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Documentation of the Convection scheme in the NCMRWF Unified Model

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10	Abstract (100 words)	Convection is a dominant process in maintaining the general circulation in tropical region through heat, moisture, and momentum transport. This report describes the representation of convection in the NCMRWF Unified Model (NCUM). A general introduction to cumulus parameterization and a detailed description of the mass flux schemes used in NCUM are documented. Additionally, the sensitivity of entrainment mass flux and Convective Available Potential Energy (CAPE) time scale in NCUM during an active monsoon spell are also discussed
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Abstract

Convection is a dominant process in maintaining the general circulation in tropical region through heat, moisture, and momentum transport. This report describes the representation of convection in the NCMRWF Unified Model (NCUM). A general introduction to cumulus parameterization and a detailed description of the mass flux schemes used in NCUM are documented. Additionally, the sensitivity of entrainment mass flux and Convective Available Potential Energy (CAPE) time scale in NCUM during an active monsoon spell are also discussed.

Contents

1. Introduction to Convection

Convection plays a major role in maintenance and evolution of the tropical atmosphere, through convective heating, moisture and momentum transport. Convection and associated precipitation in the Asian-monsoon region is very important, since it shapes the livelihood of millions of people. Accurately representing moist convection in the general circulation model (GCM) is a challenge, as the atmosphere responds in a complex manner to the thermodynamic changes and large-scale forcing and also due to its wide range of scales and mutual interactions. As convective plumes are in the order of few hundred meters to kilometers, whereas typical GCM resolutions are of the order of tens of kilometers or more, convection will remain an unresolvable process. The convective parameterization scheme is designed, to represent the sub-grid scale transport of heat, moisture and momentum associated with cumulus clouds within a model grid-box. In general, the convection scheme discusses three components; a trigger, which determines whether the atmosphere can support convection, a cloud model, which describes the distribution of convective activity in the vertical, and a closure, that deduces the overall feedback of convection.

The deep convection in the monsoon belt is significant with higher occurrence of frequency and larger horizontal spatial structure. Kalpana INSAT Outgoing Longwave Radiation (OLR, Mahakur et al., 2013) is a good convective proxy in the monsoon belt, albeit with a mix of high clouds not directly of convective origin. Occurrence frequency of deep convection is indicated from the count of OLR value less than 200 Wm^{-2} from the total OLR count during the June-September period. Deep cumulus convection is a frequently observed phenomenon over the Southern Asian monsoon region with a frequency of $\simeq$ 70% during June-September period, whereas lesser frequency is observed in Tibetan plateau and south tropical oceanic region (Fig. 1). High amount of convective heating during monsoon period makes it a big source of heating over the tropical belt.

This report presents the convection scheme used in the NCMRWF Unified model (NCUM) in a simplified format and for the minute details of the schemes and the formulations the readers are advised to refer to Stratton et al. (2013). This report is ordered in the following manner: Section 1 gives a brief review of moist convection and cumulus parameterisation, a detailed description of the convection scheme in NCUM is given in Section 2, and Section 3 describes the sensitivity study on entrainment mass flux and Convective Available Potential Energy (CAPE) time scale during an active monsoon spell.

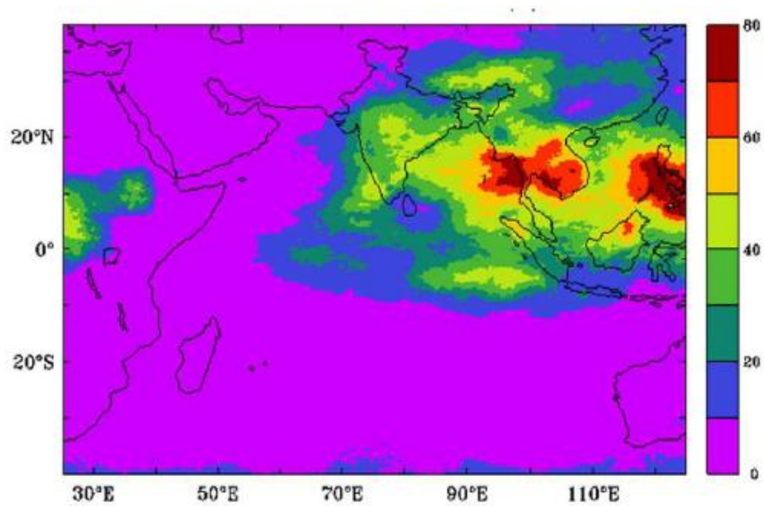

Fig. 1: Occurrence frequency (in %) calculated by taking ratio of count of daily Kalpana INSAT OLR value less than 200 W/m^2 to the total OLR count during the June-September for the period 2008-2013.

1.1 Moist convection

Convection occurs generally, when density gradient results in non-zero buoyancy forces in some regions of a fluid. This force imbalance accelerates the flow, and facilitates the transport of energy and mass we generally associate with convection. In the field of meteorology, convection tends to be framed in the language of parcel theory (see e.g. Wallace & Hobbs, 2006). We consider a hypothetical 'parcel' of air existing in some background environment. The parcel is displaced infinitesimally in such a way that it exchanges no energy with the environment. The parcel buoyancy at the new position is calculated, and if the resultant force serves to amplify the initial disturbance, the atmosphere is called unstable. This instability causes a rearrangement of fluid in the vertical which reinstates atmospheric stability. This rearrangement is called atmospheric convection.

However, an important characteristic of atmospheric convection that sets it apart from convection in other fluids is the presence of water vapour. When water changes phase, a large amount of latent heat is either absorbed or released. This fundamentally changes the thermodynamics of the fluid. Incorporating this effect into the parcel description of convection can result in a situation termed conditional instability, in which the atmosphere is stable to small perturbations, but unstable to displacements of finite amplitude (Emanuel, 1997). Rising thermals entrain environmental air and thus modify the cloud air through mixing. Thermals are nonhydrostatic, nonsteady, and highly turbulent. The buoyancy of an individual thermal depends on a number of factors, including the environmental lapse rate, the rate of dilution by entrainment, and drag by the weight of liquid water in cloud droplets. The level at which lifting by free convection starts is the level of free convection (LFC). As warm air near the surface rises, it expands and cools dry-adiabatically. If the buoyant parcel contains water vapor, and if the parcel temperature cools to the isentropic condensation temperature, the vapor condenses, forming a cloud base. The altitude of the cloud base is the lifting condensation level (LCL). Release of latent heat during condensation at the cloud base provides buoyancy for further lifting, cooling and cloud development. This spontaneous release of energy continues to accelerate the parcel upwards until it again becomes negatively buoyant at the level of neutral buoyancy (LNB). Thus, with the inclusion of the thermodynamic effects of water vapour, it can be seen that the atmosphere can be stable to small displacements of fluid, but still contain large amounts of potential energy that is released with a sufficiently large disturbance. The amount of potential energy that can be released in this way is called the Convective Available Potential Energy (CAPE) and it is a good measure of atmospheric buoyancy. Following the method by described by Williams and Renno (1993), CAPE for the parcel undergoes pseudo adiabatic ascent as,

$$CAPE = -\int_{P_{LFC}}^{P_{LNB}} (T_{\nu p} - T_{\nu e})\, R_d d(lnP) \tag{1}$$

In reality, much of this liquid water, and indeed solid ice, is carried upwards along with the parcel, and reduces its buoyancy, altering the amount of energy available. Thus, CAPE as we have defined it is sometimes called irreversible CAPE, since mass is irreversibly removed from the system via precipitation. The negative buoyancy is considered in the parameter called Convective Inhibition Energy (CINE), and is calculated as,

$$CINE = -\int_{P_{LFC}}^{P_{SFC}} (T_{\nu p} - T_{\nu e})\, R_d d(lnP) \tag{2}$$

A hypothetical sounding gives an idea of CAPE and CINE as given in Fig. 2.

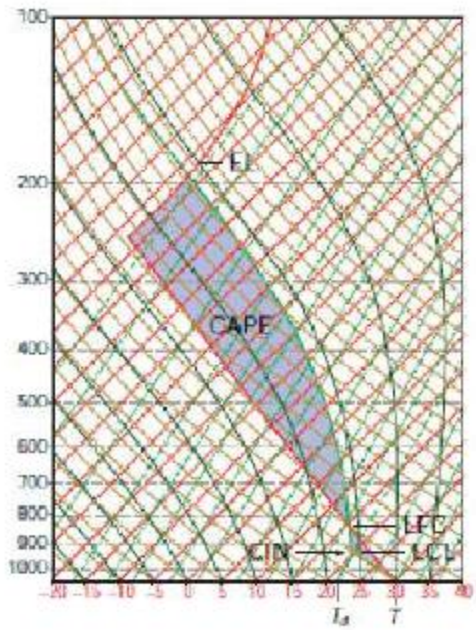

Fig. 2: A hypothetical sounding illustrating the concept of CAPE and CINE. The CAPE and CINE in this sounding is shaded.

1.2 Cumulus parameterization

Atmospheric stability above the LCL affects cloud type. If a cloud base is below 2 km, and the air is stable above the LCL, the cloud that forms is a cumulus humilis (cumulus cloud of slight vertical extent), stratus, or stratocumulus. If the atmosphere is unstable up to the tropopause, the cloud that forms is a cumulonimbus.

Boundary layer structure is a part of active convection, a schematic diagram showing the atmospheric profile is given in Fig. 3. The first level at which convective mass, momentum and thermodynamic fluxes are estimated is cloud base. The convective air motions generate intense turbulent mixing. This tends to generate a mixed layer, which has potential temperature and humidity nearly constant with height. When buoyant turbulence generation dominates the mixed layer, it is called a convective boundary layer. The lowest part of the boundary is called the surface layer. To represent the effects of convective updraughts on the sub-cloud layer a simple scaling of cloud base fluxes is applied in which they decrease to zero at the surface through the sub-cloud layer. Cloud layer is conditionally unstable, where the ensemble of active clouds are created here. Inversion is a stable layer above the mixed layer, acts as a lid to rising thermals. Above inversion, the planetary layer is called free atmosphere.

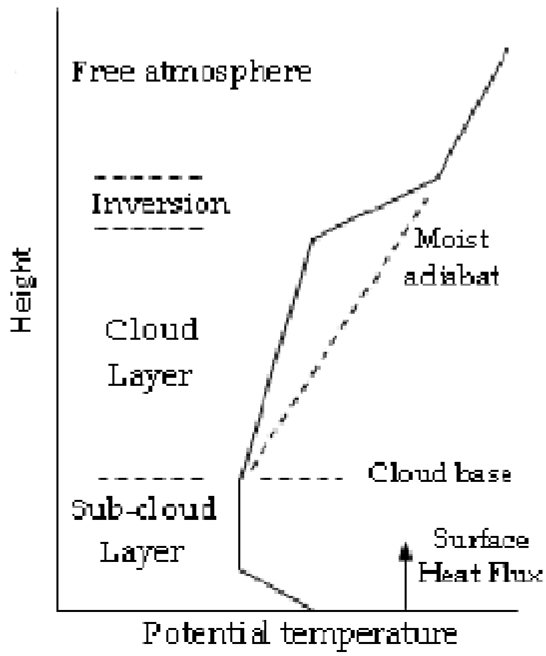

Fig. 3: Boundary layer structure of cumulus cloud layer

The horizontal scale of individual cumulus cloud is of the order of 0.1-10 km. Therefore, models whose grid sizes are of the same order can directly resolve the cumulus clouds without the need of cumulus parameterisation. On the other hand, grid spacing of synoptic forecast models are much greater than the sizes of individual cumulus clouds. Therefore, it becomes totally impracticable to resolve them in any numerical model of large-scale circulation. Instead, the collective influence of clouds within a larger area is formulated or parameterised in terms of the large scale environmental variables. This is called the cumulus parameterisation. The goal is to predict the changes in the large scale variables due to cumulus convection. Important model-scale variables used to predict sub-grid effects are horizontal and vertical wind speeds, potential temperatures, and total water mixing ratios. Cumulus parameterisation uses these variables to adjust potential temperature, total water, and momentum fields and to predict precipitation rates.

