11. Instead, crude assumptions can be found in the literature. For example, the evaporation of hydrometeors is ignored in Yanai et al. (1973), while Tiedtke (1989) assumed an instantaneous evaporation of detrained cloud water. Other authors have related the evaporation in the downdraft to the precipitation rate (Betts and Miller, 1986) or avoided any microphysical formulation by assuming that the evaporation of rain acts to maintain a constant RH at each level (Fritsch and Chappell, 1980; Zhang and McFarlane, 1995). This allows evaporation to be calculated backwards. More sophisticated formulations include those of Kreitzberg and Perkey (1976) based on Kessler (1969), and Song and Zhang (2011) based on Sundqvist (1988).

Table 11: A sample of empirical values and assumptions used in the evaporation in the downdraft.

Empirical value or assumption	Choices in the literature	Reference
Detrained liquid water takes place at the same level where water detrains		Arakawa and Schubert (1974)
Related to the precipitation efficiency PE	$EVP = C_{evap}PE$, with $C_{evap} = -0.25$	Betts and Miller (1986)
Detrained cloud condensates evaporate immediately		Tiedtke (1989)
Function of the precipitation mixing ratio q_{prec} and environmental thermodynamic properties	$EVP = \frac{(1 - q_d^i / q_{sat}^i)\sqrt{q_{prec}^i}}{2 \cdot 10^3 + 10^4 / (p^i q_{sat}^i)}$ where q_d is the mixing ratio in the downdrafts, and q_{sat} the saturation mixing ratio	Emanuel (1991)
Assumed to maintain a constant RH at each level	$RH = 100\ \%$ (ZM95)	Zhang and McFarlane (1995)
Takes place when the RH is smaller than a certain threshold value	$RH < 90\ \%$	Bechtold et al. (2001)
Function of RH and the conversion of cloud water to rainwater r	$EVP = C_{evap}(1 - RH)r$, where $C_{evap} = 2.0 \cdot 10^{-4} (\mathrm{km\ m^{-2}\ s^{-2}})^{-1/2}\ \mathrm{s}^{-1}$	Wu (2012)

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4.3.3 Aerosols

Aerosols play a key role in the climate system due to their influence on the Earth's energy budget through absorption and scattering of solar radiation. Focused on microphysical processes, aerosols serve as cloud condensation nuclei (CCN) and ice nuclei (IN) and thus affect cloud properties, dynamics, and precipitation. However, aerosol-convection interactions are very complex processes, seldom included in convection microphysics. Zhang et al. (2005) developed a new parameterization accounting for the effects of aerosols in stratiform and convective clouds. This was later modified by Lohmann (2008) to include droplet activation by aerosols in terms of the updraft velocity *w*, temperature, aerosol number concentration, and size distribution, while ice nucleation is a function of *w*, aerosol properties, and air temperature. More recently, Grell and Freitas (2014) developed a new convective parameterization that includes an interaction with aerosols through an autoconversion of cloud water to rainwater dependent on CCN, parameterized in terms of the aerosol optical thickness (AOT) at 550 nm, as well as an aerosol dependent evaporation of cloud drops. The authors also included tracer transport and wet scavenging in their parameterization. This convection scheme is currently available in WRF.

5 Closure: strategies to close the budget equation

Closure consists in defining the intensity or strength of convection, i.e., the amount of convection regulated by large-scale variables. Therefore, it is essential to close the budget equations (Eq. (2)). Despite the number of hypotheses proposed in the literature, it is still considered an unresolved problem (Yano et al., 2013). The following subsections discuss the main closure types, as well as their main assumptions and empirical values. The impact of the closure formulation in convective model concludes the section.

5.1 Closure types

Existing convective closures can be classified into diagnostic, prognostic, and stochastic. While diagnostic closures relate cumulus effects to the large-scale dynamics at a particular time scale, prognostic closures perform a time integration of explicitly formulated transient processes. Stochastic closures include randomness elements to closure schemes, such as the first-order Markov process in Lin and Neelin (2003) or the Gaussian white noise in Stechmann and Neelin (2011). In the following, we focus only on deterministic closures.

