Moist processes in the atmosphere:

From simple concepts to sophisticated parameterizations

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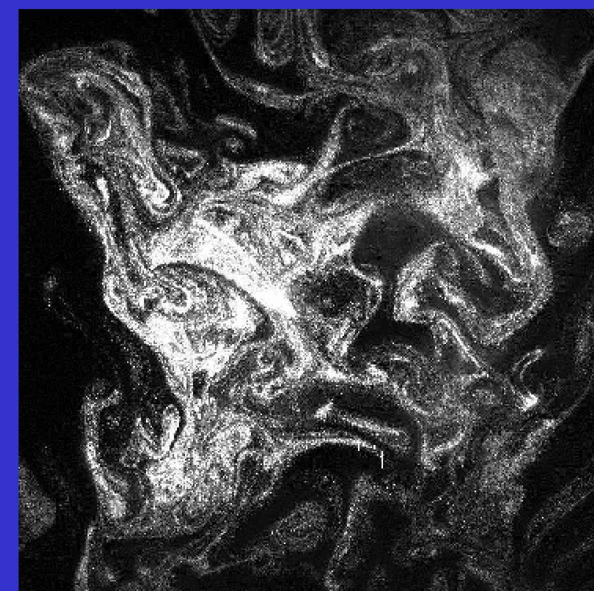
Earth
in visible light

1,000 km

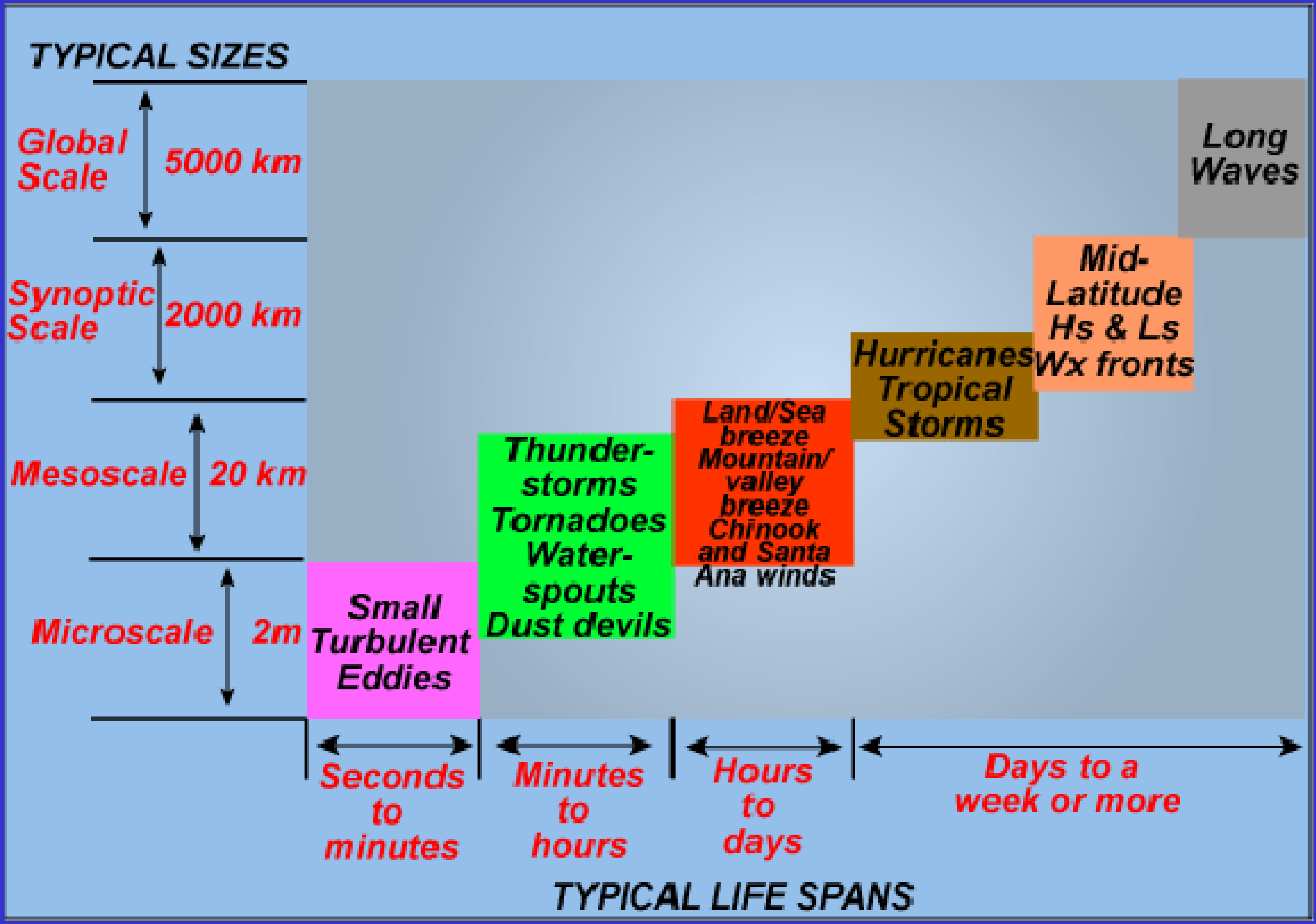
Small cumulus
clouds

Mixing in laboratory
cloud chamber

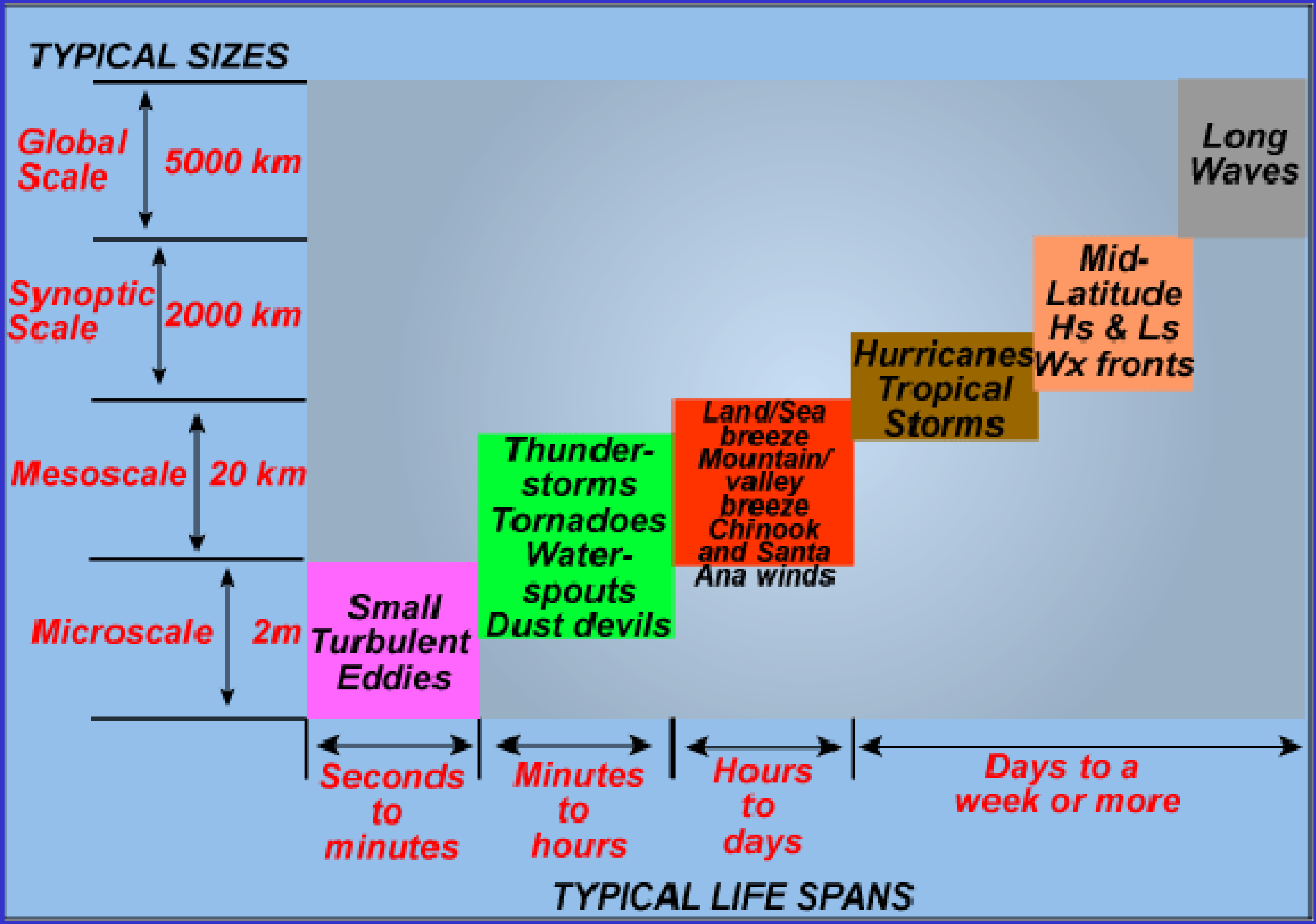


10 cm

Why is it so hard to simulate the Earth climate system?



Why is it so hard to simulate the Earth climate system?



Because some of the key processes are even not on this diagram….

