Instituto Nacional de Pesquisas Espaciais

Parametrização de chuva e nuvens

Chou Sin Chan

chou@cptec.inpe.br

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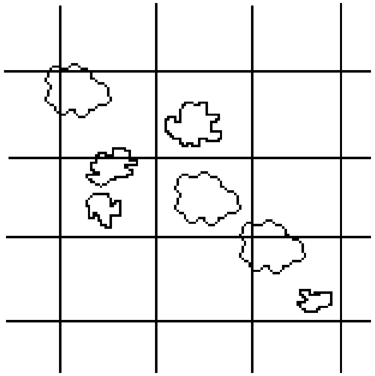
The need for cumulus parameterization

*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.

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):

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

Variável composta por uma média resolvida e uma perturbação

'–' se refere à média na grade do modelo

ϕ' perturbação subgrade

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

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

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

$$\boxed{\overline{\frac{\partial u}{\partial t}} = \frac{\partial \overline{u}}{\partial t}} \qquad \boxed{\overline{\frac{\partial u}{\partial x}} = \frac{\partial \overline{u}}{\partial x}}$$

$$\boxed{\overline{\phi' \overline{u}} = 0} \qquad \boxed{\overline{\phi' w'} \neq 0}$$

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

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

Equations of heat, moisture and continuity

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

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

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

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.

Effects of cumulus convection on large scale thermodynamic fields

$$Q_1 \equiv \frac{\partial \bar{s}}{\partial t} + \bar{v} \bullet \nabla \bar{s} + \bar{\omega} \frac{\partial \bar{s}}{\partial p} = Q_R + L(\bar{c} - \bar{e}) - \nabla.\overline{s'v'} - \frac{\partial}{\partial p}\overline{s'\omega'}$$

Q_1: Apparent Heat source

$$Q_2 \equiv -L\left(\frac{\partial \bar{q}}{\partial t} + \bar{v} \bullet \nabla \bar{q} + \bar{\omega} \frac{\partial \bar{q}}{\partial p}\right) = L(\bar{c} - \bar{e}) + L\nabla.\overline{q'v'} + L\frac{\partial}{\partial p}\overline{q'\omega'}$$

Q_2: Apparent Moisture sink

Assume horizontal fluxes are negligible compared to vertical fluxes

$$\boxed{Q_1 - Q_R - Q_2 = -\frac{\partial}{\partial p}\overline{h'\omega'}}$$

Eddy (sensible and latent) heat fluxes
h is the moist static energy

$$\frac{c_p}{g}\int_{p_t}^{p_s}(Q_1 - Q_R)dp = LP + H_s$$

Verify against sfc observations

$$\frac{c_p}{g}\int_{p_t}^{p_s} Q_2 dp = L(P - E)$$

Yanai et al., 1973
Yanai and Johnson, 1993
Fig 4.1, Fig 4.3, F4.17

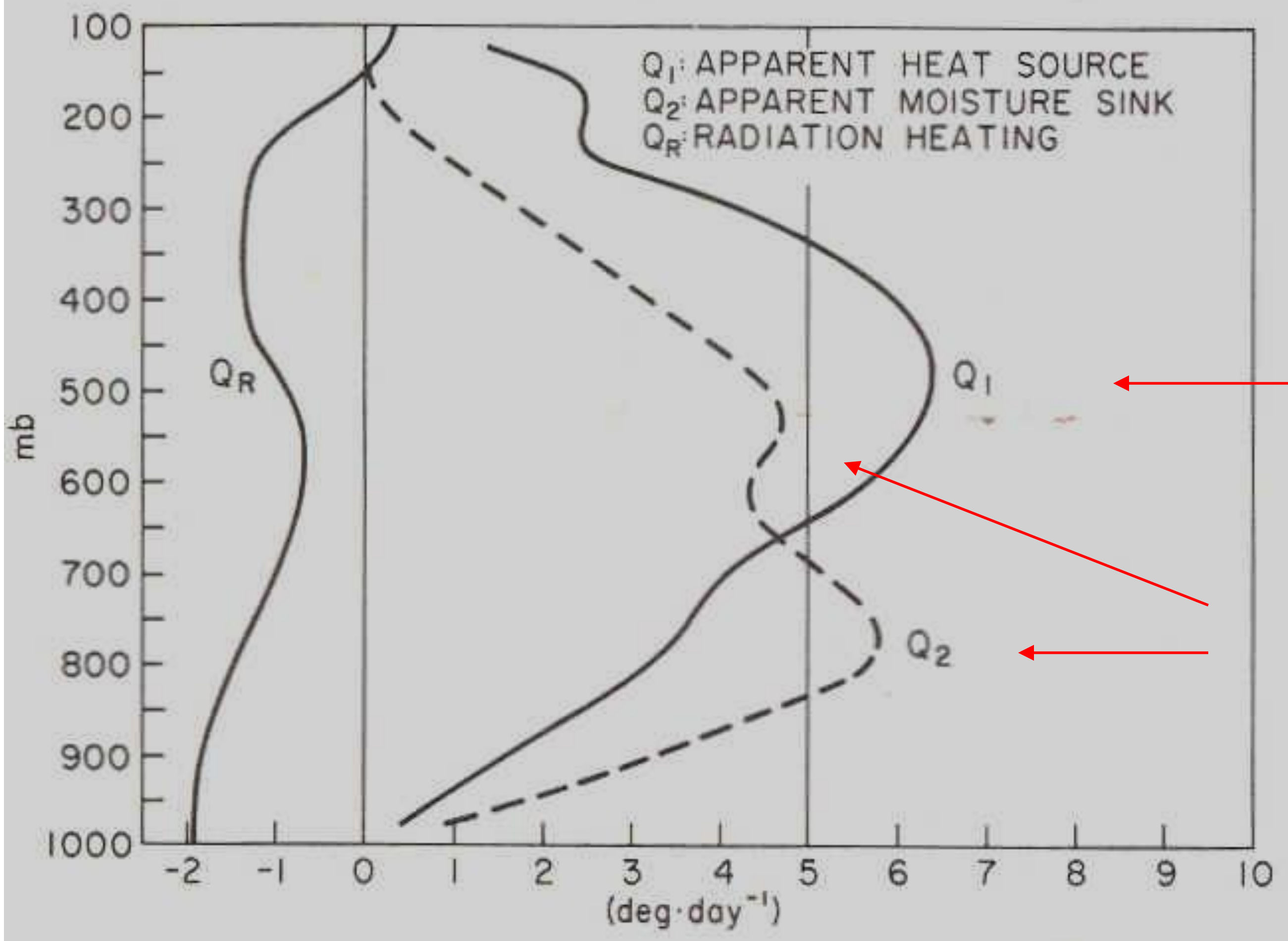


FIG. 4.1. The mean apparent heat source Q_1 (solid) and moisture sink Q_2 (dashed) over the Marshall Islands. On the left is the radiational heating rate given by Dopplick (1972) (from Yanai et al. 1973).

