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Parametrização Cumulus

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V WorkEta
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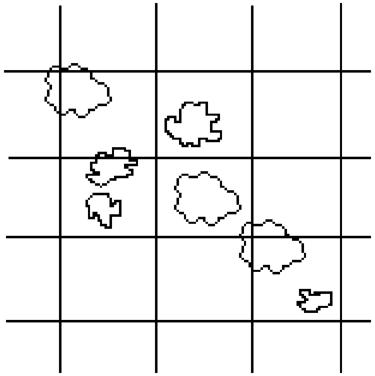
The need for cumulus parameterization

*Convective clouds can organise in clusters and show their collective effects in model grid-box.



*Large-scale destabilizes the environment >>> cumulus parameterization scheme acts to remove the convective instability

*The upward fluxes of heat, moisture and momentum in the cloud can be seen by means of an area average over the equations of mass continuity and heat energy.

*Up to which resolution should the parameterization act in a model? No clear agreement.

General assumption:

- This area is assumed to be large enough to contain the cluster or the ensemble of clouds, but still regarded as a small fraction of the large scale model, the grid size.

AKIO ARAKAWA AND WAYNE HOWARD SCHUBERT

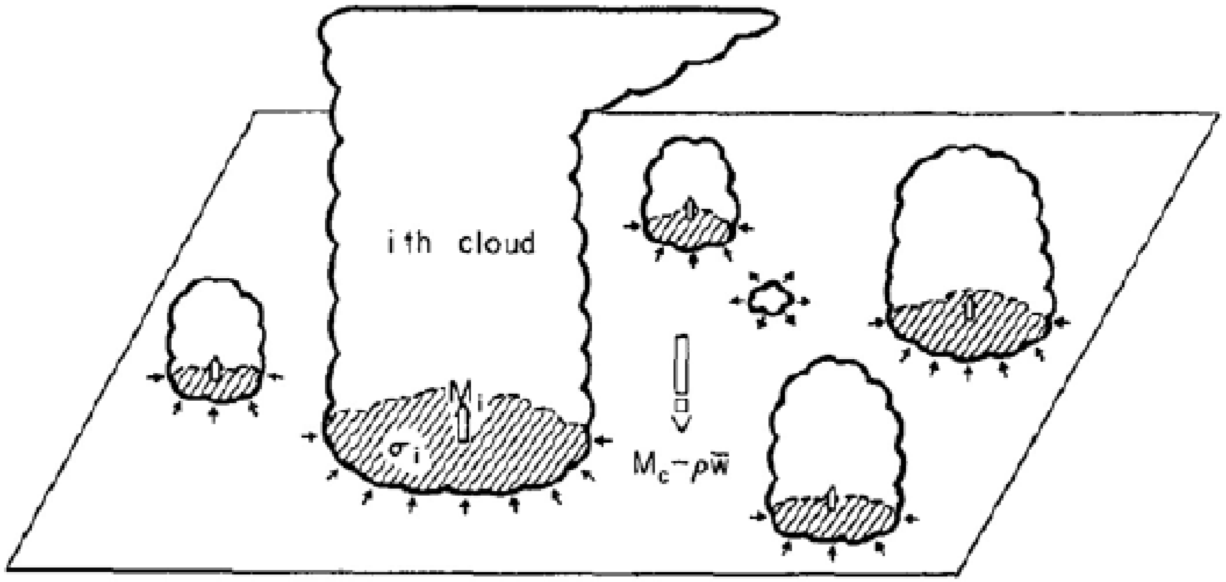


FIG. 1. A unit horizontal area at some level between cloud base and the highest cloud top. The taller clouds are shown penetrating this level and entraining environmental air. A cloud which has lost buoyancy is shown detraining cloud air into the environment.

Médias

•Os movimentos atmosféricos existem em várias escalas espaciais. Em um modelo numérico há processos resolvidos pela grade do modelo e outros processos “sub-grade”.

•Há necessidade de descrever os processos **resolvidos** pelo sistema de observações e aqueles **não-resolvidos** e designados por perturbação (“eddy”).

•Para identificar as propriedades estatísticas de um sistema, utiliza-se médias.

•1 **Média no volume da grade (Grid-volume averaging):**

$$\phi = \overline{\phi} + \phi'$$

Variável composta por uma média resolvida e uma perturbação

'–' se refere à média na grade do modelo

ϕ' perturbação subgrade

$$\overline{\phi} = \int_t^{t+\Delta t} \int_x^{x+\Delta x} \int_y^{y+\Delta y} \int_z^{z+\Delta z} \phi\, dz\,dy\,dx\,dt / (\Delta t \Delta x \Delta y \Delta z)$$

Δx, Δy, Δz são as dimensões da grade do modelo e Δt o passo de tempo.

$$\overline{\overline{\phi}} = \overline{\phi}$$

$$\overline{\phi'} = 0$$

$$\overline{\frac{\partial u}{\partial t}} = \frac{\partial \overline{u}}{\partial t}$$

$$\overline{\frac{\partial u}{\partial x}} = \frac{\partial \overline{u}}{\partial x}$$

$$\overline{\phi' \overline{u}} = 0$$

$$\overline{\phi' w'} \neq 0$$

$$\overline{T' w'} \neq 0$$

A média do produto da correlação das perturbações resulta em valor diferente de zero!!

Equations of heat, moisture and continuity

$$\frac{\partial\bar{\theta}}{\partial t}+\overline{\nabla.\theta\mathbf{v}}+\frac{\partial\bar{\theta}\bar{w}}{\partial z}=\frac{Q_R}{c_p\bar{\pi}}+\frac{L}{c_p\bar{\pi}}(c-e)-\frac{1}{\bar{\rho}}\frac{\partial}{\partial z}\bar{\rho}\overline{\theta`w`}$$

$$\frac{\partial\bar{q}}{\partial t}+\overline{\nabla.q\mathbf{v}}+\frac{\partial\overline{qw}}{\partial z}=-(c-e)-\frac{1}{\bar{\rho}}\frac{\partial}{\partial z}\bar{\rho}\overline{q`w`}$$

$$\overline{\nabla.\mathbf{v}}+\frac{\partial\bar{w}}{\partial z}=0$$

These subgrid terms need to be parameterized because their effects contribute to model grid scale

The vertical eddy fluxes are due mainly to the cumulus convection and turbulent motions in the boundary layer.

