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Phenomenology of convection-parameterization closure

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Abstract

Closure is a problem of defining the convective intensity in a given parameterization. In spite of many years of efforts and progress, it is still considered an overall unresolved problem. The present article reviews this problem from phenomenological perspectives.

The variables that are expected to contribute in defining the convective intensity are listed, and their statistical significances identified by observational data analyses are reviewed. A possibility is discussed for identifying a correct closure hypothesis by performing a linear stability analysis of tropical convectively-coupled waves with various different closure hypotheses. Various individual theoretical issues are considered from various different perspectives. Finally, it is emphasized that the dominant physical factors controlling convection differ between the tropics and extra-tropics, as well as between oceanic and land areas.

Both observational as well as theoretical analyses, often focused on the tropics, do not necessarily lead to conclusions consistent with our operational experiences focused on midlatitudes. Though we emphasize the importance of the interplays between these observational, theoretical and operational perspectives, we also face challenges for establishing a solid research framework that is universally applicable.

1 Introduction

The importance of convection parameterization both in numerical weather forecasts as well as climate projections can hardly be overemphasized. There are various fundamental issues to be addressed in order to make it more robust.

In general, convection parameterization can be considered as consisting of the two major parts: (1) regulation of the amount of convection by large-scale (grid scale) variables and (2) regulation of the large-scale variables by convection. These two problems are usually called closure and the cloud model, respectively. Putting it differently,

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convective closure is a problem of finding a relation of the intensity of the subgrid-scale convective activity to large-scale variabilities (model-resolved variables). As emphasized by Arakawa and Schubert (1974), "*The real conceptual difficulty in parameterizing cumulus convection starts from this point*". Although extensive progress has been made since that time (e.g. Arakawa and Chen, 1986; Xu, 1994), the statement is still valid even today (cf. Arakawa, 2004; Yano et al., 2005a). The problem may furthermore include the issues of conditions for triggering and suppression of convection under a given scheme.

The present paper reviews the closure problem from phenomenological perspectives, rather than as being a formulational issue. Typically, a closure is based on a steady-state budget of a certain physical variable that is expected to control the convective evolution. Thus the key physical question in the present review is: what controls convection?

The present paper consists of the three major sections. The following two sections (Sects. 2 and 3) examine the closure problem from a perspective of observational analysis and the tropical large-scale dynamics, respectively. It may be argued that the first step for constructing a closure is to identify a variable that controls convection observationally. Section 2.3 reviews such data analysis after identifying the role of convection in global observational perspectives in Sect. 2.1, then listing potential control variables for convection in Sect. 2.2.

Section 3 examines the closure problem from a perspective of the tropical large-scale dynamics. Various types of large-scale convective variabilities (so-called convectively-coupled waves) over the tropics may be understood in terms of coupling between convection and large-scale dynamics. Under these theories, a right closure must be chosen in order to explain these variabilities properly. Conversely, the consistency of a theory with observations provides an ample test for an adopted closure for a theory. Thus, a review of tropical wave theories provides good general insights on the closure problem.

Section 4, in turn, examines the individual closure hypotheses from theoretical perspectives by focusing on individual physical processes in concern. This section intends to provide a good contrast with general discussions in the precedent two sections.

Under the mass flux formulation (Arakawa and Schubert, 1974), the closure problem can be more formally stated in the following manner. A core of the mass flux formulation is in the quantity called, mass flux, M. Once the mass flux at cloud base is defined, the convective mass flux through the remainder of the atmosphere is computed using the chosen cloud model. From the resulting mass flux profile, more or less all of the convective tendencies required for a large-scale model can then be obtained with certain caveats to be referred to Donner (1993), and Yano and Plant (2012a). Here, the mass flux, M, is defined as an upward momentum flux associated with convection averaged over a given horizontal area. The area in mind, which is expected to correspond to the grid box size, is traditionally taken to be much larger than individual clouds (i.e. scale separation principle). However, as the resolution of the model increases, this assumption also begins to break down. Issues of the closure in the high-resolution limit will more specifically be discussed in Sect. 4.6 (see also Bister, 1998).

As a standard procedure, the mass flux, M, is divided into the two components, a normalized vertical profile, $\eta(z)$, and a time-dependent amplitude, $M_B(t)$:

$$M = \eta(z)M_B(t).$$

Here, the convective amplitude (or convective intensity), M_B, is usually defined as the mass flux at the convection base. The convective vertical profile, η, is defined by a *cloud model*, which usually constitutes of a specification of entrainment and detrainment, or mixing rate of convection with the environment (cf. de Rooy et al., 2012). The closure, in turn, defines the convective amplitude, or convective intensity, M_B. This is the main concern of the present review.

Many related issues are not discussed. For example, in the original mass-flux formulation by Arakawa and Schubert (1974), a simple entraining plume is adopted as a cloud model. To be realistic, the cloud model should also include various additional

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elements, including downdrafts. In practice, it is not possible to consider the closure problem of specifying the convective intensity, or more specifically cloud-base convective updraft mass flux, without specifying a cloud model, because the behavior of a prameterization would simply be changed by a choice of the latter even under a same closure. For example, if the cloud model does not give the sensitivity to moisture by entrainment and does not contain an unsaturated downdraft, then the closure has to compensate for the crudeness of the cloud model. Note that the downdraft provides a further process for drying and cooling the boundary layer in addition to the convective updraft. Thus, broadly speaking, the absence of the former must be compensated by increasing the convective updraft in the closure.

In order to suggest a richness of the closure problem, Sect. 5 is devoted to differences in the environments controlling convection around the globe. The paper is concluded in Sect. 6.

2 Observational perspectives

We first consider the role of moist convection from a perspective of global heat budget in the first subsection. A list of variables potentially contributing for control of convection is presented next. We may argue that from an observational perspective, the closure problem reduces to that of identifying a nonlinear function of these control variables that defines convective intensity. Existing data analyses are reviewed in the last subsection from this perspective.

2.1 Moist convection in global perspectives

Climatological energy balance of the globe (Newell et al., 1974) shows that latent heat release is needed in opposing the radiative cooling. Even though there is transfer of heat from lower to higher latitudes by the atmosphere and oceans, still the effect of condensation (and freezing/deposition) is crucial in balancing the heat loss by radiation

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in the free troposphere over most of the globe. This does not mean that in order to maintain climatological energy balance, condensation itself should occur everywhere. Atmospheric convection is often very localized especially in the tropics. The adiabatic heating associated with compensating subsidence can locally balance with radiative cooling without invoking local latent heating. Excess latent heating associated with local convection would be balanced with adibatic cooling associated with convective ascent. Thus, the latent heating in localised convection and radiative cooling over much wider areas are together balanced by compensating subsidence. Gravity waves are expected to redistribute the convectively-generated heat anomalies horizontally on fast time-scales. However, gravity waves can propagate straight only up to the scale of the Rossby deformation radius, and in general less, when dissipation is accounted for. As a result, the dispersion of heating depends on the latitude: the Rossby deformation radius is about 1200 km at 15° N and 400 km at 50° N, using values from Bretherton et al. (2005).

On the other hand, in the middle latitudes, the situation is different due to the dominance of the baroclinic waves and horizontal advection, that can transport heat from one region to another. Outside regions of forced ascent, being induced either by baroclinic waves or topography, deep convection plays an important role in vertical transport.

The importance of latent heating can be understood if one considers that, for example, over the tropics, CAPE (convective available potential energy) would increase by about 700 $\mathrm{J\,kg^{-1}}$ every day if radiative cooling (assumed to be 1.4 $\mathrm{K\,day^{-1}}$ from 900 to 150 hPa) were not compensated for by latent heating (cf. Emanuel et al., 1994). In order to prevent steady increase of CAPE, that is not observed, moist convection must consume CAPE in compensation. A basic state of the free troposphere and the boundary layer must be, in turn, such that a rather small change can easily trigger moist convection frequently enough. Alternatively, an externally-imposed large-scale ascent would be required in order to maintain the required condensation rate. A major exception to

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this rule is high-latitude regions in the free troposphere of the summer hemisphere where the total radiative heating can be even positive.