~5ºC/dia,
~400-500 hPa

~4ºC/dia,
~500 hPa,
~800 hPa

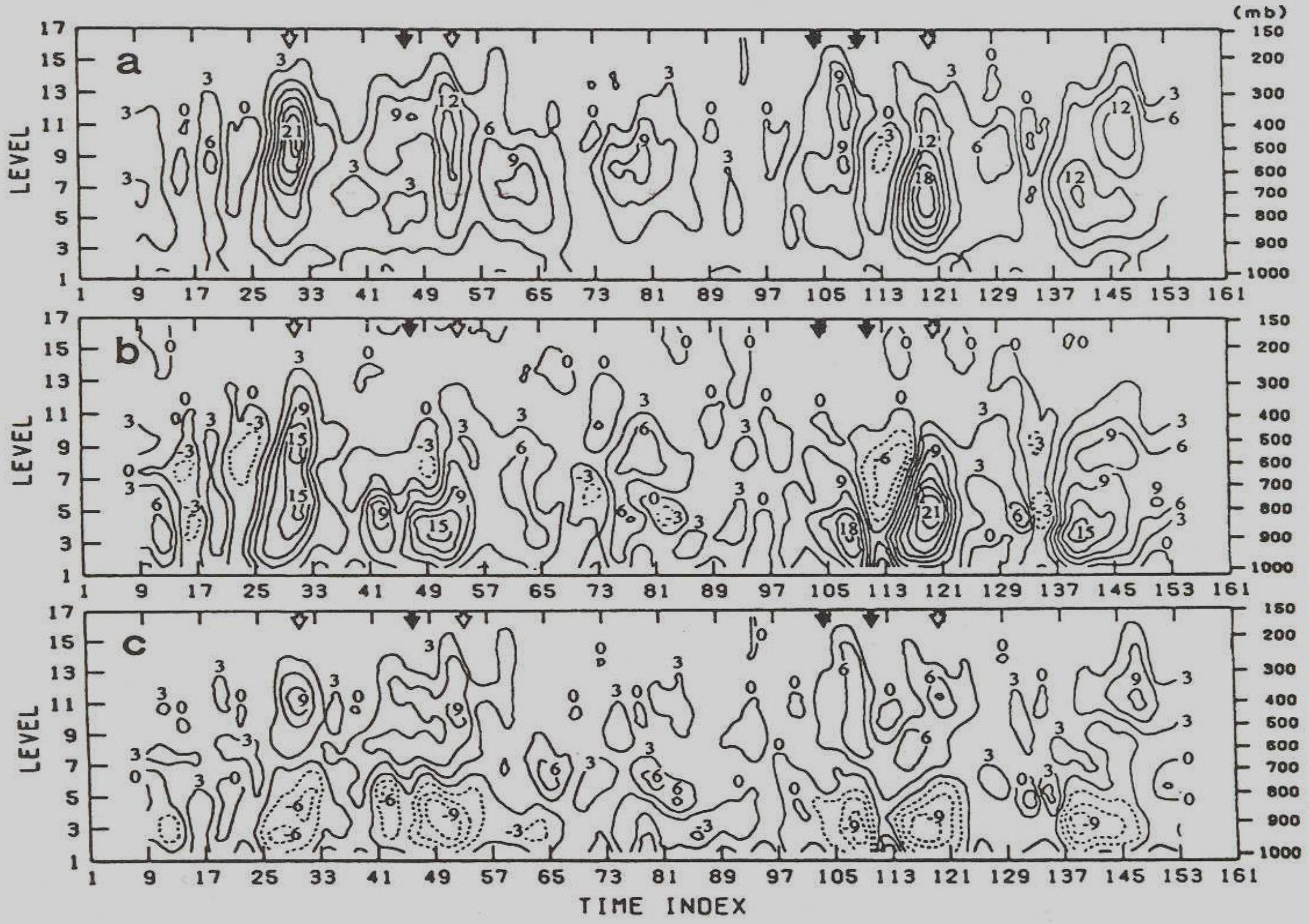


FIG. 4.3. Time–height sections of the observed (a) $Q_1 - Q_R$, (b) Q_2, and (c) $Q_1 - Q_2 - Q_R$, averaged over the 3° × 3° area at the center of the GATE network during a period from 0000 UTC 31 August (index 9) to 0000 UTC 18 September 1974 (index 153). Units are kelvins per day (from Cheng and Yanai 1989).

Q_1

Q_2

Q_1-Q_2-Q_R

warming

cooling

$$\boxed{Q_1 - Q_R - Q_2 = -\frac{\partial}{\partial p}\overline{h'\omega'}}$$

Downward motion

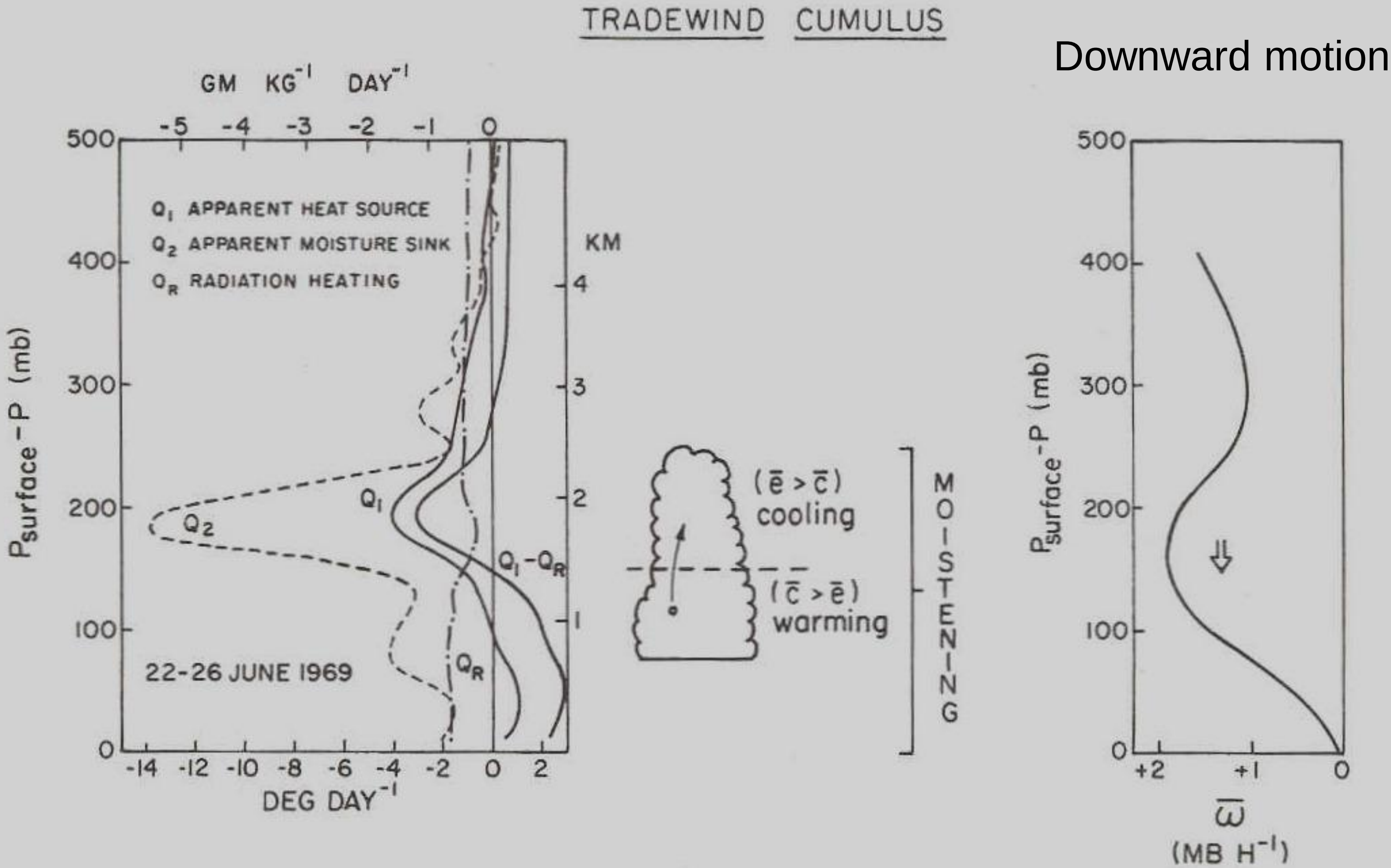
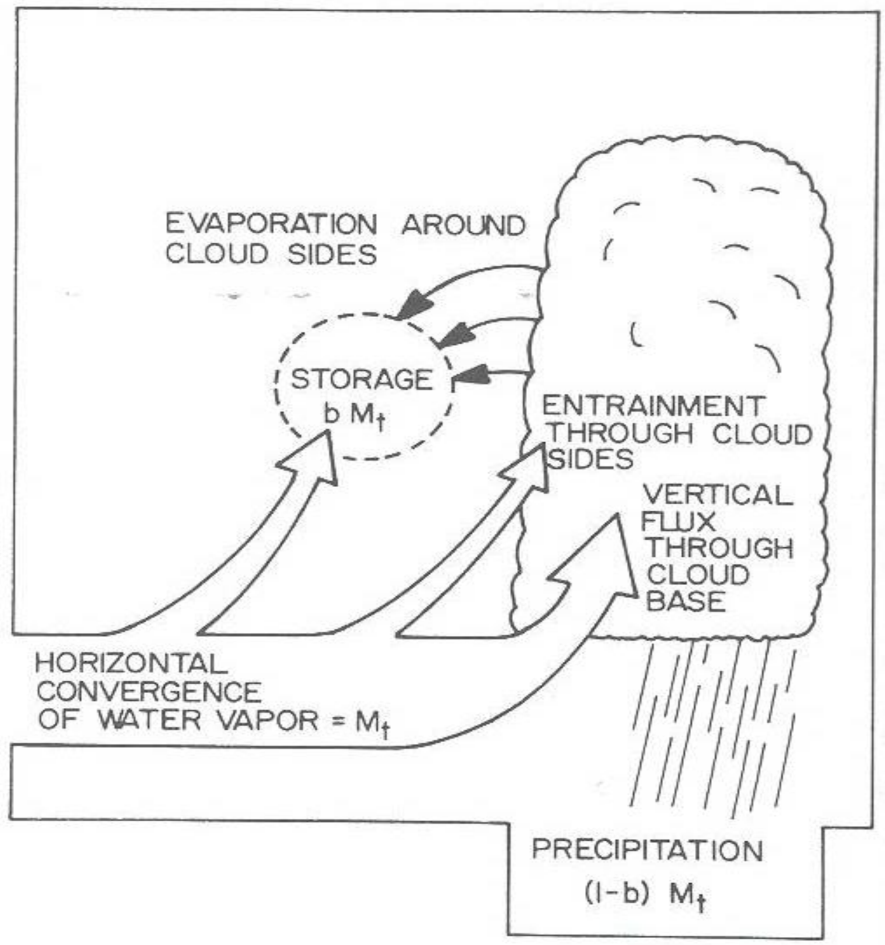


FIG. 4.17. (Left) The observed Q_1, Q_2, Q_R, and $Q_1 - Q_R$ for the undisturbed BOMEX period 22–26 June 1969 (from Nitta and Esbensen 1974). (Center) Schematic of trade-wind cumulus layer showing effects of condensation and evaporation on the heat and moisture budgets. (Right) Mean vertical p velocity $\bar{\omega}$ over budget area.