Deep convection cloud structure

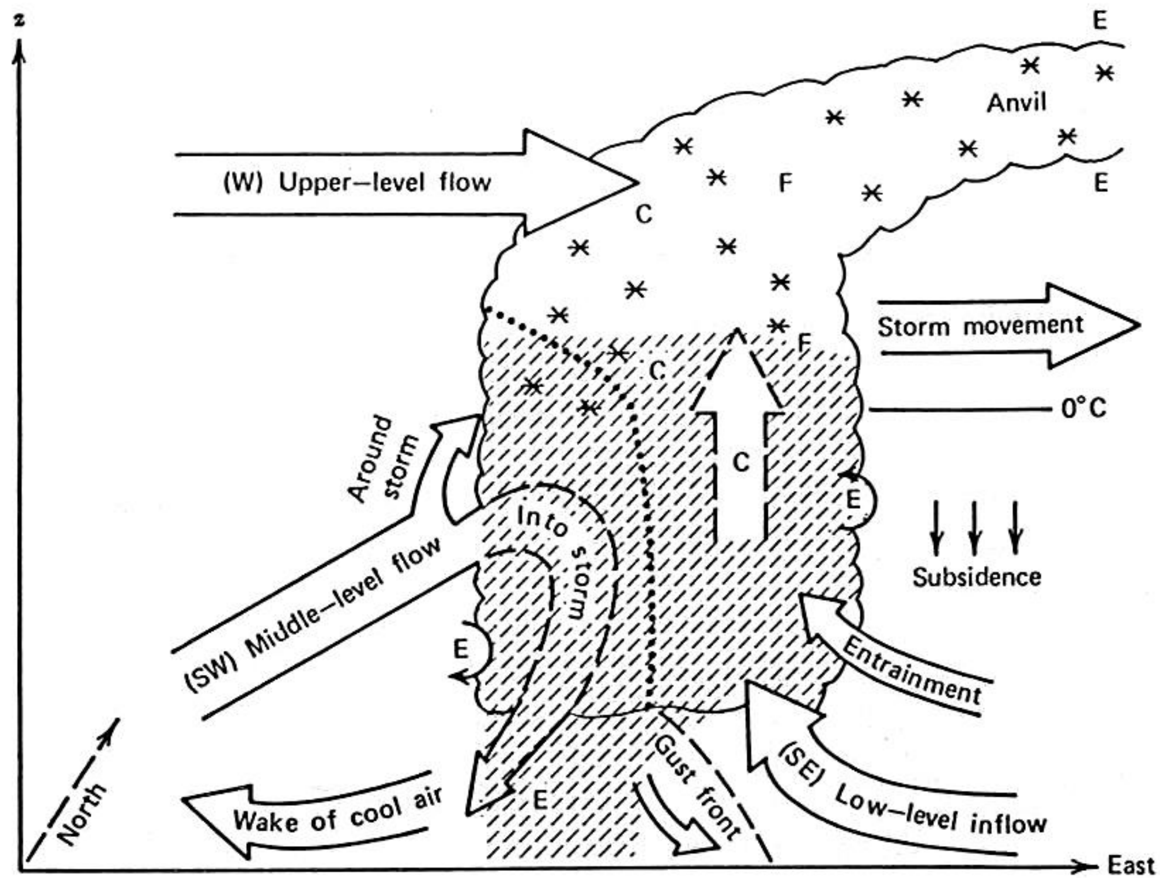


Figure 9-1 Schematic diagram of a mature thunderstorm. (*After Anthes, 1978.*) *C* refers to condensation; *E*, evaporation; and *F*, freezing. Letters in arrows refer to wind direction. (From R. A. Anthes, H. Orville, and D. Rayword. Chapter XIX of *The Thunderstorm: A Social Scientific and Technological Documentary*. E. Kessler, ed. University of Oklahoma Press, 1978.)

The contributions from the cumulus convection to the large scale moisture and heat are:

$$\left(\frac{\partial\bar{\theta}}{\partial t}\right)_{cv} = -\frac{1}{\bar{\rho}}\frac{\partial}{\partial z}\bar{\rho}\overline{\theta`w`} + \frac{L}{c_p\bar{\pi}}(c-e)$$

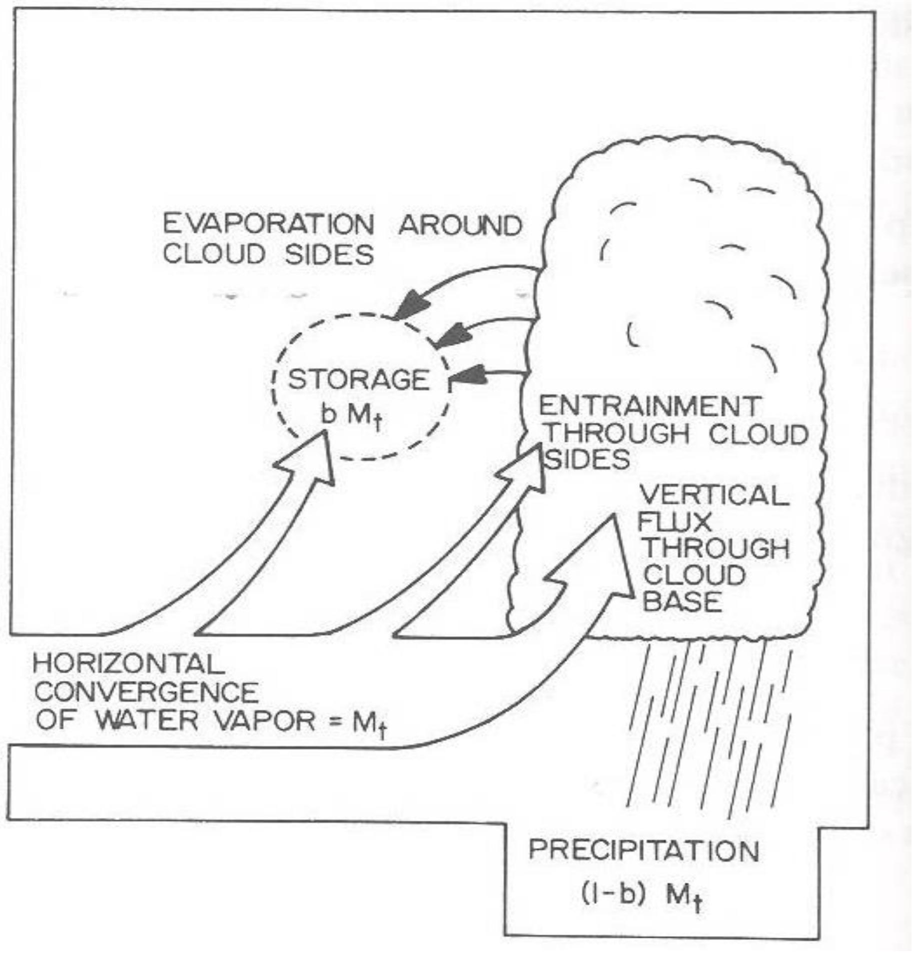
$$\left(\frac{\partial\bar{q}}{\partial t}\right)_{cv} = -\frac{1}{\bar{\rho}}\frac{\partial}{\partial z}\bar{\rho}\overline{q`w`} - (c-e)$$

- Cumulus parameterization schemes are designed to represent the convective fluxes and the associated condensation or evaporation.
- Model temperature and moisture profiles are altered by scheme due to the parameterized convective fluxes and precipitation.

Types of convective scheme:

Kuo:	Kuo (1974)
Adjustment:	Betts-Miller (1986), Janjic (1994)
Mass-flux:	Arakawa-Schubert (1974), Fritsch-Chappell (1980), Tiedtke (1989), Grell, Kain-Fritsch (1993)

The Kuo Convective Parameterization Scheme



The Kuo Method

In an attempt to represent the convective cloud, Kuo (1965; 1974) introduced the parcel concept into parameterization schemes. The temperature and moisture profile inside the cloud follows the moist adiabat. The cloud is assumed to occupy a fraction a of the grid area, so the heat and moisture changes are determined by the difference between cloud air and the large scale environment air, such as:

$$\Delta T = a(T_c - \bar{T})$$
$$\Delta q = a(q_c - \bar{q})$$

Where suffix 'c' means in cloud from now on. The estimate of fraction a is the essence of this scheme. It is calculated in terms of the production of cloudy air over the time interval Δt. This cloudy air is assumed to be produced from a net moisture convergence into the column, and from moisture sources from the underlying surface, E_s, so the total energy supplied by the large scale flow to form the cloud is:

$$C_t = -\frac{L}{g}\int_0^{p_s} \nabla.(q\mathrm{v})dp + LE_s$$

C_t refers to the total moisture source, L is the latent heat, and p_s is the surface pressure. The surplus of total energy computed by the cloud profile over the environment air is given by:

$$W = \frac{1}{g}\int_0^{p_s}\left[c_p(T_c - \bar{T}) + L(q_c - \bar{q})\right] dp$$

the fractional cloud coverage is given by the ratio, $a = C_t/W$.

The scheme is activated only in the presence of moisture convergence and conditional instability. In this formulation, the clouds dissolve immediately through horizontal mixing, and no moisture detrainment occurs above the cloud top. This rather simple formulation, however, does not describe the observations well. Some further improvements have been carried out on this scheme, most of them were involved in obtaining an optimal factor b. This factor is a fraction of the large scale converged moisture, stored in the column which increases humidity locally, thus it is given as $a = (1-b)C_t/W$, where the factor $(1-b)$ falls out as rain or is carried away by advection. Anthes (1977) included a cloud model into this scheme and obtained better rainfall estimates. This scheme is largely used in forecast and climate models.

Betts-Miller-Janjic scheme

Betts-Miller-Janjic

- The Betts-Miller scheme (Betts and Miller, 1986) uses reference profiles of T and q to relax the model profiles in convective unstable conditions. Profiles derived from campaigns GATE, VIMEX, etc

- The reference T and q profiles are based on observational studies of convective equilibrium in the tropics.

- Treats deep and shallow convection.

- (Modification by Janjic, 1994)

1. Determine type of cloud

- Parcel lift: determine cloud base and cloud top
- Check: cloud depth > 0.2*Psfc, then deep convection, else shallow convection.