5.1.1 Diagnostic closures

Diagnostic closures include different types of closures based on a certain physical variable that expresses the intensity of convection. Table 12 shows a sample of empirical values and assumptions used in the closure in the updraft. In moisture convergence schemes, moisture convergence or vertical advection of moisture are selected as the closure variable (Kuo, 1974;

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Anthes, 1977; Krishnamurti et al., 1980, 1983; Kuo and Anthes, 1984; Molinari and Corsetti, 1985; Tiedtke, 1989), therefore assuming that convection consumes the moisture supplied by the large-scale processes.

Table 12: A sample of empirical values and assumptions used in the closure in the updraft.

Main closure variable	Empirical value or assumption	Choices in the literature	Reference
Moisture convergence	Convection is controlled by the column-integrated water vapor		Kuo (1974); Tiedtke (1989); Gerard (2007)
CWF	QE assumption		Arakawa and Schubert (1974); Grell (1993)
	Relaxed at a certain time scale τ		Pan and Wu (1995); Lim et al. (2014) (includes a factor depending on the vertical velocity at the cloud base)
	Relaxed at a certain time scale τ and towards a CWF reference value	$CWF_{ref} = 10$ J kg^{-1}	Zhao et al. (2018)
CAPE	Consumed by convective activity at a certain time scale τ		Fritsch and Chappell (1980); Betts (1986); Betts and Miller (1986) (deep convection is suppressed if the precipitation rate is negative), Nordeng (1994); Gregory et al. (2000); Bechtold et al. (2001)
	Consumption proportional to heat and moisture sources		Donner (1993); Donner et al. (2001); Wilcox and Donner (2007)
	Consumed at an exponential rate by cumulus convection		Zhang and McFarlane (1995)
	Modified by the vertical velocity		Stratton and Stirling (2012)
Boundary-layer QE (CAPE)	QE between increased boundary layer moist entropy and decreased entropy due to moist downdrafts		Emanuel (1995); Raymond (1995)
	Cloud-base upward mass flux is relaxed toward subcloud-layer QE. Includes a fixed relaxation rate α and a convection buoyancy threshold δT_k	$\alpha = 0.02$ kg (m^2 s K)$^{-1}$ and $\delta T_k =$ 0.65 K (EZ99), 0.90 K (BE01)	Emanuel and Živković-Rothman (1999); Bony and Emanuel (2001)
Free tropospheric QE (dCAPE)	Convective and large-scale processes in the free troposphere above the boundary layer are in balance. Contribution from the free troposphere to changes in CAPE is negligible.		Zhang (2002); Zhang and Mu (2005a); Zhang and Wang (2006); Song and Zhang (2009); Zhang and Song (2010); Song and Zhang (2018)
Dilute CAPE	Consumed by convective activity at a certain time scale τ		Kain (2004); Neale et al. (2008); Wang and Zhang (2013); Walters et al. (2019)
PCAPE	Relaxation of an effective PCAPE that includes the imbalance between BL heating and convective overturning		Bechtold et al. (2014); Baba (2019)

The first parameterizations based on moisture convergence were too crude to produce results similar to those observed in nature, which led to the formulation of mass flux schemes. Early parameterizations lacked a theoretical framework to explain the interactions between the large-scale dynamics and convection or were incomplete, such as in Ooyama (1971). In an attempt to overcome this drawback, Arakawa and Schubert (1974) proposed a closed theory based on the QE of the CWF, which is similar to CAPE. Since then, many CPs use CAPE-like closures, generally assuming that the adjustment occurs at a relaxed

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time scale in contrast to the instantaneous adjustment proposed in Arakawa and Schubert (1974), among others. Table 13 lists the most important choices made for the relaxation time scale.

Table 13. A sample of the empirical values and assumptions in the relaxation time scale.