Types of convective scheme:

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

The Kuo Convective Parameterization 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 > 290 hPa: deep convection, else shallow convection.

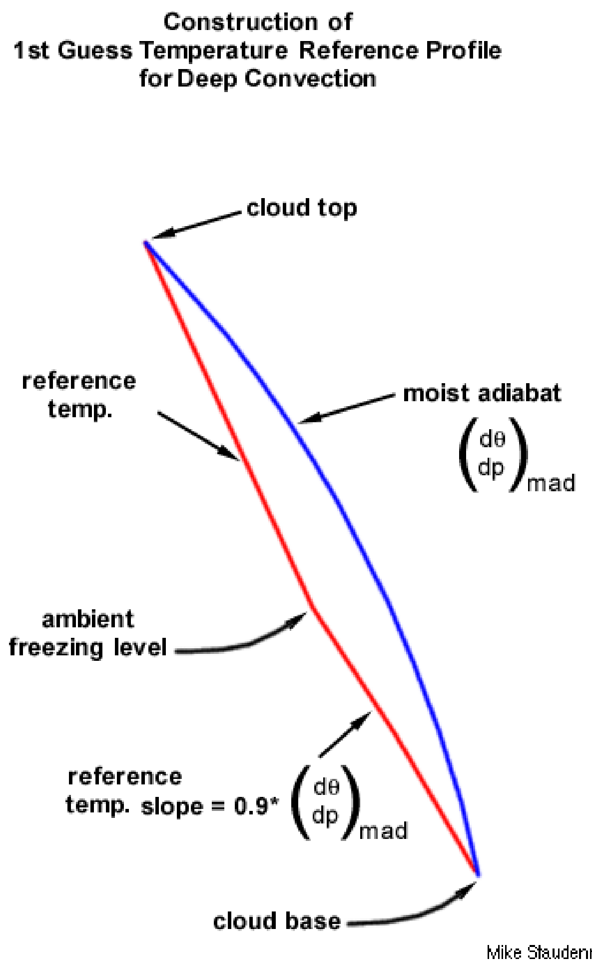
2. Determine reference profiles

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

Make sure enthalpy is conserved

Deep Convection

Temperature Reference Profile



To draw the temperature profile 3 levels are important:

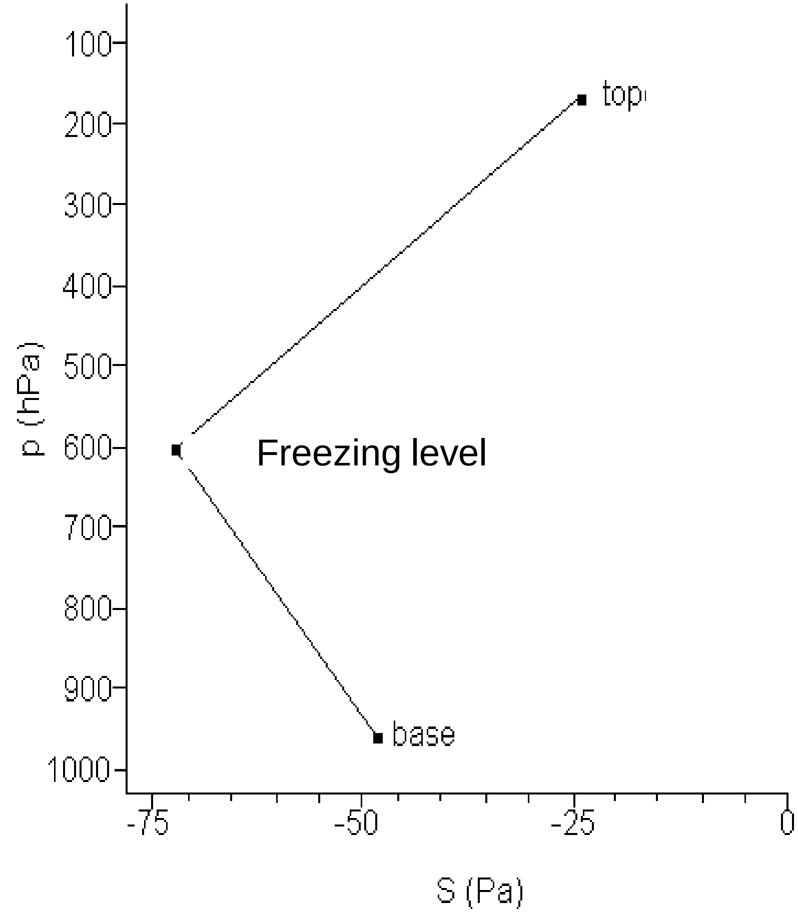
- cloud base
- freezing level
- cloud top

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

- 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

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

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

$$0.7 \leq F \leq 1.0 \;\; for \;\; 0.2 \leq E \leq 1.0$$

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

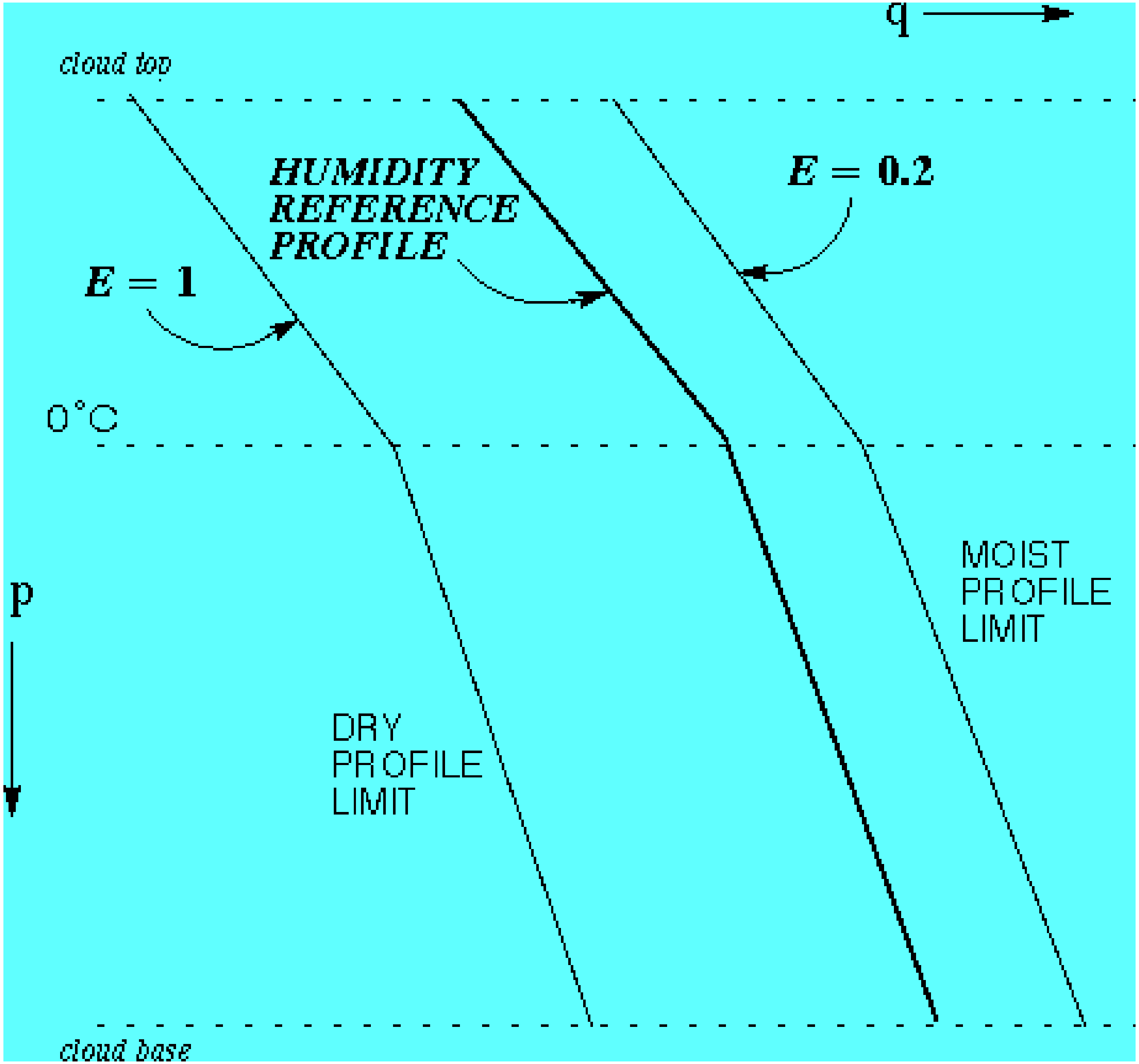
$$T_{new} = T_{old} + \frac{\Delta t_{cnv}}{\tau'}\left[T_{ref} - T_{old}\right]$$