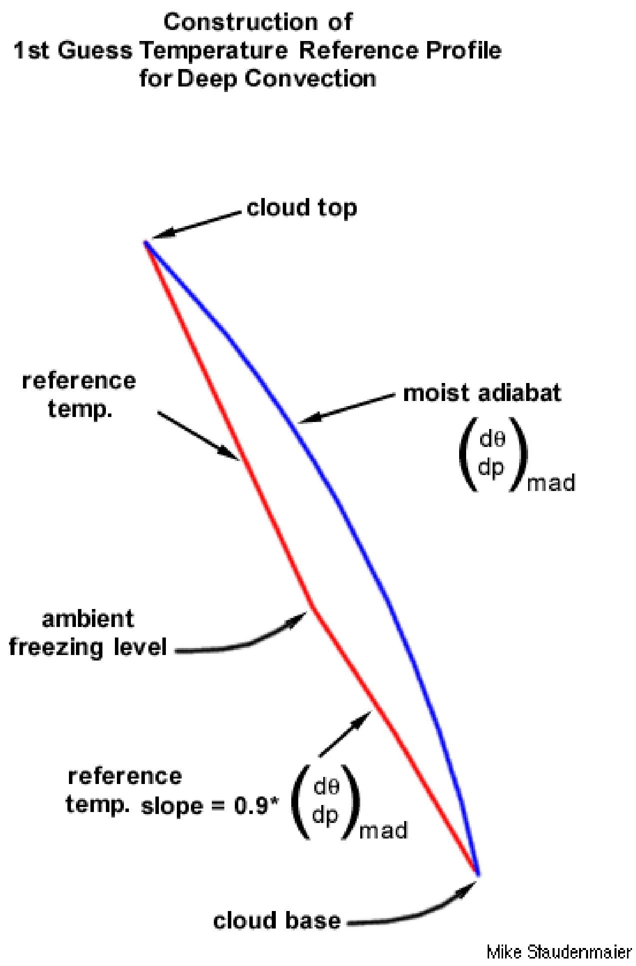
2. Determine reference profiles

$$\int_{base}^{top} (c_p \Delta T - L \Delta q) = 0$$

Make sure enthalpy is conserved

Deep Convection

Temperature Reference Profile



To draw the temperature profile 3 levels are important:

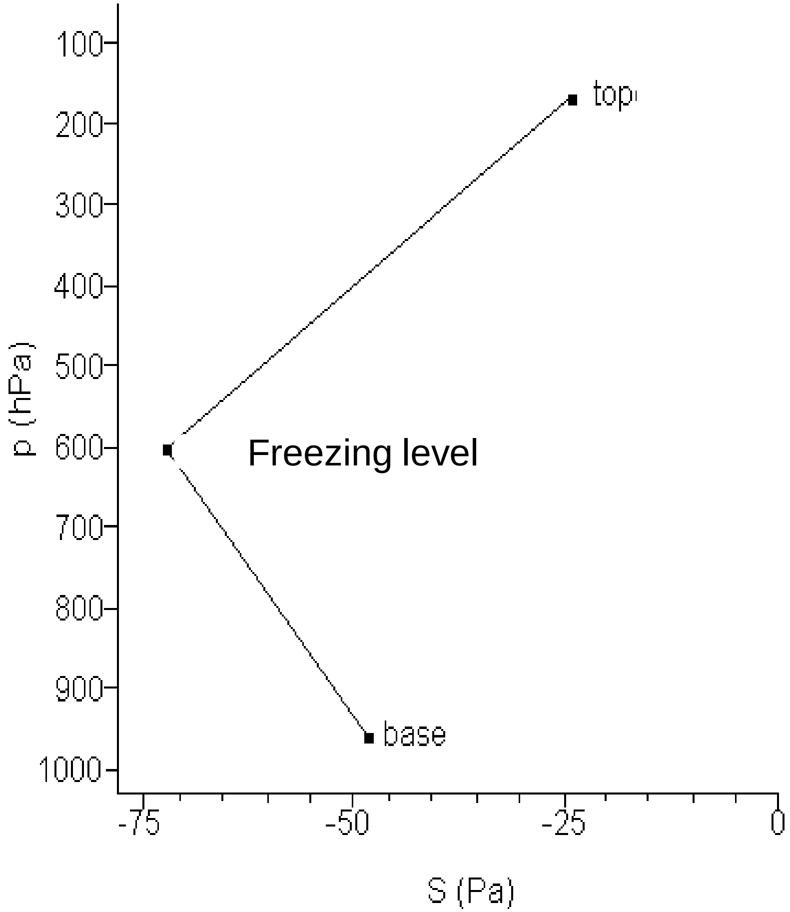
- cloud base
- freezing level
- cloud top

Some instability is left in the lower part of the cloud.

Profile is linearly interpolated from the freezing level to the cloud top

Deep Convection

Moisture Reference Profile



DSP= p_{sat} - p

Deficit saturation pressure (DSP) is defined at the 3 levels.

- cloud base: **DSPb**
- freezing level: **DSP0**
- cloud top: **DSPt**

Values are linearly interpolated between the levels

$$T_{new} = T_{old} + \frac{\Delta t_{cnv}}{\tau}\left[T_{ref} - T\right]$$

$$q_{new} = q_{old} + \frac{\Delta t_{cnv}}{\tau}\left[q_{ref} - q\right]$$

$$\Delta t_{cnv} = 4 * \Delta t$$

$$\tau = 3000s$$

$$P = \frac{1}{\rho_w g} \frac{\Delta t_{cnv}}{\tau} \sum_{base}^{top} (q_{ref} - q)(p_s - p_t)$$

1. Cloud Eficiency

(Janjic,1994)

- Eficiency related to the precipitation production
- Proportional to the cloud column "entropy" change, per unit precipitation produced.
- Eficiency varies from 0.2 to 1.0
- Modifies reference profiles
- Modifies relaxation time

$$\tau' = \frac{\tau}{F(E)}$$

$$F(E) \ is \ linear$$

$$0.7 \le F \le 1.0 \ \ for \ \ 0.2 \le E \le 1.0$$

Thus,

$$\textit{larger } E \Rightarrow \textit{less mature system}$$
$$\textit{smaller } E \Rightarrow \textit{more mature system}$$

USE E TO MODERATE HEAVY RAIN IN LONG–LIVED MATURE SYSTEMS

(A) Modify the humidity reference profile

(B) Modify the relaxation time τ

$$T_{new} = T_{old} + \frac{\Delta t_{cnv}}{\tau}\left[T_{ref} - T_{old}\right]F(E)$$

$$q_{new} = q_{old} + \frac{\Delta t_{cnv}}{\tau}\left[q_{ref} - q_{old}\right]F(E)$$

$$OR$$

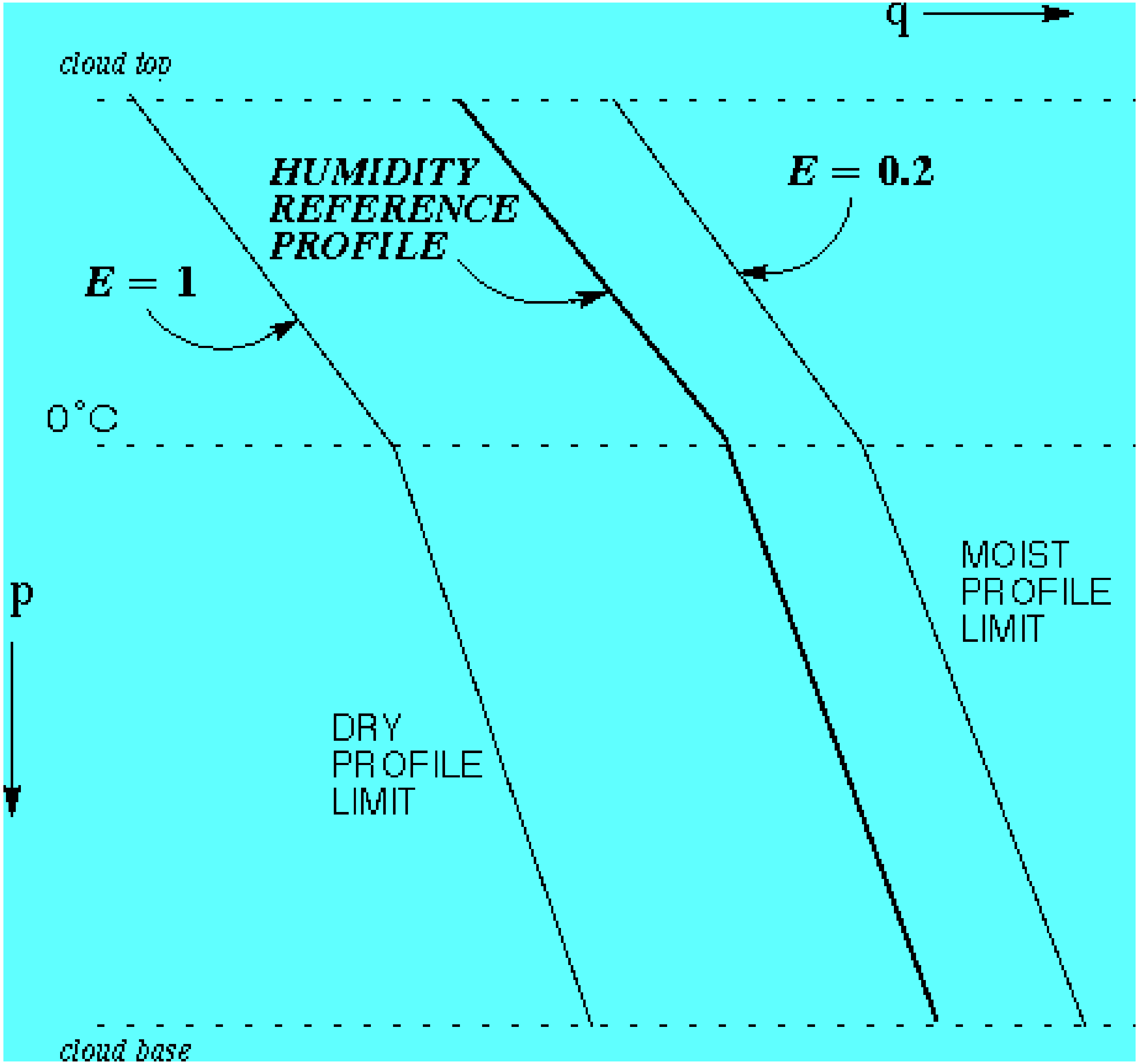
$$T_{new} = T_{old} + \frac{\Delta t_{cnv}}{\tau'}\left[T_{ref} - T_{old}\right]$$