It is evident that a large amount of latent heat is released in cumulus clouds and the released heat is transported upward and how this heat is utilised in warming the large scale environment is determined by the sub-grid scale cumulus parameterisation. Ensemble of cumulus clouds with differing heights are embedded in a horizontal area, which is assumed to cover only a small fraction of the grid-box area.

The governing equations of the atmosphere which require the parameterization of physical processes are equations of mass continuity, heat energy and moisture continuity averaged over a hypothesized area are given by

$$\overline{\nabla.V} + \frac{\partial \omega}{\partial p} = 0 \tag{3}$$

$$\frac{\partial \overline{s}}{\partial p} + \overline{\nabla.sV} + \frac{\partial \overline{\omega s}}{\partial p} = Q_R + L(c-e), \tag{4}$$

$$\frac{\partial \overline{q}}{\partial p} + \overline{\nabla.qV} + \frac{\overline{\omega q}}{\partial p} = e - c \tag{5}$$

where S = (C_pT + gz) is the dry static energy and h = S +Lq is the moist static energy, ω is the average vertical velocity, Q_R the heating rate due to radiation, c is the rate of condensation per unit mass of air, and e is the rate of re-evaporation of cloud droplets.

The basic assumption behind parameterization is that the unresolved motions have no projection on to the grid-scale. Equations 4 and 5 can be rearranged, after linearization into the grid-mean quantity and the unresolved perturbations, to give

$$Q_1 = \frac{\partial \overline{s}}{\partial p} + \overline{\nabla.sV} + \frac{\partial \overline{\omega s}}{\partial p} = Q_R + L(c-e) - \frac{\partial \overline{\omega' s'}}{\partial p}, \tag{6}$$

$$Q_2 = -L\left(\frac{\partial \overline{q}}{\partial p} + \overline{\nabla.qV} + \frac{\partial \overline{\omega q}}{\partial p}\right) = L(c-e) + L\frac{\partial \overline{\omega' q'}}{\partial p} \tag{7}$$

Most of the notations are conventional, equation 6 shows that the apparent heating Q_1 of the large- scale motion system consists of the heating due to radiation, the release of latent heat by net condensation and vertical convergence of the vertical eddy transport of sensible heat (Yanai et.al., 1973). Equation 7 being measure of the apparent moisture sink which is due to the net condensation and vertical divergence of the vertical eddy transport of moisture. From equations 3, 6 and 7, we have

$$Q_1 - Q_2 - Q_R = -\frac{\partial \overline{\omega' h'}}{\partial p} \qquad (8)$$

Where $\frac{\partial \overline{\omega' h'}}{\partial p}$ is a measure of the vertical eddy transport of total heat and may be used to measure the cumulus convection activity. This will be related to the product of the cloud mass flux and temperature excess of the cloud.

1.2.1 Parameterisation of Shallow Convection

Shallow convection starting from the boundary layer and stops below freezing level or below an inversion with descending or weak ascent above. When there is no synoptic scale low level convergence, only shallow cumulus can form above the mixed layer. The shallow convection originates in the form of ascending current from the warm surface layer (the layer from ground to 10 or 100 m where the lapse rate is super-adiabatic). As these ascending convection currents penetrate above the condensation level, cumulus clouds form. The depth of the cumulus clouds produced by the convective cell depends on the temperature excess δT and the vertical velocity ω of the ascending current above the condensation level. Shallow cumuli are produced if δT becomes negative and ω vanishes due to the stable stratification above the mixed layer, whereas deep cumulus clouds form when both δT and ω are positive above the condensation level so that convective instability activates the further rise of the currents to higher levels. The parameterization of shallow cumulus convection has been discussed by Kuo (1974) and Tiedtke (1989). Assuming that these types of convection are controlled by the surface heat and moisture fluxes and are not related to the large scale flow, we can parameterize them by taking the sensible and latent heat fluxes across the surface. These fluxes can be expressed either in terms of the temperature and humidity gradients or in terms of the temperature and moisture differences in accordance with the turbulent theory.

1.2.2 Parameterization of Deep Convection

These clouds form when there is a strong low level convergence of moisture in a conditionally unstable atmosphere. The success of parameterising deep convective processes depends on the scale of events and is greatly influenced by the dynamics of the circulation to be modelled. There have been numerous efforts of cumulus parameterisation in the large scale models in the last four decades. These schemes assume that the cloud effects can be specified in terms of the large scale variables at the grid point. The schemes can be broadly divided into two groups; (i) parameterisation without involving cloud model and (ii) parameterisation involving cloud models. The first type of schemes does not use any explicit cloud model to parameterise the effects of cumulus clouds. Major type of schemes discussed are 1) Kuo-type scheme (based on moisture budget, Kuo, 1974), 2) adjustment scheme (e.g., Betts-Miller Scheme). The second categories of schemes are based on Mass-flux scheme, where the parameterisation is accomplished by using explicit cloud models. The cloud models are used to determine the cloud physical and dynamical properties and the clouds are assumed to modify the environment by their interaction. The most popular scheme used in this category is the Arakawa-Schubert type scheme (Arakawa and Schubert, 1974) based on Mass-flux spectral model. However, the mass-flux approach based on ”bulk cloud” type schemes are used in the ECMWF forecast model (Tiedtke, 1989) and the UM (Gregory and Rowntree, 1990).

Moist convective adjustment schemes are the most basic cumulus parameterisation schemes. In these schemes, the model-scale vertical temperature profile is adjusted to a critical, stable profile when the relative humidity exceeds a specified value and the temperature profile is unstable with respect to moist air. During adjustment, the temperature profile is adjusted to the pseudo-adiabatic rate, the large-scale relative humidity is unchanged, condensed water vapor precipitates, and total moist enthalpy is conserved. In the Kuo type schemes, rainfall from cumulus convection is assumed to occur when there is a low level moisture convergence. Part of the moisture condenses, releasing latent heat and increasing rainfall. The rest is used to increase the relative humidity of the environment. Cloud dynamics and microphysics are not computed in Kuo schemes and the cloud types are not classified.

One such approach, which has had a large impact on subsequent work in the field, is the mass-flux representation of convection. These types of models were first developed by a number of authors in the 1970s (see e.g. Yanai et al., 1973), however, a complete mass-flux parameterisation was presented by Arakawa and Schubert (1974). Modern implementations of this parameterisation have three components, a trigger model which determines whether the thermodynamic state of the atmosphere is able to support convection, a cloud model which calculates the vertical distribution of convective heating, and a closure which governs the overall strength of the convective response.

1.3 Mass-flux theory of parameterization

The most common way to parameterise the vertical transport of heat, moisture and momentum is through the use of so-called mass flux schemes. In such schemes the updraught strength of a cumulus ensemble is characterized by a mass flux which quantifies the amount of mass that is transported in the vertical. Right hand side terms from thermodynamic equation of the large scale flow of Eq. 8 represents the impact of radiation and small-scale motions upon the large scale flow. The perturbations represent the unresolved cumulus updraft aliased onto the grid, and are non-zero. We assume that the unresolved motions always consist of a closed circulation within the grid-box, so that the grid mean of the perturbations will be zero. Cumulus parameterisation estimates part of these terms, where as there are also contributions from boundary layer turbulence and sub-grid scale phase changes.

It is assumed here that horizontal subgrid-scale fluxes should be represented by another parameterisation scheme and hence convective subgrid-scale motions result only in the net flux transport in the vertical. Thus convection is seen to affect the large-scale flow through condensational heating and the vertical eddy transport of heat. Cloud-environment decomposition in the eddy transport term of the Eq. 8 is obtained by considering the cloud element in the fractional area (σ) and environment with $(1 - \sigma)$, within some area A, taken to be the grid point of a numerical model. We use φ instead of s as the thermodynamic variable, for taking general expression of eddy transport term in Eq. 8. Hence total area average, assuming hydrostatic balance, is given by, $\omega\varphi = \sigma\omega\varphi^{c} + (1 - \sigma)\ \omega\varphi^{\ e}$. Using Reynolds averaging for cumulus elements and environment separately and further simplification by taking small area approximation $\sigma \ll 1 \Rightarrow (1 - \sigma) \approx 1;\ \omega^{\ c} \gg \omega^{\ e}$,

$$\overline{\omega'\phi'} = \sigma\overline{\omega^c}\left(\overline{\phi^c} - \overline{\phi^e}\right) \tag{9}$$

and convective mass flux is defined as: $M_c = -\frac{\sigma\omega^c}{g}$, where ω^c is mean ascent in cumulus cloud, then Eq. 9 is modified as

$$-\overline{\omega'\phi'} = gM_c\left(\overline{\phi^c} - \overline{\phi^e}\right) \tag{10}$$

Hence convective heat source term of the Eq. 4, i.e., right hand side term except radiative heating term, is rewritten as,

$$Q_{1c} = L\left(\overline{c} - \overline{e}\right) + g\frac{\partial\left[M_c\left(\overline{s^c} - \overline{s}\right)\right]}{\partial p} \tag{11}$$

where c and e denotes the rates of condensation and evaporation. To predict the influence of convection on the large-scale with this approach we now need to describe the convective mass-flux, the values of the thermodynamic (and momentum) variables inside the convective elements and the condensation/evaporation term. This requires, as usual, a cloud model and a closure to determine the absolute (scaled) value of the mass flux.

1.3.1 Cloud model for the cumulus ensemble

Model of a convective cloud is obtained either by a one-dimensional steady state entraining plume model of the cloud or bulk mass-flux approach (one representative updraught). Here we hypothesize that, over time scale describing the behavior of the large-scale motion system, the cloud ensemble maintains the heat and moisture balance with the environment.

Using the cloud model, the vertical distribution of convective heating and drying can be deduced in terms of the cloud base mass-flux. It is this mass-flux that must be determined in order to close the model. The convective closure is generally expressed in terms of the energy released by clouds in the process of convection. However, at its root is the concept of convective quasi-equilibrium (QE). The QE hypothesis asserts that the consumption of conditional instability by convective processes close to balances its production by non-convective processes (Emanuel, 1994). Specifically, Arakawa and Schubert (1974) assumed that the change in the ‘cloud work function’ with respect to time was much smaller than the changes due to the large-scale forcing alone. Thus this approximation amounts to the assumption that the timescale over which convection adjusts to the large scale forcing, is much smaller than the timescale over which the forcing changes. Thus the convection is always in a state near to equilibrium with the ‘slow’ processes (Arakawa and Schubert, 1974). The validity of this assumption is obviously dependent on what the temporal and spatial scales which we are considering, as well as exactly what we meant by ‘non-convective processes’ (Randall et al., 1997).

