Méso-NH shallow convection scheme for AROME

CNRM/GMME/Méso-NH

23 novembre 2005

Contents

A shallow convection scheme for Arome ?

current situation

turbulence | **shallow convection** | **dynamics + explicit microphysics**

sub–grid | **explicit** | **scales**

eddy diffusive fluxes | **non precipitating convection** | **deep precipitating convection**

generalised turbulence | **dynamics + explicit microphysics**

« ideal » futur situation

The convection parameterization of Méso-NH has been developped on the basis of existing frameworks, essentially the rather general framework proposed by Kain and Fritsch (1993). The parameterization is intended to provide an efficient representation of atmospheric shallow and deep convection for both mesoscale and global applications.

A mass-flux convection scheme for regional and global models
Bechtold et al, 2001, QJRMS

In practice in Méso-NH, shallow and deep convection are treated by two « twin » schemes. We focus on the shallow scheme only (no microphysics of precipitation, no downdraft).

The mass flux equation for shallow convection

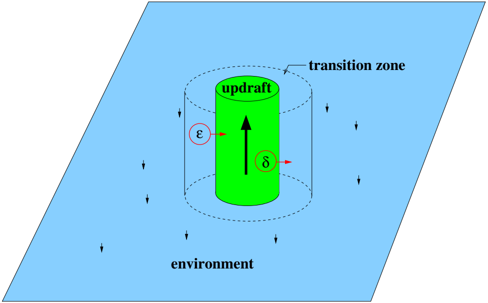
The effect of a small convective cloud population on its environment can be approximated using a mass flux equation of the form (see e.g. Arakawa and Schubert (1974), Gregory and Miller (1989), Betts (1997) for various derivations)

$$
\begin{aligned}
\left.\frac{\partial \overline{\Psi}}{\partial t}\right|_{\mathrm{conv}} &= -\frac{1}{\overline{\rho}}\frac{\partial(\overline{\rho w' \Psi'})}{\partial z} \\
&\approx -\frac{1}{\overline{\rho} A}\frac{\partial}{\partial z}\left[M^u(\Psi^u - \overline{\Psi}) + \tilde{M}(\tilde{\Psi} - \overline{\Psi})\right] \\
&\approx -\frac{1}{\overline{\rho} A}\frac{\partial}{\partial z}\left[M^u \Psi^u - M^u \overline{\Psi}\right]
\end{aligned}
$$

Mass flux vertical profile

The mass exchange of the cloud ensemble with its environment is described with an entrainment rate ϵ and detrainment rate δ :

$$\frac{\partial}{\partial z}(M^u \Psi^u) = \epsilon^u \overline{\Psi} - \delta^u \Psi^u$$



Final mass flux equation

$$\left.\frac{\partial\overline{\Psi}}{\partial t}\right|_{\text{conv}} = \frac{1}{\overline{\rho}A}\left[\frac{\partial}{\partial z}(M^u\overline{\Psi}) - \epsilon^u\overline{\Psi} + \delta^u\Psi^u\right]$$

In the Méso-NH paramatrization of shallow convection, the mass fluxes equations are written for the enthalpy or « liquid water static energy » h_{il} and the total water mixing ratio r_w

$$\begin{aligned} h_{il} &= C_{ph}T - L_v r_c - L_s r_i + (1 + r_w)gz \\ r_w &= r_v + r_c + r_i \end{aligned}$$

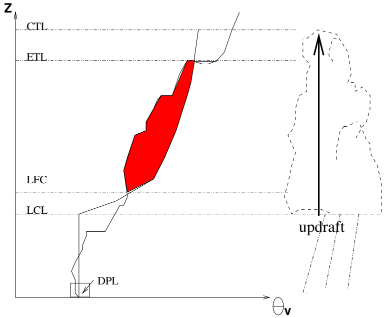
The cloud model

The cloud model is designed to represent shallow convective clouds (no precipitation, no downdraft) that are characterized by their respective cloud radius.

The trigger condition

The «test » parcel

- Construction of a caracterictic mixed parcel (40 hPa deep, mean potential temperature $\overline{\theta}^{mix}$, mean vapor mixing ratio $\overline{r}_v^{mix}$)
- this mixed air parcel is lifted without entrainment to its LCL.



Instability condition

The air parcel is unstable with respect to moist convection if at the LCL

$$\overline{\theta}_v^{mix} - \overline{\theta}_v + \underbrace{\Delta T/\Pi}_{-CIN} > 0,$$

For shallow convection the temperature increment ΔT is set to 0.2 K.

Detection of shallow convection

The shallow convection scheme is activated only if the mixed parcel is unstable and lifted by buoyancy of at least 500 m and if the avaible CAPE is not insignificant.