$$q_{new} = q_{old} + \frac{\Delta t_{cnv}}{\tau'}\left[q_{ref} - q_{old}\right]$$

where

$$\tau' = \frac{\tau}{F(E)}$$

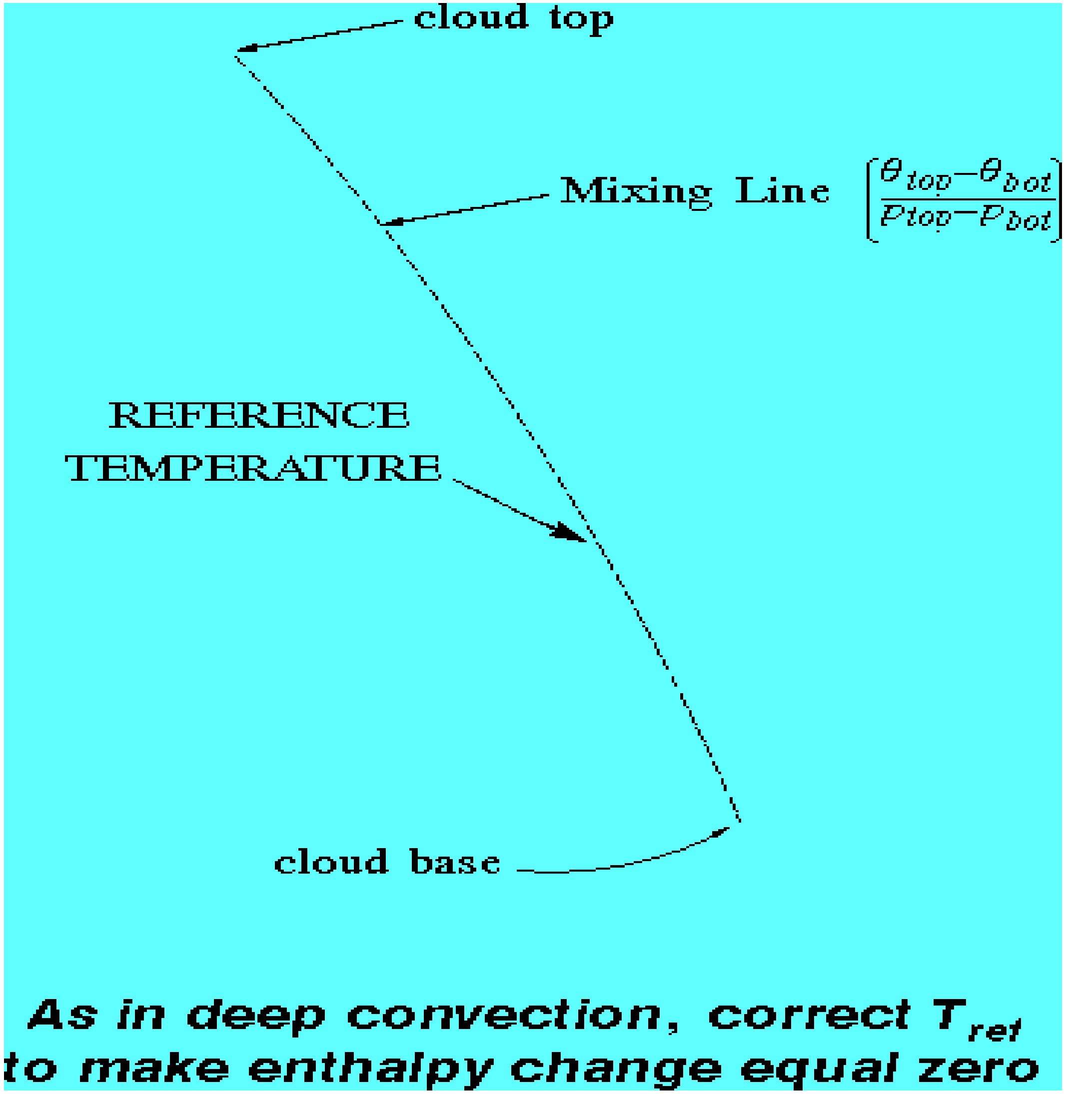
fss



q
cloud top
HUMIDITY
REFERENCE
PROFILE
E = 0.2
E = 1
0°C
MOIST
PROFILE
LIMIT
p
DRY
PROFILE
LIMIT
cloud base

Shallow Convection

Temperature Reference Profile



- *Applied to points where cloud depth is larger than 10hPa and smaller than 0.2Psfc*
- *At least two layers*
- swap points:
  - precipitation < 0
  - entropy change < 0