In most operational models, a single entrainment rate is used rather than a spectrum of cloud types. Thus the variable to which the QE hypothesis is applied is the CAPE. Additionally, strict QE, in which the amount of CAPE is constant in not enforced. Instead convection is assumed to relax the atmosphere back to equilibrium over a specified timescale. Thus the statement of quasi-equilibrium becomes,

$$\frac{\partial CAPE}{\partial t} = -\frac{CAPE}{\tau_{CAPE}} \tag{12}$$

Here, CAPE is the grid-box mean convective available potential energy, the tendency is that due to convective processes, and τ_{CAPE} is the CAPE timescale: the timescale over which convection brings the environment back to equilibrium. While the QE hypothesis has been verified in observational studies with some success (see Arakawa and Schubert, 1974; Lord and Arakawa, 1980), the CAPE timescale is difficult to constrain. Gregory and Park (1997) note that this

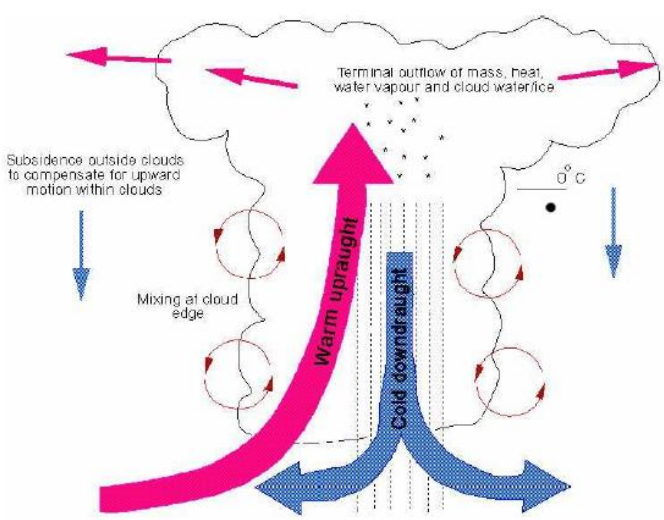

Fig. 4: Convective cloud model

timescale must be resolution dependent, as the mean grid-box vertical velocity is also resolution dependent. Following the analysis of Gregory and Miller (1989), consider the thermodynamic equation averaged over an individual cloud, the fluxes across the cloud boundary as it varies with height and the horizontal growth of the cloud with time by entrainment and detrainment fluxes.

$$\frac{\partial\left(\sigma_i s_i\right)}{\partial t}+D_i s_i-E_i \bar{s}-g \frac{\left(\partial M_i s_i\right)}{\partial p}=L c_i \tag{13}$$

where E_i entrainment rate (the rate at which air gets into the cloud through its side balance the increase of vertical mass flux in the cloud with height), and D_i detrainment rate (the rate at which air leaves the cloud as the cloud vertical mass flux decrease with height). A cloud model for the convective plume in mass flux scheme is illustrated in (Fig.4). It has also been assumed that the whole ensemble is in a steady state, and so remaining time dependent terms can be neglected, therefore $E = \Sigma E_i$ and $D = \Sigma D_i$, $c = \Sigma c_i$ i.e., summation is carried out for over all clouds.

Hence Eq.11 is modified as after some algebra,

$$Q_{1c}=-g M_c \frac{\partial \bar{s}}{\partial p}+D\left(\overline{s^c}-\bar{s}\right)-L e \tag{14}$$

This can be interpreted as: Convection affects large scale by **(a)** heating through compensating subsidence between cumulus elements (term 1), **(b)** the detrainment of cloud air into the environment (term 2) and **(c)** evaporation of cloud and precipitation (term3). A similar analysis can be applied on moisture and horizontal momentum equation.

2 NCMRWF Unified Model convection

The mass flux convection scheme based on Gregory and Rowntree (1990) with various other extensions for GA3.0 and GA4.0 are as given in Walters et al. (2011) and Walters et al. (2014) respectively. NCMRWF UM (NCUM) uses the hybrid height as the vertical coordinate, with temperature, moisture variable, vertical velocity and tracers are held on main model levels, whereas cloud-base and cloud top are defined in model half levels (Rajagopal et al., 2012). The convection scheme (along with the boundary layer scheme) in the NCUM is run multiple times for every model timestep, and we refer to this shorter time interval as the sub-timestep. Sub-time step used in this set up is 2. Top level control routine ATMOS PHYSICS2 calls fast physics for the convection and boundary level schemes and ATMOS PHYSICS1[1] is the slow physics part of NCUM (Fig.5).

In general, NCUM Convection scheme consists of three stages: **(i)** convective diagnosis to determine whether convection is possible from the boundary layer; **(ii)** a call to the shallow or deep convection scheme for all points diagnosed deep or shallow by the first step; and **(iii)** a call to the mid-level convection scheme for all grid points.

Now the field seen by convection in the m_{th} sub-step are:

$$F_{n^*}^{m} = F_n + \Delta F_{atmosphys1} + \Delta F_{dyn} + \sum_{i}^{m-1} \Delta F_{conv}^{i} \qquad (15)$$

where ΔF_{conv}^{i}=increment to the field, F on the i^{th} convection sub-step
where ΔF_{dyn} = increments to F from the semi-Lagrangian dynamics,
$\Delta F_{atmosphys1}$= increments to F from the slow physics part of the UM (ATMOS_PHYSICS1)

[1]ATMOS_PHYSICS1 interfaces to physics schemes: microphysics (cloud and large scale precipitation schemes); radiation; gravity wave drag is called before Semi-Lagrangian in atmosphere timestep. ATMOS_PHYSICS2 interfaces to physics schemes: boundary layer; convection; hydrology and is called after Semi-Lagrangian in atmosphere timestep.

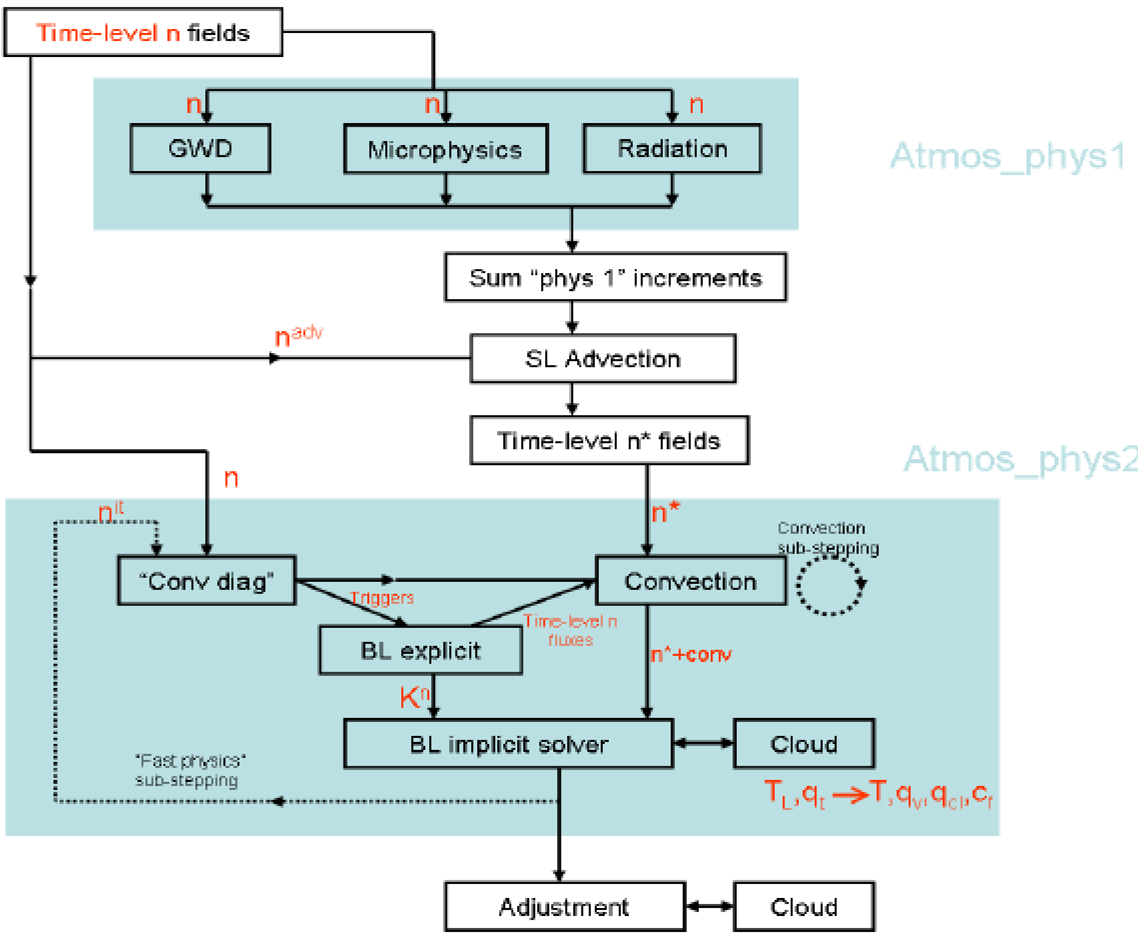

Fig. 5: Schematic diagram of the NCUM timestep showing the location of the convection scheme

2.1 Convective diagnosis

Convective diagnosis calculates information required for the convective and boundary layer scheme. The mass-flux cloud model routine will only be called, if either the convection trigger finds the boundary layer suitable for shallow/deep convection, or the mid-level convection scheme finds instability in the free troposphere. The triggering of surface forced convection is based on the boundary layer parameterisation of Lock et al. (2000). Convective diagnosis checks surface layer instability by calculating the surface buoyancy flux using the bulk aerodynamic formulae.

$$\overline{\omega' b'_{surf}} = g \frac{\overline{\omega' \theta'_{surf}}}{\rho_* \theta_v (k = 1)} \tag{16}$$

The surface buoyancy flux is calculated according to whether the surface layer is land/sea/sea ice. The parcel calculations are done for those locations, where the surface layer is unstable i.e. $\overline{\omega' b'_{surf}} \geq 0.0$. The parcel is initialised with the properties of the environment above the surface layer. Before the parcel is lifted, however, the pressure and temperature at LCL are calculated using an approximate formula derived by Bolton (see Lock et al., 2000) using vapour pressure 'e' at the start of the undiluted ascent i.e. k = kplume.

$$e_{kplume} \approx \frac{q^{E}_{kplume} p_{kplume}}{100.\epsilon} \tag{17}$$

where ε= 0.621 in NCUM.

If the surface buoyancy flux is positive, the parcel is lifted reversibly in order to find the top of the boundary layer. In clear air the top of the boundary layer is assumed to be the level at which the parcel becomes negatively buoyant by an amount greater than a threshold θ_{pert}. Like the threshold for forced detrainment b_{min}, this is a variable, but has a minimum value of 0.2 K. If the parcel reaches above the LCL, the boundary layer top can also be taken as the level at which the excess buoyancy of the parcel over the environment is maximum. Thus instead of using the absolute value of the buoyancy excess, the cloud and environment buoyancy gradients are compared. The diagnosis of shallow and deep convection is based on an undiluted parcel ascent from the near surface for grid boxes where the surface layer is unstable and forms part of the boundary layer diagnosis (Lock et al., 2000).