After the trigger condition determination, the computation are done only for « shallow convective columns ».

The updraft computation

- Starting from the LCL, the mass flux at each level is computed for each convective time step
- At each level, the thermodynamic characteristics of the updraft are computed assuming conservation of the enthalpy h_{il} and of the total water mixing ratio r_w by the undiluted part of the updraft
- Entrainment and detrainment are the only source/sink of the « conservative » variables of the updraft

The initial updraft mass flux is set to a unit value of

$$M^u(\mathrm{LCL}) = \overline{\rho}\, w_{\mathrm{LCL}}\, S$$

- w_{LCL} is an «arbitrary» vertical velocity of 1 m s^{-1}
- S is the updraft area defined as a fraction of the total area A

h_{il}^u and r_w^u are computed at the LCL by :

$$\begin{aligned} h_{il}^u &= C_{ph} T(\mathrm{LCL}) + (1 + r_v^{mix}) g z(\mathrm{LCL}) \\ r_w^u &= r_v^{mix} \end{aligned}$$

The mass flux variation between level $k+1$ and level k is due to the mean entrainment and detrainment of cloudy air in the layer between these two levels :

$$M^{u^{k+1}} - M^{u^{k}} = \epsilon^u - \delta^u$$

The flux of h_{il}^u and r_w^u also change between two levels because of the mean entrainment and detrainment :

$$\begin{aligned} \Delta(M^u h_{il}^u) &= \epsilon^u \overline{h}_{il} - \delta^u h_{il}^u \\ \Delta(M^u r_w^u) &= \epsilon^u \overline{r}_w - \delta^u r_w^u \end{aligned}$$

Modelisation of entrainment and detrainment

The performance of a plume cloud model critically depends on the specification of updraft entrainment/detrainment which are functions of the cloud radius.

In Méso-NH, we adopt the mixing formalism proposed by Kain and Fritsch (1990, KF90).

Estimation of entrainment and detrainment rate

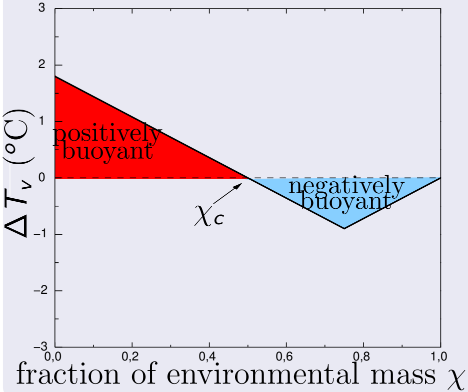
The exchanges between the cloudy updraft and its environment are supposed to occur near the periphery of the updraft. In this region, the air is a mixture of clear and cloudy air of various proportion.

Total rate of mass entering the transition zone

$$\delta M_t = \delta M_e + \delta M_u = M^u (c_{\mathrm{etr}} \Delta z / R_0)$$

with $c_{etr} = 0.2$.

KF90 supposed that the mixed parcels which are positively buoyant follow the cloudy updraft (entrain) and the ones which are negatively buoyant detrain.



χ_c is the fractional amount of environmental mass that just yields a neutrally buoyant mixture.

Hypothesis

KF90 supposed that

- the turbulent mixing processes show some propensity to mix updraft and environmental air masses in equal proportions
- the relative frequency distribution of parcel mixtures may be reasonably estimated by a gaussian-type distribution $f(\chi)$

Let δM_t be the total rate at which mass enters the transition region between clear and cloudy air :

$$\delta M_t = \delta M_e + \delta M_u = \delta M_t \int_0^1 f(\chi) d\chi$$

- δM_e is the mass rate coming from the environment to the transition region
- δM_u is the mass rate coming from the cloudy updraft to the transition region

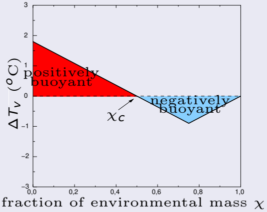
$$\delta M_e = \delta M_t \int_0^1 \chi f(\chi) d\chi \qquad \delta M_u = \delta M_t \int_0^1 (1-\chi) f(\chi) d\chi$$

By definition, ϵ^u is the proportion of δM_e which is positively buoyant, and so will be entrain in the updraft
Symetrically, δ^u is the proportion of δM_u which is negatively buoyant, and so will be detrain from the updraft

$$\epsilon^u = \delta M_t \int_0^{\chi_c} \chi f(\chi) d\chi \qquad \delta^u = \delta M_t \int_{\chi_c}^{1} (1-\chi) f(\chi) d\chi$$

Computation of χ_c in Méso-NH

The only missing parameter necessary to compute the updraft mass flux is then χ_c.