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

where

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

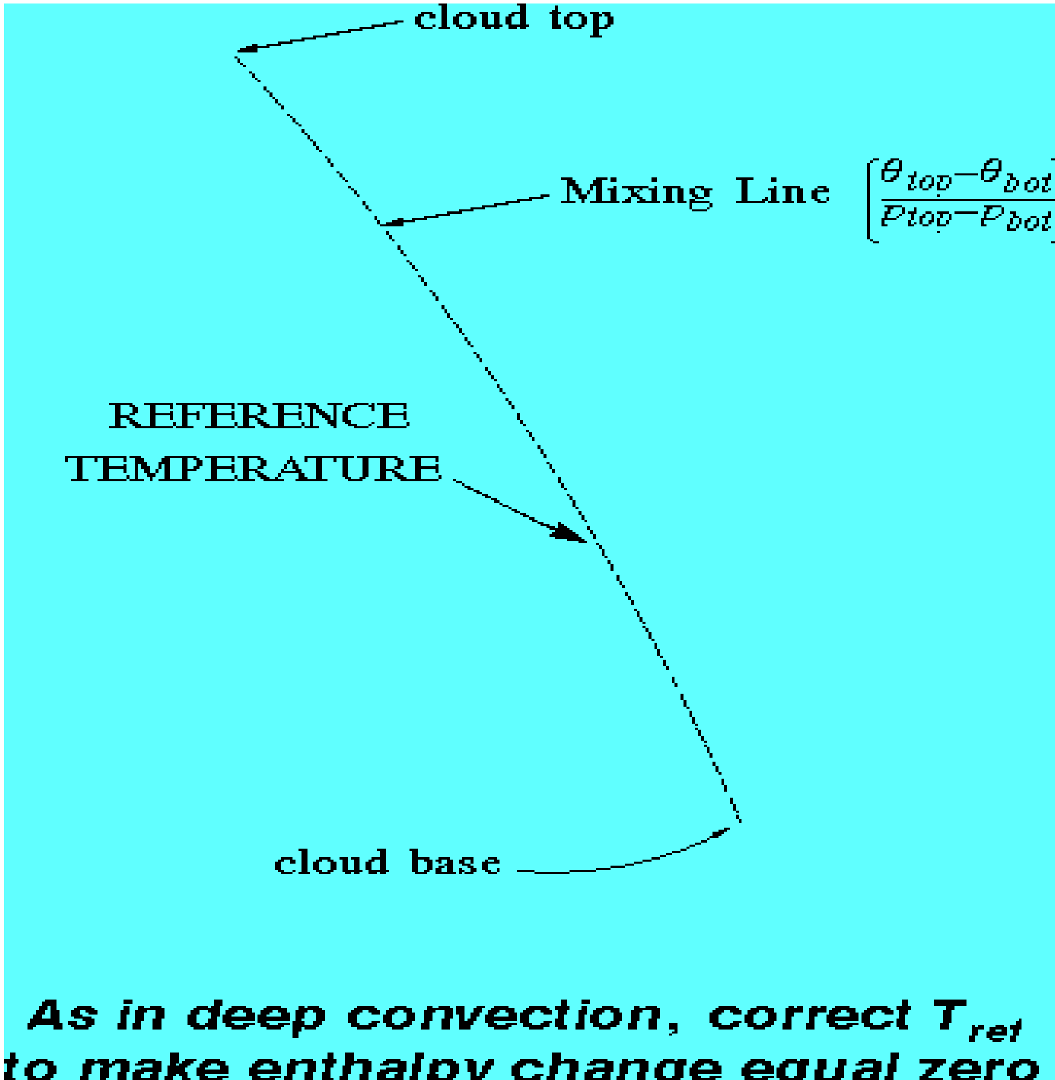
Fefi, fss



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 290hPa
- At least two layers
- swap points:
  - precipitation < 0
  - entropy change < 0

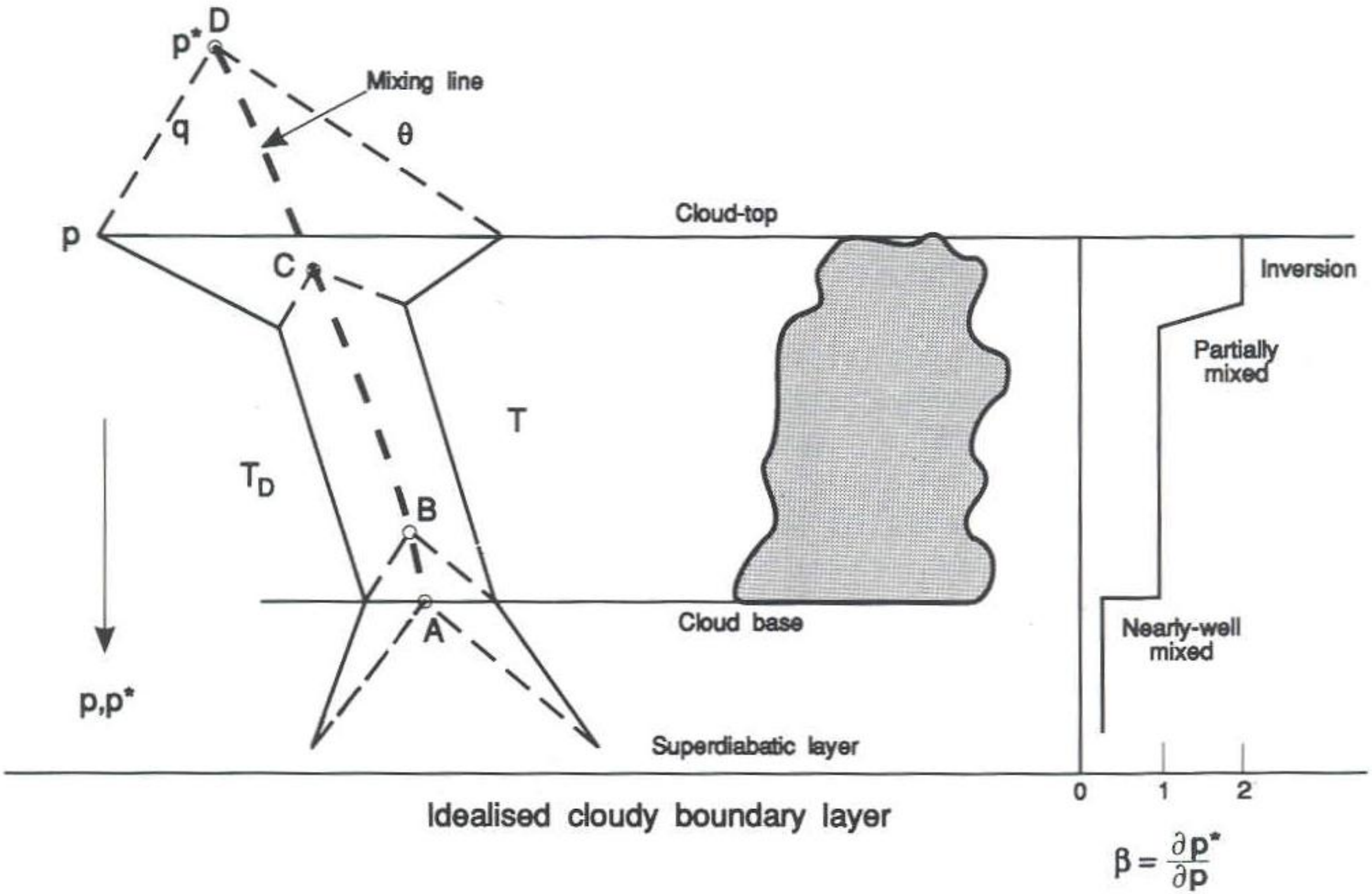


FIG. 9.1. Relationship between mixing line, temperature, and dewpoint, and a mixing parameter β for an idealized convective boundary layer. The light dashed lines are lines of constant potential temperature θ and mixing ratio q (from Betts 1986).