Parcel ascent is started from the level, kplume ($\approx 0.1Z_h$, where Z_h is the boundary layer depth) and it assumed to conserve moist static energy and total water (q_T) from the start of the parcel ascent. The parcel potential temperature is calculated by assuming the parcel ascends following a moist adiabat above the LCL and conserving initial parcel theta. The parcel condensate at level k is estimated using Taylor expansion of q_{sat} about the parcel reference temperature and hence parcel temperature T will be recalculated based on the partial differential equation,

$$\frac{\partial q_{sat,k}}{\partial T} = \epsilon L \frac{q_{sat}\left(T_{ref}, p_k\right)}{R T_{ref}^2} \tag{18}$$

Similarly, environment liquid water (ql_k^E) is also estimated and it will be used for the buoyancy measure. The undiluted parcel ascent assume, all the parcel condensate remain in the plume adding to water loading. Top of the parcel ascent (k = ntpar), CAPE and CINE are the quantities calculated during the parcel ascent. Boundary layer cloud is diagnosed if the parcel reaches above the LCL. If this cloud is determined to be cumulus, the convection scheme is called, otherwise it is assumed to be stratocumulus, and mixing is performed by the diffusion profiles of the boundary layer parameterisation. A cumulus capped boundary can be diagnosed, if (i) cloud depth > 400 m and greater or equal to three model levels and the cloud base (LCL) is below model level ten, and (ii) Top of the parcel ascent≥ 3000m. Additionally, the lifting condensation level must be above the surface layer, otherwise a fog layer is diagnosed, and convection does not occur. If all these conditions are satisfied, stratocumulus decks are differentiated from cumulus regions by comparing gradients of water content in the cloud and sub-cloud layer. Based on the assumption that stratocumulus cloud decks are relatively well-mixed in comparison to regions of convection, it is assumed that in cloud gradients of moisture are larger in cumulus clouds than in stratocumulus (Lock et al., 2000). That is, cumulus convection is diagnosed if,

$$\frac{\partial q_T}{\partial z_{cloud}} > 1.1 \frac{\partial q_T}{\partial z_{sub-cloud}} \tag{19}$$

where q_T is the total water content and gradients averaged over the cloud and sub-cloud layer. Finally, one more condition must be satisfied if the convection scheme is to be called in a given time step. The integral of the parcel buoyancy within the cloud layer is required to be positive. That is, a reversible, undiluted parcel ascent must result in the net release of available potential energy for the boundary layer to be seen as able to support cumulus convection. These are the different checks available in the convective diagnosis whether convection scheme called or not for the particular parcel ascent. Cumulus boundary layer is then diagnosed for shallow, mid-level and deep convection.

Shallow convection is then diagnosed if the following conditions are met: (i) the parcel attains neutral buoyancy below 2.5 km or below the freezing level, whichever is higher, and (ii) the air in model levels forming a layer of order 1500 m above this has a mean vertical velocity less than 0.02 m/s. Otherwise, convection diagnosed from the boundary layer is defined as deep. The mid-level scheme can be triggered at any sub-timestep by a simple stability criterion. The mid-level scheme operates on any instabilities are found in a column above the top of deep or shallow convection or above the LCL.

2.2 Deep convection

As discussed in the earlier section, mass flux approximation is used in the NCUM convective parameterisation, and bulk-cloud model is for representing the ensemble of convective cloud with different characteristics. Equations governing cloud mass flux (M, Pa/s), potential temperature (θ, K), moisture (q, kg/kg) and cloud liquid water (l, kg/kg) for a cloud I within the ensemble are given by,

$$-\frac{\partial}{\partial p}(M_I^c) = E_I - N_I - D_I \tag{20}$$

$$\frac{\partial}{\partial p}(M_I^c\theta_I^c) = E_I\theta_I^e - N_I\theta_I^N - D_I\theta_I^R + \frac{L'}{c_p\Pi}Q_I \tag{21}$$

$$\frac{\partial}{\partial p}(M_I^c q_I^c) = E_I q_I^e - N_I q_I^N - D_I q_I^R - Q_I \tag{22}$$

$$\frac{\partial}{\partial p}(M_I^c l_I^c) = E_I\left(qcl_I^e + qclf_I^e\right) - N_I l_I^N - D_I l_I^R - Q_I - PREC_I \tag{23}$$

Entrainment and detrainment rates are given by,

$$E_I = \epsilon_I M_I^c;\ \ N_I = \mu_I M_I^c;\ \ D_I = \delta_I M_I^c \tag{24}$$

Above equations are similar to those presented by Yanai et al.(1973), albeit difference in the treatment of detrainment processes. In addition to the entrainment rates, mixing detrainment

(N_I) represents the detrainment of cloud air through the turbulent mixing at the edge of the cloud and the forced detrainment (D_I), occurs near the cloud top and thus integrating all cloud types leads to bulk updraught cloud model.

The discretized forms of equation (20 - 23) can be derived for the profiles of temperature, humidity and wind within the cloud model. These equations depend only on (1) the entrainment and detrainment rates, and (2) the cloud base mass flux. The cloud base mass flux is calculated via the convective closure, while entrainment and detrainment are parameterised directly. These are discussed in the following subsections.

2.2.1 Entrainment and detrainment

The interaction between updraughts and their environment is represented in cloud models by the mixing of environment air into the convective plume, and this process is known as entrainment. Air entrained into the plume is assumed to have grid-mean (environment) properties. In rising a distance -Δp, the plume entrains a mass of - ε M Δp of environmental air (M is negative in upward) and changes the thermodynamic property φ and mass of the plume. The entrainment rate is determined as,

$$-M^{-1}\frac{\partial M}{\partial p} = \epsilon \simeq \frac{0.2}{R} \tag{25}$$

Changes in the thermodynamic property φ of the plume by the entrainment ε is given as,

$$\frac{\partial \Phi^c}{\partial p} = \epsilon(\Phi^c - \overline{\Phi}) \tag{26}$$

Entrainment can be further divided to dynamical entrainment, due to large scale organised flow and turbulent entrainment, caused by turbulent mixing at the cloud edge. The original entrainment rate (ε) given in Gregory and Rowntree (1990) is determined as

$$\epsilon = f_{dp} 3 A_E \frac{p(z)\rho(z)g}{p_*^2} \tag{27}$$

where f_{dp} is the entrainment factor, and $A_E = 1.5$ for all model levels, except lowest model level(k = 1) where $A_E = 1.0$. Sensitivity to the different entrainment profile in the GA3.0 and GA4.0 will be discussed in the section 4.1.However, within an ensemble updraughts will reach different final heights and, at any one time, they may be at different stages in their life cycles. Such changes in the mass flux are known as detrainment and to model it a fractional detrainment rate (δ) is introduced. Detrainment is divided into two components, namely, mixing detrainment and forced detrainment. The evaporation of convective plume air due to the mixing with drier environmental air is the counterpart of mixing entrainment. Mixing detrainment is dependent on the relative humidity as,

$$\mu = 1.5(1 - RH)\epsilon \tag{28}$$

Thus total change in the mass flux with height is,

$$-M^{-1}\frac{\partial M}{\partial p} = (\epsilon - \delta) \tag{29}$$

2.2.2 Forced detrainment and termination of updraught

When an individual member of the ensemble of convective plumes reaches zero buoyancy its cloudy air will detrain into the environment. For the weaker members this effect is included in the fractional detrainment rate. Forced detrainment occurs when the buoyancy of the parameterised plume, which is effectively an ensemble mean, becomes less than a prescribed minimum value. Sufficient cloudy air is then detrained into the environment to restore this minimum buoyancy to the remaining plume. Effectively the fractional detrainment in Eq. 29 increases such that $(\varepsilon - \delta)$ becomes negative and the mass flux decreases as the convective plume ascends. Forced detrainment of all cloudy air occurs at a height where an undiluted plume would have zero buoyancy.

2.2.3 Precipitation processes

The potential temperature θ, humidity mixing ratio q, and liquid water mixing ratio ql (and/or ice) are used as thermodynamic variable in Eq. 26 within the cloud model and accounted for the conversion between phases and release of latent heat. In rising $-\Delta p$ there is a conversion of cΔp of water vapour to liquid/water/ice and $\Delta p \frac{\partial P_R}{\partial p}$ is lost through precipitation. Eq. 26 then becomes,

$$\frac{\partial \theta^c}{\partial p} = \epsilon \left(\theta^c - \overline{\theta}\right) + \frac{Lc}{c_p \Pi M} \tag{30}$$

$$\frac{\partial q^c}{\partial p} = \epsilon \left(q^c - \overline{q}\right) - \frac{c}{M} \tag{31}$$

$$\frac{\partial q_l^c}{\partial p} = \epsilon q_l^c + \frac{c}{M} - \frac{1}{M}\frac{\partial P_R}{\partial p} \tag{32}$$

Precipitation occurs within the plume model, when certain thresholds are exceeded, mainly by the coalescence and Bergeron processes. Typically there will be a minimum depth of updraught for both continental (3-5 km) and maritime air masses (1-2 km) for precipitation to be initiated, through difference in the characteristics of cloud condensation nuclei. Liquid water exceeding the threshold value is converted to precipitation. The cloud condensate within the parcel if, $l_k^p > l_{\min}$, where $l_{\min}$ is the minimum value of convection over land/sea. If the criteria are met, the amount of precipitation produced (kg/m2/s) in taking the ensemble from level k to k + 1 is

$$PREC_{k+1} = (l_{k+1}^p - l_{min})\frac{M_{k+1}}{g} \tag{33}$$

A proportion of precipitation created in the updraught is assumed to fall through the downdraught, undergoing evaporation and sublimation. The downdraft is a single entraining plume that begins at the column minimum of moist static energy or Level of Free Sinking (LFS). The mass flux though the downdraft plume is determined as a fraction of the updraft mass flux at cloud base and the precipitation efficiency. The vertical distribution of the downdraught mass flux, dry static energy, moisture and horizontal momentum below the LFS are determined by entraining/detraining plume equations similar to those for the updraught.

The downward motion is driven by the evaporation of the falling precipitation and also by the downward drag of the water. It is based upon a saturated inverted entraining plume, whose negative buoyancy is maintained by the evaporation and sublimation of the falling precipitation and changes of phase. Downdraft is initialised at a level where the saturated updraught terminates, the depth of the cloud layer exceeds 150 hPa, the column accumulated precipitation from the updraught is greater than 1.e-12 and the starting level is above 150 hPa of the surface. The initial downdraught mass flux is given by

$$M_{init}^{DD} = \alpha\, M_{ref}^{\ UD} \quad , \text{where } \alpha = 0.1$$

The reference level is chosen to be ¾ the way up the depth of the cloud, the mass flux of which is used for downdraught calculations irrespective of the level from which downdraught is initialised. The diagnosis for the downdraught initialisation starts from the lowest layer above the cloud top and works down to within 150 hPa of the surface. The downdraught air is made up of equal mixture of cloudy updraught and unsaturated environmental air, which is then brought to saturation through the evaporation/sublimation of the precipitation. The heat released by saturating the downdraught is used to find out the saturation point. Taylor expansion of qsat about a first guess potential temperature is used with Clausius-Clapeyron equation to get the initial saturation potential temperature of the downdraught. If the resultant buoyancy of the saturated parcel is negative, a downdraught is initialised from that level and the corresponding mixing ratio is set as its saturation value at that level temperature.

Parcel descend is driven by the maintenance of negative buoyancy by evaporation/sublimation into the sub-saturated downdraught and water loading, which offset for the positive buoyancy by the adiabatic warming due to the dry descent. If the buoyancy of the downdraught exceeds 0.5K, then the downdraught is totally detrained and the descent is terminated. The downdraught modifies the cloud environment through the en/detrainment of heat and moisture and upward motion within the clear air, compensating for the downward motion within the cloud. A proportion of the precipitation created in the updraught is assumed to fall through the downdraught, undergoing evaporation and sublimation as it does so. Also some of the precipitation is allowed to enter the downdraught during its descend in the cloudy air. As the precipitation falls through the downdraught, it accounts for limited phase-changes in the form of metling and freezing processes although it is possible for rain and snow to coexist within the downdraught. Below the cloud base, the precipitation undergoes evaporation or sublimation. It is also assumed that non-downdraught precipitation occupies a fractional area of the grid-box (convective cloud area) calculated in the updraught part. Evaporation/sublimation is calculated using precipitation rates divided by the downdraught fractional area, which is taken as the 50% of the updraught area. Part of the rain and snow which does not undergo evaporation is allowed to fall on to the ground.