Kiehl and Trenberth, BAMS 1997

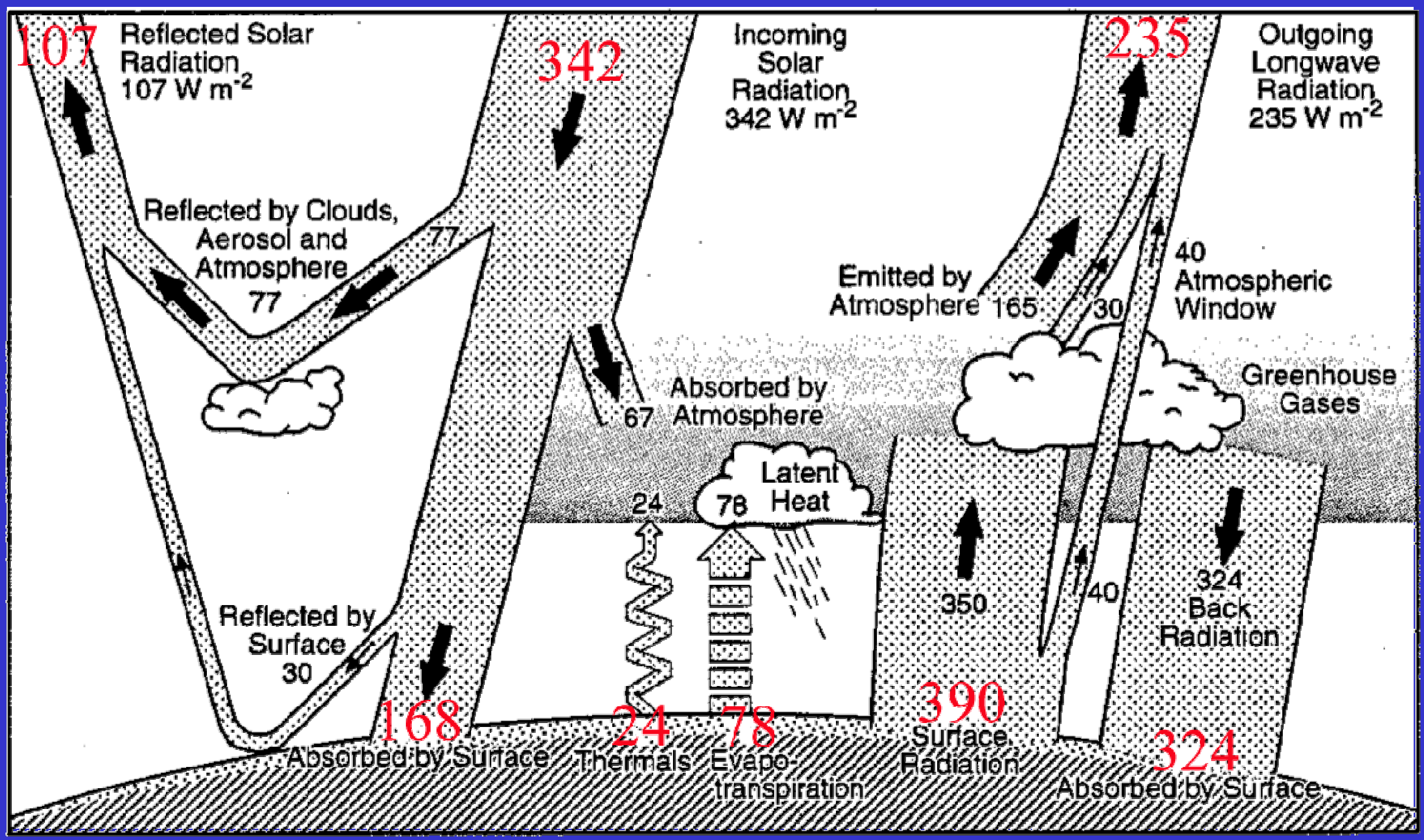
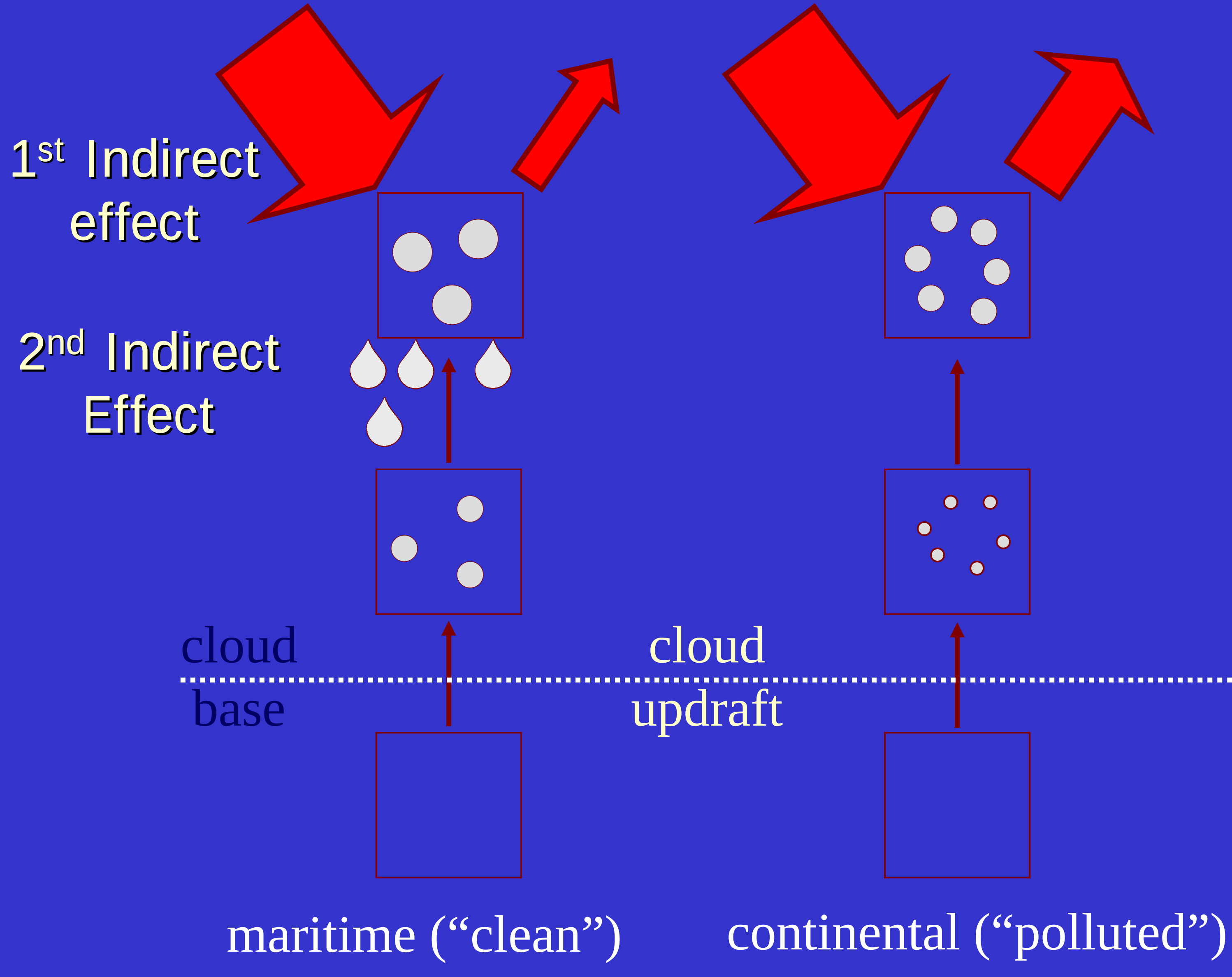
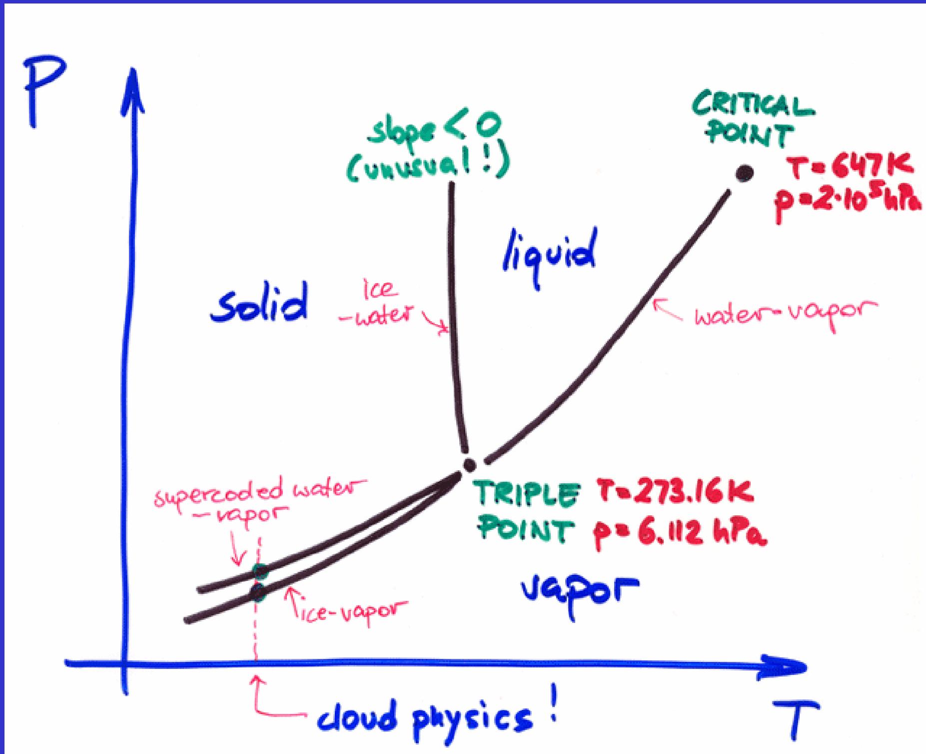