Empirical value or assumption	Choices in the literature	Reference
Varies within a specified range	$\tau = 10^3 - 10^4$ s	Arakawa and Schubert (1974)
	$0.5\ \mathrm{h} < \tau < 1\ \mathrm{h}$	Bechtold et al. (2001)
Set to a constant value	$\tau = 2$ h	Betts (1986); Betts and Miller (1986); Zhang and McFarlane (1995); Zhang (2002, 2003); Zhang and Mu (2005b); Zhang and Wang (2006); Song and Zhang (2009); Zhang and Song (2010); Stratton and Stirling (2012)
	$\tau = 3600$ s	Nordeng (1994)
	$\tau = 1$ h	Pan and Wu (1995)
	$\tau = 8$ h	Zhao et al. (2018)
Inversely proportional to cloud efficiency		Janjić (1994)
Function of the cloud depth CD, the vertical average updraft velocity $\overline{w}$ and an empirical scaling function f that decreases with horizontal resolution	$\tau = \frac{CD}{\overline{w}} f$. In B14 the minimum allowed value for τ is 12 min	Bechtold et al. (2008, 2014); Baba (2019)
Varies with a bulk RH over the cloud layer		Derbyshire et al. (2011)
Varies according to the large-scale velocity ω within the range 1200−3600 s	$\tau = \max\left\{min\left[\Delta t + max(1800 - \Delta t, 0) \times \left(\frac{\omega-\omega_4}{\omega_3-\omega_4}\right), 3600\right], 1200\right\}$, with $\omega_3 = -250/\Delta x$, $\omega_4 = 0.1 \cdot \omega_3$, Δx the grid size (in m) used in the model, and Δt the real model integration time step (in s)	Lim et al. (2014)
Dynamic formulation. Depends on the cloud depth *CD*, the grid resolution *Dx* and the in-cloud vertical velocity *w*	$\tau = \frac{CD}{w}\left[1 + ln\left(\frac{25}{Dx}\right)\right]$	Zheng et al. (2016)

Following Lin et al. (2015), CAPE-like closures can be classified into two types according to the decomposition and constraints applied to the closure variable: the flux type and the state type. In the flux type, the change of the CAPE-like variable is decomposed into its large-scale and convective components, with a much smaller change in CAPE compared to any of the flux terms. Of these types of closures, CAPE is the most commonly used closure variable in CPs (Fritsch and Chappell, 1980; Kain and Fritsch, 1993; Zhang and McFarlane, 1995; Gregory et al., 2000; Bechtold et al., 2001) with adjustment time scales varying from constant values to functional forms (Bechtold et al., 2008). Another CAPE-related closure is dilute CAPE, which adds dilution effects due to entrainment to the definition of CAPE. It is currently available in an updated version of the Kain scheme in WRF (Kain, 2004), as well as in CAM5 (Neale et al., 2008; Wang and Zhang, 2013), CAM6, and the Met Office Unified Model Global Atmosphere 7.0 (GA7.0) (Walters et al., 2019). While the preceding schemes applied convective closure to the full troposphere, Emanuel (1995) and Raymond (1995) proposed the so-called boundary-layer QE, where only the boundary layer component of the CAPE closure is considered. On the other hand, Zhang (2002) introduced a modified version of the

QE assumption, in which only dCAPE is employed as the closure variable, without considering the effect of boundary layer forcing. This type of closure, known as the free tropospheric QE or the parcel-environment QE, provides a better simulation of the diurnal cycle of precipitation than the boundary-layer QE (Zhang, 2003), as well as a better representation of MJO and ITCZ than the QE assumption used in the Zhang-McFarlane scheme (Zhang and Mu, 2005b; Zhang and Wang, 2006; Song and Zhang, 2009; Zhang and Song, 2010). More recently, Bechtold et al. (2014) developed a modified version of the free tropospheric QE hypothesis by adding a convective adjustment time scale for the free troposphere, as well as a time-scale-based coupling coefficient between the free troposphere and the boundary layer. The authors also replaced the dCAPE closure variable for PCAPE, defined as the integral over pressure of the buoyancy of an entraining ascending parcel with density scaling. The implementation of this closure in the ECMWF IFS led to a better representation of the diurnal cycle of precipitation.

In contrast to the previous flux-type closures, state-type closures decompose the change of CAPE-like variable into its boundary layer component and free troposphere component, with a much smaller change in CAPE compared to any of the state terms. The main representatives of state-type closures are the convective adjustment schemes of Betts (1986) and Emanuel (1994). Differences between these adjustment schemes are in the adjustment time scale and reference profiles selected for the adjustment. More recently, authors such as Khouider and Majda (2006, 2008) and Kuang (2008) applied this scheme only to the lower troposphere.