2.2.4 Cloud-base closure

The convection scheme which uses an adjustment type closure, the initial mass flux at the base of the updraught is calculated from relaxing the atmosphere back to an equilibrium state. In the CAPE closure, for instance, the cloud base mass flux is calculated based on the reduction to zero of CAPE over a given timescale. The cloud-base mass flux is obtained from a simplified turbulence kinetic energy (TKE) budget for the sub-cloud layer, in which cumulus convection is assumed to be associated with a transport of TKE from the sub-cloud layer to the cloud layer. The rate of change of θ and q in terms of cloud mass flux will be relaxed back to the atmosphere at a CAPE time scale (τ) due to convective activity. The mass fluxes (and grid-mean convective tendencies) are proportional to CAPE and inversely proportional to CAPE time scale. In the termination option, some thermodynamic quantity (e.g. humidity, moist static energy) is assumed constant in the sub-cloud layer when convection is present. Changes due to all other processes are first calculated in the sub-cloud layer (e.g. advection, radiation) and then the bottom level flux is set equal to these changes. Sensitivity of the variation of CAPE closure in the NCUM according to the CAPE timescale (τ) to the convective rain will be discussed detailed in Section 4.2.

2.2.5 Impact on environment

As with the mass flux scheme, the plume interacts with its environment through en/detrainment of heat, moisture and cloud liquid water , as well as subsidence induced in the cloud environment which compensate for the parcel upward mass flux which are summarized in the apparent heat source and moist sink in the Eq. 6 and 7. Using the mass flux approximation, these equations become,

$$Q_1 - \frac{1}{\Pi}Q_R = -\frac{\partial \overline{M}}{\partial p}\left(\theta_l^c - \overline{\theta}\right) - M\left(\frac{\partial \theta_l^c}{\partial p} - \frac{\partial \overline{\theta}}{\partial p}\right) + \frac{L}{c_p \Pi}\frac{\partial P_R}{\partial p} + \frac{L}{c_p^2 \Pi p}\frac{\partial P_R}{\partial p} M q_l^c \tag{34}$$

$$-\frac{Q_2}{L} = -\frac{\partial M}{\partial p}\left(q_t^c - \overline{q}\right) - M\left(\frac{\partial q_t^c}{\partial p} - \frac{\partial \overline{q}}{\partial p}\right) - \frac{\partial P_R}{\partial p} \tag{35}$$

Equations 26 and 29 can then be used to give

$$Q_1 - \frac{Q_R}{\Pi} = M\frac{\partial \overline{\theta}}{\partial p} - M\delta\left(\theta_l^c - \overline{\theta}\right) \tag{36}$$

$$-\frac{Q_2}{L} = M\frac{\partial \overline{q}}{\partial p} - M\delta\left(q_t^c - \overline{q}\right) \tag{37}$$

so that the liquid water content of the environment is zero. The first terms on the right-hand sides of Eq. (36) and (37) can be identified as compensating subsidence which is warming and drying the environment (M is negative in pressure coordinates). The second terms describes the detrainment of cloudy air and will tend to cool and moisten the environment, particularly in regions of forced detrainment near the top of the updraught. Rewriting Eq. (36) and (37) in terms of θ, q and q_l , the equations become

$$Q_1 - \frac{Q_R}{\Pi} = M\frac{\partial \overline{\theta}}{\partial p} - \delta M\left(\theta^c - \overline{\theta}\right) + \delta M\frac{L q_l^c}{c_p \Pi} \tag{38}$$

$$-\frac{Q_2}{L} = M\frac{\partial \overline{q}}{\partial p} - M\delta\left(q^c - \overline{q}\right) - \delta M q_l^c \tag{39}$$

The last terms arise from the detrainment of liquid water into the environment giving rise to a cooling through evaporation and a moistening of the environment. These are less effective in deep precipitating convection due to the loss of liquid water through precipitation. This is consistent with the observed warming and drying associated with deep convection.

2.2.6 Convective Momentum Transport

The parameterisation of convective momentum transports (CMT) for deep convection is based upon the mass-flux convection-scheme discussed by Gregory et al. (1997). Since it is a parameterised process, the effect of the convection upon the large scale momentum field is estimated from the vertical eddy fluxes of horizontal momentum as

$$\frac{\partial u^c}{\partial p} = \varepsilon(u^c - \overline{u}) + C_u \overline{\frac{\partial u}{\partial p}} \tag{40}$$

The extra term on the right-hand side represents the horizontal pressure gradient in the environment which will influence the momentum within the plume. C_u is a constant (=0.7) in the NCUM.

2.3 Shallow convection

Shallow cumulus convection predominantly occurs in undisturbed flow i.e. in the absence of large-scale convergent flow. This type of convection seems to be effectively controlled by sub-cloud layer turbulence. The shallow convective scheme is triggered when an initial parcel ascent indicates convection will terminate below the freezing level, and the large scale flow implies subsidence. The main differences between the shallow scheme and the deep scheme are that the shallow scheme includes more mixing with the environment, and it uses the boundary-layer turbulence kinetic energy closure of Grant (2001). In fact, most of the diagnostic studies carried out for trade-wind cumuli show that the net upward moisture flux at cloud-base level is nearly equal to the turbulent moisture flux at the surface. If cumulus convection is diagnosed then the shallow scheme is triggered from the top level of mixed layer (k=ntml).

The updraught in the shallow convection is treated with different rate of entrainment/detrainment based on similarity theory developed by Grant and Brown (1999).The net mixing detrainment is also sensitive to the RH, and mass flux decrease with the height. The cloud base closure for shallow convection is based on Grant (2001), the cloud mass flux m_b is given by,

$$m_b = 0.03\omega_* \tag{41}$$

where ω_* is the sub-cloud convective velocity scale. A check is applied that the initial mass flux does not exceed the mass of the layer, where the convection starts and that the mass flux at higher levels in the ascent does not break CFL criterion.

Convective momentum transport for shallow convection is based on turbulence based Grant CMT scheme. This is actually derived for the shallow convection but this approach is applied for deep convection, as it significantly improved low level tropical wind in the model. For shallow convection, the momentum flux is given by

$$\overline{u'\omega'} = -K\frac{\partial U}{\partial Z} + \overline{u'\omega'}_{cb}F(\frac{z'}{z_{cld}}) \qquad (42)$$

where K is the eddy viscosity, $\overline{u'\omega'}_{cb}$ is the cloud base momentum flux and F is the similarity function.

2.4 Mid-level convection scheme

The mid level convection scheme acts either above deep or shallow convection to remove instabilities in the model, or above the boundary layer in columns where the surface layer is stable and below stratosphere. The mid-level scheme can be triggered at any sub-time step by a simple stability criterion. A parcel from model level k is lifted to level k + 1 pseudo-adiabatically (i.e. irreversibly). Upon reaching level k +1 the mid-level convection scheme is allowed to run if its equivalent potential temperature satisfies,

$$(\theta_e)_c - (\theta_e)_E > -\triangle\theta \qquad (43)$$

The maximum stability, $\Delta\theta$ is set to 1.5 K in the standard scheme. This stability test is applied to all levels above the boundary layer and above any surface driven convection to determine the occurrence of mid-level convection. Similar deep scheme en/detrainment rate setting is used in the mid-level, except mixing detrainment currently has no dependence on RH. Mid-level mixing detrainment follows the formula,

$$\mu = (1 - \frac{1}{1.5f_{dp}})\epsilon \qquad (44)$$

where f_{dp} is the same adjustable parameter as in Eq. 27. Gregory-Kershaw CMT scheme is used for setting initial winds in the values of the plume being set to the winds at the level mid-level convection initiated.

3. Convection Subroutines

Interface to Atmospheric Physics convection code is done through Subroutine NI_CONV_CTL. That is, one call to NI_CONV_CTL from ATMOS_PHYSICS2.

The control level routine of the convection scheme is illustrated in the Fig. 6. The convective state of the boundary layer is determined by calculating surface buoyancy flux by CONV SURF FLUX. If there is more than one convection step per model physics time-step then re-diagnose cumulus points by recalling CONV DIAG before each convective sub-step. Convection occurrence and type of unstable point is determined by CONV DIAG COMP. According to CONV_DIAG diagnosis, the convection scheme is called in GLUE_CONV, either for deep, shallow, mid or congest modes. UPDATE routine updates convective cloud diagnostics and prognostics after call to GLUE each sub-step of convection. Prognostic condensate scheme PC2_HOM_CONV calculates the change in liquid content, liquid cloud fraction and total cloud fraction based on detraining condensate and cloud fraction updates from GLUE directly into the large scale.

3.1 Subroutine NI CONV CTL

All working arrays other than prognostics are defined without surface level values, i.e. the convection scheme only operates on model levels above the surface.

The main subroutines called by NI_CONV_CTL are listed below:

1. CONV_PC2_INIT - This subroutine is called to apply logic checks based on the options selected in the run cloud namelist.