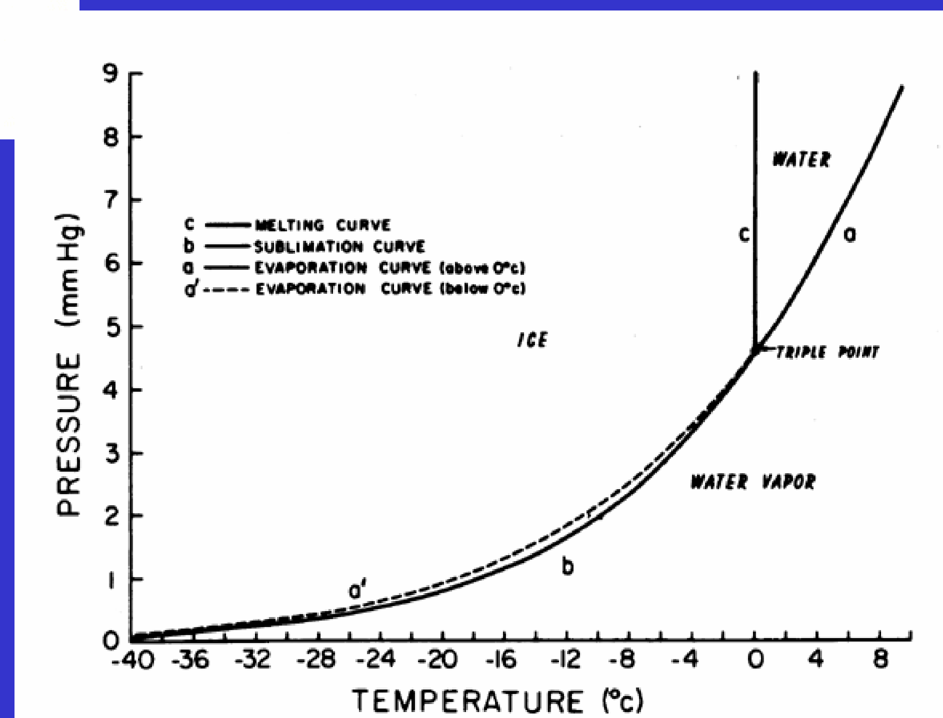
FIG. 7. The earth's annual global mean energy budget based on the present study. Units are W m^{-2}.

Indirect aerosol effects



p-T phase equilibrium diagram for water substance





ELEMENTARY CLOUD PHYSICS:

clouds form due to cooling of air (e.g., adiabatic expansion of a parcel of air rising in the atmosphere)

- *condensation:* water vapor $\longrightarrow$ cloud droplets

heterogeneous nucleation on atmospheric aerosols called Cloud Condensation Nuclei (CCN); typically highly soluble salts (sea salt, sulfates, ammonium salts, nitrates)

only a very small percentage of CCN used by clouds (i.e., water clouds form just above saturation)

ELEMENTARY CLOUD PHYSICS, cont.:

- *formation of ice particles*

heterogeneous nucleation on atmospheric aerosols called Ice-forming Nuclei (IN); dominates for temperatures higher than about -40 deg C (233 K); dominated by the so-called contact mode where a supercooled droplet freezes upon contact with IN;

IN are typically silicate particles (clays) or other compounds with crystallographic lattice similar to ice, highly insoluble

IN are scarce, their number depends exponentially on temperature (typically, 1 per liter at -20 deg C, 10 per liter at -25 deg C).

homogeneous freezing is possible once droplet temperature is smaller than about -40 deg C.

From cloud droplets and ice crystals to precipitation:

WARM RAIN:

$\longrightarrow$ gravitational collision and coalescence between cloud droplets

ICE PROCESSES:

$\longrightarrow$ Findeisen-Bergeron process: water vapor pressure at saturation is lower over ice than over water; it follows that once ice crystal is formed from supercooled droplet, it grows rapidly through diffusion of water vapor at the expense of cloud droplets

$\longrightarrow$ riming: falling ice crystal collects supercooled droplets that freeze upon contact (graupel, hail, etc).

Modeling moist processes in the atmosphere:

- Gas dynamics for the air with moisture (i.e., containing water vapor, suspended small cloud particles, falling larger precipitation particles);
- Thermodynamics for the air containing water vapor (i.e., phase changes, latent heating, etc).

Gas dynamics for moist air:

- Water vapor is a minor constituent:

mass loading is typically smaller than 1%; thermodynamic properties (e.g., specific heats etc) only slightly modified;

- Suspended small particles (cloud droplets, cloud ice):

mass loading is typically smaller than a few tenths of 1%, particles are much smaller than the smallest scale of the flow; multiphase approach is not required

- Precipitation (raindrops, snowflakes, graupel, hail):

mass loading can reach a few %, particles are larger than the smallest scale of the flow; multiphase approach needed only for very-small-scale modeling (e.g., DNS).

In the spirit of the Boussinesq approximation, moisture and condensate affect gas dynamics equations only though the buoyancy term

$$\frac{d\mathbf{u}}{dt} = -\frac{1}{\rho}\nabla p - g\mathbf{k} + \ldots(\text{Coriolis, turbulence, etc})$$

$$\rho = \rho_o(z) + \rho'$$

$$p = p_o(z) + p'$$

$$(\rho_o + \rho')\frac{dw}{dt} = -\frac{\partial p_o}{\partial z} - \rho_o g - \frac{\partial p'}{\partial z} - \rho' g + \ldots$$

$$\frac{d\mathbf{u}}{dt} = -\frac{1}{\rho_o}\nabla p' - g\mathbf{k}\frac{\rho'}{\rho_o} + \ldots$$

For small-Mach number flows ($|\mathbf{u}| \ll c_s$; c_s - speed of sound):

$$\frac{\rho'}{\rho_o} \approx -\frac{T'}{T_o}$$

$$\frac{d\mathbf{u}}{dt} = -\frac{1}{\rho_o}\nabla p' + g\mathbf{k}\frac{T'}{T_o} + \ldots$$

Density temperature T_d: the temperature dry air has to have to yield the same density as moist cloudy air

$$T_d = T\frac{1 + q/\epsilon}{1 + q + Q}$$

T - air temperature

q - water vapor mixing ratio ($\sim 10^{-3}$)

Q - condensate mixing ratio (cloud water, rain, ice, snow, etc.; $\sim 10^{-3}$)

$$\epsilon = \frac{R_d}{R_v} \approx 0.622$$

$$T_d \approx T\left[1 + \left(\frac{1}{\epsilon} - 1\right)q - Q\right]$$

$$T_d \approx T\,(1 + 0.61q - Q)$$

T, *q* and *Q* – thermodynamics (and much more!)

Thermodynamics:

Moist air is treated as a perfect gas

Phase changes lead to the release of latent heat and formation of condensed (liquid or solid) phase of the water substance (cloud droplets, raindrops, ice crystals, snow, etc)

Condensed phase is treated as continuous medium, i.e., described as density (of cloud droplets, raindrops, etc).

In practice, variables most often used to describe water vapor and condensate are not densities, but mixing ratios, i.e., densities normalized by the air density.

$$\frac{\partial \rho_a}{\partial t} + \nabla(\rho_a \mathbf{u}) = 0 \quad \text{or} \quad \frac{d\rho_a}{dt} + \rho_a \nabla \mathbf{u} = 0$$

$$\frac{\partial \rho_v}{\partial t} + \nabla(\rho_v \mathbf{u}) = 0 \quad \text{or} \quad \frac{d\rho_v}{dt} + \rho_v \nabla \mathbf{u} = 0$$

$$\text{mixing ratio}: \quad Q = \frac{\rho_v}{\rho_a}$$

$$\frac{dQ}{dt} = \frac{1}{\rho_a}\frac{d\rho_v}{dt} - \frac{\rho_v}{\rho_a^2}\frac{d\rho_a}{dt} \equiv 0$$

First Law of Thermodynamics:

$$dq = du + p\,dv \tag{1}$$

dq - heat (per unit mass) added to the system
du - increase of internal energy (per unit mass)
$p\,dv$ - work (per unit mass) performed by the system

$$du = c_v\,dT, \quad pv = RT, \quad v = 1/\rho, \quad c_v + R = c_p$$

$$dq = c_p\,dT - \frac{RT}{p}dp \tag{2}$$

Indtroducing *potential temperature* as:

$$\theta = T\left(\frac{p_{oo}}{p}\right)^{R/c_p} \tag{3}$$

where p_{oo}=const (typically 1000 mb), (1) can be written as:

$$d\theta = \frac{\theta}{c_p T}dq \tag{4}$$

$$\frac{d\theta}{dt} = \frac{\theta}{c_p T} S$$

where $S = \dfrac{dq}{dt}$ in the heat source per unit mass [in J kg^{-1} s^{-1}]

$S = 0$ - adiabatic motions

$S \neq 0$ - motions with diabatic processes (heating due to radiative transfer, phases changes, chemical reactions, etc)

For phase changes of water substance:

$$S = L\frac{dQ}{dt}$$

where L is the latent heat (of condensation, freezing, or sublimation), and $\dfrac{dQ}{dt}$ is the change of corresponding water mixing ratio

BULK MODEL OF CONDENSATION:

$$\frac{d\theta}{dt} = \frac{L_v\theta}{c_pT}C_d$$

$$\frac{dq_v}{dt} = -C_d$$

$$\frac{dq_c}{dt} = \quad C_d$$

θ - potential temperature
q_v - *water vapor* mixing ratio
q_c - *cloud water* mixing ratio
L_v - latent heat of condensation/evaporation
C_d - condensation rate
Note: θ/T function of pressure only ($\approx \theta_o/T_o$)

C_d is defined such that cloud is always at saturation, which is a very good approximation:

$$q_c = 0 \quad \text{if} \quad q_v < q_{vs}$$

$$q_c > 0 \quad \text{only if} \quad q_v = q_{vs}$$

where $q_{vs}(p,T) \approx 0.622\frac{e_s(T)}{p}$ is the water vapor mixing ratio at saturation

WARM RAIN BULK MODEL (Kessler 1969):

$$\frac{d\theta}{dt} = \frac{L_v \theta}{c_p T}(C_d - EVAP)$$

$$\frac{dq_v}{dt} = -C_d + EVAP$$

$$\frac{dq_c}{dt} = C_d - AUT - ACC$$

$$\frac{dq_r}{dt} = \frac{1}{\rho}\frac{\partial}{\partial z}(\rho q_r v_t) + AUT + ACC - EVAP$$

θ - potential temperature
q_v - *water vapor* mixing ratio
q_c - *cloud water* mixing ratio
q_r - *rain water* mixing ratio
C_d - condensation rate
$EVAP$ - rain evaporation rate
AUT - "autoconversion" rate: $q_c \rightarrow q_r$
ACC - accretion rate: $q_c, q_r \rightarrow q_r$

$v_t(q_r)$ - rain terminal velocity (typically derived by assuming a drop size distribution; e.g., the Marshall-Palmer distribution $N(D) = N_o exp(-\Lambda D)$, $N_o = 10^7$ m^{-4}).