Recall that convection is defined as "thermally direct circulations which result from the action of gravity upon an unstable vertical distribution of mass, with vertical taken to mean along the gravitational vector" (Emanuel, 1994). It follows that the important variables for convection are: (1) the boundary moist entropy (2) free tropospheric temperature and (3) moisture in the low-to-middle troposphere (LTMT). The last variable may not be obvious directly from the above definition, but it comes from the known effects of entrainment on convection and indirectly from feedback of convection on subsequent convection (cf. Sect. 5, ii).

2.2 List of variables and physical basis

Atmospheric moist convection is locally controlled by various physical processes and associated physical variables. By further elaborating the discussions of Sect. 2.1, here, we list a more complete list of these variables in the order of stability measures, thermodynamic and dynamics variables:

1. the vertical temperature gradient: the simplest stability measure in analogy with dry-convective instabilities (cf. Stone and Carlson, 1979; Zhang and Klein, 2010).

2. CAPE (convective available potential energy): by definition, convection is driven by buoyancy, thus it is natural to expect that the convective intensity is controlled by strength of buoyancy acting on a convective flow. The easiest way to measure the strength of buoyancy forcing on convective flows is to take a simple lifting parcel argument, which leads to a definition of CAPE. There are various ways for lifting an air particle from the surface (or a middle of boundary layer), leading to various different definitions for CAPE.

2a. Undiluted CAPE: the simplest, and the standard procedure for diagnosing CAPE from sounding is to lift an air particle from the surface without any mixing with the environment.

2b. Diluted CAPE: a more "realistic" estimate can be made by mixing the air particle with the environment at a certain rate. An updated version of the Kain-Fritsch scheme (Kain, 2004) takes it for closure.

2c. CIN (Convective INhibition): it is defined by a vertical integral of negative part of parcel-lifted buoyancy (cf. Sect. 4.4). Raymond et al. (2003) propose DCIN (Deep Convective INhibition) as a similar measure of inhibition focussing on deep convection.

3. Cloud-work function (Arakawa and Schubert, 1974): it is work performed by convective buoyancy per convective mass flux defined at the convection base. It is equal to a vertical integral of vertical buoyancy flux normalized by a convection-base convective vertical momentum (mass flux). The cloud work function provides a more accurate estimate than CAPE for an efficiency of convective kinetic energy generation by buoyancy forcing (cf. Sect. 4.5).

3a. PEC (potential energy convertibility): the cloud work function can be estimated both from cloud-resolving modes and large-eddy simulations by replacing the convection-base mass flux by an alternative normalization factor for the convective vertical momentum, as proposed by Yano et al. (2005b).

4. GCAPE (Generalized Convective Available Potential Energy: Randall and Wang, 1992): it is defined as part of the available potential energy (APE) calculated for the moist atmosphere. APE is defined as the difference between the total enthalpy of the given atmospheric state and that of the reference state (Lorenz, 1978, 1979). Generally, the reference, defined as the minimum enthalpy state, is sought by re-shuffling the air parcels both horizontally and vertically. Here, more precisely,

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GCAPE is defined as an APE considering only the vertical redistribution of air parcels in order to define a reference state.

5. GMS (Gross Moist Stability): this concept originally introduced by Neelin and Held (1987) is a measure of moist convective instability based on a vertical integral of the moist static energy (cf. Sect. 3).

6. Boundary moist entropy.

7. Free tropospheric temperature.

8. Water vapor:

8a. Saturation fraction: the column-integrated water divided by that of the saturated column with the same temperature profile.

A focus may, furthermore, be placed to a particular vertical layer, leading to various more specific measures:

8b. Water-vapour mixing-ratio in the PBL (planetary boundary layer): this is where the majority of water vapor is found in the atmosphere. Under a standard argument of the lifting parcel theory, it is air lifted from this layer saturated and leading to condensation of water vapor originated from PBL.

8c. Water-vapor deficit in lower troposphere : defined as a difference between the saturated specific humidity for the given observed (or modeled) temperature and the actual observed (or modeled) specific humidity (Redelsperger et al., 2002).

8d. Water vapor in the low-to-middle troposphere (LTMT).

9. Vertical wind shear: it is observationally well known that the vertical wind shear tend to organize convection and to increase its longevity (cf. Klemp, 1987), and this tendency can also be demonstrated by numerical modelling (e.g. Weisman and Klemp, 1982). Various theories are developed (e.g. Moncrieff and Green,

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1972; Thorpe et al., 1982; Rotunno et al., 1988). Especially, the first two works link the wind shear with the energy-cycle of convection described by CAPE (cf. Sect. 4.3).

Among those, CAPE and water vapor (moisture) are the two most commonly adopted variables for closures.

2.3 Observational identification of convection-controlling variables

We have argued that from an observational perspective, the closure problem reduces to that of identifying a nonlinear function of these control variables that defines convective intensity. Clearly a first step towards this goal is to identify the variables that control convective intensity observationally. Existing data analyses can be reviewed from this perspective.

A "statistical" analysis by Sherwood (1999) would probably be the first example of such systematic analyses over the tropics. A multivariate analysis is performed on the sounding and the satellite data sets over the Tropical Western Pacific. Low- to midtropospheric moisture is identified as the dominant factor regulating convection under this statistical analysis. No other variables, including CAPE, presents any significance.

A similar extensive statistical analysis is performed by Zhang and Klein (2010) for afternoon showers over the North American Southern Great Plains. A particularly interesting piece of information is found in their Fig. 9. This figure shows a statistical test of control parameters for triggering afternoon convective shower: longer bars have more statistical significance with positive and negative correlations to the left and the right. This analysis again shows that most of the above quantities do not show any significant role in defining a shower over a certain afternoon over the Southern Great Plains. An only noticeable correlation is found with the relative humidity both in the boundary layer and the lower free troposphere, a mid-level meridional wind, and surface temperature standard deviation. CAPE has no significant role, and the role of CIN is negligible. On the other hand, interestingly, a mean temperature gradient (cf.

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Sect. 2.2, 1) over 2–4 km represents a marginal significance. However, we should be cautious with the role of moisture in controlling convection, because according to their Table 1, the lower-level relative humidity is not significantly correlated either with the total rain or maximum rain.

Barkidija and Fuchs (2012) based on measurements for 20 European stations for the period of 1972–2009 and the results of Global Forecasting System (GFS) model further shade light on this issue. It shows that CAPE does not correlate with precipitation rate at any place of the globe, including the tropics, the middle latitudes, as well as the higher latitudes. CIN shows a clear positive correlation, but the authors tend to suspect this is an artifact from the analysis (cf. Sect. 4.4). On the other hand, the saturation fraction (cf. Sect. 2.2, 8a) has a nice correlation with precipitation rate although in higher latitudes it does not represent a well-defined function as found in the tropics by Raymond et al. (2007).

As the above relatively systematic studies show, the water vapor (or saturation fraction) is almost the only variable that shows a clear correlation with convection. Notably, increases in lower-tropospheric water vapor leading increase in precipitation as shown by Raymond (1995); Brown and Zhang (1997); Sherwood (1999); Sherwood and Wahrlich (1999); Sobel et al. (2004); Mapes et al. (2006, 2009). Two theories are proposed by Raymond (2000) and Peters and Neelin (2006: see also Yano et al., 2012). However, we emphasize that this correlation should necessarily not be taken as a causality as it becomes clearer as we discuss further.

On the other hand, observations in the tropics generally find that heavy rainfall is not correlated with CAPE in an obvious manner, but with decreased CAPE (Ramage, 1971; McBride and Frank, 1999). Figure 8 of Xu and Randall (1998) also shows a similar point. Here, they take GCAPE rather than CAPE for this purpose: it shows that GCAPE is almost completely out of phase with convective precipitation. Thus, (G)CAPE does not appear as a good measure of convective instability. A similar point is also made by Thompson et al. (1979); Wang and Randall (1994). We refer to Sects. 4.3 and 4.5 for a possible explanation for such CAPE behavior.