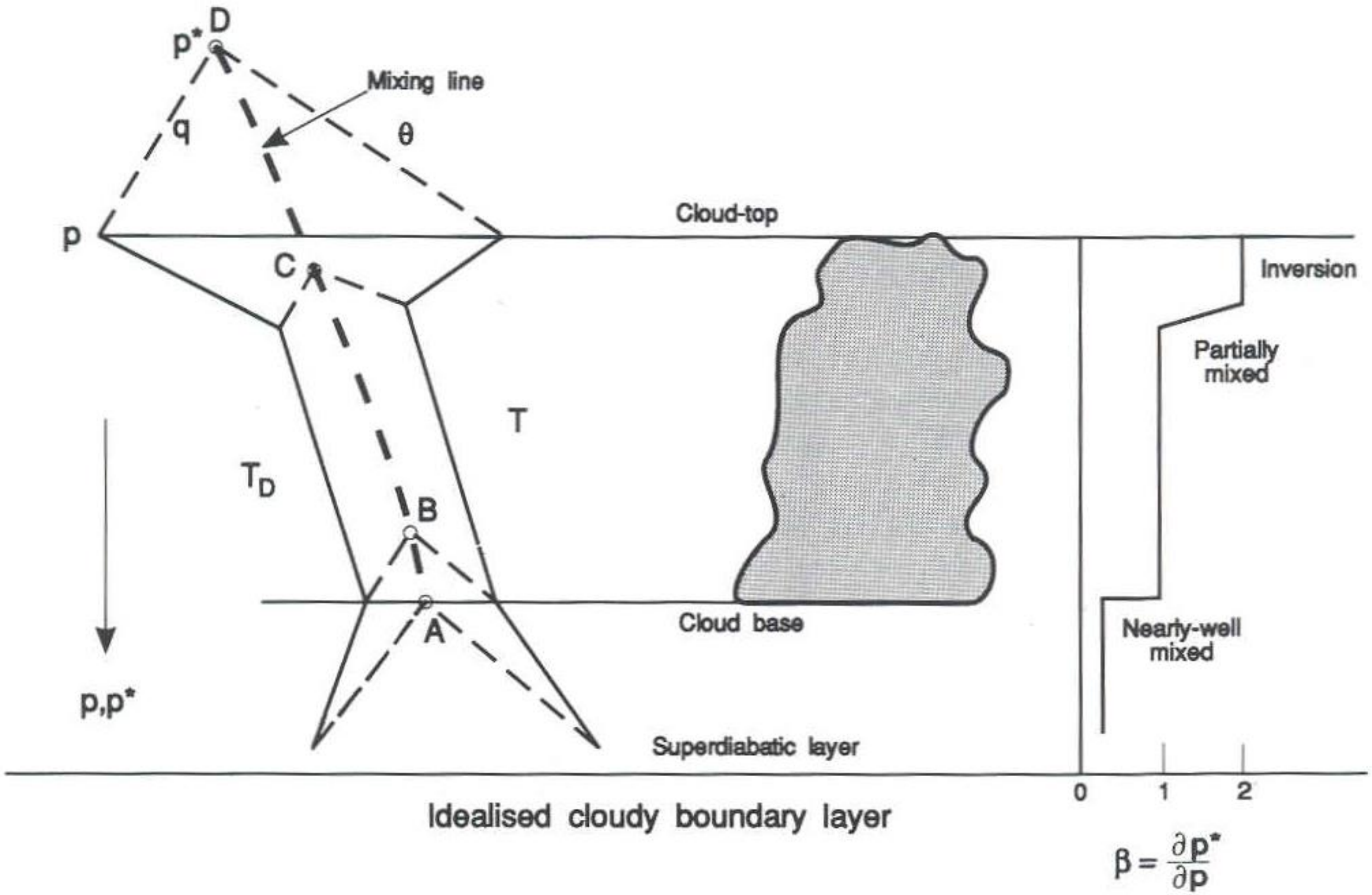
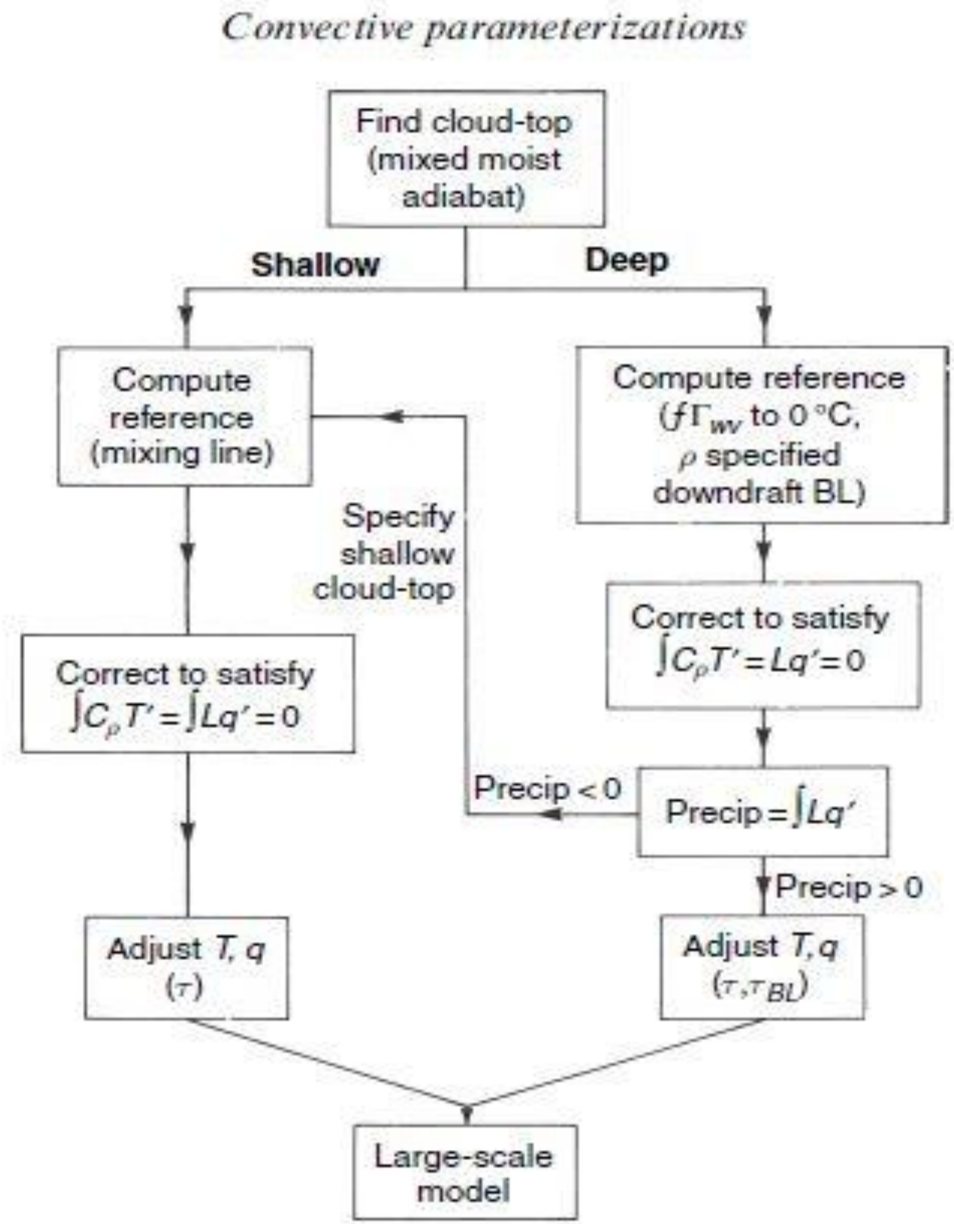


FIG. 9.1. Relationship between mixing line, temperature, and dewpoint, and a mixing parameter β for an idealized convective boundary layer. The light dashed lines are lines of constant potential temperature θ and mixing ratio q (from Betts 1986).

Betts-Miller scheme



Kain-Fritsch scheme

(Kain, 2004)

KAIN-FRITSCH CONVECTION PARAMETERIZATION

- Based on Fritsch-Chappell Scheme (1980)
- Based on Mesoscale Convective Systems
- Mass Flux Type
- Cloud Model to estimate the convective mass fluxes

Mass Flux Scheme

- The single component of the cloud is not treated individually but as bulk effects produced by an ensemble of clouds.

- The large scale area average mass flux, $\overline{M}$, is assumed to contain an environment part, represented as M_e, and a cloudy part, M_c, which represents the contribution from all clouds.

$$\overline{M} = M_c + M_e$$

- The clouds occupy a fractional area σ, and the environment $1-\sigma$

- Vertical mass fluxes can be rewritten as

$$\overline{\rho}\,\overline{w} = \sigma \overline{\rho} w_c + (1-\sigma)\overline{\rho} w_e$$

Mass Flux Scheme

• The observed $\overline{w}$ is generally small, which implies that the strong ascent within the cloud is compensated by the descent between clouds.

From the assumption that $\sigma << 1$, and that $\overline{w} \approx 0$, we have then

$$\overline{\rho}\,\overline{s'w'} \approx \sigma \overline{\rho} w_c (s_c - s_e) = M_c (s_c - s_e)$$

s : dry static energy

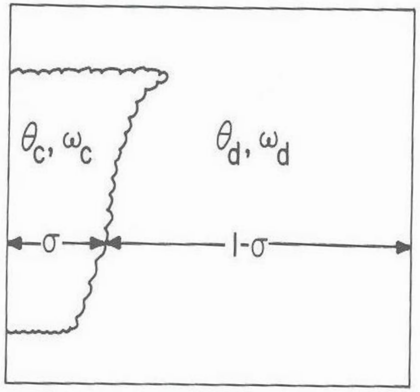


Fig. 3.1. Schematic diagram of a deep cumulus cloud and the percentage of the area covered by the cumulus cloud. [From Anthes (1977).]

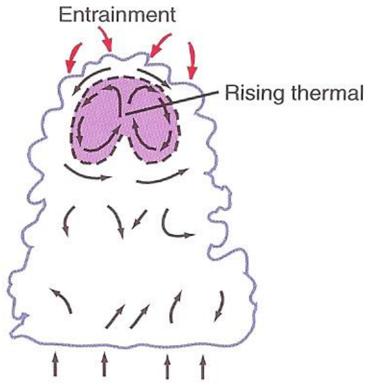


Fig. 6.12 Schematic of entrainment of ambient air into a small cumulus cloud. The thermal (shaded violet region) has ascended from cloud base. [Adapted from *J. Atmos. Sci.* **45**, 3957 (1988).]