BIN-RESOLVING WARM RAIN MODEL:

$$\frac{d\theta}{dt} = \frac{L_v \theta}{c_p T} \sum_{i=1}^{N} C_d^{(i)}$$

$$\frac{dq_v}{dt} = -\sum_{i=1}^{N} C_d^{(i)}$$

for $i = 1, N$:

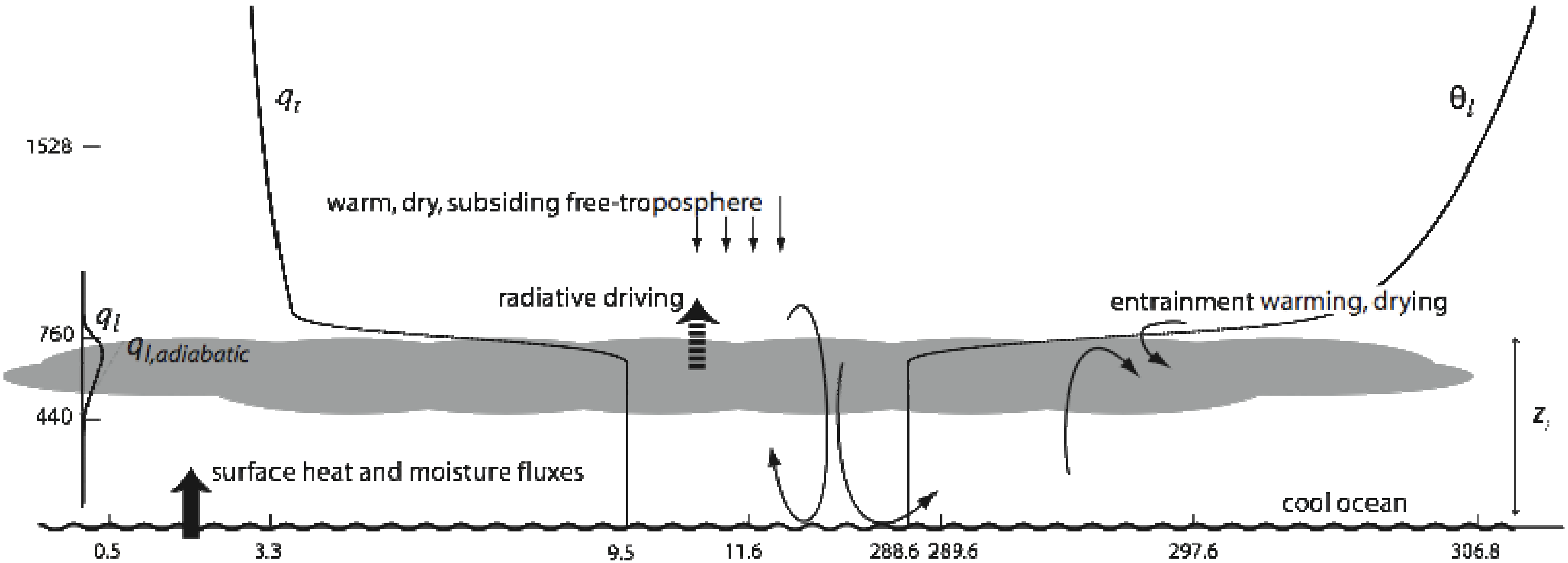
$$\frac{dq_c^{(i)}}{dt} = \frac{1}{\rho}\frac{\partial}{\partial z}\left[\rho q_c^{(i)} v_t(r^{(i)})\right] + \sum_{i=1}^{N} C_d^{(i)} + F_+^{(i)} - F_-^{(i)}$$

θ - potential temperature
q_v - *water vapor* mixing ratio
$q_c^{(i)}$ - *cloud water* mixing ratio for drops in size bin i ($i = 1.N$, $N \sim 100$)

$C_d^{(i)}$ - condensation/evaporation rate for drops in size bin i; depends on super/undersaturation $S = q_v/q_{vs} - 1$ and drop size $r^{(i)}$.

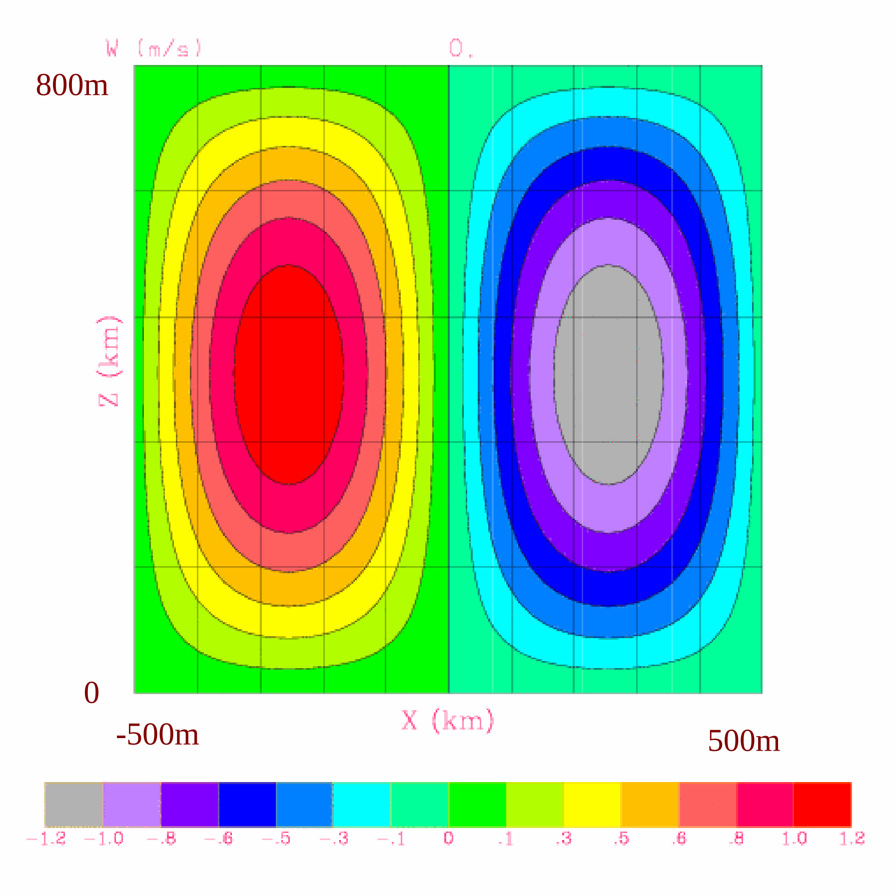
$F_+^{(i)}$ - source due to collisions between j and k resulting in drops in i
$F_-^{(i)}$ - sink due to collisions between i and all other bins

Modeling of indirect effects in Stratocumulus

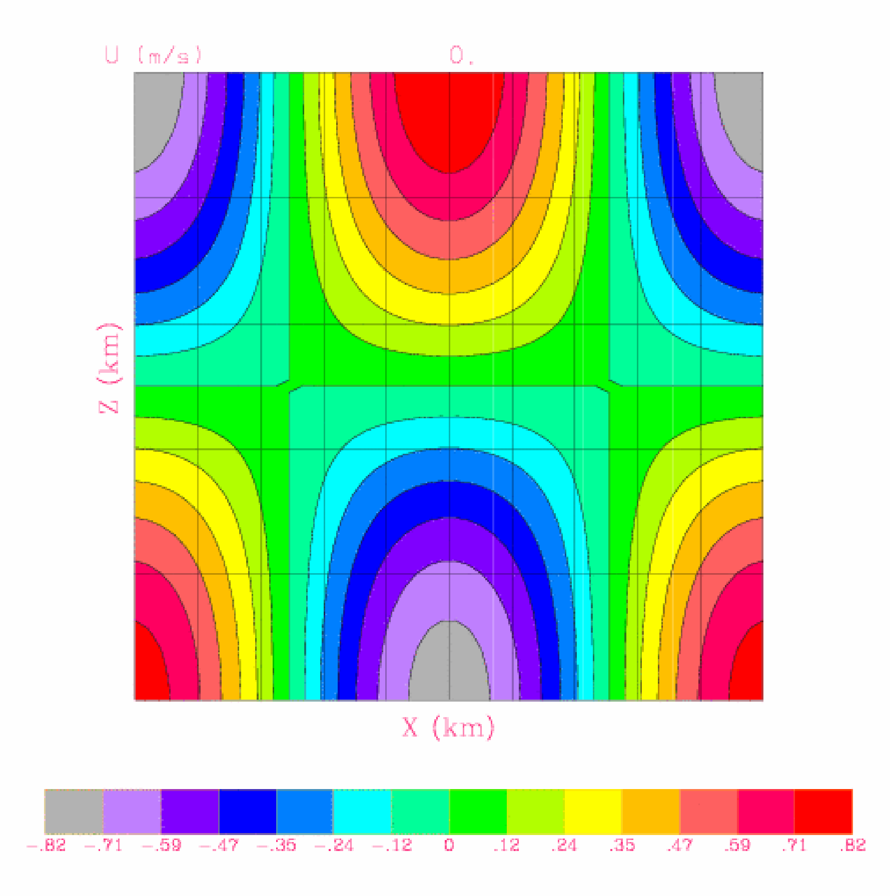


Stevens et al., 2006 , MWR

Kinematic (prescribed-flow) model of microphysical processes in Stratocumulus (2D: x-z)