$$\theta_v^{mix}(\chi) - \overline{\theta}_v = a\chi + b$$

- in $\chi = 0$, $\theta_v^{mix} = \theta_v^u \longrightarrow b = \theta_v^u - \overline{\theta}_v$
- for a choosen χ_s, $\theta_v^{mix}(\chi_s) - \overline{\theta}_v = a\chi_s + b$, then

$$a = \frac{\theta_v^{mix}(\chi_s) - \overline{\theta}_v - b}{\chi_s} = -\frac{\theta_v^u - \theta_v^{mix}(\chi_s)}{\chi_s}$$

Computation of χ_c in Méso-NH

For $\chi = \chi_c$, $a\chi_c + b = 0$, then

$$\chi_c = \frac{\theta_v^u - \overline{\theta}_v}{\theta_v^u - \theta_v^{mix}} \chi_s$$

In Méso-NH, we use $\chi_s = 0.1$ and $\theta_v^{mix}(\chi_s)$ is computed from

$$\begin{aligned} h_{il}^{mix} &= \chi_s \overline{h}_{il} + (1 - \chi_s) h_{il}^u \\ r_w^{mix} &= \chi \overline{r}_w + (1 - \chi_s) r_w^u \end{aligned}$$

The closure (scaling) hypothesis

Finally, a closure assumption is needed to control the intensity of convection (recall : we used an arbitrary w_{LCL} for the cloud base updraft mass flux).

The reference code of Méso-NH adopt a Fritsch Chappell type closure which is based on the assumption that all convective available potential energy (CAPE) in a grid element is removed within an adjustment period τ.

For shallow convection an adjustment time τ of 3 h is used.

The idea is to compute iteratively the mean state of the atmospheric column which should be reached after a convective ajustment due to the shallow convective processes.

The esssential point of the present adjustment procedure is that only the environmental values $\Psi = h_{il}, r_w, r_c, r_i$ are updated and the mass fluxes are adjusted in the closure adjustment procedure, but no updraft or downdraft computations are repeated so that the updraft and downdraft values of the thermodynamic variables keep unchanged.

In an iterative procedure, the conservative variables at step $n+1$ are computed from the conservative variables at step n :

$$\overline{\Psi}^{(n+1)} = \overline{\Psi}^{(0)} + (\tau/m_A)\left[-\Delta(\tilde{M}^{(n)}\overline{\Psi}^{(0)}) - \epsilon^{u(n)}\overline{\Psi}^{(0)} + \delta^{u(n)}\Psi^u\right]$$

$$\tilde{M} = -M^u = \overline{\rho}\tilde{w}A; \quad \tilde{w} = \int(\partial\tilde{w}/\partial z)dz; \quad \left(\frac{\partial\tilde{w}}{\partial z}\right) = \frac{\epsilon^u - \delta^u}{m_A}$$

A new value of CAPE is computed at each iteration by using undilute parcel ascent

$$\mathrm{CAPE}^{(\mathrm{n+1})} = \int_{\mathrm{LCL}^{(\mathrm{n+1})}}^{\mathrm{ETL}} g \left[\frac{\overline{\theta}_e^{(n+1)}(\mathrm{DPL})}{\overline{\theta}_{es}^{(n+1)}} - 1 \right] dz,$$

Scaling of the convective mixing

At all model levels the updraft mass fluxes and the entrainment/detrainment fluxes are multiplied by the adjustment factor

$$F_{adj}^{(n+1)} = F_{adj}^{(n)} \frac{\mathrm{CAPE}^{(0)}}{\mathrm{CAPE}^{(0)} - \mathrm{CAPE}^{(n+1)}},$$

where $\mathrm{CAPE}^{(0)}$ is the initial value of CAPE and $F_{adj}^{(0)} = 1$.

The above described procedure is repeated until $\mathrm{CAPE}^{(n+1)} < 0.1\ \mathrm{CAPE}^{(0)}$.

Convective tendencies

At the end of the adjustment procedure the final convective tendencies are simply evaluated as

$$\left.\frac{\partial\overline{\Psi}}{\partial t}\right|_{\mathrm{conv}} = (\overline{\Psi}^{(n)} - \overline{\Psi}^{(0)})/\tau,$$

where Ψ now stands for either θ, r_v, r_c, r_i.

A new closure

Convective vertical velocity w^*

Instead of the CAPE iterative closure, we directly compute the cloud bottom mass flux with a convective vertical velocity proportionnal to the shallow convective activity in the boundary layer (for exemple, Grant et al, 1999)

$$M^u(\mathrm{LCL}) = \overline{\rho}\, w^*\, S$$

with w^* computed in the turbulence scheme as a function of the surface fluxes and the depth of the boundary layer.