An alternative principle to QE is the so-called activation control proposed by Mapes (1997), in which the intensity of deep convection is controlled by inhibition and initiation processes at low levels, and closure is formulated in terms of CIN and the turbulent kinetic energy (TKE) (Mapes, 2000; Fletcher and Bretherton, 2010). However, as highlighted in (Yano and Plant, 2012b) this formulation is not self-consistent, which is a must, as models are intended to test physical hypotheses.

This section presented the assumptions and empirical values used in the formulation of the closure for updrafts. However, the magnitude of the downdrafts should also be addressed. In the schemes where it is included, it is commonly expressed as a fraction γ_d of the closure of the corresponding updraft, setting γ_d as a certain value (Johnson, 1976; Tiedtke, 1989; Baba, 2019). Alternatively, other authors have related γ_d to precipitation efficiency (Emanuel, 1995; Bechtold et al., 2001), the RH in the LFS (Kain, 2004) or proposed a formula for γ_d in terms of the total precipitation rate within the updraft (Zhang and McFarlane, 1995). Table 14 lists some of the empirical values and assumptions used in closure in the downdraft.

Table 14: A sample of empirical values and assumptions used in the closure in the downdraft.

Empirical value or assumption	Choices in the literature	Reference
Proportional to the updraft mass flux Mu	$M_d = \gamma_d M_u$, where $\gamma_d = 0.2$	Johnson (1976); Tiedtke (1989); Nordeng (1994)
	$\gamma_d = 0.1 - PE$	Emanuel (1989, 1995); Bechtold et al. (2001)
	$\gamma_d = 0.1 - RH$	Kain (2004)
	$\gamma_d = 0.3$	Baba (2019)

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Function of updraft mass flux Mu and re-evaporation of convective condensate		Grell (1993); Grell et al. (1994); Pan and Wu (1995)
Function of updraft mass flux Mu, height z, and maximum downdraft entrainment rate ε_{max}^{d}	$M_d(\mathrm{z}) = -\alpha\, M_b \frac{\exp[\varepsilon_{max}^{d} \cdot (z_{LFS}-z)]-1}{\varepsilon_{max}^{d} \cdot (z_{LFS}-z)}$, where α is a proportionality factor that depends on the total precipitation and evaporation rates	Zhang and McFarlane (1995) (downdraft ensemble is constrained both by the availability of precipitation and by the requirement that the net mass flux at cloud base be positive)
	$M_d(\mathrm{z}) = -\alpha\, M_{d(LFS)} \frac{\exp[\varepsilon_{max}^{d} \cdot (z_{LFS}-z)]-1}{\varepsilon_{max}^{d} \cdot (z_{LFS}-z)}$, with $M_{d(LFS)} = 2\,(1-\overline{RH_{LFS}})\, M_{u(LFS)}$, where RH_{LFS} is the mean (fractional) RH at LFS, $M_{u(LFS)}$ is M_u at LFS, and $\varepsilon_{max}^{d} = 5 \cdot 10^{-4}\ \mathrm{m}^{-1}$	Wu (2012)

5.2.2 Prognostic closures

Compared to the QE assumption used in the majority of the diagnostic closures mentioned above, prognostic closures do not distinguish between large-scale and convective processes and substitute the QE assumption with time integration of prognostic equations. These equations explicitly account for the time changes of different physical variables, i.e., convective kinetic energy or h, which are related to the cloud-base mass flux through a dimensional parameter. Energy dissipation rate is also included in this type of closure through a dissipation term, either determined by a second dimensional parameter called dissipation time (Randall and Pan, 1993; Pan and Randall, 1998; Yano and Plant, 2012a) or expressed in terms of the entrainment rate and an aerodynamic friction coefficient (Gerard and Geleyn, 2005).