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In concluding this subsection, however, we have to recognize a major difficulty in identifying "convection" from conventional measurements including satellite. Sherwood (1999) simply uses the satellite-measured infrared brightness temperature as a measure of convection. On the other hand, most of the other analysis mentioned above use the precipitation as a measure of convection. Both the cloud height (as measured by the brightness temperature) and precipitation can be considered a good measure of convective activity over the tropics. However, a straight use of precipitation as a measure of convection in midlatitudes and higher latitudes become questionable, because the precipitation is strongly controlled by the synoptic-scale processes. Zhang and Klein (2010) partially avoid this difficulty by limiting their analysis to the summer-time of the Southern Great Plains, where the precipitation is known to be predominantly convective during this season.

Unfortunately, more direct measurements of, say, convective updrafts, are not possible with a conventional observational network. For this purpose, we need a special-type of radar with a Doppler capacity, or probably better still, direct measurements of updrafts by flight penetrations into convection. Both measurements are, however, rare. This is probably the most fundamental difficulty for identifying a closure formula from the observations.

3 Perspective from the tropical large-scale dynamics: inferences from linear stability analysis

As the discussions of the last section shows, it is not at all easy to identify a nonlinear function of convection-controlling variables that defines convective intensity from observations. Thus, a more theoretically-based general approach must be taken.

As an example of such an alternative theoretical approach, in the present section, we consider a theory for the tropical large-scale dynamics (cf. Yano and Bonazzola, 2009). A guiding principle to adopt for this purpose is a hypothesis that tropical large-scale convective variability is controlled by linear instabilities arising from coupling between convection and large-scale dynamics (but see also Yano et al., 2009; Delayen and Yano, 2009). Under this theoretical framework, various closure hypotheses can be introduced, and the obtained instability characteristics of a theoretical system under a given closure can be examined. We expect that a consistency of the obtained instability characteristics with observations provide a measure of physical relevance of the adopted closure.

We would like to expect that these theoretical analyses lead to more robust conclusions than the afore-mentioned observational studies. However, it would be fair to say that one has to inevitably take a particular theoretical approach, especially for a cloud model, so a question of the generality of the results always remains.

Fuchs and Raymond (2002, 2005, 2007), and Raymond and Fuchs (2007) have performed extensive work on large-scale waves in the tropics. These studies are specifically reviewed here. As it turns out, from a point of view of coupling of the large-scale disturbances and convection in the tropics CAPE closure is irrelevant. Especially, they found that a CAPE closure does not produce any interesting large-scale solutions. On the other hand, they found that for convectively coupled Kelvin Waves, convective inhibition (CIN) plays an important role.

Furthermore, a moisture closure makes an impact. This result suggests that the so-called moisture mode might be an underlying mechanism for many tropical disturbances such as the Madden-Julian oscillations (MJO), easterly waves, westerly wind bursts. This moisture closure includes the dependence on gross moist stability (GMS): negative gross moist stability acts in a way that produces the instability in the moisture modes. If the models don't have the moisture closure, many disturbances in the tropics cannot be reproduced.

The instability mechanism of those large-scale disturbances is examined. Those disturbances include the Kelvin waves, Rossby waves, inertio-gravity waves, moisture mode. Different precipitation closures are applied, and examined their impact. In this manner, it is found that CAPE didn't affect the linear dry or moist modes, but they were decayed by the CAPE closure.

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Here, GMS can be understood in terms of moist static energy as it was originally introduced by Neelin and Held (1987). Alternatively, it can also be defined in terms of moist entropy: a definition adopted here. Arguably, GMS is an important part of moisture closure as a way of destabilization. The other possibility is to invoke radiative-convective instability (RCI). In the latter case, a moisture closure can be considered without GMS. In that case the moisture modes are equivalent to the modes obtained by using the weak temperature gradient approximation (Sobel et al., 2001), i.e. to assume that moisture perturbation is larger than the temperature perturbation. If there is no RCI or GMS, the moisture modes are stable. Another important ingredient is to take into account the two vertical modes in stability analysis.

A CAPE-based adjustment description only gives a damping mode (i.e. moist convective damping: Emanuel et al., 1994). It is found that even when an extra effect of wind-induced surface heat exchange (WISHE: Yano and Emanuel, 1991) is added, no instability is found for realistic wavelengths. Though the result virtually contradicts with an earlier study on WISHE instabilities assuming a single vertical mode (Emanuel, 1987; Neelin et al., 1987; Yano and Emanuel, 1991), the result is also consistent with Mapes (2000) and Majda and Shefter (2001). Nevertheless, WISHE plays a role in providing a propagation mechanism to moisture mode.

The linear stability analysis reviewed in this section tends to confirm the observational diagnosis reviewed in the last section: the moisture closure predicts instabilities consistent with observations, whereas the CAPE closure only leads to damping modes. The WISHE mechanism does not contribute to an instability either. An important role of CIN and GMS is suggested.

4 Further theoretical considerations

In this section, attempts are made to discuss pros and cons for both moisture and CAPE based closures from more general perspectives. Issues with the high-resolution limit are discussed separately in Sect. 4.6.

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4.1 Moisture-based closure

The moisture-based closure loosely assumes that a large-scale supply of moisture balances with a consumption by convective processes. However, the definition of “large-scale supply” varies from a scheme to scheme. The simplest choice is to consider only the large-scale convergence, but many schemes also include the surface flux effect. One may argue that moisture is central, because deep convection is moist; its triggering involves bringing moist air to saturation. Deep convection is part of the water cycle.

It is fair to say that the moisture-based closure has been a popular approach since its original proposal by Kuo (1974). This idea is intimately related in people’s mind with the notion of large-scale uplifting or lower-level convergence leading to moist deep convection. Such a process leads to water condensation that triggers convection. For this reason, the moisture-based closure may even be conceptually replaced by an assumption of convection proportional to the low-level large-scale convergence. The latter is a popular idea originated from CISK (conditional instability of the second kind: Charney and Eliassen, 1964), then it is generalized into wave-CISK (Hayashi, 1970, 1971; Lindzen, 1974).

However, it is rather within a narrow context of tropical deep convection in which low-level large-scale convergence is invoked as a triggering mechanism. A trivial example for demonstrating the irrelevance of low-level large-scale convergence for convection is shallow convection (e.g. stratocumulus-topped boundary-layer convection: Lilly, 1968; Schubert et al., 1979). Those non-precipitating convective clouds are typically maintained under large-scale descent. In the literature, as far as the authors are aware, large-scale uplifting is never mentioned as a mechanism for transformation of stratocumulus-topped boundary-layer convection into trade cumuli, for example.

It is true that non-precipitating convective clouds are also maintained by moisture supplied from the surface. However, the moisture transport is almost exclusively taken care by eddy vertical transport, an aspect that is totally neglected in standard moisture

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closures. In this respect, a picture of convection driven by large-scale convergence only has a limited applicability, and it may not always true even in deep-convection contexts.

As discussed in Sect. 2.3, the availability of water vapor in a given atmospheric column is a good indicator of convective precipitation over the tropics. Theoretical studies on tropical convectively-coupled linear waves in Sect. 3 also favor the moisture-based closure. On the other hand, as emphasized by Emanuel et al. (1994), at a very philosophical level, this closure has a causality problem by assuming convection is driven by moisture rather than by buoyancy, as already emphasized in Sect. 2.1. As a result, convection is made to depend on something that is the result of convection

However, the causality may not be a real issue, because the closure treats the large-scale average conditions for deep convection, dominated by availability of moisture (but not CAPE as observations show, cf. Sect. 2.3). We should also keep in mind that the parameterization does not attempt to resolve the individual buoyancy-driven motions. This last point may be understood by taking, as an example, a prediction of the motions and patterns of sand-dunes driven by the wind without predicting the motion of each grain of sand.

On the other hand, our operational experiments (P. Bechtold, personal communication, 2012) tell that moisture-based closure works less well than the CAPE-based closure that is going to be discussed next. A well known problem with the moisture-based closure is its tendency for grid-point storms associated with a spurious increase of CAPE, as demonstrated by an idealized analysis (Yano et al., 1998). Also, a moisture convergence closure can cause an artificial CISK (cf. Ooyama, 1982).