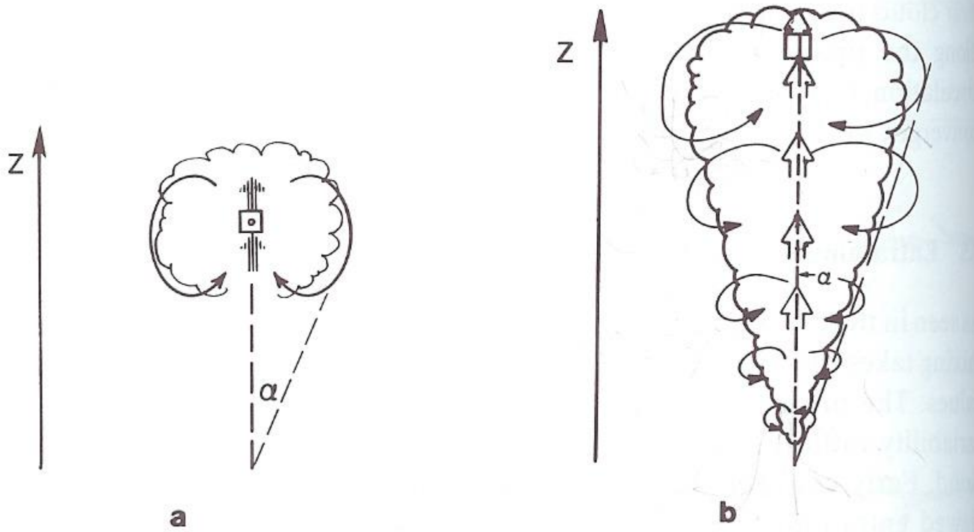


Fig. 8.33. (a) Schematic view of the "bubble" or "thermal" model of lateral entrainment in cumuli. (b) Schematic view of the "steady-state jet" model of lateral entrainment in cumuli.

•**Lateral entrainment** : injection of environmental air into the cloud. Dilution from cloud top downwards.

•**Detrainment:** cloud water lost to the environment. Cloud droplets evaporate in the unsaturated environment, cloud environment is cooled and bouyancy is increased.

Mass Flux Scheme

A cloud model is necessary to give values of M_c and s_c

Based on the mass conservation and assuming steady state

$$\frac{\partial \rho_d}{\partial t} + \nabla.(\rho_d V) = 0 \qquad \therefore \quad \frac{\partial \rho w_c}{\partial z} = E - D$$

E and D are the entrainment and detrainment rates

$$\frac{\partial M_c}{\partial z} = E - D$$

Extension to other typical cloud conservative properties:

$$\frac{\partial (M_c s_c)}{\partial z} = E\bar{s} - D s_c$$

In a moist atmosphere: $$\frac{\partial (M_c s_c)}{\partial z} = E\bar{s} - D s_c + L\rho c$$

Mass Flux Scheme

The conservation of water substance can be split into vapour, q, and liquid, l, phases.

$$\frac{\partial(M_c q_c)}{\partial z} = E\bar{q} - Dq_c - \rho c$$

$$\frac{\partial(M_c l)}{\partial z} = -Dl + \rho c - \rho k l$$

k is the rate of conversion of liquid water into precipitation.

By writing the convective eddy transports in the flux form, the energy conservation in the column is assured.

Mass Flux Scheme

The contribution from cumulus activity to the large scale heat and moisture are thus,

$$\left(\frac{\partial\bar{\theta}}{\partial t}\right)_{cu} = -\frac{1}{\bar{\rho}}\frac{\partial\left[M_c(\theta_c-\bar{\theta})\right]}{\partial z} + \frac{L}{c_p\bar{\pi}}(c-e)$$

$$\left(\frac{\partial q}{\partial t}\right)_{cu} = -\frac{1}{\bar{\rho}}\frac{\partial\left[M_c(q_c-\bar{q})\right]}{\partial z} - (c-e)$$

Observations show that cloud mass flux is larger than the vertical mass flux forced by large scale convergence.

The representation of cloud mass flux from large scale convergence is not enough to reproduce the warming in cloud free area. There is need for an explicit representation of mass transports, or of other quantities, within the cloud.

Trigger Function:

1. $$T_{LCL} + \Delta T - T_{ENV} \begin{cases} >0 \Rightarrow unstable \\ \le 0 \Rightarrow stable \end{cases}$$

→ Deep convection

$$\Delta T = k(w_g - c_z)^{\frac{1}{3}}$$

w_g : background vertical motion at LCL

$c_z = 2$, $z_{LCL} > 2000$

$c_z = 2*(z_{LCL}/2000)$, $z_{LCL} < 2000$

Parcel is given an extra temperature perturbation

2. $$w_{po} = 1 + 1.1\left(\frac{\Delta T}{T} Hpbl\right)^{1/2}$$

Dilute parcel ascent

$$w_p > 0$$ Within cloud depth (3-4km)

3. $$D_{\min} = \begin{cases} 4000 & , \quad T_{LCL} > 20^O C \\ 2000 & , \quad T_{LCL} < 0^O C \\ 2000 + 100 \times T_{LCL} & , \quad 0 \le T_{LCL} \le 20^O C \end{cases}$$

Minimum cloud depth

Cloud base Mass flux

$$M_{u0} = \rho w_{p0} A$$

- Deep convection : buoyancy
- Shallow clouds: TKE

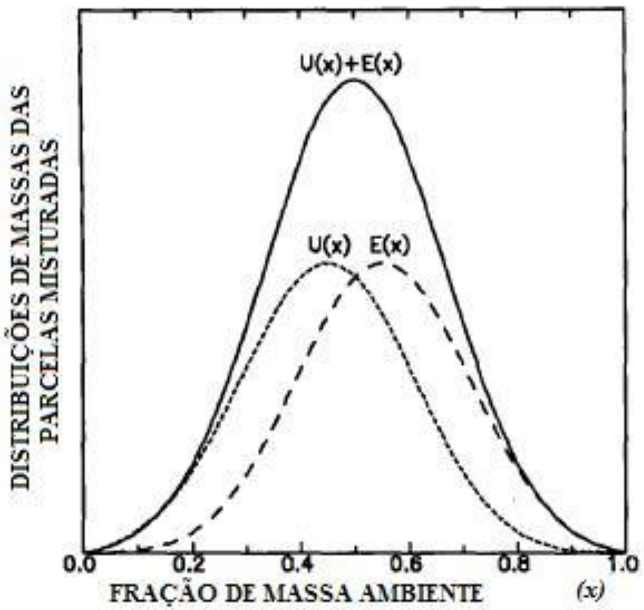
Updraft:

$$\frac{\partial M_u}{\partial z} = \varepsilon_u - \delta_u$$

$$\frac{\partial M_u \theta_u}{\partial z} = \varepsilon_u \theta - \delta_u \theta_u + L(c-e)$$

$$\frac{\partial M_u q_u}{\partial z} = \varepsilon_u q - \delta_u q_u - L(c-e) - P$$

$$\frac{\partial M_u l}{\partial z} = -\delta_u l - P + L(c-e)$$



increase ε in high buoyancy and/or moist environment

increase δ in low buoyancy and/or dry environment

E(x): environmental mass distribution

U(x): updraft mass distribution

f(x): gaussian mass distribuition

Entrainment rate

$$M_{ee} = \delta M_t \int_0^{x_C} x f(x) dx$$

Detrainment rate

$$M_{ud} = \delta M_t \int_{x_C}^{1} (1 - x) f(x) dx$$

Updraft:

Variable cloud radius, *R*: control of entrainment rate

$$\delta M_e = M_{uB} \frac{(-0.03 * \delta p)}{R}$$

δM_e : maximum possible entrainment rate

R dependent on large scale forcing through grid-scale vertical motion

$$R = \begin{bmatrix} 1000, W_{KL} < 0 \\ 2000, W_{KL} > 10 \\ 1000 + \frac{W_{KL}}{10}, 0 \leq W_{KL} \leq 10 \end{bmatrix}$$

W_{KL} is proportional to w_g

Weaker dilution when low-level forcing is stronger

Downdraft:

- LFS: Level of Free Sink (DSL : downdraft source level) occurs about 150-200 hPa above cloud base

- downdraft source is environmental air only

- downdraft ends when it becomes warmer than the environment or reaches the surface.