The need for modelling cloud microphysics

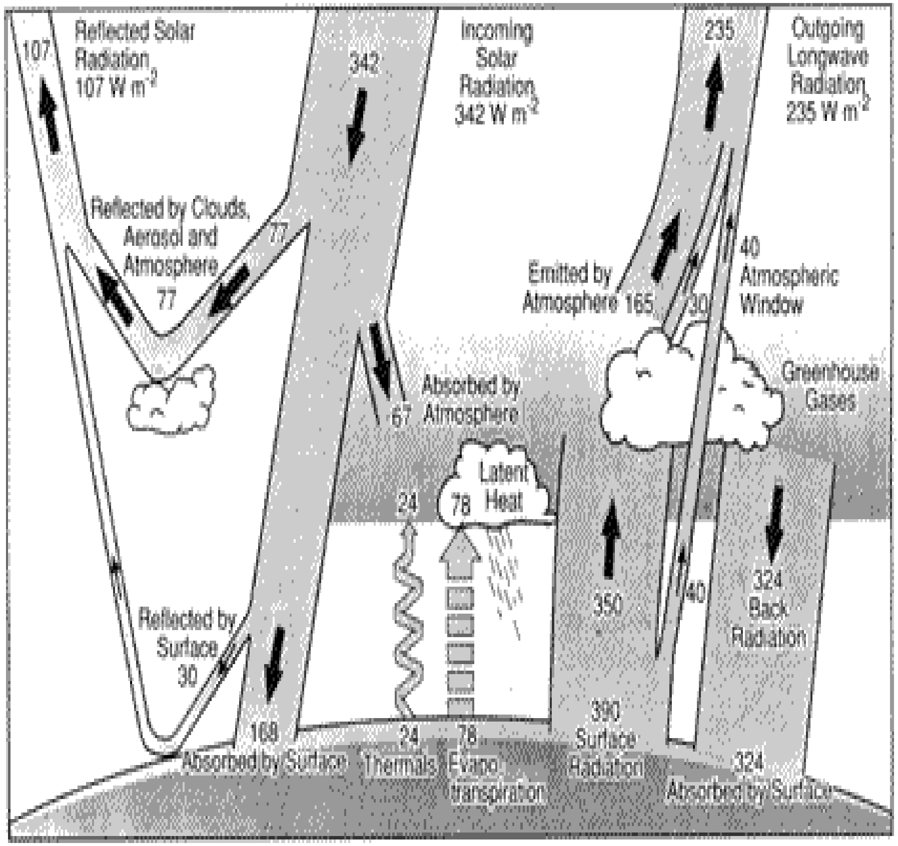
*Presence of clouds change the radiation budget

*Moisture and temperature adjustments in convectively stable points.

*Precipitation production for model hydrology

*Identification of precipitation types.

Excess from grid-point moisture excess in large scale models (large scale precipitation)



Cloud Formation

- Air rises and cools to saturation at LCL,
- Most effective Cloud Condensation Nuclei, CCN, are activated,
- Saturation vapor pressure decrease as parcel continues to rise and cool - the parcel becomes supersaturated,
- More CCN activate at the higher humidity.

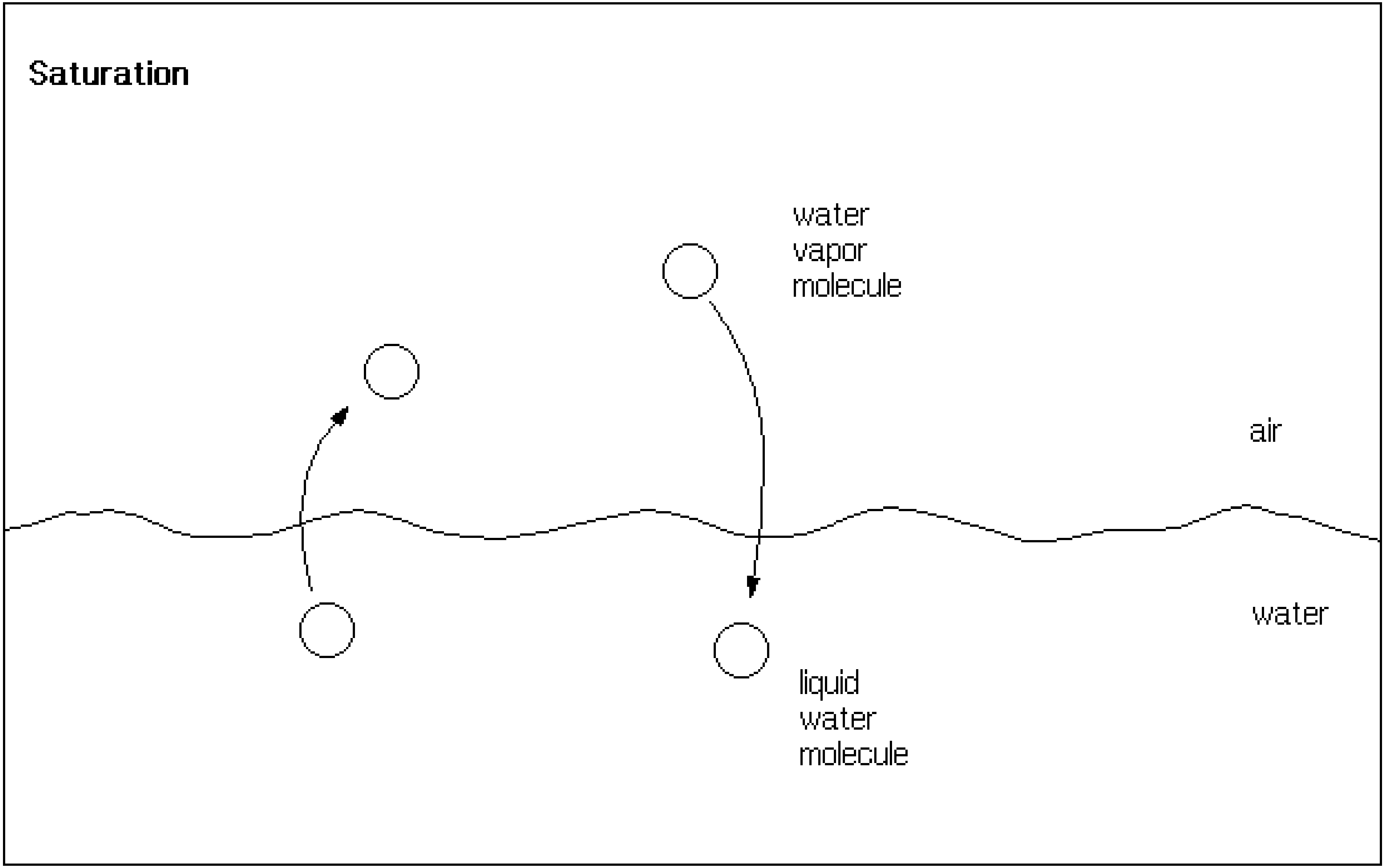
Cloud Condensation Nuclei

Particles suspended in the air which support the growth of cloud droplets or ice on their surface

soils, sand, dust, salt from sea spraying, volcanic debris, particles emitted from urban factories, natural aerosols

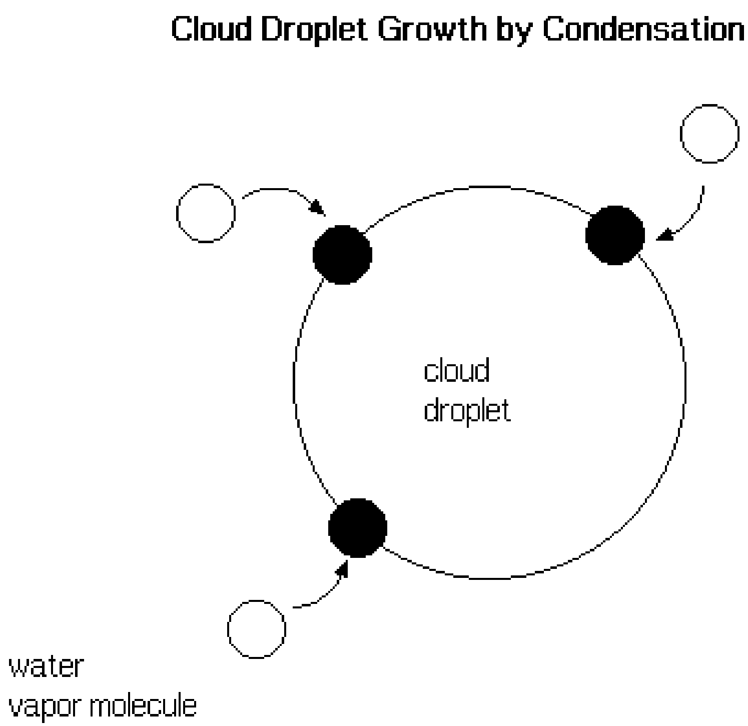
Cloud droplet Growth

- Condensation - Diffusion
- Collision and Coalescence
- Melting from ice


Saturation
water
vapor
molecule
air
water
liquid
water
molecule

Cloud Droplet Growth by Condensation (Diffusion)

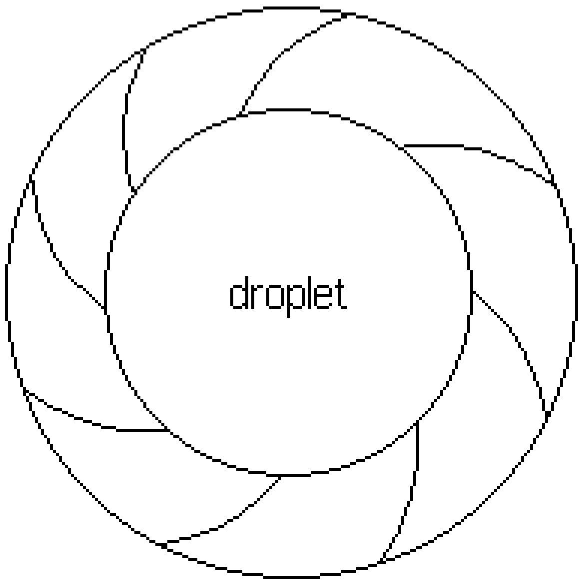
- Driven by the saturation vapor pressure difference
- Vapor is transported from higher to lower saturation vapor pressure.