Nevertheless, drawback of the moisture closure should not be overemphasized, either. It would be important to keep in mind that both moisture and thermal structure of the atmosphere (e.g. CAPE) are altered by convection. Ultimately, the real problem with both types of closure may be that the subgrid-scale variability of humidity, forced ascent, and environmental temperature that is not statistically parameterized properly, so that the true dependency on local CAPE and on the degree of forced lifting up to the moisture-dependent LFC (level of free convection) are not included in the closure.

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4.2 CAPE

To repeat the point, convection is ultimately driven by buoyancy. By taking a vertical integral of buoyancy, arguably, CAPE is a physically much relevant measure of convective instability, compared to the moisture. Fundamental importance of CAPE in the energy-cycle of convection is hardly overemphasized. However, various limitations of the CAPE concept must also be recognized.

First of all, CAPE is based on a Lagrangian trajectory analysis of air motions. Recall that a work, dW, performed on a unit mass of air by bouyancy, b, by lifting over a distance, dz, is given by

$$dW = b dz. \tag{1}$$

The vertical integral of the above leads to CAPE. However, we should keep in mind that this integral is useful only when we strictly follows a Lagrangian framework moving along an individual air parcel. Our standard dynamical formulation rather follows an Eulerian description.

As well demonstrated by Rennó and Ingersoll (1996), the role of CAPE is best established in the convective energy cycle when a closed circulation is assumed. Recall that in the original paper on CAPE (Moncrieff and Miller, 1976), it is introduced as a part of Bernoulli integral (see e.g. their Eq. 8). Thus, the role of the counter-acting dynamic pressure, Δp, must be properly taken into account when the circulation is not closed.

Aside from these caveats, CAPE is no doubt a very appealing quantity for measuring a degree of moist-convective instability of a system based on a notion of a conditional instability induced under a parcel-lifting process. One may even argue that CAPE provides maximum work that can be extracted from a vertical ascent. On the other hand, recall that CAPE does not provide a degree of convective intensity (e.g. tropical precipitation) by observational diagnosis as already discussed in Sect. 2.3. At the same time, we also emphasize that the CAPE closure works the best among the available options in our operational experiences (P. Bechtold, personal communication, 2012). In the following three subsections, we consider variants of the CAPE closure with a hope of filling a gap of the two opposite perspectives: observational and operational.

4.3 Parcel-environment CAPE closure

A major shortcoming of taking "total" CAPE as a measure of convection comes from the fact that it varies almost simultaneously with convection, so observations of correlation between CAPE and rainfall do not tell us much about causality, as already discussed in Sect. 2.3. Zhang (2002) proposes an alternative formulation called parcel-environment CAPE closure. Note that, in fact, Arakawa and Schubert (1974)'s original demonstration for quasi-equilibrium (their Fig. 13) is also based on this formulation. This closure states that the component of the CAPE change due to changes in the free tropospheric environment are related to the convection, but not a part coming from the boundary layer. Under the same spirit, Donner and Phillips (2003) examine CAPE closures by analysing observations from Oklahoma and the Tropical Pacific and Tropical Atlantic. They observe that the parcel-environment closure is closest to the observations, and that the relaxed- and strict quasi-equilibrium closures involving total CAPE are generally not satisfied, especially over land. So, the paradox or challenge for us is to explain why the parcel-environment closure is observed to be well satisfied and why total-CAPE closures are observationally less satisfied.

Donner and Phillips (2003) find that the large-scale average of total CAPE is irrelevant for deep convection over a mesoscale area. The total CAPE varies very rapidly in noisy manner, because it is controlled by fast-varying boundary layer processes. Whenever convection happens, for example, cold pools from downdraft air into the boundary layer disturb it. As a result, CAPE on the convective scale has much spatial variability.

There are two possibilities for explaining the observed validity of parcel-environment closure: first, mesoscale systems of deep convection evolve more slowly with timescales of half a day or so for them to respond to the fast-varying total CAPE. Instead, they can only respond to the similarly slow variation of the environmental free-troposphere's thermodynamic structure. Second, only the amplitude of "subgrid-scale"

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variability of total CAPE on the convective scale may be relevant for the buoyancy of updraft parcels.

We may further ask whether the triggering criteria, determining forced vertical lifting of parcels to their level of free convection, are part of the closure problem. Once convection has been triggered somewhere inside a mesoscale "grid-box", the convective intensity over the grid-box is presumably determined by a local "triggering" condition. Each and every convective cloud only occurs because thermals in the boundary layer rose to their level of saturation. So, the depth of the boundary layer and amplitude of variability of humidity and temperature inside it, which determine how often parcels are lifted by chance beyond saturation so as to form a convective cloud, may be part of the closure problem determining the overall large-scale convective mass flux. So, a approach for probability distribution of convective-scale variability of total CAPE inside each global model's grid-box may be useful.

The convective mass flux must depend on the longevity of individual cells of convection. Partly, the cells' longevity determines the dehumidification of the environment by removal of moisture by precipitation, thus increasing the cloud-base level and CIN. For example, a more wind shear (cf. Sect. 2.2, 9) means more tilt which means more separation of downdraft air from updraft inflow, which means each cell lasts longer.

4.4 PBL-based closure

A counterpart approach against the parcel-environment closure discussed in the last subsection (Sect. 4.3) is to try to close a convection parameterization based on a PBL (planetary-boundary layer) process. Here, CIN is expected to play a key role.

It is a commonly accepted view that a convection parameterization requires a trigger, and CIN is a typically adopted quantity for a trigger. Intuitively, a trigger is required based on the fact that convection rarely happens in spite of the fact that a finite value of CAPE always exits. It is normally interpreted that convection is triggered in order to initiate its life cycle, then it is terminated when its life cycle is over. Thus, in any

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convection parameterization, both triggering and termination conditions are required (however, see also Sect. 4.5).

Mapes (1997) emphasizes that a relative importance of trigger depends on the scale of convective systems. In order to elaborate this point, he proposes two major convective regimes depending on whether the convection activity is controlled by the increase of instability (equilibrium control), or by the processes overriding or suppressing CIN (activation control). When convective activity is on a large scale (say the whole tropical band), convection responds to radiative destabilization quite continuously, maintaining a state of near radiative-convective equilibrium, without any apparent role for CIN. On the other hand, when convective activity is on the mesocale, the spatial organization of convection is to be found into arcs, or lines, related to the existence of various PBL processes overriding the CIN, such as gust fronts, sea and land breezes, dryline convergence, etc.

Mapes (2000), in turn, applies the concept of activation-control principle in order for defining the convective intensity. His main proposal is to set the convective intensity proportional to exp(−CIN/TKE) with TKE the turbulence kinetic energy of PBL. More precisely, Mapes (2000) adds a pre-factor proportional to $\mathrm{CAPE}^{1/2}$ to this closure, but this detail is not followed by subsequent works. A closed expression for TKE must also be supplied, but none of the work following Mapes (2000) invokes CAPE for this purpose, either. As a result, essentially, we arrive at a closure based on CIN instead of CAPE. Though Mapes (2000)'s idea is also followed by Bretherton et al. (2004); Hohenegger and Bretherton (2011) for shallow-convection parameterizations, our operational experiences (P. Bechtold, personal communication, 2012) tell that CIN is too unreliable to be used for a closure

However, more importantly, we should note the inconsistency of the logic here: the original activation-control principle (Mapes, 1997) simply says triggering under low CIN is important, but it says nothing about a closure, whereas Mapes (2000) re-interprets this principle for defining a closure. We should clearly distinguish between the use of CIN as a trigger, say with a CAPE closure, and as a closure condition by itself.

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It may be argued that CIN is over-abused in the convection community. Extensive discussions from perspectives against CIN are given in Yano (2011). Here, main points are summarized succinctly with some supplementary remarks.

- Inconsistency between Mapes (1997) and Mapes (2000) may stem from the fact that Mapes (1997)'s activation control principle is fundamentally not consistent with the basic premises of mass flux parameterizations based on steady-plume hypothesis. Under this formulation, ensemble of convective plumes is assumed always in equilibrium with the large-scale environment. Transient behavior of individual plumes, especially their individual triggering process, which may well be conceptually considered by a parcel-lifting process associated with a presence of CIN, is not at all in concern. As a result, the issues of "triggering" is, from this strict point of view, beyond the scope of the given parameterization formulation.