Vertical velocity

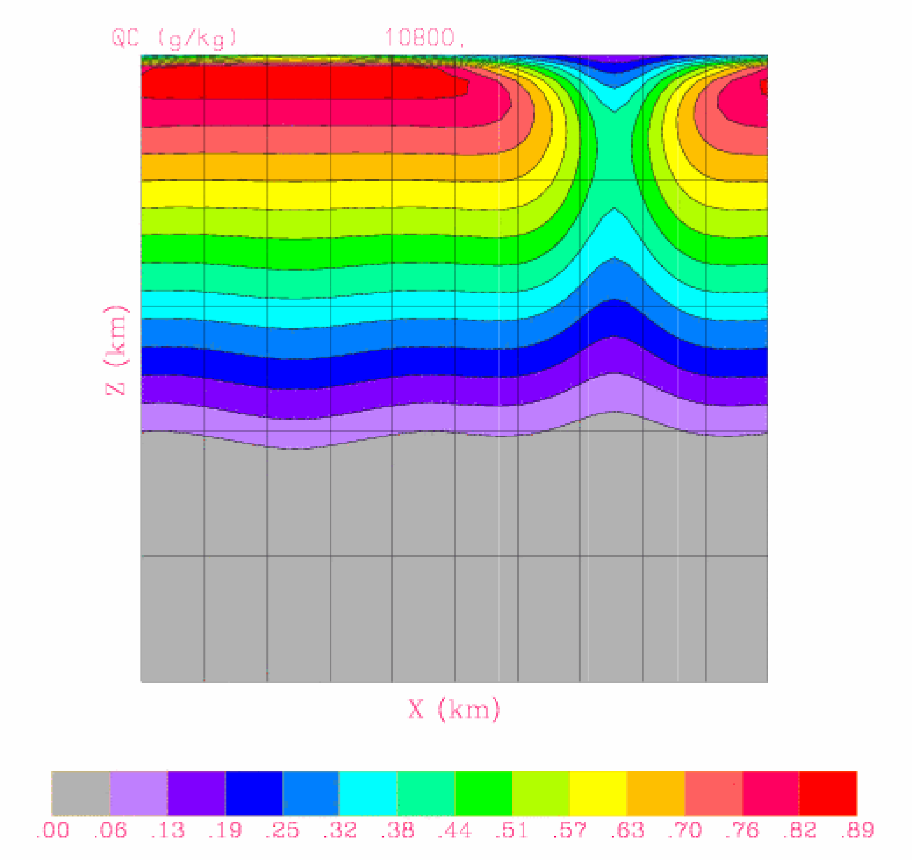


Horizontal velocity

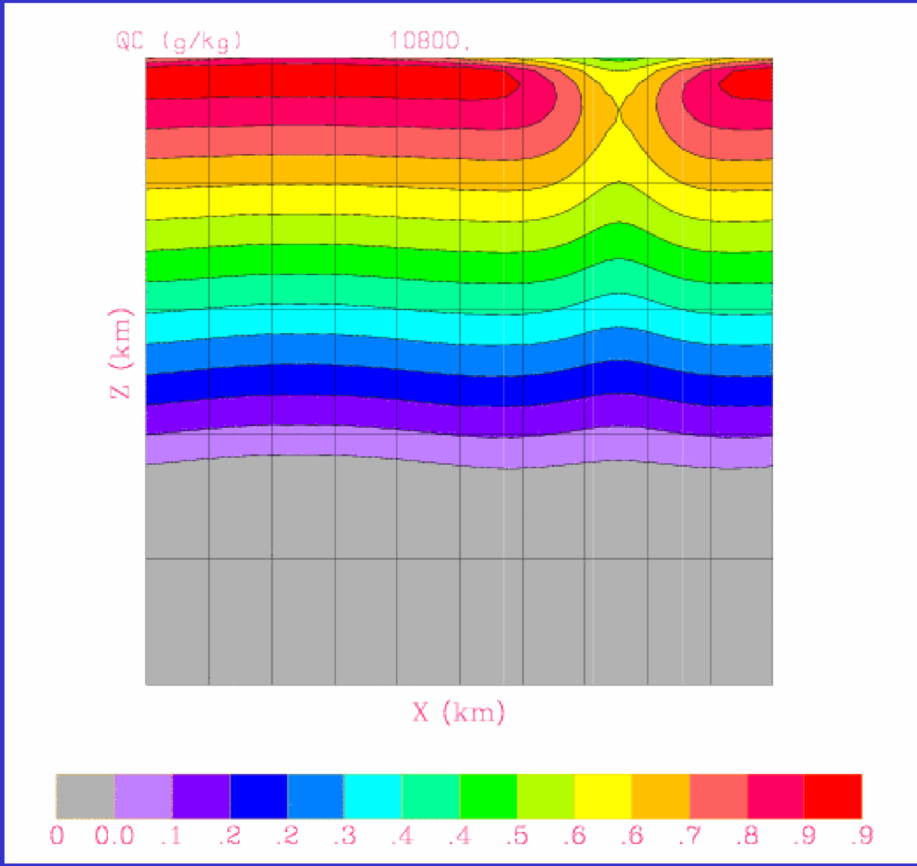
Run up to quasi-steady-state is obtained (typically couple hours)…

Cloud water (after 3hrs)

Maritime (clean)

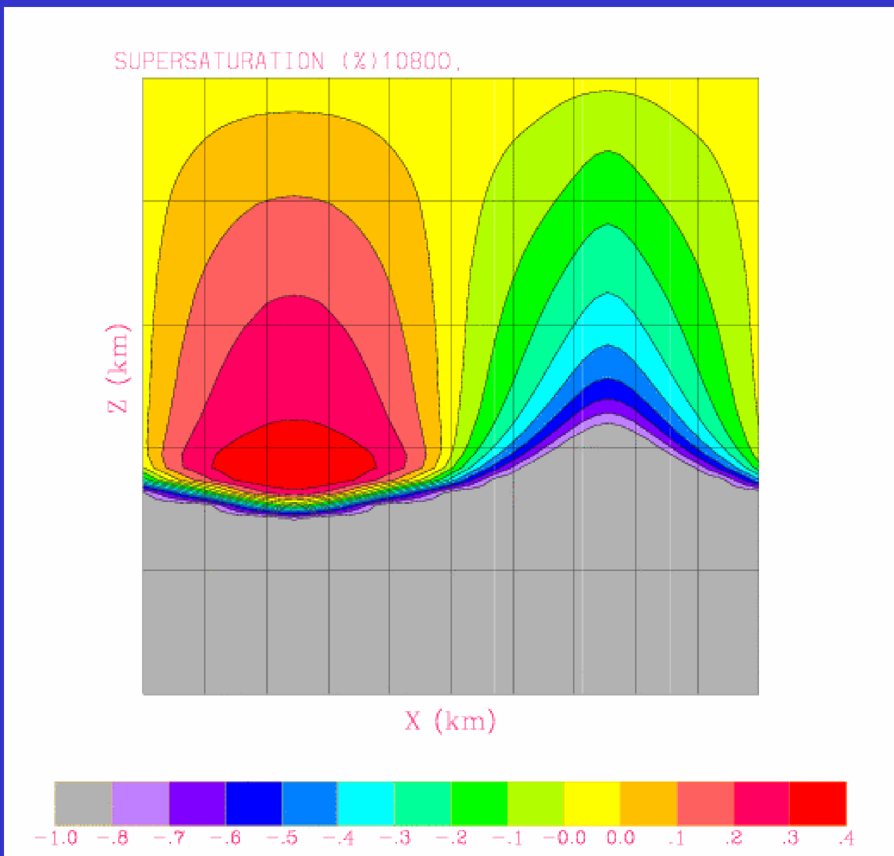


Continental (polluted)

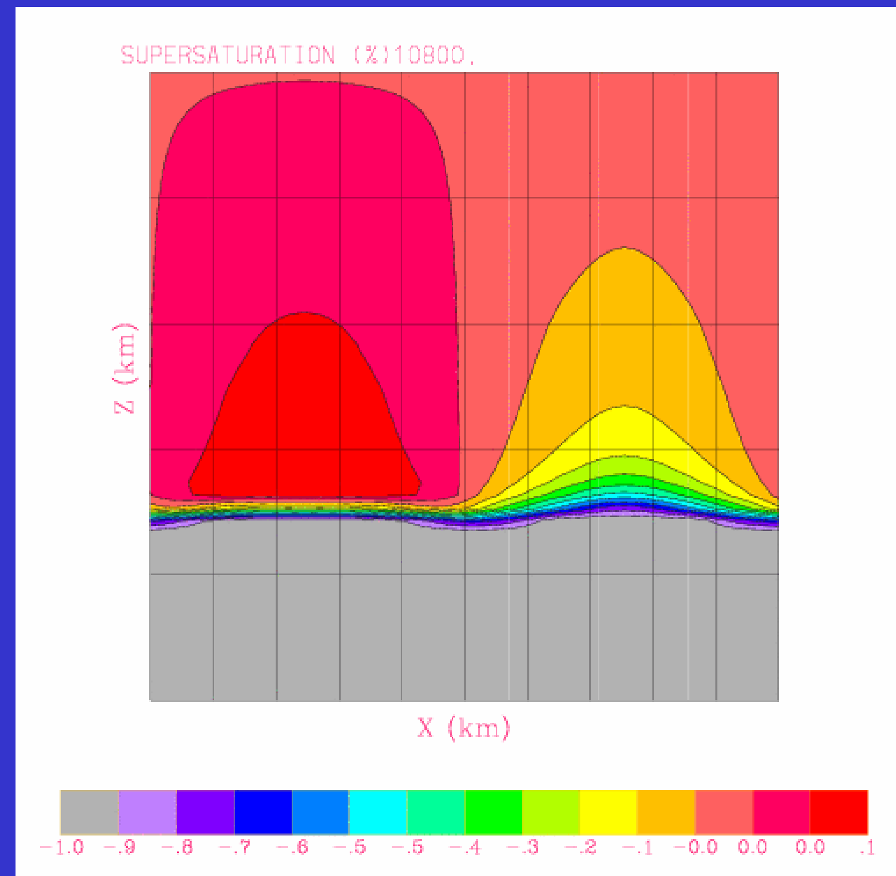


Supersaturation (after 3hrs)