Comparisons between e_s of a Droplet and its Environment

the environment and the air surrounding the droplet
are assumed to be saturated

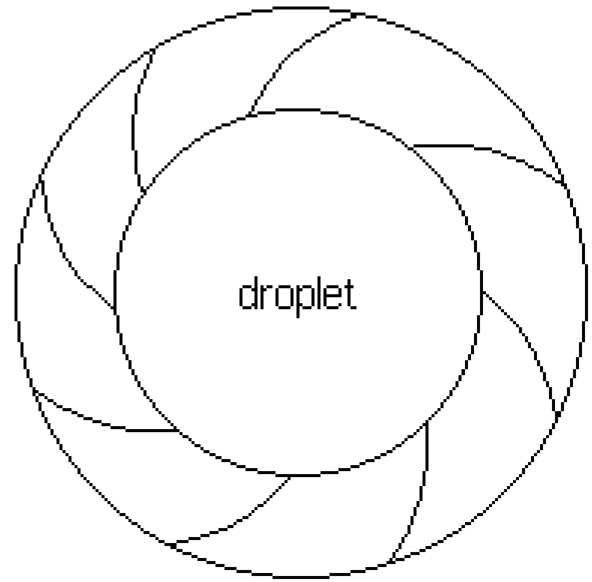
$e_{s_{environment}}$



$e_{s_{env}} > e_{s_{droplet}}$

The droplet grows by condensation

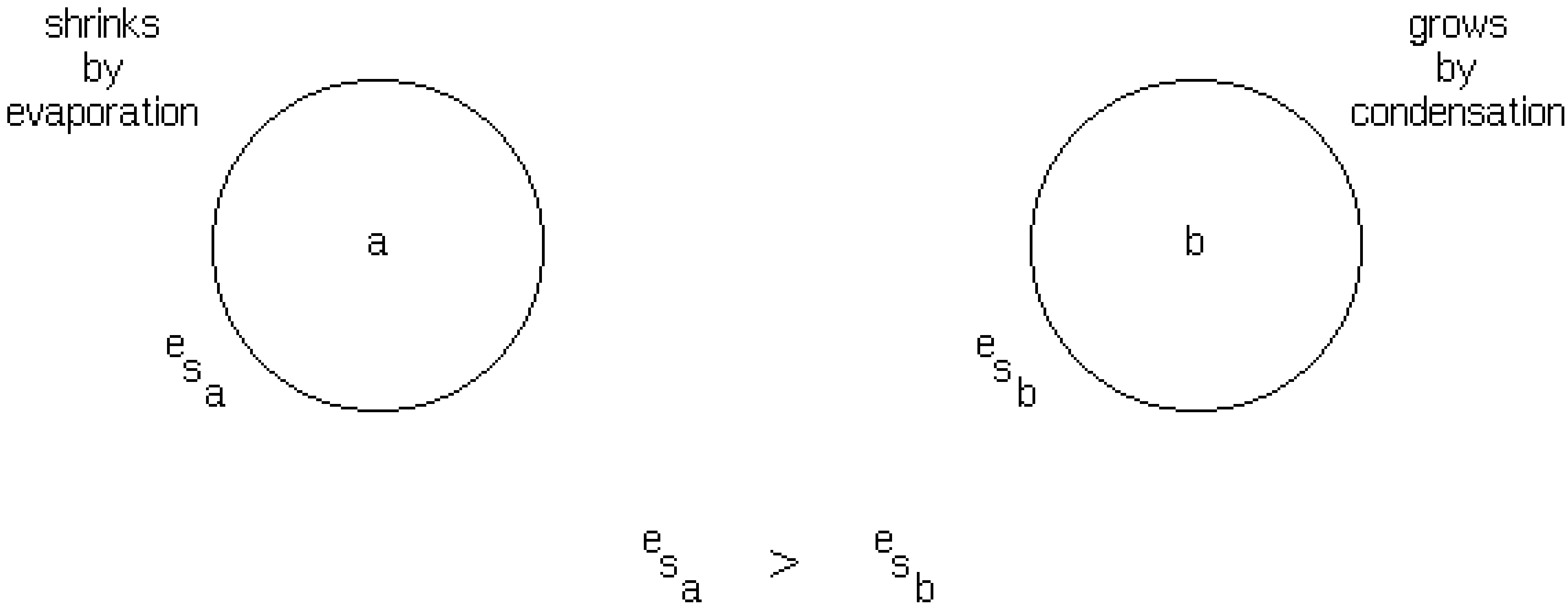
$e_{s_{environment}}$



$e_{s_{env}} < e_{s_{droplet}}$

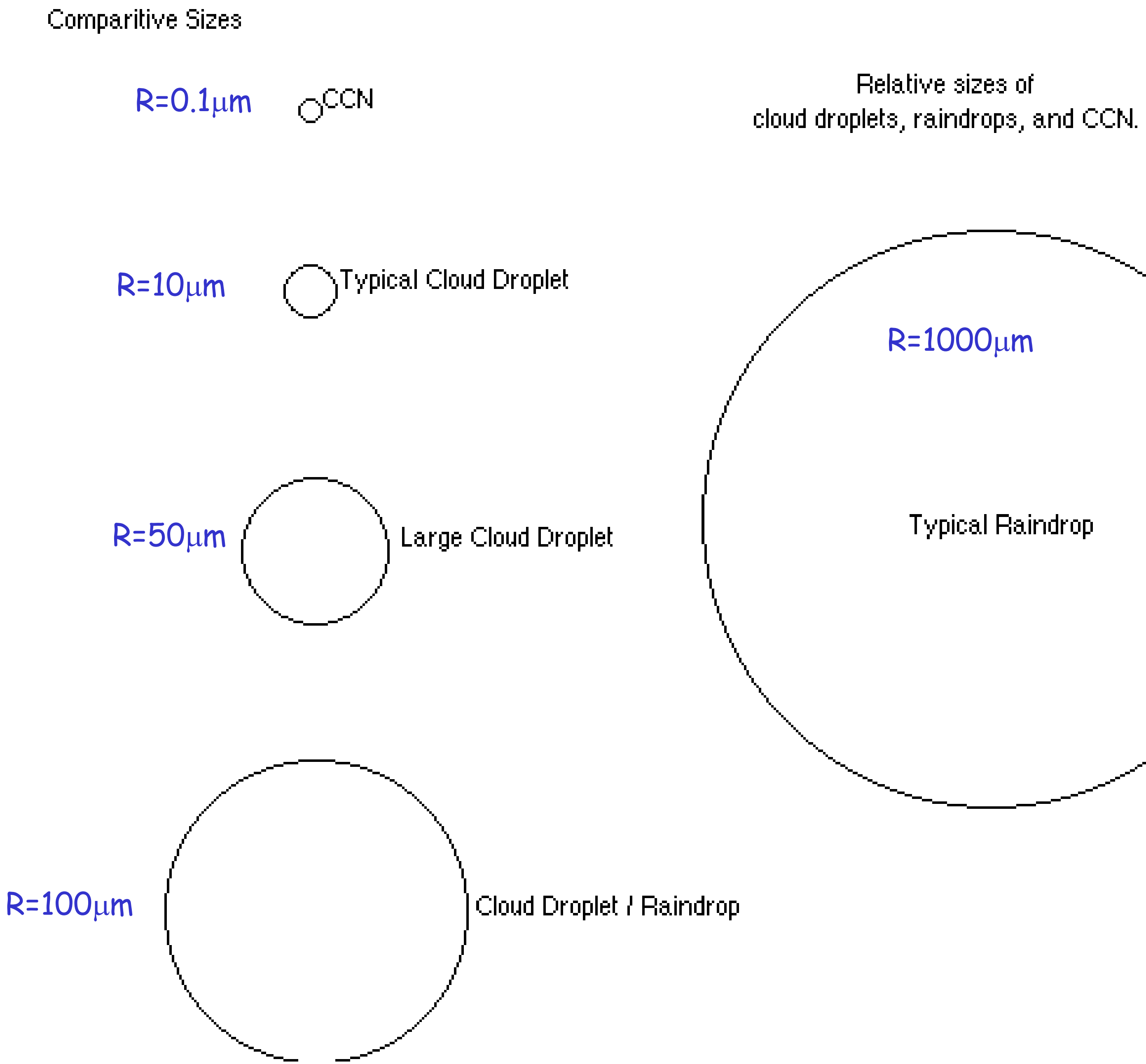
The droplet shrinks by evaporation

Interaction of Droplets with different e_s



vapor is transported from a to b

Fig. 8


Comparitive Sizes
R=0.1μm
CCN
Relative sizes of
cloud droplets, raindrops, and CCN.
R=10μm
Typical Cloud Droplet
R=1000μm
R=50μm
Large Cloud Droplet
Typical Raindrop
R=100μm
Cloud Droplet / Raindrop