  Thus, we need a radical modification of mass-flux parameterization in order to introduce Mapes (1997)'s activation control principle. Though we may well need a condition to turn on a conventional convection parameterization for practical purposes, this issue must be carefully distinguished from that of the triggering of individual convective plumes.

- A steady plume, that the standard mass-flux parameterization is based on, is, by definition, driven by buoyancy (or more precisely vertical buoyancy flux, cf. Sect. 4.5) integrated vertically from the bottom to the top of the given plume. In other words, only the total CAPE defined as a sum of both positive and negative contributions matters for its evolution.[1] CIN may become an issue only when a transient initial phase of a single plume is considered. For a steady plume, there is no reason to single out a role of CIN away from the other part of CAPE. As discussed in Sect. 4.5, the convective energy-cycle is better defined in terms of the

[1] Here, recall that buoyancy is defined in terms of a difference of the virtual tempeature between convection and the environment in evaluating CAPE. Note that the scale for the environment is not specified in this argument.

could work function or PEC (cf. Sect. 2.2, 3, 3a), in which a contribution of negative buoyancy is in no place considered separately under a standard mass-flux formulation.

- More physically speaking, CIN is a misleading quantity arising from an artificial use of a lifting parcel. In well-mixed boundary layers both over tropics and mitlatitudes, physically such a barrier should not exist, because the motions of parcels in a layer are buoyancy-driven by nature. In other words, actual individual parcels typically feel buoyancies positively correlated with the vertical velocity, unlike an artificially-defined lifting parcel. If CIN is re-defined, energetically more consistently, as a negative contribution of the buoyancy flux (cf. Eq. 2 in Sect. 4.5 below), we do not see such a negative buoyancy barrier as shown in Fig. 2 of Yano (2003) and Fig. 1 of Yano (2011). As going to be explicitly derived in Sect. 4.5, it is the vertical buoyancy flux, rather than buoyancy itself, that defines the generation rate of convective kinetic energy.

- The role of CIN in convective triggering is often interpreted rather in arbitrary manner. For example, Chaboureau et al. (2004) argue that convection is triggered under a diurnal cycle when CIN becomes sufficiently small (their Fig. 3b). However, they do not explain why convection is maintained afterwards in spite of the fact that CIN increases again. It is well possible that continuity of convection is due to a compensation effect by a presence of more turbulent kinetic energy in the boundary layer, for example. However, the reference in concern does not verify this point.

- Not all the existing theories agree upon the point that CIN inhibits deeper convection, especially when it is more properly re-interpreted as a vertically-integrated normalized buoyancy flux (cf. Sect. 4.5). An example is the "decoupling" theory proposed by Bretherton and Wyant (1997) for explaining the transition of the cloud-topped boundary layer into deeper convection (trade cumuli). Here, the decoupling is realized as a development of negative buoyancy flux at the top of the

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subcloud layer as the sea surface temperature increases. Thus, under this theory, CIN (i.e. negative buoyancy flux) positively contributes for triggering deeper convection rather than suppressing it.

4.5 Cloud work function and energy cycle of convective system

How important it may be in convective processes, CAPE is ultimately only a cheap substitute for a true energy conversion process. CAPE, being based on a simple lifting parcel theory, has the two major limitations: (i) use of an undiluted parcel buoyancy, (ii) absence of a vertical momentum factor.

The first point appears to be widely appreciated: CAPE is calculated by assuming a lifting parcel without any mixing with the environment. In a more realistic situation, a rising air parcel is more likely to experience substantial mixing with the environment. As a result, buoyancy is "diluted" compared to the standard lifting parcel value. This effect becomes critical, especially when the lower free troposphere is extremely dry as aftermath of a dry intrusion from midlatitudes in the Western Pacific (cf. Redelsperger et al., 2002). Under this situation, a conventional CAPE suggests huge convective instability while deep convection is completely suppressed observationally. However, Donner and Phillips (2003) find that the diluted-CAPE (cf. Sect. 2.2, 2b) is still often not a good indicator of convection.

The second point may appear less obvious, but it is simply understood by noting that CAPE is only an approximate substitute for a true energy conversion process, unless we take a Lagrangian description of the motions (cf. Sect. 4.2) as adopted by Moncrieff and Miller (1976); Rennó and Ingersoll (1996): CAPE provides a work performed on a particular air parcel when an integral is performed along the parcel trajectory (cf. Eq. 1).

However, we are usually interested with a generation rate of the kinetic energy at a fixed spatial point. For this purpose, we re-write Eq. (1) by noting that the Lagrangian

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parcel lifting, dz, is related to the vertical velocity, w by

$$dz = w dt$$

where t is the time. By substituting the above into Eq. (1), a local generation rate of kinetic energy is given by

$$\frac{dW}{dt} = bw \tag{2}$$

Clearly the right hand side is the vertical buoyancy flux. Thus, from an Eulerian point of view, it is the vertical buoyancy flux that controls the generation rate of kinetic energy, rather than the buoyancy itself (cf. Yano et al., 2005b).

A more formal consideration of the convective energy cycle along this line leads to the notion of cloud work function as originally introduced by Arakawa and Schubert (1974). It provides a more consistent measure for the convective-kinetic energy generation efficiency. The cloud work function can be estimated as PEC (potential energy convertibility, cf. Sect. 2.2, 3a) from cloud-resolving modelling as discussed by Yano (2003), and Yano et al. (2005b). Both works show from CRM experiments that PEC is indeed much better correlated with convective precipitation than CAPE.

As shown in Arakawa and Schubert (1974), the generation rate of kinetic energy, K_λ, for a given convection type, λ, is proportional to $A_\lambda M_{\lambda,\mathrm{B}}$, where A_λ and $M_{\lambda,\mathrm{B}}$ are the cloud work function and the cloud-base mass flux for the given convective type designated by λ. In turn, the cloud work function, A_λ, is modified by a rate

$$\sum_{\lambda'} \gamma_{\lambda,\lambda'} M_{\lambda'} \tag{3}$$

defined in terms of a matrix $\gamma_{\lambda,\lambda'}$ that characterizes the efficiency that convection type, λ', modifies the convection type, λ, with a given cloud-base mass flux, $M_{\lambda'}$.

Note that this energy-cycle system (as more precisely given by Eqs. 132 and 142 in Arakawa and Schubert, 1974), consisting of three variables, K_λ, A_λ, and M_λ, for each

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convective type, is closed once a certain relationship is established between two of the above variables. It is most logical to link K_λ to M_λ, and we may in general set:

$$K_\lambda \propto M_\lambda^p \tag{4}$$

with an unspecified exponent p. Randall and Pan (1993), and Pan and Randall (1998) close the above system by assuming $p = 2$, and propose to use this prognostic formulation in place of conventional closure for running Arakawa and Schubert (1974)'s convection parameterization.

On the other hand, Yano and Plant (2012a) suggest $p = 1$ is a more consistent choice based on statistical behavior of idealized cloud-resolving simulations (Emanuel and Bister, 1996; Shutts and Gray, 1999; Parodi and Emanuel, 2009). A beauty of this alternative choice is that as a result, the system spontaneously represents a life cycle of convective systems consisting of discharge and recharge under constant external forcing due to its nonlinearity, as shown in Yano and Plant (2012a). For example, this model provides a very simple explanation for delay of convection trigger under a diurnal cycle against solar forcing. Most importantly, here, a life cycle of convective ensemble is described without trigger and termination conditions (cf. 2nd paragraph, Sect. 4.4).

However, unfortunately, Randall and Pan (1993)'s and Pan and Randall (1998) do not consider a full implementation of the above prognostic formulation, but only consider the diagonal terms in the matrix, $\gamma_{\lambda,\lambda}$. This restriction physically means that the individual convection types, labeled by λ, evolve by themselves without interacting with the other convective types. This is the major restriction of their implementation.

These off-diagonal terms with $\gamma_{\lambda,\lambda'}$ ($\lambda \neq \lambda'$) represent interactions between different convection types in the convective energy cycle. Interactions between different convection types are expected to be important in many problems. For example, the transformation of shallow convection into deep convection can easily be described under the interactions between shallow and deep convection (Yano and Plant, 2012b).