2. CONV_DIAG_4A

3. CONV_DIAG_5A

4. CONV_DIAG_6A

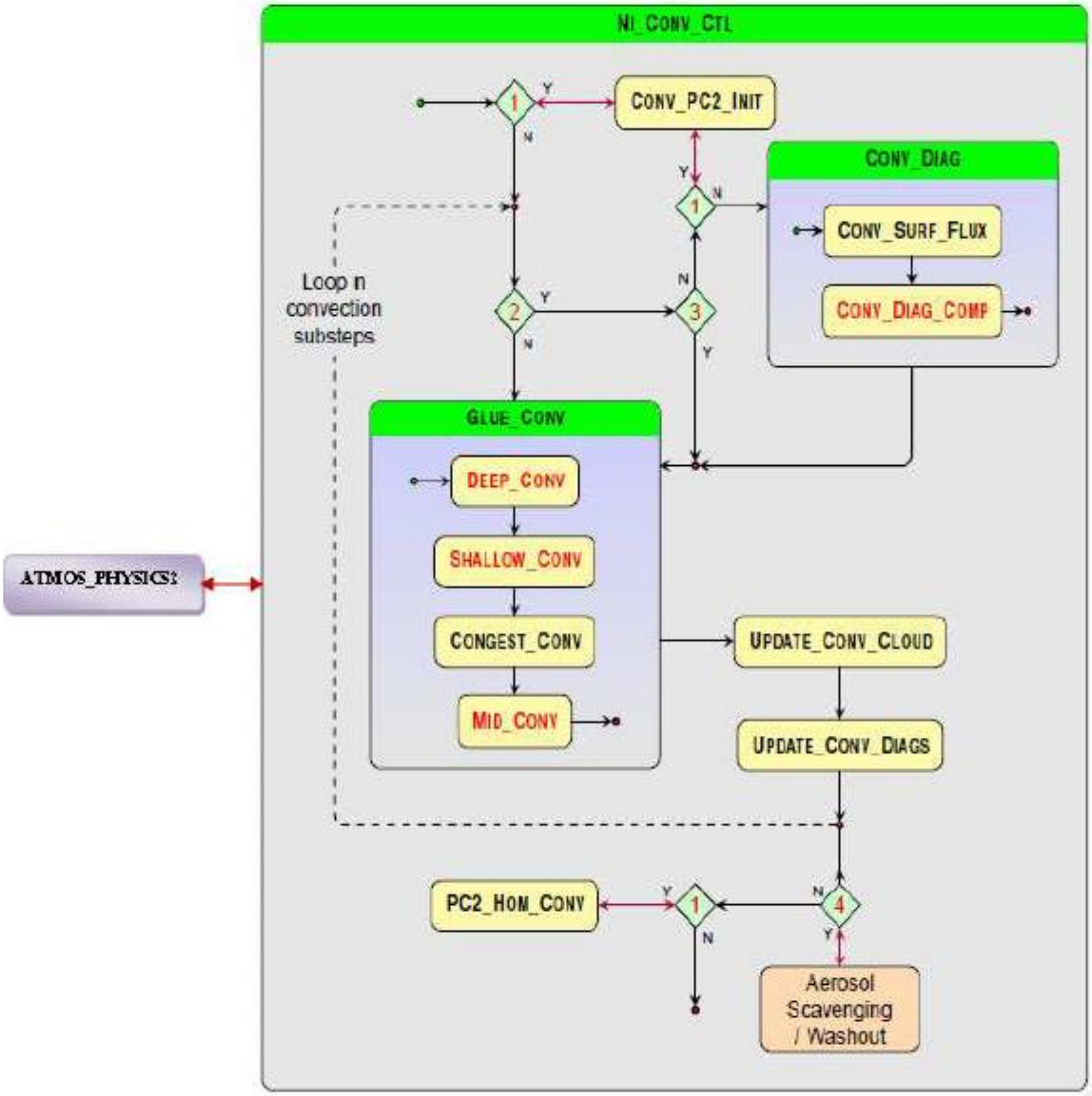

Fig. 6: Flowchart for NI_CONV_CTL

Subroutines 2-4 are to diagnose convective occurrence and type which are called by ATMOSPHYSICS2. This routine works on the model levels and not the surface level so all level loops start at 1.

• CONV_SURF_FLUX - Calculate surface fluxes for convective diagnosis. There are calls to CONV SURF FLUX from CONV DIAG 6A; CONV DIAG 5A

• CONV_DIAG_COMP - To diagnose convection occurrence and type on just unstable points. It is called by CONV_DIAG_6A & CONV_DIAG_5A.

5. GLUE_CONV_4A

6. GLUE_CONV_5A

7. GLUE_CONV_6A

Subroutines 5-7 are gather-scatter subroutines for deep and shallow convection points. Interface to deep, shallow and mid level convection.

• DEEP_CONV: Deep convection scheme - works on points diagnosed as deep in subroutine CONV_DIAG. Called by GLUE_CONV.

• SHALLOW_CONV: Shallow convection scheme - works on points diagnosed as shallow in subroutine CONV_DIAG. Called by GLUE_CONV.

(NOTE: Flowchart for DEEP_CONV is given in Fig. 7 and similar flow chart can be drawn for SHALLOW_CONV and MID_CONV, not included here.)

• MID_CONV: Mid level convection scheme - works on all points. Called by GLUE_CONV

8. UPDATE_CONV_CLOUD

9. UPDATE_CONV_DIAGS

These subroutines update convective cloud diagnostics and prognostics after call to GLUE at each sub-step of convection.

10. PC2_HOM_CONV

This subroutine calculates the change in liquid content, liquid cloud fraction and total cloud fraction as a result of homogenous forcing of the gridbox with temperature, pressure, vapour and liquid increments and turbulent mixing effects.

11. CV_PARAM_MOD: Module containing parameters used by the convection code.

12. CV_RUN_MOD: Module containing runtime logicals/options used by the convection code. All switches/options which are contained in sub-namelist (&RUN_Convection) in the SHARED control file are declared in this module. Default values have been declared where appropriate.

13. CV_STASH_FLG_MOD: Module containing stash flags used in top level convection routine. Declares all flags and contains a subroutine used to set the flags.

14. LEVEL_HEIGHTS_MOD: This module is used to hold levels values set in atmos init. The height levels arrays are calculated in routine control/top level/set levels called from atmos init, and used in dynamics advection and other routines.

15. CON_SCAV: Scavenge aerosol by convective precipitation.

16. C_BM_CON_MOD: Convective scavenging coefficients for biomass smoke

17. C_OCFF_CON_MOD: Convective scavenging coefficients for fossil-fuel organic carbon

18. C_ST_CON_MOD: Convective scavenging coefficients for soot

19. NCNWSH2: Scavenge ammonia (NH3) by convective precipitation

20. SCNSCV2: Scavenge Sulphur Cycle tracers by convective precipitation

21. SCNWSH2 : Scavenge SO_2 by convective precipitation

22. SCMOUTPUT

Create output diagnostic based on given inputs. This routine cannot be called more than once per time step with the same ”sname” unless the call is inside a sub-stepped part of the model and the sub-stepping has been delimited by calls to scm sub-step start and scm sub stepping end. The order of calls to SCM output should not change between time steps.

23. SCM SUBSTEP START

In order to be able to output diagnostics that are defined within sub-stepped parts of the model, the diagnostic system needs to know which parts of the model are sub-stepped and which sub-step the model is currently on, if any. This routine is called to let the system know that the specified sub-step has just started. The system will assume that it is on the specified sub-step until this routine is called again to indicate the start of the next sub-step, or SCM SUBSTEPPING END is called to indicate that the sub-stepping has now ended. Any call to SCM output from a sub-stepped part of the code that is not delimited by calls to these two routines (the first inside the sub-stepping loop and the second outside once it has finished) will result in an error message.

3.2 Deep Convection

1. CONVEC2_4_5A

This routine completes lifting of the parcel from layer k to k+1 and calls subroutines parcel and environ. Subroutine parcel calculates an initial mass flux, carries out the detrainment calculation, tests to see if convection is terminating and calculates the precipitation rate from layer k+1. Subroutine environ calculates the effect of convection upon the large-scale atmosphere.

• PARCEL_4A5A: Completes lifting of the parcel from layer k to k+1. It calls DETRAIN, TERM_CON and CLOUD_W. An initial mass flux is calculated in this routine.

DETRAIN:- This carries out the forced detrainment. TERM_CON:- This tests for any convection which is terminating in layer k+1. CLOUD_W:- This carries out the cloud microphysics calculation.

• ENVIRON Calculate the effect of convection upon the large-scale atmosphere.

• CAPE Closure & CFL scaling - adjust initial mass flux so that CAPE is removed by convection over timescale (τ)

2. DD_ALL_CALL_4A5A: Calculates initial downdraught mass flux. Reset en/detrainment rates for Downdraught.

3. DQS_DTH: Calculates gradient of saturation mixing ratio with potential temperature from the Clausius-Clapeyron equation

4. EVAP_BCB_NODD_ALL

To calculate the convective precipitation reaching the surface. The evaporation below cloud base follows that done in the downdraught routine for the environmental part of the column. The points which are gathered here are those points which have an updraught, but no downdraught.

5. FLAG_WET: Calculates a mask for when condensation is liquid Conditions.

If $0.5 \ast (T\ K + T\ K + 1) > T\ ICE$ then any condensation in layer k+1 is liquid

If $0.5 \ast (T\ K + T\ K + 1) < T\ ICE$ then any condensation in layer k+1 is ice

6. LAYER_CN: Calculates layer dependent constants for layer K i.e. pressure, layer thickness, entrainment coefficients, and detrainment coefficients

7. LIFT_PAR: Lifts Parcel from layer k to k+1 taking environment and moist processes into account

8. MIX_IPERT_4A5A: To well-mix convective increments from the initial parcel perturbation in shallow/deep convection throughout the boundary layer.

9. MIX_INC: To well mix convective increments in the boundary layer.

10. CMT_MASS: Calculates the mass flux profile for deep convection to be used in CMT calculations. Uses the cloud-base mass flux from the plume scheme, but profile is not the same as used for the thermodynamic part of the convection scheme.

11. DEEP_CMT_INCR: Calculates increments to U and V due to deep convection.

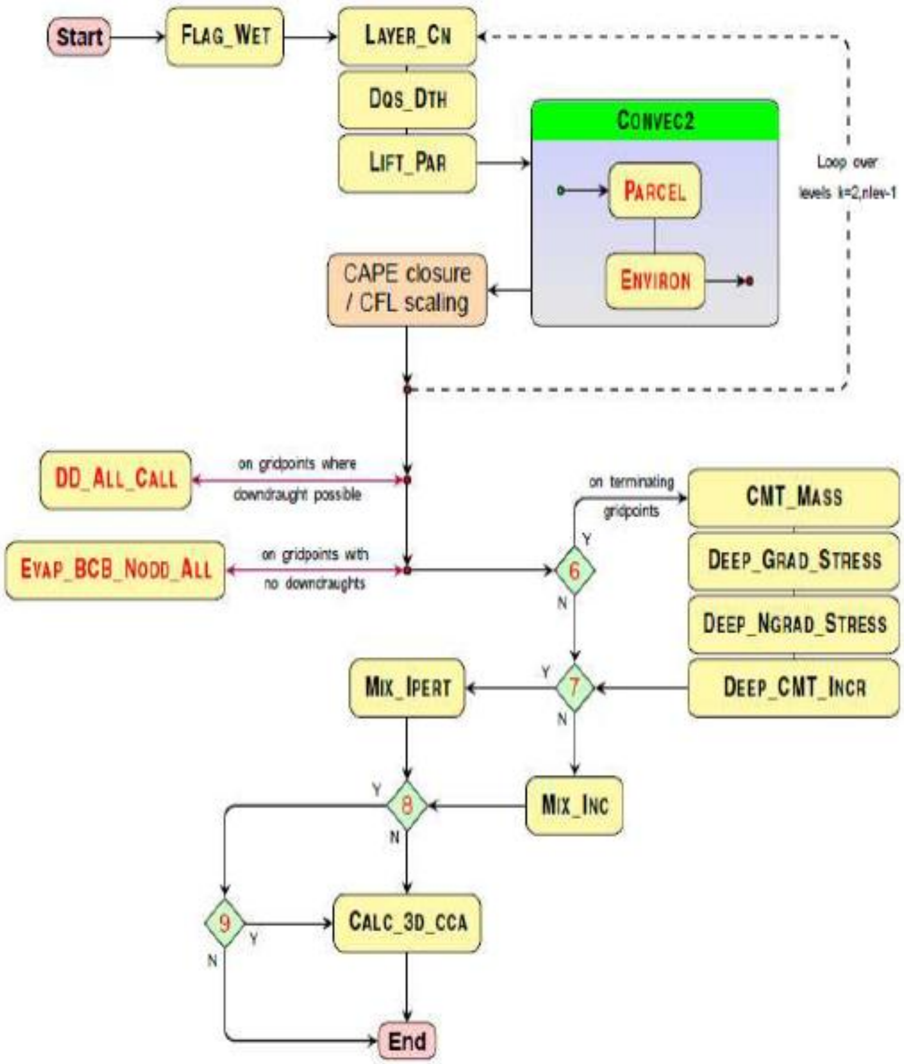

Fig. 7: Flowchart for DEEP CONV

12. DEEP_GRAD_STRESS: Calculates the gradient component of the stress profile for deep convection. Calculation can be done explicitly or implicitly.