$$\frac{\partial M_d}{\partial z} = \varepsilon_d - \delta_d$$

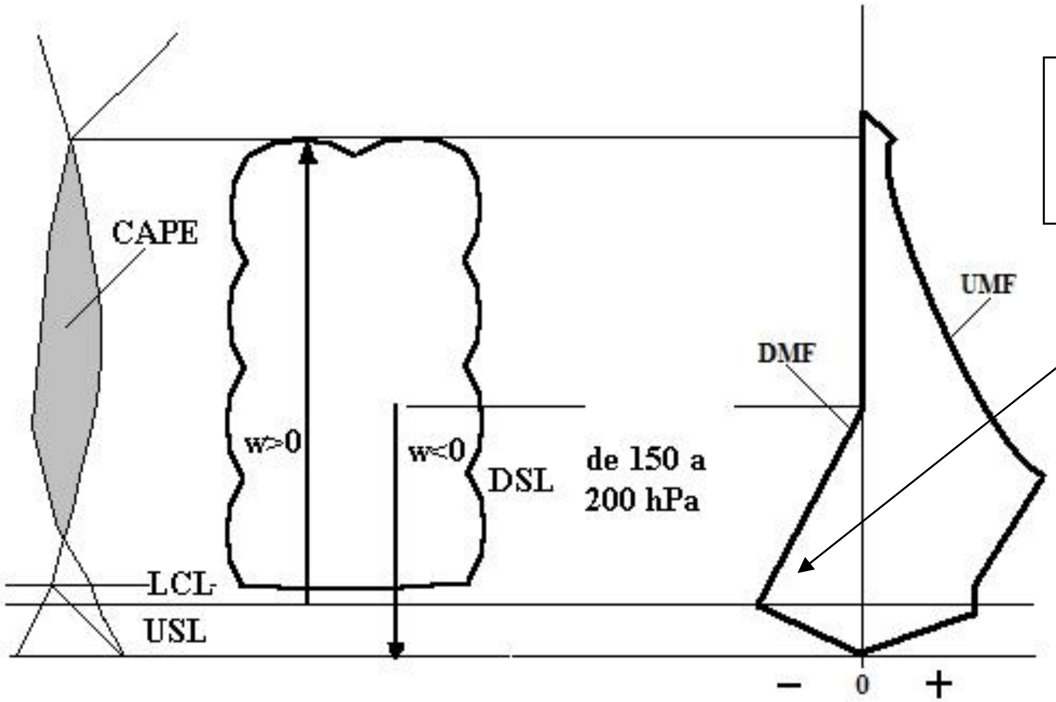
$$\frac{\partial M_d \theta_d}{\partial z} = \varepsilon_d \theta - \delta_d \theta_d - Le$$

$$\frac{\partial M_d q_d}{\partial z} = \varepsilon_d q - \delta_d q_d + Le$$

M_d at DSL to start derivation

$$M_{dDSL} = M_{uB} * (2 \times (1 - \overline{RH}_{DSL}))$$

Downdraft:



$$\frac{DMF_{USL}}{UMF_{USL}} = 2 \times \left(1 - \overline{RH}\right)$$

Shallow convection:

- Trigger: same as deep convection but cloud depth smaller than the minimum cloud depth, D_{min}
- No precipitation produced
- R is constant, entrainment rate is constant
- $M_{uB} \propto$ max TKE in subcloud layer

$$M_{u0} = \begin{cases} \left(\frac{TKE_{MAX}}{k_0}\right) \times \left(\frac{m_{USL}}{\tau_C}\right) & , \quad TKE_{MAX} < 10 \\ \left(\frac{10}{k_0}\right) \times \left(\frac{m_{USL}}{\tau_C}\right) & , \quad TKE_{MAX} \geq 10 \end{cases}$$

τ_C é o período de tempo convectivo, variando de 1800 a 3600 [s];

m_{USL} é quantidade de massa no USL [kg];

ko valor de referência [m^2 s^{-2}].

CLOSURE:

- 0.9*CAPE in the column is removed within t_c
- CAPE is calculated from dilute parcel ascent > smaller M_{uB}
- CAPE is removed by lowering θ_e in the USL and warming environment aloft

Tendencies:

$$\left.\frac{\partial \theta}{\partial t}\right|_{con} = -\frac{1}{\rho}\frac{\partial}{\partial z}\left[\left(M_u + M_d\right)\theta + \left(\varepsilon_u + \varepsilon_d\right)\theta - \left(\delta_u + \delta_d\right)\theta_u\right]$$

$$\left.\frac{\partial q}{\partial t}\right|_{con} = -\frac{1}{\rho}\frac{\partial}{\partial z}\left[\left(M_u + M_d\right)q + \left(\varepsilon_u + \varepsilon_d\right)q - \left(\delta_u + \delta_d\right)q_u\right]$$

Flux form: Conservation of moisture and energy

Extension to momentum fluxes introduced by Paulo Bastos

$$\left.\frac{\partial u}{\partial t}\right|_{con} = -\frac{1}{\rho}\frac{\partial}{\partial z}\left[\left(M_u + M_d\right)u + \left(\varepsilon_u + \varepsilon_d\right)u - \left(\delta_u + \delta_d\right)u_u\right]$$

Convective momentum fluxes (KF,93)

Zonal momentum equation

$$\left.\frac{\Delta \bar{u}}{\Delta t}\right|_{CONV} = \frac{1}{\Delta p}\left[(\omega_{u2}+\omega_{d2})\bar{u}_2 - (\omega_{u1}+\omega_{d1})\bar{u}_1 + (\varepsilon_u+\varepsilon_d)\bar{u}_m - \delta_u u_{um} - \delta_d u_{dm}\right]$$

Mass flux Entrainment Detrainment

Meridional momentum equation

$$\left.\frac{\Delta \bar{v}}{\Delta t}\right|_{CONV} = \frac{1}{\Delta p}\left[(\omega_{u2}+\omega_{d2})\bar{v}_2 - (\omega_{u1}+\omega_{d1})\bar{v}_1 + (\varepsilon_u+\varepsilon_d)\bar{v}_m - \delta_u v_{um} - \delta_d v_{dm}\right]$$

Kain Fritsch scheme improvements

1. - Inclusion of a F parameter to control precipitation production
2. - Inclusion of Cumulus friction