It is important to realize that the convective energy cycle, outlined here by invoking Eq. (2), can be derived under a formal procedure of the energy integral of an ensemble

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system of convective plumes (cf. Yano and Plant, 2012a). Equations (132) and (142) in Arakawa and Schubert (1974), obtained in this manner, thus provide a robust basis for closing a convection parameterization. Recall that Arakawa and Schubert's convective quasi-equilibrium closure is defined as a steady condition for their Eq. (142). Being based on a formal procedure, it also provides a solid basis for analyzing various convective-resolved simulations as well as interactions between convection and large-scale dynamics, in alternative manner than discussed in Sect. 3.

4.6 High-resolution limit

One of the issues to be taken into account in defining the closure assumption is a possibility that a dominant physical process defining the convective intensity may change with the scale, and henceforth also with the model resolution (cf. Bister, 1998). Here, it may be argued that as moving to higher resolutions, convective drafts must be more explicitly computed by model dynamics.

The behavior of the subgrid-scale parameterization of a model is related to a numerical algorithm adopted. Its behaviour can substantially differ from what is expected if the averaging area is gradually narrowed, from a simple statistical diagnosis. For instance, at a higher resolution, the model vertical profiles in cloudy regions are likely to become closer to moist adiabat, so that the buoyancy of a lifted parcel tends to decrease with increasing resolutions. The vertical velocity, w_c, in subgrid-scale updrafts, is estimated from a given grid-box state, based on either diagnostic or prognostic equation using a buoyancy in most of convection parameterizations. As moving towards high resolutions, the convective vertical velocity, w_c', as defined as a deviation from a grid-box average, would tend to zero.

From a purely statistical point of view with a fixed large-scale environment, the convective vertical velocity is defined by an ensemble average, thus as the model resolution becomes higher, this ensemble size reduces, leading to a more stochastic behavior. Such a stochasticity leads to a more chance to see a higher convective vertical velocity as the model resolution increases, virtually contradicting with the conclusion

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just stated in the last paragraph. Note that such a purely statistical reasoning is misleading because the "environment" itself is highly inhomogeneous in high resolution limit. A similar statistical behavior holds for the convective fraction σ: at high resolution we are likely to observe larger convective fractions than at coarse resolution.

Concerning the closure we may expect the following situations:

- The above-mentioned reduction of buoyancy makes CAPE almost vanished within convective drafts. This argument *could help* the CAPE-based closures to guarantee an extinction (suppression) of convection in subgrid scale schemes at high resolution properly without producing a perpetual grid-point storm. However, if the CAPE closure is used to determine the convective fraction, a vanishing convective fraction is produced at high resolution limit in counter-intuitive manner. Here, a difficulty behind is in properly defining an "environment" defining CAPE, which may be found beyond a grid box in the high-resolution limit. Furthermore, distributions of CAPE in the convective-scale may somehow be taken into account in the closure, as already suggested in Sect. 4.3.

- As moving to high resolutions, the moisture convergence does not always reach a maximum as a parameterized convection follows its life cycle. Instead, in a heart of a half-resolved convective updraft, an increasingly more moisture convergence may be induced associated with an increasing updraft velocity. For this reason, the schemes based on a moisture-based closure are likely to experience difficulties in terminating a parameterized convective event at high resolutions.

To illustrate some differences between CAPE and moisture convergence closures, we present a few preliminary results obtained with the Alaro model (Gerard et al., 2009) and a new prognostic deep convection scheme, based on a perturbation approach. The scheme includes a triggering similar to the Kain Fritsch scheme. The CAPE closure includes a contribution of the downdraft on the boundary layer, and a unsaturated downdraft scheme is also included in the model. Seeing very narrow thunderstorm systems in the observation (Fig. 1), we expect that the precipitation at 4 km resolution should be less than the extremes on the radar image. For this reason, we show the model-predicted precipitation (Fig. 2) in the same color code but with the half of the scales to the observed precipitation. The moisture convergence closure (MC) yields more intense precipitation than CAPE closure (CC). The MC precipitation field appears to be shifted to the north-east with respect to the radar image. In Fig. 3, the CAPE appears lower at places where the convective scheme is active. In the upper-right corner, MC gives precipitation that was not observed and reduces the CAPE, while CC allows the CAPE to sustain. In Fig. 4, we observe that MC yields more intense low level moisture convergence, associated with precipitation. With CC, moisture convergence is well correlated with precipitation areas, though a substantial difference in structure is still noticed. The preliminary results presented here appear to support the notion of positive feedback of moisture convergence by writing a condition on an effect. It can furthermore lead to an erroneous evolution of the forecast model.

5 Difference over globe: tropics and midlatitudes

Discussions so far has been implicitly based on a premise that a certain closure hypothesis is universally valid globally. This premise is necessary in order to develop a numerical model that is valid globally. Though some issues arise from regional differences, they are rather treated as side issues so far for this reason. In order to counterbalance this "universal" view, the present section summarizes some of the differences over the globe already emerged in the discussions so far.

Aspects of the large-scale tropical waves have been discussed in Sect. 3. As stressed there, in the tropics, the temperature does not vary much horizontally (weak-temperature gradient: Sobel et al., 2001). However, the situation is very different in the midlatitudes, where the temperature variations are much larger. As a result, the effect of lower tropospheric moisture is more important in the tropics in a relative sense. If, however, we are interested in forecasting thunderstorms and mesoscale convective systems in midlatitudes, where the temperature has a large contribution to the variation of CAPE, it may not be a good idea to use a moisture convergence closure.

It has been stressed in Sect. 2.3 that precipitation in the tropics is highly correlated with lower-tropospheric water vapor (or saturation fraction). This makes sense even from a point of view of convective energy cycle discussed in Sect. 4.5, because low- to mid tropospheric moisture has much to do with buoyancy of moist rising parcels (especially when the temperature variations are small) and also greatly affects the downdrafts.

As already remarked in Sect. 2.3, "rainfall is correlated with decreased CAPE (Ramage, 1971; McBride and Frank, 1999)", when convection takes place instability must decrease. Raymond and Sessions (2007) paraphrase this issue as follows: "*Regions of heavy rain in the tropics tend to have environments which are both moister and more stable than in less disturbed areas (Ramage, 1971; McBride and Frank, 1999). The moisturization and stabilization are likely due to the convection itself, so nonlinear feedbacks are involved in this process.*" Thus, we must be careful that we do not make convection depends on something that is the result of convection itself. See Mapes (1998) and Yano et al. (2000) for further causality issues behind.

As discussed in Sect. 3, Fuchs and Raymond (2002), and Donner and Phillips (2003) give an evidence against the CAPE closure, by showing that it does not work well in the tropics. As discussed in Sect. 4.6, in midlatitude mesoscale forecasts, the CAPE closure works rather well, if it is used properly. From one of authors' experiences, for example, a scheme by Bechtold et al. (2001) based on a CAPE closure works well in the midlatitudes. Donner and Phillips (2003), in turn, find that convective quasi-equilibrium (steadiness of the CAPE budget) is a poor approximation at subdiurnal timescales in midcontinent North America.

Convective instability, which may be crudely measured by CAPE (but see Sect. 4.5), is controlled by three major factors: (1) the boundary moist entropy (2) free tropospheric temperature and (3) moisture in the low-to-middle troposphere (LTMT: cf. Sect. 2.1). Which factor dominates in what situation? Here, we may hypothesize that the main

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limiting factor of convective intensity is the one that represents the largest temporal and spatial variation. Based on the dominance among these factors, we can divide the globe into three major regimes: (i) tropical oceans, and land areas in the (ii) absence or (iii) presence of large free-tropospheric temperature variations. Let us consider these regimes each by each.

i. Over tropical oceans, free-tropospheric temperature variations are small due to large Rossby deformation radius. Also, the diurnal variation of boundary layer entropy is relatively small. However, the low-to-middle troposphere (LTMT) moisture can vary significantly over various scales. As a result, the variations of undiluted CAPE are small, and consequently the LTMT moisture becomes the limiting factor for convection. This is consistent with many studies showing the relation of LTMT moisture, or column water vapor, with convection as discussed in Sect. 2.3.

ii. Land areas without large free-tropospheric temperature variations can be found in the tropics or sometimes also in the middle-to-higher latitudes. Over the land areas, there is typically a large-amplitude diurnal cycle in boundary layer variabilities. The LTMT moisture can also vary. For this reason, Zhang (2002), and Donner and Phillips (2003) note that quasi-equilibrium is not valid to model diurnal cycles of convection and the net rate of change of CAPE can be comparable to changes of the boundary-layer air, as discussed in Sect. 4.3.