13. DEEP_NGRAD_STRESS: To calculate the cloud base stress for deep convection and complete the calculation of the stress profile.

14. DEEP_TURB_CMT: This routine calculates convective momentum transport for deep convection using turbulence ideas. This version is designed for use with the mass flux convection scheme.

15. CALC_3D_CCA: Calculates a 3D convective cloud amount (i.e. on theta model levels) from the 2D convective cloud amount array according to parameters specified in the UMUI and the position of cloud base, cloud top and freezing level.

Method: The 2D convective cloud amount is expanded into the vertical by applying it between the cloud base and top with the additional constraints that:

(ia) If the cloud base is in the boundary layer (Original), (ib) If the cloud base is below the freezing level (CCRad), (ii) Cloud top is above the freezing level and (iii) The cloud is more than 500 mb deep

Then the cloud below the freezing level will be multiplied by

TOWER_FACTOR, and the cloud above the freezing level will be linearly increased to cloud top where it will be equal to the 2D fraction * ANVIL FACTOR.

3.3 Namelist

Namelist for GA4.0 job at a code base vn8.5 at a resolution of N216, where only convection part is labeled below

&RUN Convection

i_convection_vn = 5,	Turbulence and mass flux scheme (Option for convective scheme versions)
l_safe_conv=.TRUE.,	Safety check for the input q profile are above q_min_conv and negative CAPE
L_MOM=.TRUE.,	Logical Switch to allow convection to transport momentum in the vertical (in addition to heat & moisture)
icvdiag=1,	Convective diagnosis method; corrected undilute parcel diagnosis schemes with improve diagnosis, c_b atleast 400m, LCL atleast 1500m
tv1_sd_opt=2, use the BLD	Calculate std. dev of the level 1 virtual temperature from previous time step improved by parameterized roughness length
plume_water_load=0,	water loading option kept for whole of parcel ascent.
dil_plume_water_load=0,	same option kept for dilute plumes

cvdiag_sh_wtest=0.020,	recommended w test value of 0.02 m/s
cvdiag_inv=0	Inversion test for determining shallow convection
l_rediagnosis=.TRUE.,	A new method of sub-stepping convection in the diagnosis routine.
deep_cmt_opt=5,	deep convective momentum transport, option uses a quadratic shape factor to reduce CMT increments at upper levels.
mid_cmt_opt=0,	Gregory-Kershaw mass flux scheme.
DD_opt=1,	Downdraught scheme option , 1 for revised code
mid_cnv_pmin=10000.00,	Minimum pressure for mid level convection (Pa) to test for the initiation.
bl_cnv_mix=1,	Sub cloud mixing method
l4a_kterm=.FALSE.,	Use kterm instead of NTPAR for Deep CMT
CCW_FOR_PRECIP_OPT=4,	Critical convective cloud condensate option. Condensate above this limit is converted to precipitation
qlmin=3.0000e-04,	Min. value of critical cloud condensate profile
fac_qsat=0.500,	This sets fac_qsat which is the critical cloud condensate profile expressed as a proportion of the saturated specific humidity of environment
L_CLOUD_DEEP=.TRUE.,	Use depth criterion for convective anvils
A_CONVECT_SEGMENTS=-99,	No of batches (or "segments") used in convection
A_CONVECT_SEG_SIZE=80,	Size of convection segments (alternative to No. of convection segments)
MPARWTR=1.5000e-03,	Maximum value of the critical cloud condensate profile
L_anvil=.TRUE.,	Apply anvil scheme to 3D convective cloud
ANVIL_FACTOR=1.0000,	The radiative representation of anvils modifies the convective cloud amount (CCA) to vary with height during deep convection.
TOWER_FACTOR=1.0000,	The CCA is multiplied by the tower factor from the cloud base to the freezing level, and increased linearly from the freezing level to the cloud top
UD_FACTOR=1.0000,	Updraught factor for reducing cloud water
CAPE_TIMESCALE=1800,	This determines the e-folding time for the dissipation of CAPE (seconds).
CAPE_OPT=1,	Choice of Cape closure scheme
ADAPT=7,	Adaptive detrainment option; option 7 is the improved version of smoothed adaptive detrainment
R_DET=0.9000,	Adaptive detrainment control parameter
amdet_fac= 3.00,	Deep mixing detrainment control parameter
ENT_FAC=1.0,	Controls mid-level and deep entrainment in 4A convection.
CAPE_MIN=0.5,	Value of CAPE below which parametrized convection is reduced (J/kg).
W_CAPE_LIMIT= 0.4,	CAPE Threshold vertical velocity used in the Vertical velocity dependent CAPE timescale option.
CAPE_BOTTOM=5,	Lowest model level for rescaling parametrized convection
CAPE_TOP=50,	Highest model level for rescaling parametrized convection
TICE=263.1500,	Specify phase change temperature /K
QSTICE=3.5000e-03,	Estimate of saturation specific humidity at this phase change T.

anv_opt= 0,	The anvil profile basis selection; 0 for pressure.
limit_pert_opt=2,	Limit initial parcel perturbation option
cnv_wat_load_opt=0,	Water Loading Option
cca2d_sh_opt= 1,	Shallow convective cloud settings: CCA 2D basis
cca2d_md_opt= 1,	Mid level convective cloud settings
cca_sh_knob= 0.50,	Shallow convective cloud settings: CCA scaling
cca_dp_knob= 0.50,	Deep convective cloud settings: CCA scaling
cca_md_knob= 0.50,	Mid convective cloud settings: CCA scaling
ccw_sh_knob= 0.00,	Shallow convective cloud settings: CCW Scaling
ccw_dp_knob= 0.00,	Deep convective cloud settings: CCW Scaling
ccw_md_knob= 0.00,	Mid convective cloud settings: CCW Scaling
cld_life_opt= 0,	Cloud decay lifetime option
ICONV_SHALLOW=1,	Shallow convection option ; 1 for mass-flux scheme
ICONV_CONGESTUS=0,	Congestus convection option; 0 for no congestus
ICONV_DEEP=1,	Deep convection option ; 1 for mass-flux scheme
ICONV_MID=1,	Mid level convection option ; Gregory-Rowntree mass flux scheme
ent_opt_dp=3,	Entrainment profile for deep convection
ent_dp_power=1.00,	Power for deep entrainment option 3 in 5A and 6A convection
ent_fac_dp=1.13,	Controls deep-level entrainment in 5A and 6A convection.
ent_opt_md=0,	Entrainment profile for mid-level convection
ent_fac_md=0.90,	Controls mid-level entrainment in 5A and 6A convection.
l_cv_conserve_check= .FALSE.,	Switches on energy conservation in convection
L_CCRad= .TRUE.,	Logical switch for Convective cloud for Radiation
L_3D_CCA=.TRUE.,	Convective cloud amount to use 3D field so that vertical structure are retained from timestep to timestep.
L_EMAN_DD=.FALSE.,	Logical switch to use Emanuel downdraught scheme
L_fix_udfactor=.FALSE.,	The updraught factor is only applied to grid-levels where convective precipitation is formed, unless this switch selected
Rad_cloud_decay_opt=2,	Time decay of convective clouds ; for cloud decay, but decays on the convective substep
Sh_pert_Opt=1,	Initial perturbation method for shallow cumulus
termconv=1,	Use termination condition for deep and mid-level convection
fixed_cld_life= 7200.00,	Fixed Convective Cloud lifetime
cca_min= 2.000e-02,	Convective Cloud Fraction threshold below this value ,cloud in that level is set to zero. Usually, equal to 0.02

4. Sensitivity of convection to the Entrainment rate and CAPE timescale

In the current report, we use both Global Atmosphere 3.0 (GA 3.0) and Global Atmosphere 4.0 (GA 4.0) configurations as the choice of dynamic core, atmospheric parameterisation and model options. The horizontal resolution of the model is set to N216 and with a model time step of 15 minutes. For testing both Entrainment rate (ε) and CAPE time scale (τ) in the NCUM, we have chosen an active monsoon spell in the model, with reference to observation rainfall data sets of IMD-NCMRWF satellite Merge data (Mitra et al., 2009). The active spell is identified as period during which the standardised rain anomaly is more than the +1.0 consecutively for three days or more over monsoon core zone (Rajeevan et.al. 2010). We have chosen July 24-27, 2013 as a peak phase of active spell. 72 hour model forecast is used in the study to avoid spin-up contamination. GA4.0 includes a revision to the convective entrainment and detrainment rates used in deep convective ascents in GA3.0. Upgradation of convection in GA4.0 includes the increase of entrainment in the lower level, which has shown some benefits in the tropical simulations as discussed in detail by Walter et al. (2014).

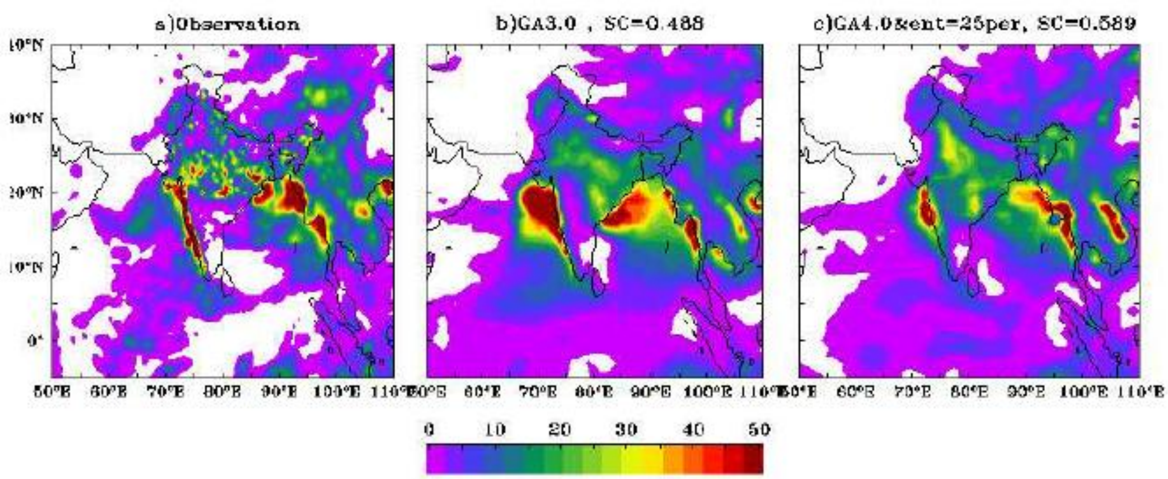

Fig. 8: Rain (mm) for the active spell from a) Observation and 72 hour forecast valid for the same date from b) GA3.0 c) GA4.0 with entrainment increased by 25%. The spatial correlations are indicated above the panels.

4.1 Entrainment rate

Increasing the entrainment rate has effect on intraseasonal variability, the changing mean state response observed in converging zone (Kim et al., 2011), and also improvement in precipitation bias of the Asian monsoon region (Bush et al., 2014). Increase of entrainment rate has positive impact on the representation of Madden Julian Oscillation through suppression of precipitation in drier environment and allowing moistening in the preconditioning stage of deep convective phase of MJO (Klingaman and Woolnough; 2014). In light of the importance of afore-mentioned studies using UM, we have looked at the sensitivity of entrainment rate for an active monsoon spell.