5.2 Impact of closure on convective models

The closure problem is one of the major challenges in CPs. As well as being essential to close the budget equations (Eq. (2)), it plays an important role in the performance of CPs. For instance, replacing the CAPE closure used in the Zhang–McFarlane scheme by a dCAPE closure, together with the addition of an RH threshold for convection trigger and the removal of the restriction in the convection originating level, the simulated MJO is more consistent with the observations in terms of variability in precipitation, outgoing longwave radiation and zonal wind, and exhibits a clear eastward propagation (Zhang and Mu, 2005b). However, the precipitation signal and the time period of the MJO differ from the observations. This revision of the Zhang–McFarlane scheme used in the NCAR CCSM3 also alleviates the biases related to the double ITCZ in precipitation and cold tongue in Sea Surface Temperature (SST) over the equator, among other benefits (Zhang and Wang, 2006; Song and Zhang, 2009; Zhang and Song, 2010). Other processes related to the ENSO and the diurnal cycle of precipitation are also known to be sensitive to the convective closure used in CPs (Zhang, 2002; Neggers et al., 2004; Wu et al., 2007; Bechtold et al., 2014; Yang et al., 2018).

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6 Conclusions

Numerical models need simplifications to be able to cope with the complexity of the physical processes actually ocurring in the atmosphere. The degree of simplification in the physics is evolving inversely to the availability of computational power. Thus, early convective parameterizations (as well as parameterizations of radiation, turbulence, microphysics, etc.) were based on very simple assumptions, such as the conditional instability of the second kind (CISK), first presented by Charney and Eliassen (1964) and Ooyama (1964). CISK states that cyclones provide moisture that maintains cumulus clouds, and cumulus clouds provide the heat that cyclones need. Despite its simplicity, this parameterization achieved acceptable results in the simulation of the life cycle of tropical cyclones (Ooyama, 1969). Simulations improved with further refinements of the interaction of cumulus clouds with the large-scale environment by, for instance, Ooyama (1971) (a statistical ensemble of bubbles represent cumulus convection), Yanai et al. (1973) (detrainment and cumulus-induced subsidence), and Arakawa and Schubert (1974) (cloud work function and adjustment towards quasi-equilibrium). With the increase in computational power, more complex parameterizations and new variables based on observations can be used to achieve better spatial and temporal resolutions within models. Thus, convective parameters require fine tuning, but there is no explicit methodology to do so. In some cases, the authors use the variables that are easiest to measure. In others, mean values describe processes that cannot be modeled in sufficient detail, or the values represent particular conditions for certain locations and atmospheric events (Mauritsen et al., 2012). For instance, Bony and Emanuel (2001) adjusted their water vapor and temperature prediction using the TOGA-COARE data measured in Western Pacific Ocean in 1993, while Betts and Miller (1986) used GATE datasets measured over the tropical Atlantic Ocean in 1974 to develop their deep convection scheme. Hence, empirical values and assumptions selected this way might yield good results when compared to observations from certain locations and less good results for others. Commonly, manual tuning of convective parameters is used, although various automatic methods have recently been used to estimate parameters, including the variational method (Emanuel and Živković-Rothman, 1999), Bayesian calibration (Hararuk et al., 2014; Wu et al., 2018), simulated annealing method (Jackson et al., 2004, 2008; Liang et al., 2014), genetic algorithm (Lee et al., 2006), or ensemble data assimilation (Ruiz et al., 2013; Li et al., 2018), among others.

Comparisons with observations were, and still are, crucial to the development of convective parameterizations. For instance, the underprediction of large-scale precipitation by dry adiabatic models compared to observations led to the inclusion of moist adiabatic processes in NWP models (Smagorinsky, 1956), and the lake-effect snow observations (Niziol et al., 1995) forced to reduce the minimum cloud-depth threshold in Kain and Fritsch (1993) to 2 km. Although observations can be used to tune parameters in convective schemes to reduce errors, it is unclear whether these tuned parameters based on particular datasets can improve model skills across different locations, model resolutions or atmospheric events. Moreover, it is known that model results are sensitive to the empirical values in convection. Numerous sensitivity studies have reported that the location and intensity of precipitation are extremely sensitive to cumulus parameterization (Bechtold et al., 2008; Ma and Tan, 2009; Chikira and Sugiyama, 2010). For instance, Wang et al. (2007) improved the simulated diurnal cycle over land and ocean by increasing the entrainment/detrainment rates for deep and shallow convection used in the Tiedtke scheme, which tends to simulate