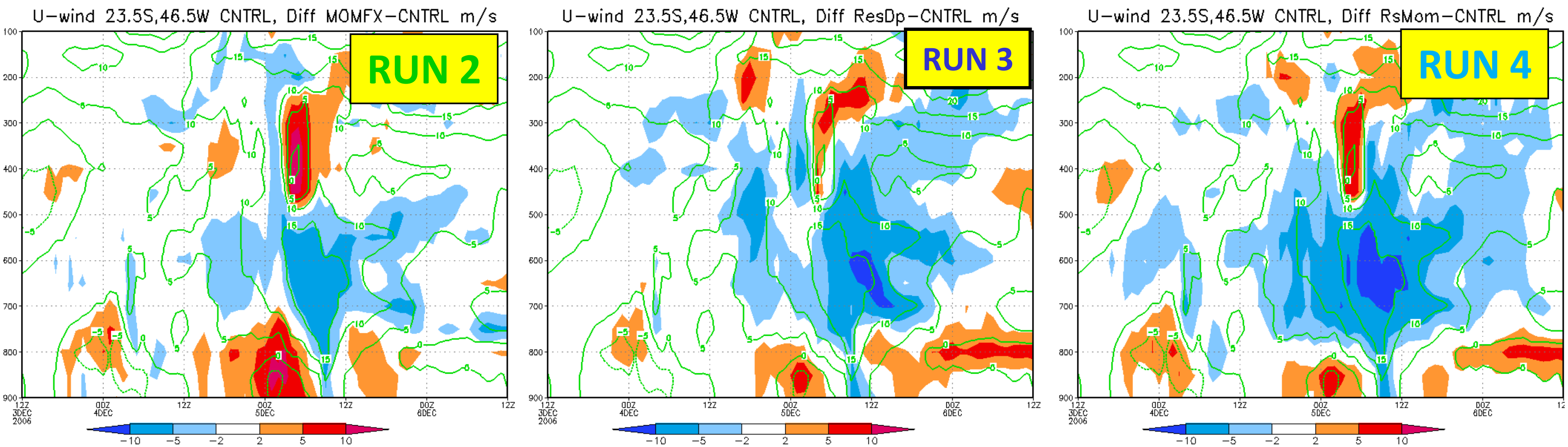
Run 1: Control

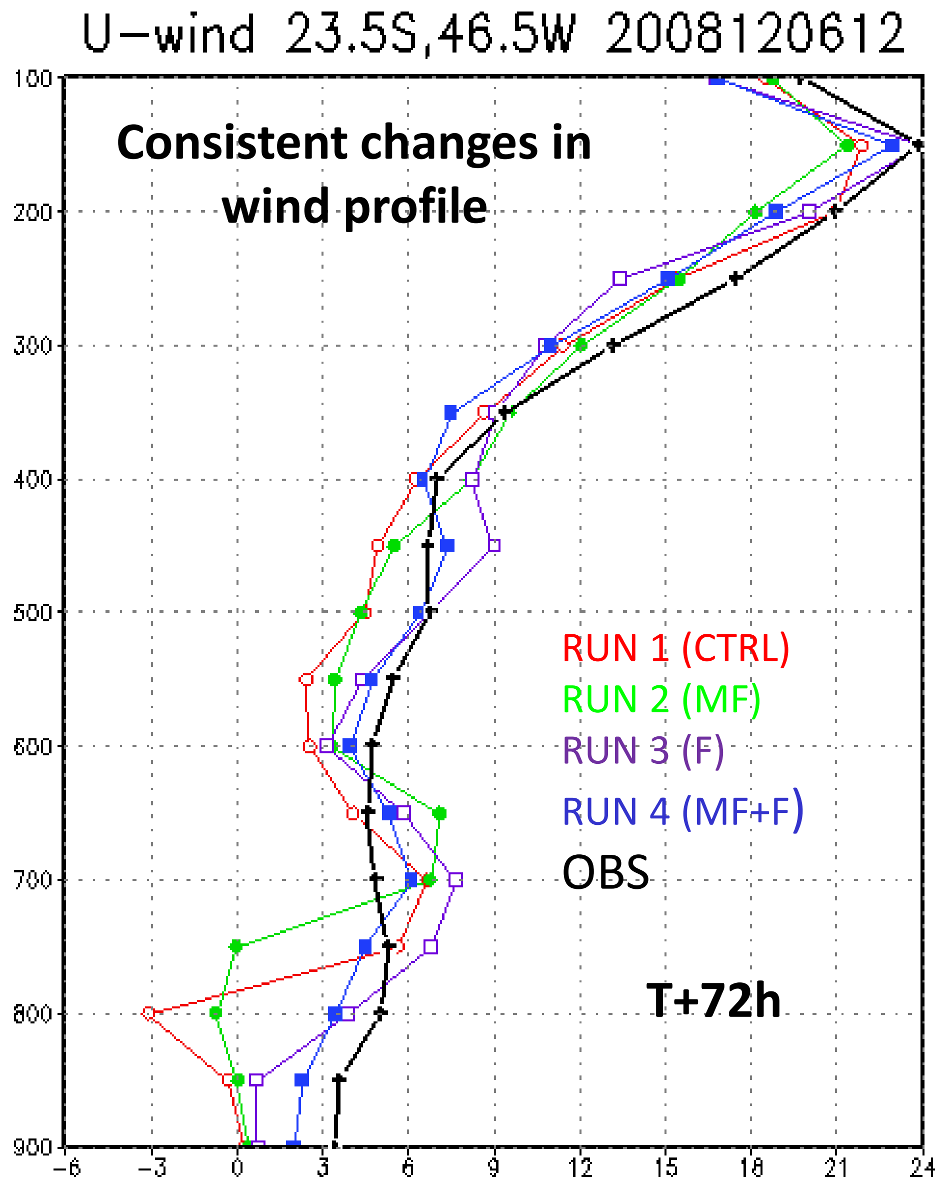
Run 2: Inclusion of cumulus friction (Bastos and Chou, 2009)

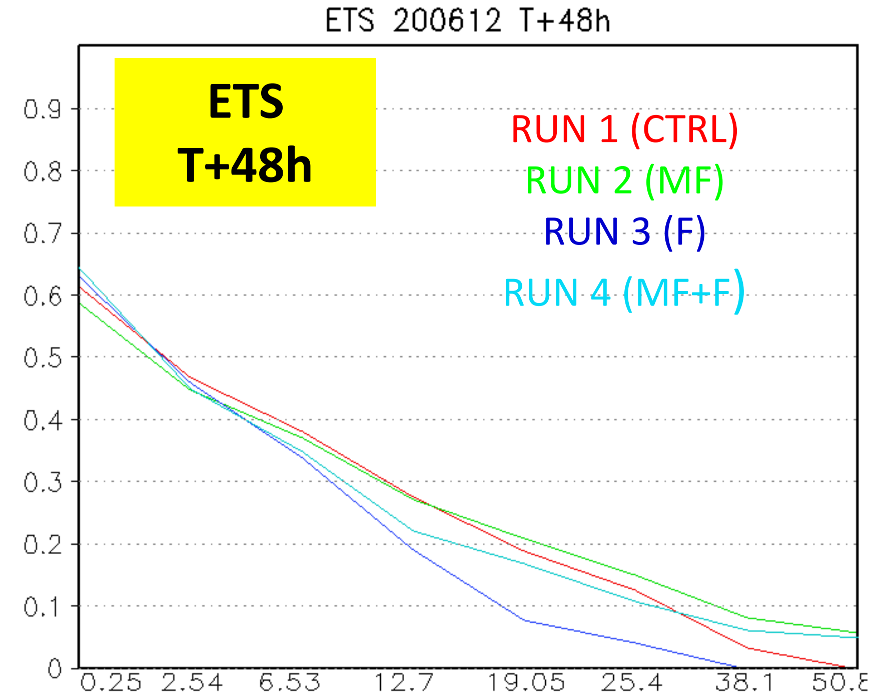
Run 3: Inclusion of F-fcn (F) (Gomes and Chou, 2010)

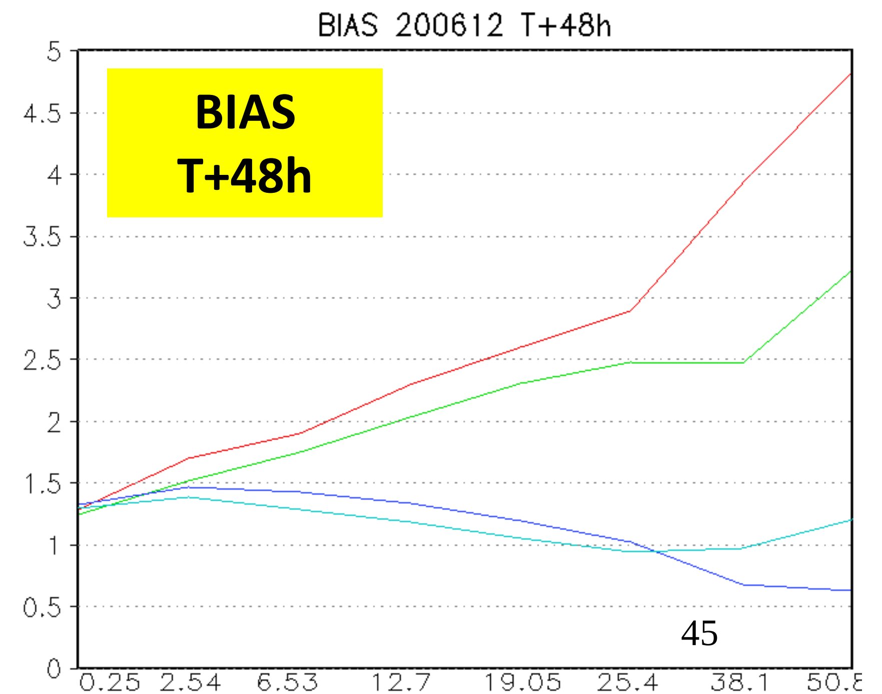
Run 4: Inclusions of cumulus friction and F-parameter (MF+F)

U-wind (m/s)




U-wind 23.5S,46.5W 2008120612
Consistent changes in wind profile
100
200
300
400
500
600
700
800
900
-6
-3
0
3
6
9
12
15
18
21
24
RUN 1 (CTRL)
RUN 2 (MF)
RUN 3 (F)
RUN 4 (MF+F)
OBS
T+72h


ETS 200612 T+48h
ETS
T+48h
RUN 1 (CTRL)
RUN 2 (MF)
RUN 3 (F)
RUN 4 (MF+F)
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
0.25
2.54
6.53
12.7
19.05
25.4
38.1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5
4.5
4
3.5
3
2.5
2
1.5
1
0.5
0
0.25
2.54
6.53
12.7
19.05
25.4
38.1