Equation for the entrainment profile available for the deep convection scheme is given in the Eq 26. Range of entrainment option available in the subroutine LAYER_CN. Entrainment factor f_{dp} used here for the model setting of N216 (0.83°× 0.55°horizontal resolution, L70 vertical resolution for the current experiments are:

GA3.0 f_{dp}= 0.9

GA4.0 f_{dp}= 1.13 (25% increase)

GA4.0 f_{dp}= 1.36 (50% increase)

Total rainfall (convective and large scale) at T+72 hour forecast for GA3.0 and GA4.0 (+25%) with corresponding observed rainfall for the active spell is indicated in the Fig. 8. Spatial correlation shows higher value in the GA4.0, it includes the impact from both change in dynamic core and enhanced entrainment rate. Again using GA4.0 (+50%) slight improvement in the spatial correlation value (≈0.591) is seen (figure not shown). Enhanced rainfall in the North Bay of Bengal may be associated with strengthened westerly wind and significant moisture availability in GA4.0. In addition, rainfall spread in the Western Arabian Sea is also a noticeable feature.

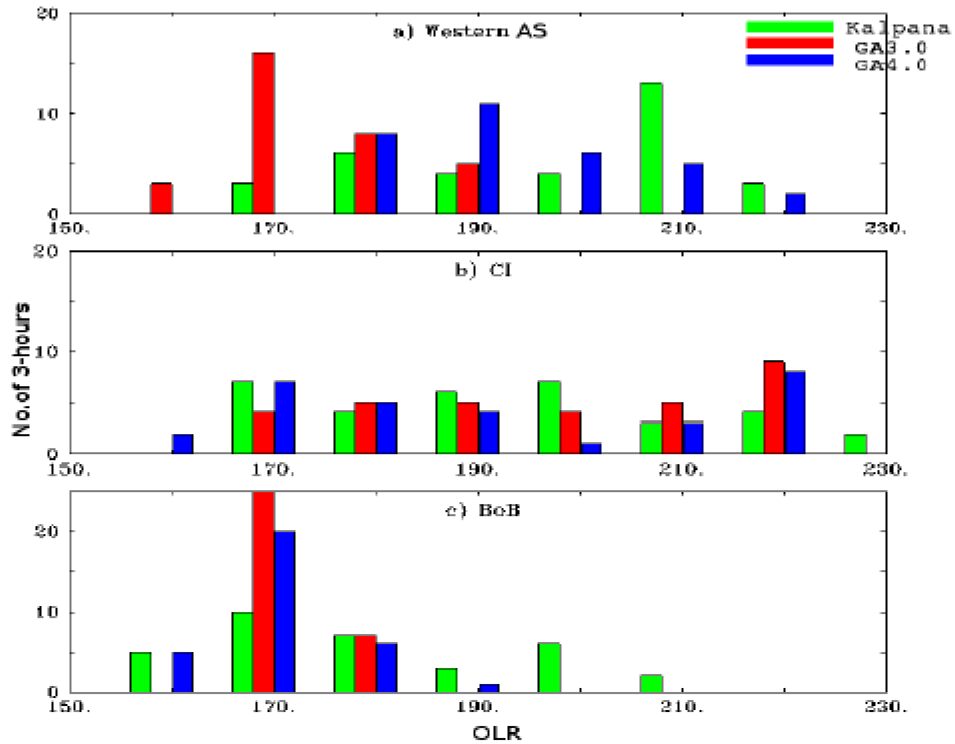

Fig. 9: Histogram of OLR (wm^{-2}) for monsoon spell by counting number of 3-hourly average over a) Western Arabian sea, b) Central India (CI) and c) Bay of Bengal (BoB) using 3-hourly Kalpana INSAT (green) and 72 hour forecast from GA3.0 (red) & GA4.0 (blue)

Lower value of OLR is a manifestation of deeper cloud (with colder cloud top), whereas higher OLR represent shallower cloud (with warmer cloud top). Frequency of occurrence of OLR over Western Arabian Sea in GA3.0 forecast shows a tendency towards more deeper type cloud than observations and GA4.0 forecast, whereas OLR histogram in GA4.0 with enhanced entrainment flux shows a better match with observations (Fig.9). Over the Northern Bay heavy rainfall area the deeper type of clouds are over-predicted in GA4.0 (50%) compared to GA4.0 (25%) (Fig.10).

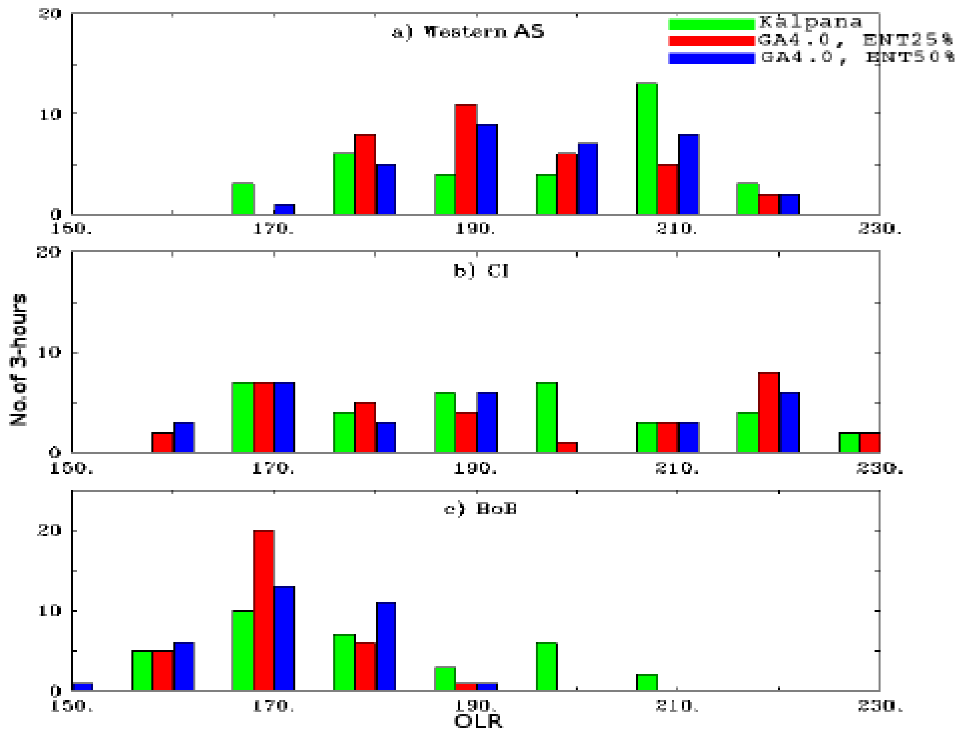

Fig. 10: Same as 9 but for Kalpana INSAT (green), GA4.0 & 25% (red), GA4.0 & 50% (blue).

4.2 CAPE timescale

Convection occurs when there is CAPE, which is subsequently removed by convection at an exponential rate using the adjustment time scale. Amount by which convection heats and dries the atmosphere per time step is effectively determined by the CAPE adjustment timescale (τ_{CAPE}). τ_{CAPE} determines the intensity and duration of convection at a given CAPE. With small τ_{CAPE}, the convection is short lived but intensity is high, on the other hand with larger τ_{CAPE}, convection is long lived, but of low intensity. Variation of CAPE closure in the convective scheme depends on the specification of τ_{CAPE} in the subroutine DEEP_CONV_5A. Different options of τ_{CAPE} are:

(a) RH-based CAPE closure

(b) Fixed cape timescale (τ_{CAPE})

(c) w based cape closure

(d) RH and w based CAPE option

We used first three options for testing the sensitivity of τ_{CAPE} in the GA4.0 setting. Fig. 11 shows the convective rain rate predicted in a peak active monsoon spell for different values of τ_{CAPE}. Spatially, overall structure looks similar, but the region with heavy rainfall such as Western Ghats, Central India and North Bay shows differences in the rain rate. Among the convective rain, significant changes are predicted in the deep convective rate (figure not shown). Rain rate intensity is reduced over the high rainfall region such as Western Ghats, Central India and BoB, when the τ_{CAPE} fixed to 1.5 hour, by limiting the deep convection. By selecting the long τ_{CAPE}, there may be adjusted enhancement in the large-scale rain.

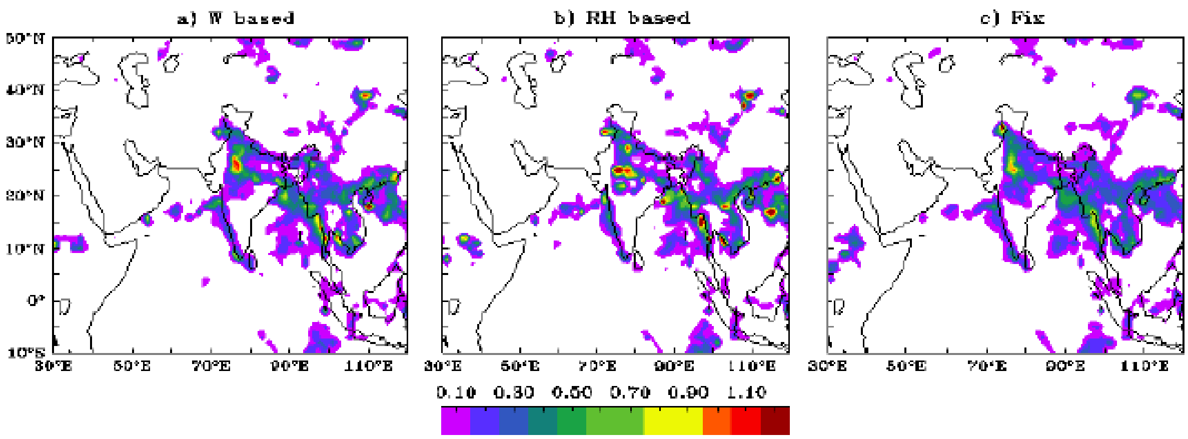

Fig. 11: Convective rain rate ($\times 10^3$mm/s) valid on 26 July 2013 03Z based on τ_{CAPE} options a) w b) RH c) Fix τ_{CAPE}=1.5hour for t+72 hour forecast

5. Concluding Remarks

NCUM convection scheme was introduced in the current report in a simplified manner as possible. For more particular details regarding the implementation in NCUM, the reader may refer to the detailed documentations and code. The sensitivity experiments were carried out to compare GA3.0 and GA4.0 configurations and are of preliminary in nature. More sensitivity studies will be carried out with the latest version of NCUM in future.

A Notation

A.1 Subscripts

b	cloud base
cld	whole cloud value
k	Model level number
surf	surface value

A.2 Superscripts

c	Convective plume value
e	Environmental value
p	Parcel value
N	for air undergoing mixing detrainment
R	for air undergoing forced detrainment

A.3 Variables

δ	Forced detrainment
ϵ	Entrainment
θ	Potential temperature
θ_l	liquid water potential temperature
Π	Exner function
ρ	Density
μ	Mixing detrainment
e	Vapour pressure (Pa)
g	Acceleration due to gravity (m/s^2)
l^p	Rain water(kg/kg)
M	Mass flux(Pa/s)
p	Pressure (Pa)
PREC	Precipitation
q	Water vapour (kg/kg)
qT	Total water (kg/kg)

qcl	Cloud liquid water (kg/kg)
qcf	Cloud ice water (kg/kg)
ql	Cloud condensate in convective plume (kg/kg)
qsat	Saturated water vapour at temperature, T and pressure, p
RH	Relative humidity
R	Gas constant for dry air
u	U component of wind (m/s)
v	V component of wind (m/s)
w	Vertical component of wind (m/s)
w_*	sub-cloud velocity scale (m/s)
w^*	Convective velocity scale (m/s)
z	height above surface (m)

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