Here, a careful specification of the entrainment rate is important, for example in transformation of shallow convection into deep convection (cf. de Rooy et al., 2012). Derbyshire et al. (2004, 2011) suggest that low-to-middle level moisture and the size of thermals controls the buoyancy, depth and rate of ascent, and entrainment rate under the transformation process. More specifically, a dry LTMT affects the strength and depth of the downdrafts and the value of θ_e in the downdrafts. Rather nontrivial aspects of this process are discussed by James and Markowski (2010). Furthermore, entrainment and upward growth of convection

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may depend on LTMT. Finally, downdrafts may enhance the development of big thermals and deep clouds (cf. Khairoutdinov and Randall, 2006).

It is likely that convection occurs when/where the boundary layer entropy reaches its maximum during the diurnal cycle due to convection. (Recall that heating due to convection occurs even outside of convection due to subsidence. Therefore, convection itself is likely to occur in preferred places and times only.) However, convection can be enhanced when the LTMT moisture is large. Hence, over these land areas the moisture can play an important role in defining a location of convection. The timing of convection might still follow the diurnal cycle. James and Markowski (2010) note by studying convection with a cloud resolving model that the influence of dry air on convection was sensitive to the value of CAPE. The detrimental effect of dry air by entrainment on convective intensity was much greater with a lower value of CAPE. So this regime, namely land areas without large free-tropospheric temperature anomalies, resembles the first regime (tropical oceans), but it differs by a prominent presence of a diurnal cycle.

iii. The third regime is land areas with large free-tropospheric variation of temperature. Therefore, also CAPE, or lapse-rate, variations can be large. James and Markowski (2010) noted that when CAPE is large then dryness of the environment is relatively inconsequential for the vigor of convection. Wu et al. (2009) showed that with more stable air, the moisture became relatively more important for the transition to deep convection. Conversely, with the drier environment, stability became relatively more important for the transition to deep convection (cf. their Fig. 6a).

Zhang and Klein (2010) studied the mechanisms affecting the transition to deep convection in the Southern Great Plains ARM site. They noted that the lower (free-)tropospheric lapse rate was related to subsequent amount of total precipitation and maximum rain rate. CAPE, however, did not seem to be important. High

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relative humidity above the boundary layer is identified as a better indicator for an earlier onset of convective precipitation.

It is intriguing why in some cases CAPE seems to be important and in other cases the lower tropospheric lapse rate (or its contribution to CAPE) is important. In this respect, the recent study of Raymond and Herman (2011) is suggestive: they show that the temperature variations in the upper troposphere are not eliminated as fast as those in the lower troposphere. Their experiment suggests that the upper troposphere is less constrained by convective quasi-equilibrium than the lower troposphere.

CAPE, however, is dominated by upper tropospheric contributions. If the difference of the temperature between the parcel and its environment is constant with height, the contribution to CAPE from the layer between 300 and 100 hPa is as large as a contribution from the layer between 900 and 300 hPa. In the cases, where the lower tropospheric buoyancy (and/or CIN) is important for convection, CAPE may not have much predictive value.

An implication from Raymond and Herman (2011) is that it may be useful to limit a range of vertical integral for CAPE to the lower troposphere, which may provide a more practically useful measure of convective instabilities. The result more generally suggests an importance of considering different instability measures for convection of different vertical scales as explicitly taken into account in the cloud work function and PEC (cf. Sect. 4.5).

Regarding the three suggested regimes for convection, the study by Stone and Carlson (1979) is revealing. Above the boundary layer, zonal mean lapse rates were observed to be within 20 % of the moist adiabatic lapse rate from the Equator up to about 30° N in January and 50° N in July and appreciably more stable in higher latitudes. This is due to moist convection being more important at lower latitudes and baroclinic eddies at higher latitudes when it comes to the average effect on the lapse rate.

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6 Conclusions

The present paper has reviewed the closure of convection parameterization from phenomenological perspectives. Loosely speaking, the closure refers to the problem of defining a convective intensity under a convection parameterization. The review has begun by listing physical variables that are expected to contribute in determining the convective intensity. One may wish that a closure formula can be developed by combining all these variables under, say, a certain statistical method. However, as it turns out, though it is easy to prepare such a list based on physical reasonings, objective statistical analyses, mostly focused on the tropics, tend to suggest that most of them do not contribute significantly for defining the convective intensity or whether a convective event happens or not.

The most disappointing such conclusions come for both CAPE and CIN. Those two major variables, that are commonly considered as controlling convection, do not present statistical significance. On the other hand, in spite of a secondary role from a point of view in energetics of convection, the water vapor tends to stand out in observational correlation analysis.

Here, problems associated with this type of observational diagnostic studies must be recognized. Most fundamentally, this type of studies is incapable of telling anything about causality. The most practical problem is a difficulty in identifying "convection" itself objectively from conventional observations. For this reason, in most of the above studies, the convective intensity is measured by the precipitation rate. Though this may be a valid assumption over the tropics, use of the precipitation rate over the midlatitudes as a measure of convection is highly questionable, where the precipitation is dominated by the synoptic-scale processes. Identification of the convection controlling parameter observationally over the midlatitudes remains a major challenge.

Moreover, a diagnostically obtained relationship does not necessarily present a useful closure relationship. A good example comes from the scalings developed by Shutts and Gray (1999) for a set of equilibrium convection simulations. Combining their

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Eqs. (8) and (15) gives a relation for the cloud-base mass flux, M_B, as being proportional to F_h/CAPE where F_h is the surface moist static energy flux. The relation provides a good estimate for CAPE if the mass flux is known (Shutts and Gray, 1999), but is too unstable to be used in the reverse sense to predict the mass flux as a practical closure condition (R. S. Plant, personal communication, 2012). Consider a small perturbation which leads to a slight excess of mass flux relative to the value required for equilibrium. The excess mass flux reduces the CAPE and so gives rise to a positive feedback that further increases the closure mass flux.

In face of difficulties in defining a closure based on a statistical data analysis, an alternative possibility of inferring a preferred closure formulation based on stability analyses of tropical convectively-coupled waves is sought. The theoretically obtained stability characteristics of the waves must be consistent with observations, if the given closure hypothesis is physically based. A particular series of work reviewed here, again, favors the water-vapor closure over the CAPE closure. However, we should not take it as a final word from the linear stability analysis. Especially, the consistency between a parameterization formulation adopted under the stability analysis and that in operational models must carefully be scrutinized.

Various theoretical reflections are presented from various different points of views. In operational implementations (P. Bechtold, personal communication, 2012), there are difficulties of making a moisture closure work, though we should not take it as a final verdict on this closure. From a point of view that convective is a dynamical process, it is more natural to base the closure on an energy cycle of convection. CAPE is a standard choice for this purpose.

Here, two caveats may be emphasized: as discussed in Sect. 4.3, the observations prove that one cannot use CAPE closures in the manner that have been used in the past. Rather, we must either follow parcel-environment contribution to CAPE change or treat the subgrid-scale variability of CAPE. Furthermore as emphasized in Sect. 4.5, we need to move to more explicit treatments of the cloud work function. Once an exponential power, p, in Eq. (4) is specified, the convective energy-cycle based on the cloud work function provides a more consistent description of the evolution of convective ensembles than any alternative closures currently available.

Finally, regional differences in factors that are contributing in defining the convective intensity are discussed, suggesting that the convective intensity depends on low-to-middle troposphere moisture when CAPE is small, and become relatively insensitive to this moisture when CAPE variation is large.

We emphasize various important background issues to be considered in order to tackle with the convection paramerization closure problem under a solid basis. However, unfortunately, the present review tends to point to difficulties of putting the operational experiences in both theoretical and observational contexts. Difficulties are further compounded by the fact that these operational experiences are not well documented in the literature. Though parameterization comparison studies are abundant (e.g. Wang and Seaman, 1997), they often fail to pin point the issues behind the closures.