Ice Crystal Growth

- Growth by **deposition**: Water vapor condenses onto an IN particle and freezes.
- Growth by **contact**: IN initiates an ice crystal at the moment it contacts a supercooled water droplet.
- Growth by **freezing**: IN floating within a supercooled droplet initiates the freezing.

- **Accretion** of cloud droplets: ice-phase particle has grown to a sufficient size to begin to fall and collect the supercooled droplet,
- **Aggregation:** ice particles coming together to form one main snowflake

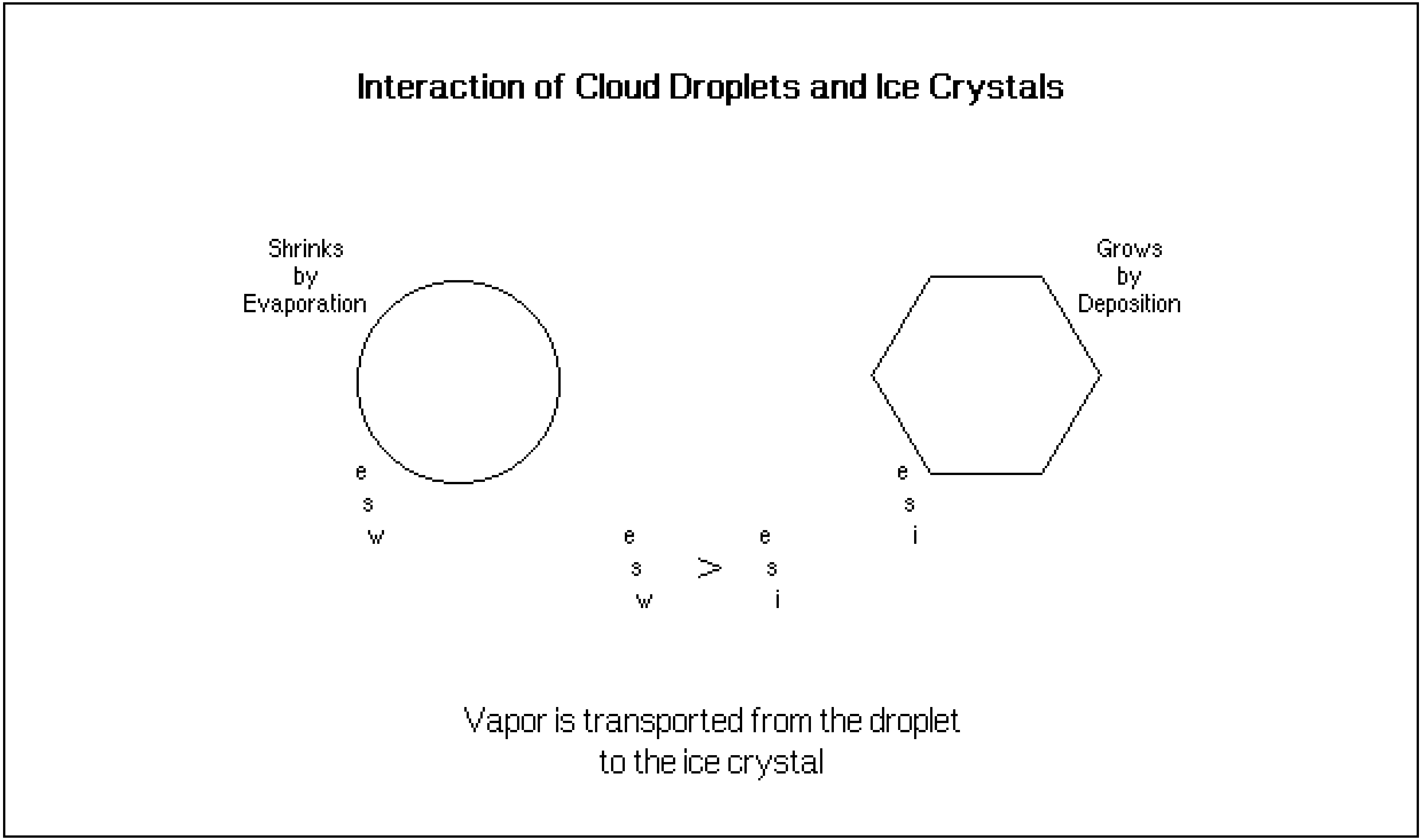
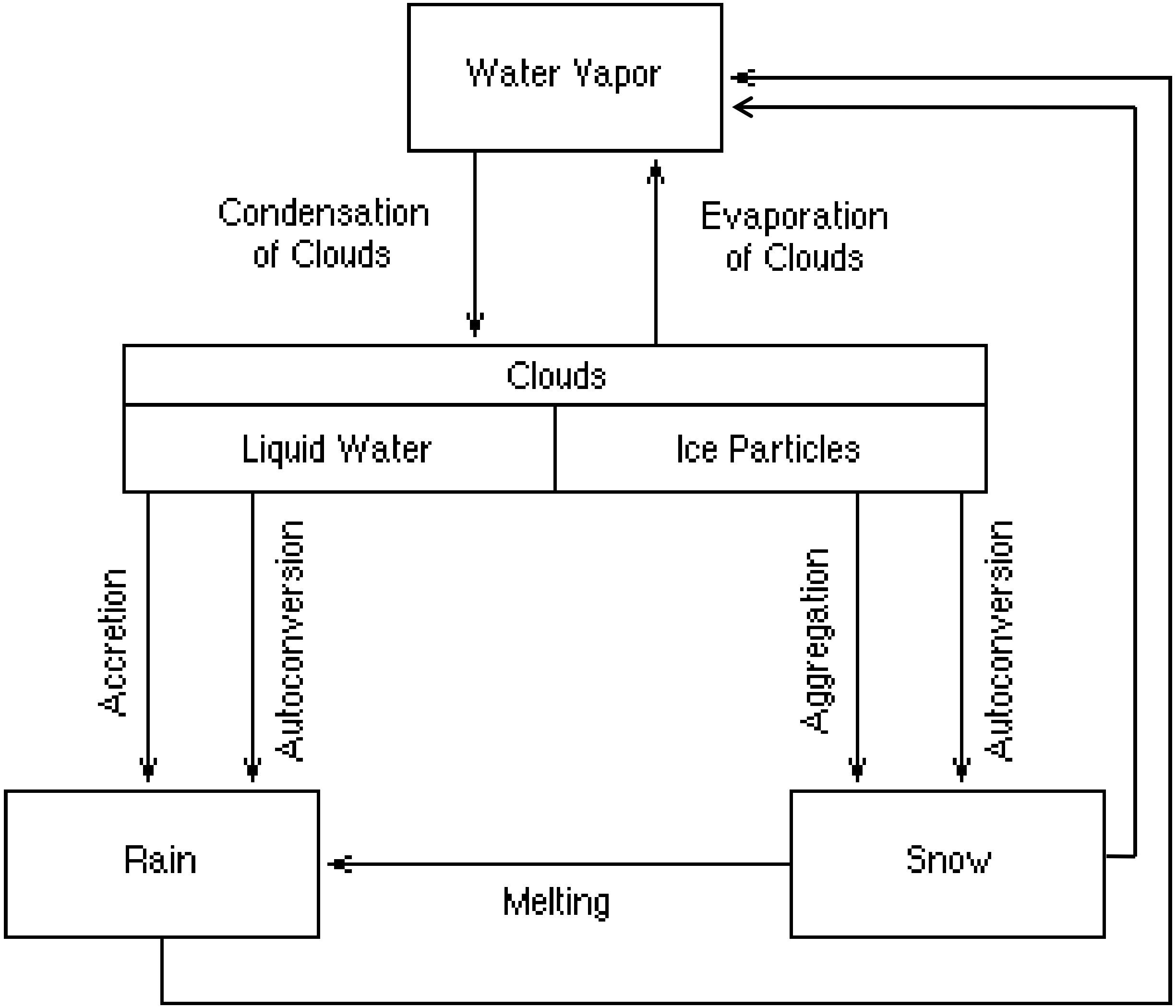

Interaction of Cloud Droplets and Ice Crystals
Shrinks by Evaporation
Grows by Deposition
e s w
e s i
e s w > e s i
Vapor is transported from the droplet to the ice crystal

Fig. 15

conversions

Water Vapor (Qv)	Snow (Qs)	Rain (Qr)	Cloud (Qc)	Ice (Qi)
condensation	sublimation	accretion	accretion	initiation
evaporation	accretion	autoconversion	autoconversion (into rain)	sublimation
ice initiation	autoconversion (from ice)	evaporation	condensation	deposition
deposition (onto ice crystals)	freezing (from rain)	freezing (to snow)	freezing (to ice)	melting (to cloud)
	melting (to rain)	melting (from snow)	melting (from ice)	freezing cloud
				accretion
				autoconversion



Zhao et al 1997

Zhao et al.