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convective precipitation too early in the day and with an unrealistic amplitude over land. Thus, the choice of a convective scheme impacts the diurnal cycle (Bechtold et al., 2004; Wang et al., 2007), as well as the simulation of monsoon precipitation in climate models (Mukhopadhyay et al., 2010), the MJO (Lin et al., 2006), the ENSO (Wu et al., 2007; Neale et al., 2008), and the ITCZ configuration (Liu et al., 2019). This topic has profound practical effects: it has also been shown that choices in the convective parameterization affect the prediction of track, intensity and associated rainfall of tropical cyclones (Mohandas and Ashrit, 2014). Indeed, timely providing the correct amount of precipitation at the right location is still a challenge for models. Figure 2 is an example of how different the precipitation field may look depending on the cumulus parameterization used. All a priori sensible methods locate the maximum and minima in different parts of typhoon Chaba and predict different areas and total accumulations. In the climate model realm, validation exercises focusing on precipitation (Tapiador et al., 2012, 2017, 2018) have shown the importance and challenges of comparing model outputs with precipitation measurements in order to improve model performance. Indeed, the difficulties of quantitative precipitation estimation suggest precipitation as a privileged metric to gauge model performance (Tapiador et al., 2019b). The "ultimate test", as has been described, makes precipitation science an active field of research. As discussed in such paper, there is no complete agreement even in the reference data, with datasets differing even in such aggregated value as the global mean value of the precipitation on Earth. Advances in satellite precipitation estimation (Kummerow et al., 1998; Joyce et al., 2004; Okamoto et al., 2005; Ushio and Kachi, 2010; Watanabe et al., 2010, 2011; Kucera et al., 2013; Hou et al., 2014; Huffman et al., 2015; Xie et al., 2017; Levizzani and Cattani, 2019; Skofronick-Jackson et al., 2019) are indispensable to advance further, since direct estimates of precipitation (pluviometers, disdrometers) and ground radars are limited to land areas. These advances need to be parallel with an explicit account of what is empirical in models in order to benefit both fields. Algorithm developers in the satellite realm are perhaps more used to specifying their assumptions through the Algorithm Theoretical Basis Documents (ATBD) but a full comparison between the physics and empirical values behind both algorithms and parameterizations is much needed to advance the field. On that note, it is clear that better access to climate models code would contribute to address scientific gaps in climate models and to improve their reliability (Añel et al., 2021). It would be also highly desirable that scientists not only specify the parameterizations they have used, but also the assumptions and empirical values they have actually selected within these. Tables 2-11 can be used to easily identify and pinpoint their choices. The benefit will be immense as some discrepancies could be readily attributed to known issues (i.e. heavy spurious rainfall over warm water in adjustment schemes) or identified as cofounding variables. As in the case of the microphysics, making transparent the codes, the assumptions and the empiricisms can only benefit the community and dispel any potential concerns.

Indeed, the focus of this paper is not comparing the publicly available convection schemes or to lean users towards one or another but to explore the Physics behind the modules, and to do that from an objective and independent point of view. Neither is the paper about criticizing the simplifications that are inherent to modeling the atmosphere, or the limitations of current methods. On the contrary, the research arises from the conviction that models are the way forward to advance climate research. Being aware of the potential misuse of the results shown here to attempt discrediting models, it is important to vaccinate

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uninformed critics and discourage futile attempts: neither this paper nor Tapiador et al. (2019a) cast any shadow on model outputs. On the contrary, they display and celebrate the delicate intricacies, nuances, precise measurements and careful choices made by the community to craft complex tools to forecast, simulate and predict precipitation.

Code and data availability

There is no code or data relevant to this paper.

Author contributions

Conceptualization, F.J.T. and A.V.P.; Funding acquisition, F.J.T.; Investigation, F.J.T. and A.V.P; Methodology, F.J.T. and A.V.P; Supervision, F.J.T.; Writing – original draft, A.V.P.; Writing – review & editing, F.J.T. and A.V.P.

Competing interests

The authors declare that they have no conflict of interest. They have not participated in the development any existing convection module or engaged in any collaboration or discussion with their developers in order to prepare this paper. Their review is an independent, purely objective analysis based on literature and stays neutral on the suitability or performances of any of the parameterizations for any alleged purpose.

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