There is still a long list of issues to be resolved concerning the convection parameterization closure. A particular example is the issues of the high-resolution limit. More generally, the scale dependence of the closure is still a wide open question. This issue is only briefly discussed in considering the PBL-based closure in the present review. A major remaining challenge is to develop both theoretical and observational studies that can positively contribute for the parameterization closure problem in operational contexts. The goal of the current COST Action ES0905 (2010–2014: http://convection.zmaw.de) is to develop a common ground for theoretical, observational, and operational researchers for identifying a much needed breakthrough.

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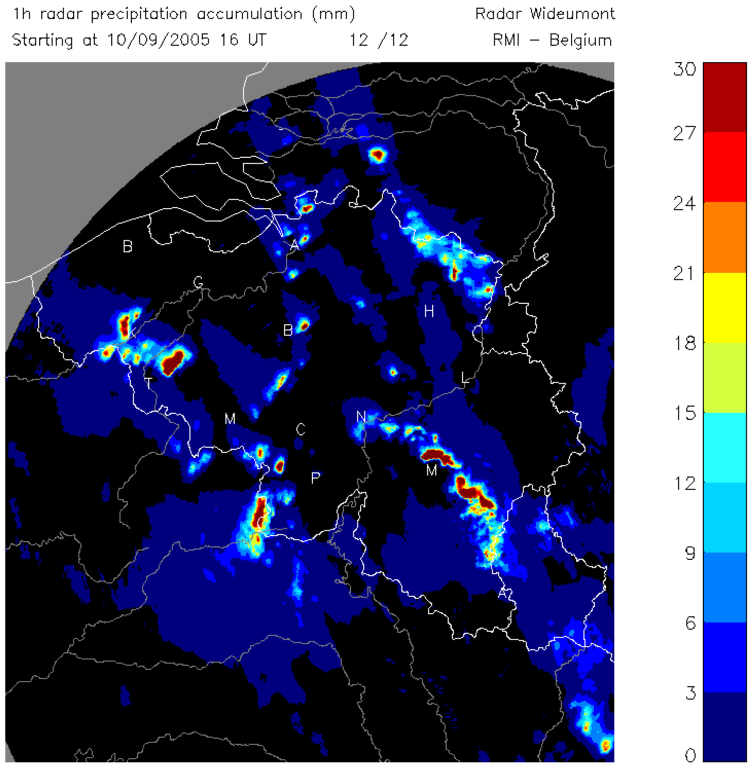

Fig. 1. 1-h accumulated precipitation, thunderstorm of 10 September 2005, Wideumont Radar, RMIB.

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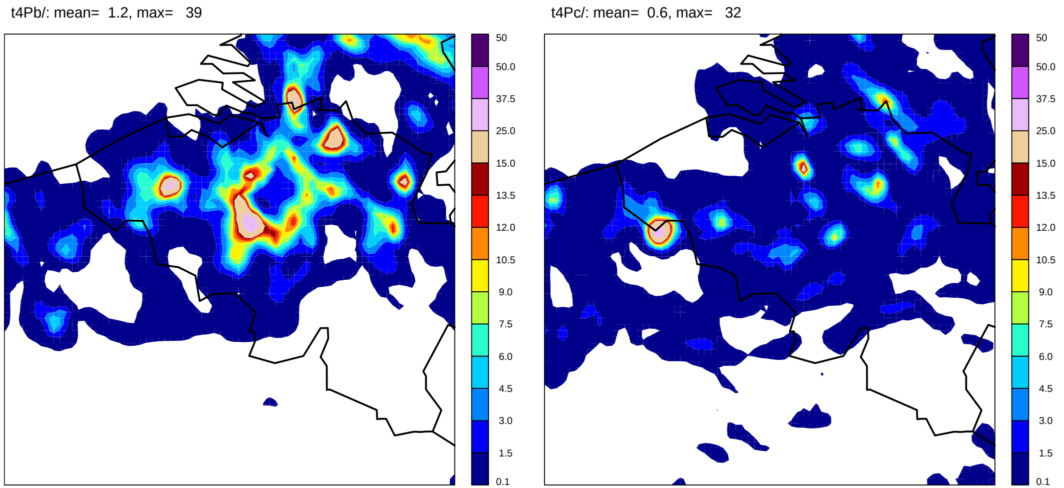

Fig. 2. Preliminary results of new prognostic convection scheme in Alaro model, 4-km resolution run on limited area, 41 vertical levels. Left: moisture convergence closure, right: CAPE closure. 1-h accumulated precipitation (mm), precipitation scale is shown by a half-scale of the radar image.

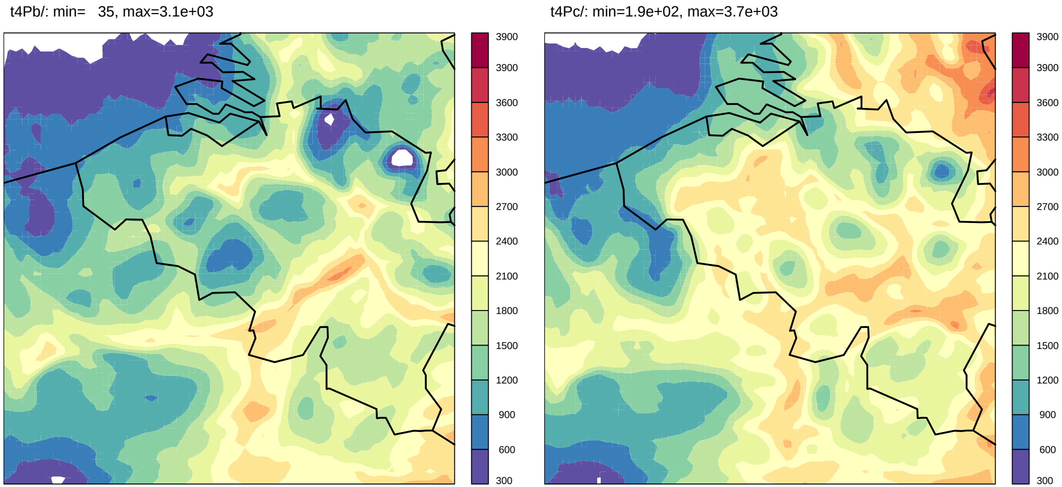

Fig. 3. The same as Fig. 2, but for a horizontal distribution of CAPE.

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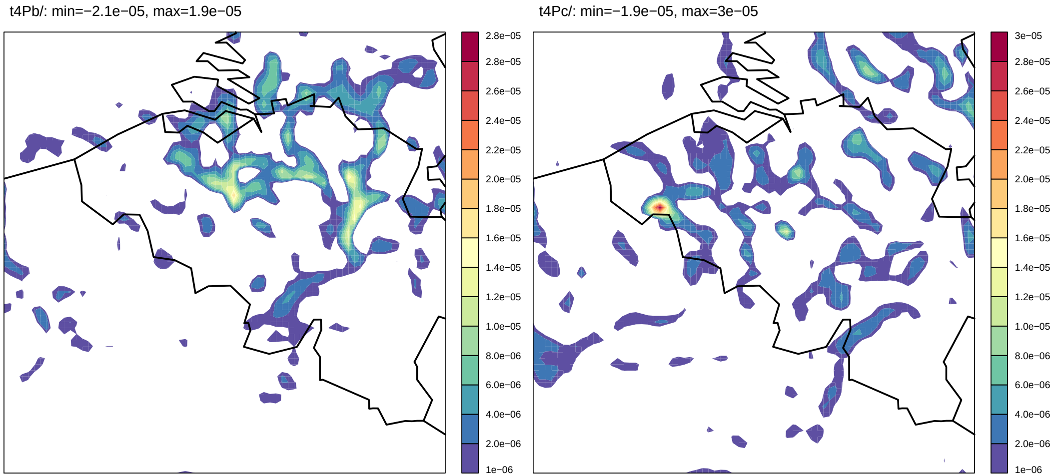

Fig. 4. The same as Fig. 2, positive low level moisture convergence field.

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