Maritime (clean)

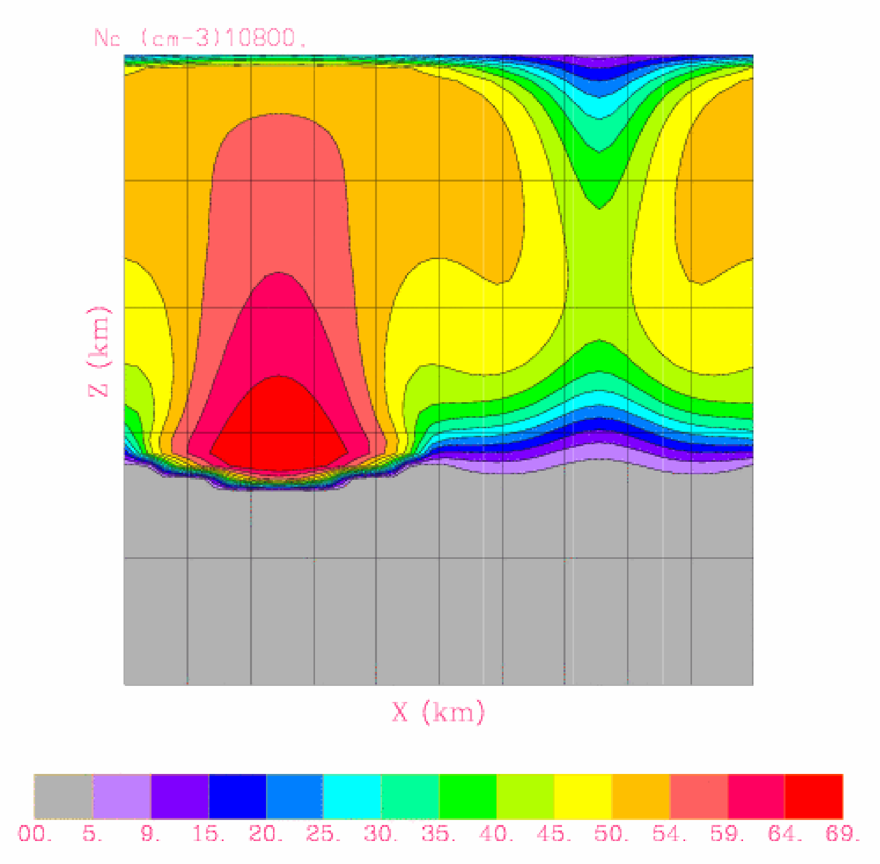


Continental (polluted)

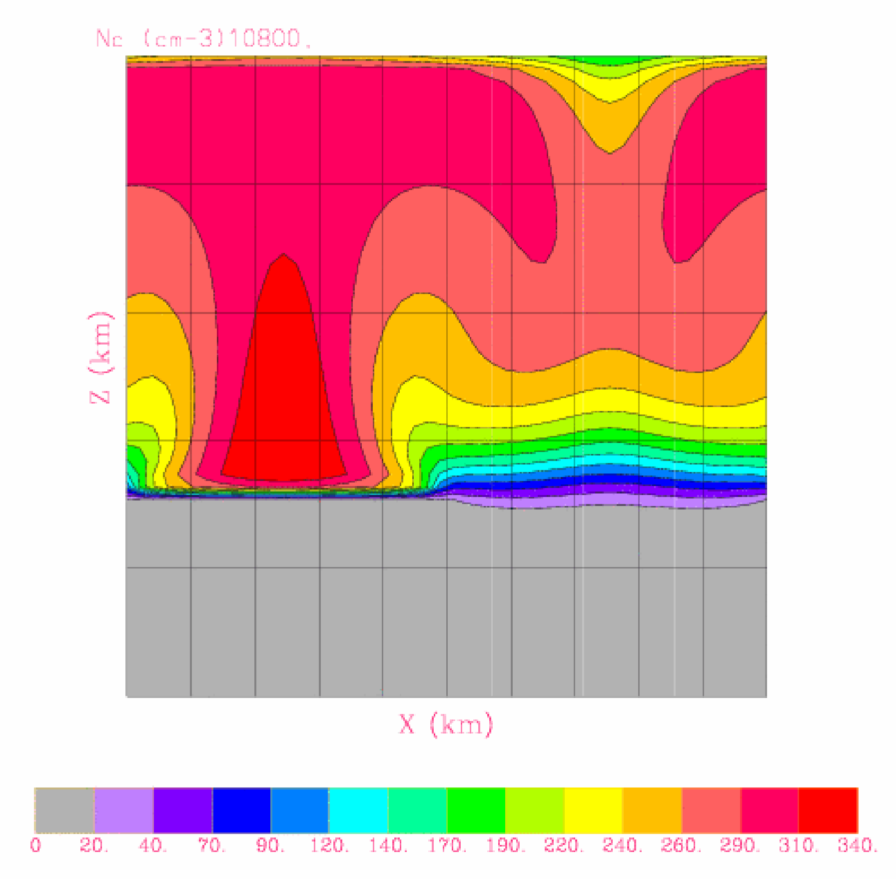


Cloud droplet (r < 20 microns) number concentration

Maritime (clean)

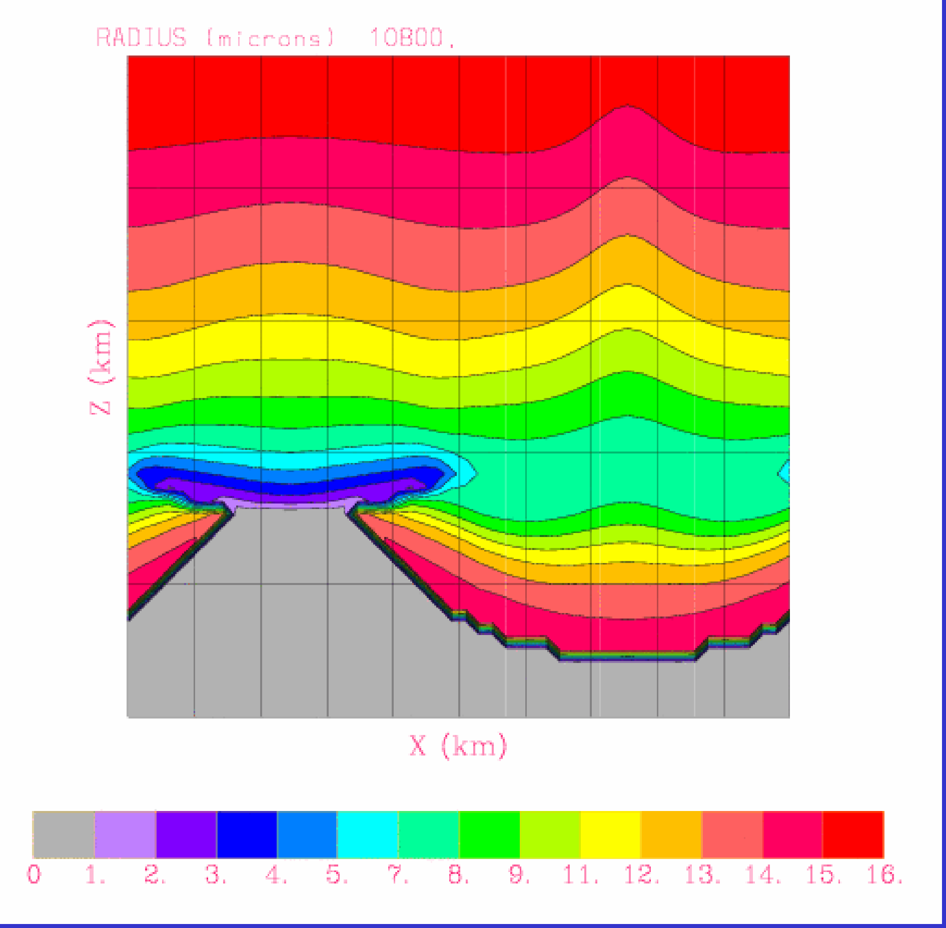


Continental (polluted)