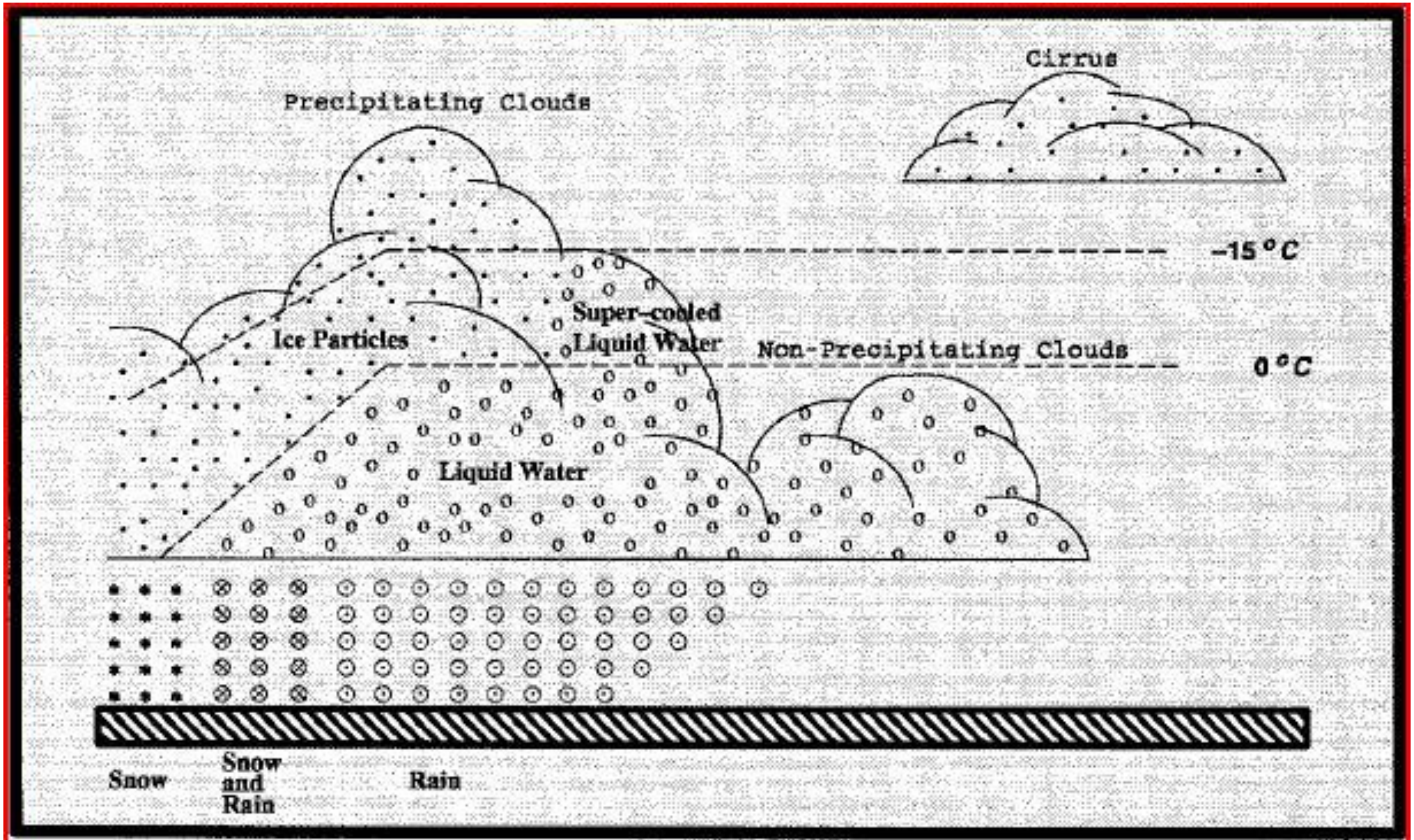


FIG. 2. Distribution of cloud water and cloud ice inside clouds.

Fração de cobertura de nuvens *b*

in Eq. (2.3). According to Sundqvist et al. (1989), cloud fraction b at a grid point can be estimated from relative humidity using the equation

$$b = 1 - \left(\frac{U_s - U}{U_s - U_{00}}\right)^{1/2} \qquad (2.11)$$

for $U > U_{00}$ and $b = 0$ for $U < U_{00}$. Since both temperature

Us=1.0

U00=0.85

Precipitação *P*

Following Sundqvist et al. (1989), the autoconversion of cloud water to rain, P_{raut}, can be parameterized from the cloud water mixing ratio m and cloud fraction b; that is,

$$P_{\mathrm{raut}} = c_0 m \left\{ 1 - \exp\left[-\left(\frac{m}{m_r b} \right)^2 \right] \right\}, \qquad (2.15)$$

Taxa de precipitação por autoconversão da água da nuvem

where constants c_0 and m_r are 1.0×10^{-4} s^{-1} and 3.0×10^{-4}, respectively. The autoconversion of cloud ice to snow is simulated using the equation from Lin et al. (1983):

$$P_{\mathrm{saut}} = a_1 (m - m_{i0}), \qquad (2.16)$$

Neve por autoconversão do gelo da nuvem

The collection of cloud liquid water by the falling rain is proportional to the cloud water mixing ratio m and the rain rate P_r and can be expressed by

$$P_{\mathrm{racw}} = C_r m P_r, \qquad (2.18)$$

Taxa de precipitação por coleta de gotas de nuvem pela chuva em queda

5.0×10^{-4} m^2 kg^{-1} s^{-1}. Similarly, the aggregation process of ice particles by the falling snow is simulated by

$$P_{\mathrm{saci}} = C_s m P_s, \qquad (2.19)$$

where P_s is the precipitation rate of snow and C_s is the collection coefficient. Unlike C_r, C_s should be a function of temperature since the open structure of ice crystals at relatively warm temperatures increases the likelihood

Taxa de neve por agregação cristais de gelo com a neve em queda

$$P_{\mathrm{sm1}} = C_{\mathrm{sm}}(T - 273.15)^a P_s,$$

ter is given by (Zhao and Carr

$$P_{\mathrm{sm2}} = C_{\mathrm{ws}} P_{\mathrm{sacw}},$$

is the collection rate of cloud li

g snow below the freezing level

$$P_{\mathrm{sacw}} = C_r m P_s$$

Chuva por derretimento da neve em queda

Chuva por derretimento da neve formada por coleta de água de nuvem

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Neve por coleta de agua de nuvem pela neve

Evaporation of rain as it falls through an unsaturated layer is calculated using the equation (Sundqvist 1988)

$$E_{rr} = k_e(U_{00} - U)(P_r)^{\beta}. \qquad (2.24)$$

The sublimation of the falling snow is also computed using the equation

$$E_{rs} = [C_{rs1} + C_{rs2}(T - 273.15)]\frac{(U_{00} - U)}{U_{00}}P_s. \qquad (2.25)$$

Evaporação *E* *da chuva ou neve, em camadas subsaturadas subnuvem*

$$P_r^{(n)} = P_r^{(n-1)} + \frac{P_{\text{sfc}} - P_{\text{top}}}{g\eta_s}(P_{\text{raut}} + P_{\text{racw}} + P_{\text{sacw}} + P_{\text{sm1}} + P_{\text{sm2}} - E_{\text{rr}})^{(n)}\Delta\eta \qquad (2.26)$$

and

$$P_s^{(n)} = P_s^{(n-1)} + \frac{P_{\text{sfc}} - P_{\text{top}}}{g\eta_s}(P_{\text{saut}} + P_{\text{sact}} - P_{\text{sm1}} - P_{\text{sm2}} - E_{\text{rs}})^{(n)}\Delta\eta, \qquad (2.27)$$

Chuva e neve