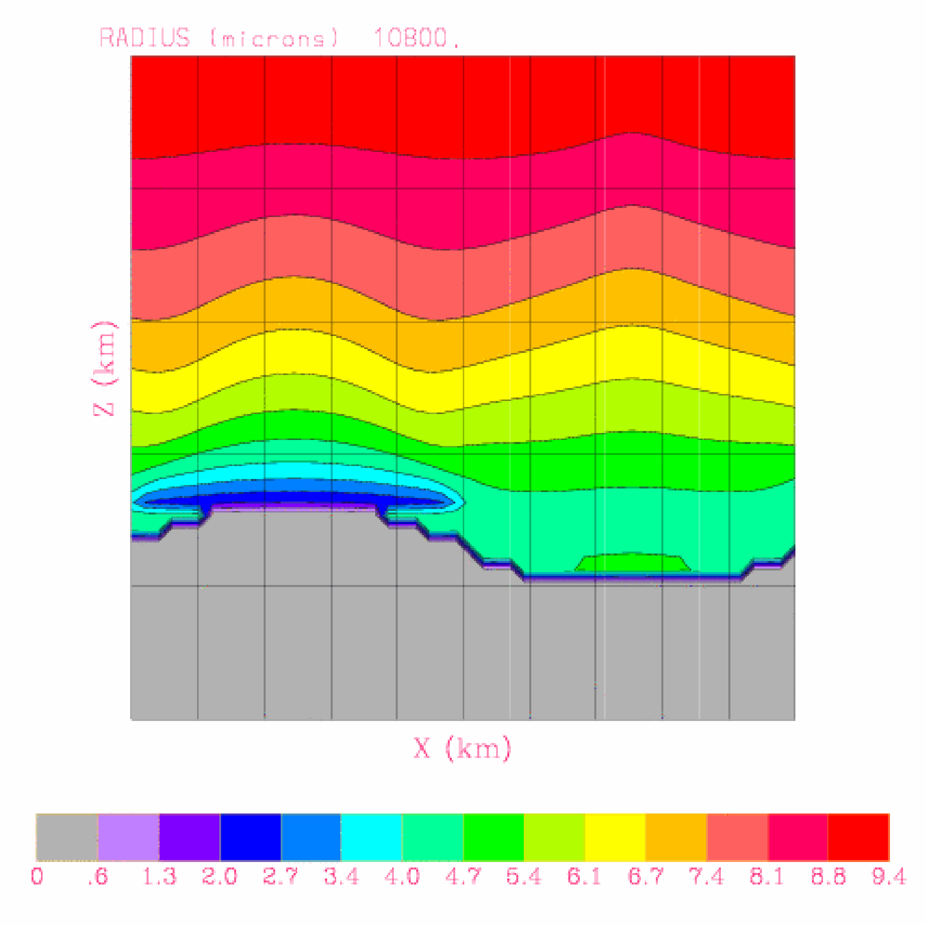


Cloud droplet (r < 20 microns) mean radius

Maritime (clean)

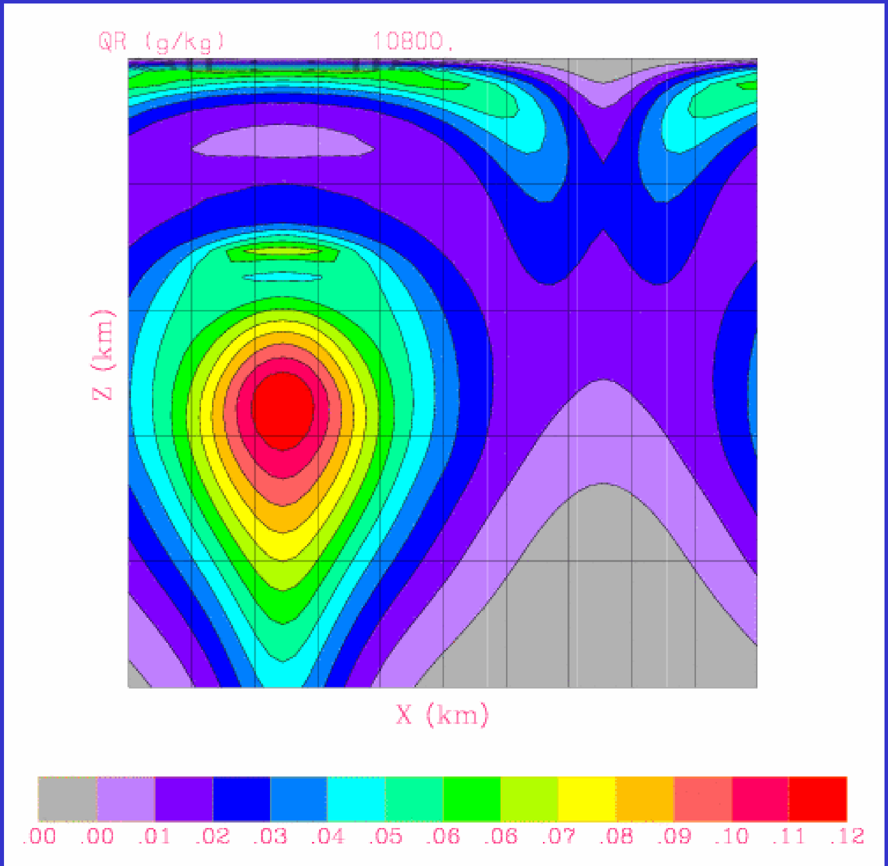


Continental (polluted)

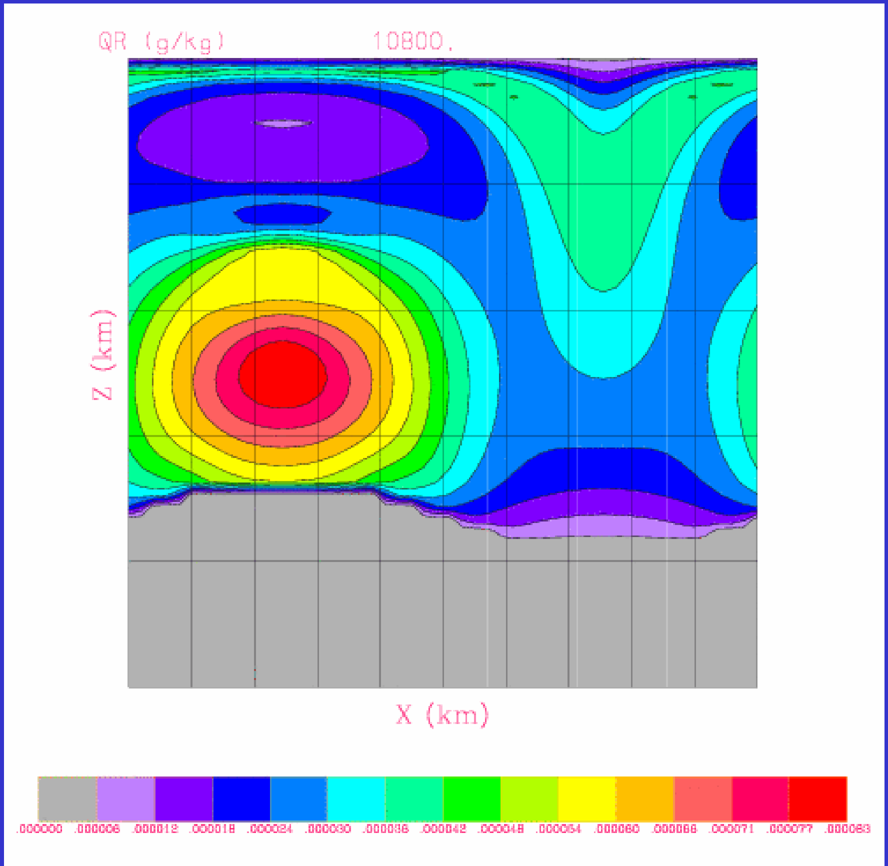


Drizzle (r > 25 microns) mixing ratio

Maritime (clean)

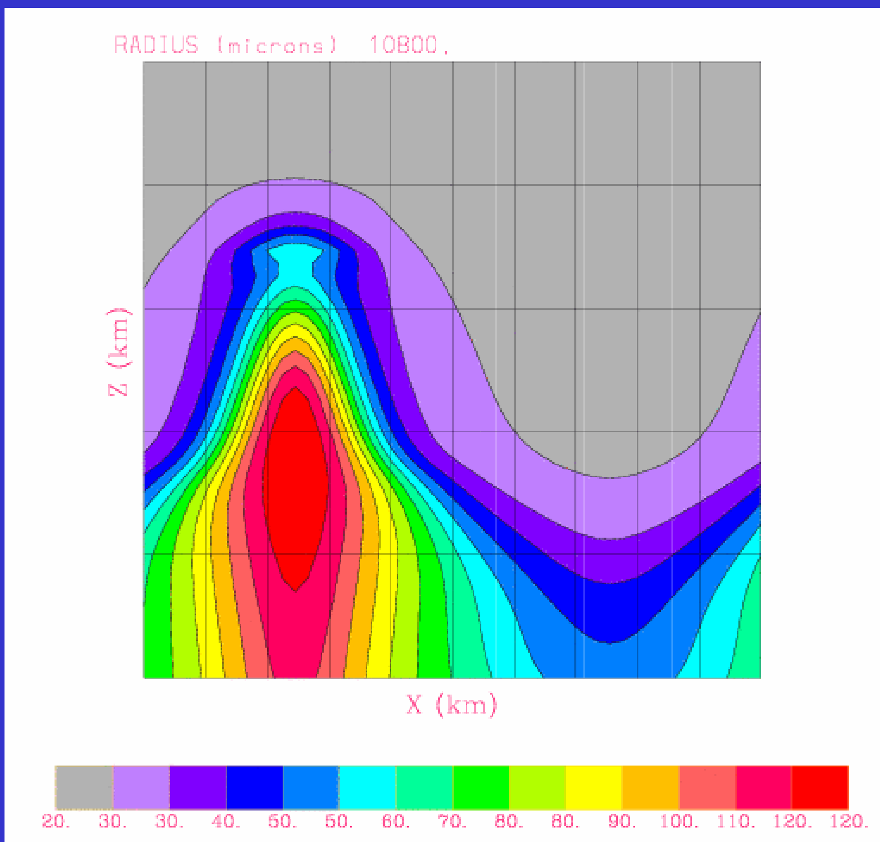


Continental (polluted)

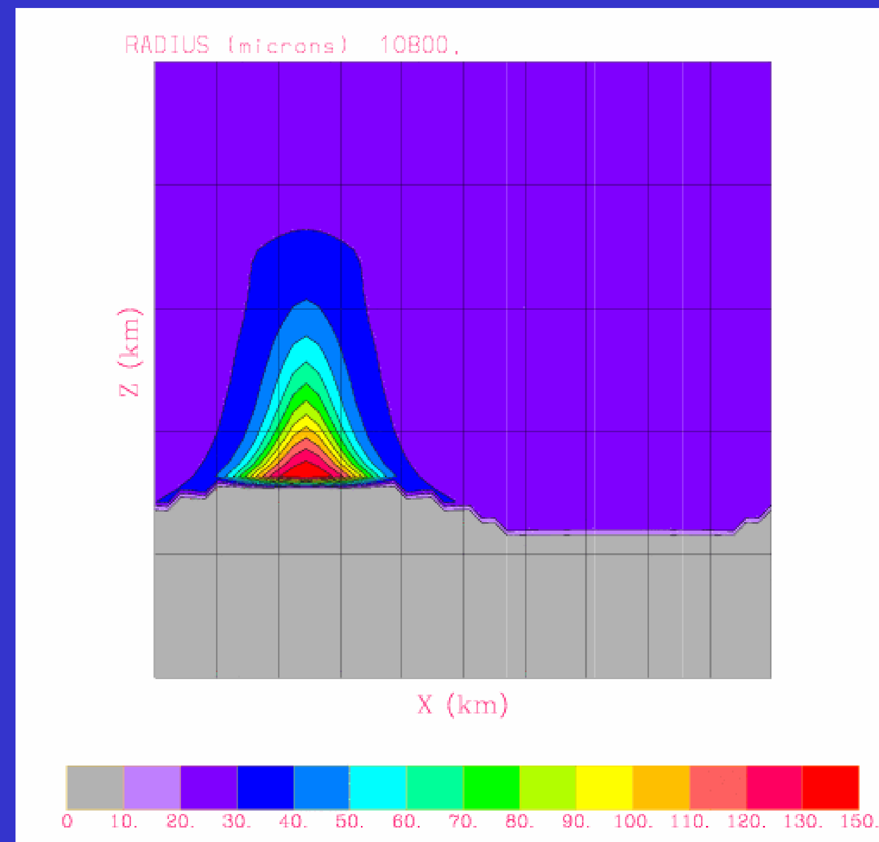


Drizzle (r > 25 microns) mean radius

Maritime (clean)



Continental (polluted)



Comments:

-Bulk warm rain microphysics is a relatively straightforward and computationally efficient approach (just 2 variables for the condensed water);

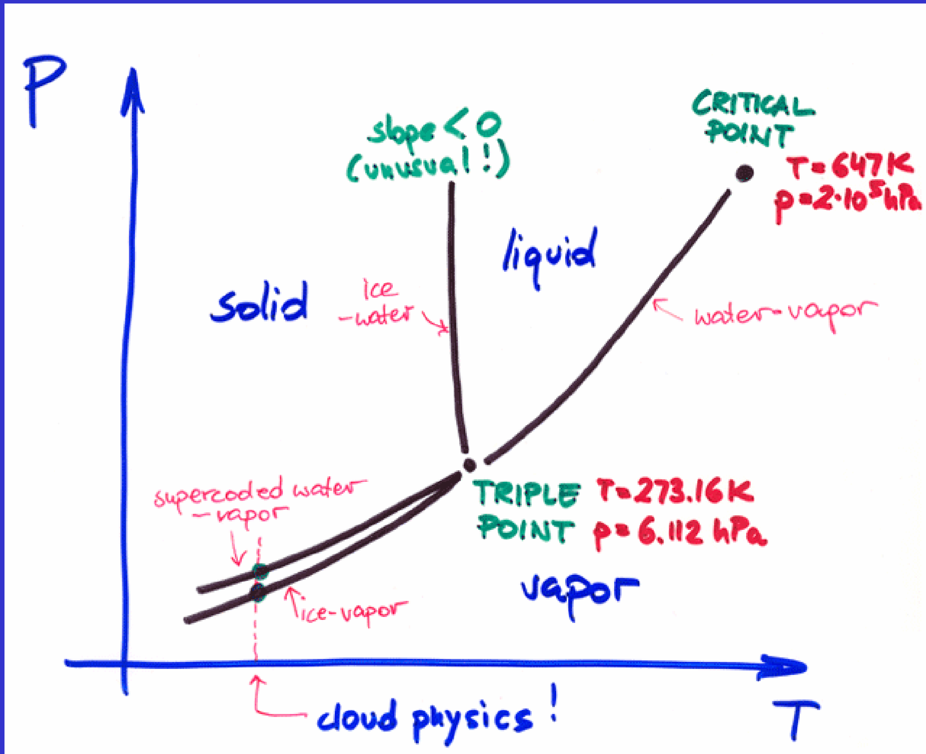
-Problems begin for shallow clouds when microphysical details decide whether precipitation develops or not (e.g., stratocumulus, shallow convection);

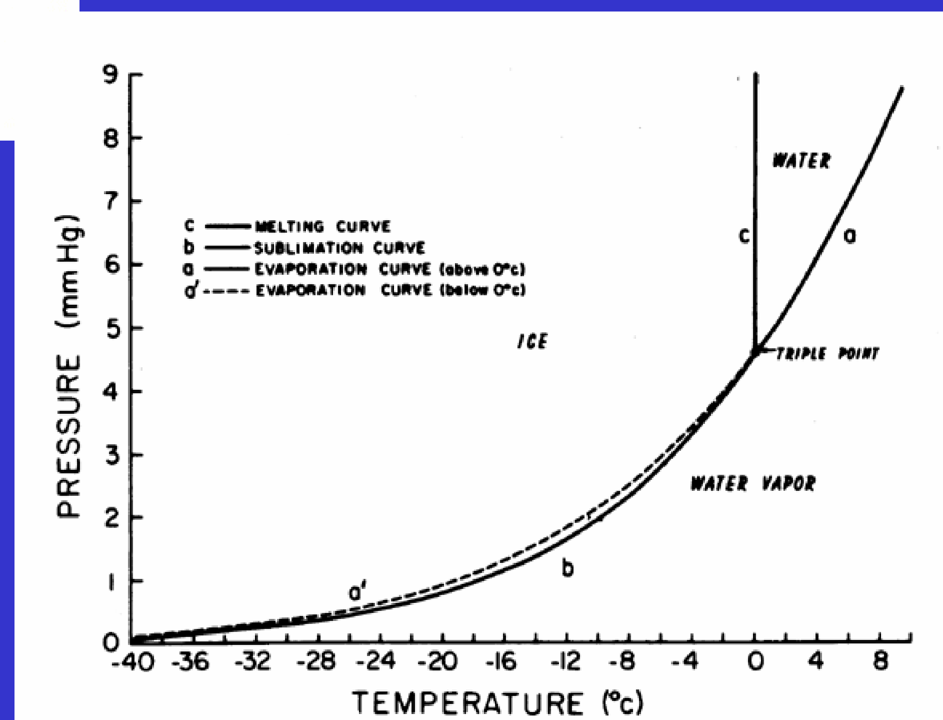
-When coupled to the radiative transfer, information about cloud droplet size is needed; bulk warm rain model is not able to provide this;

-Detailed (bin-resolved) microphysics solves the above two problems, but it is very expensive (~100 extra variables) and still leaves some issues (discussed later);

-A reasonable compromise is to predict both the mass and the number of particles (thus, 4 variables used for the condensed water); “two-moment bulk microphysics schemes”.

p-T phase equilibrium diagram for water substance





Ice processes:

Three main problems:

Ice processes:

Three main problems: ice initiation, ice initiation, and ice initiation;

(primary, secondary)

Ice processes:

Three main problems: ice initiation, ice initiation, and ice initiation;

(primary, secondary)

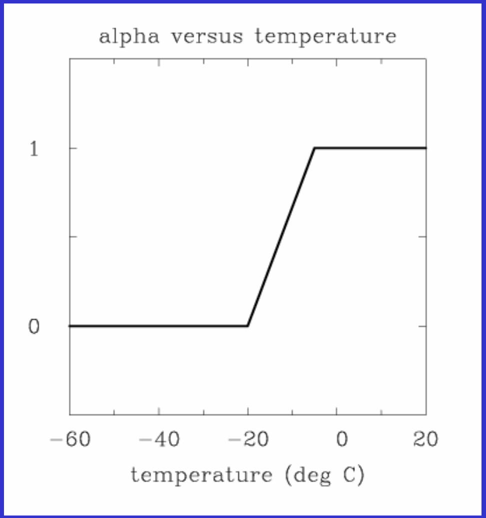
Equilibrium approach possible only in special cases (e.g., deep convection), questionable in shallow clouds (like arctic Scu);

Unlike warm-rain microphysics, where cloud droplets and rain/drizzle drops are well separated in the radius space, growth of ice phase is continuous in size/mass space. Both diffusional and acretional growth important;

Complexity of ice crystal shapes (“habits”).

Equilibrium approach

(simple extension of a warm-rain scheme)



SIMPLE BULK ICE MODEL (Grabowski 1998):

$$\frac{d\theta}{dt} = \frac{L_v \theta}{c_p T}(COND - DIFF)$$

$$\frac{dq_v}{dt} = -COND + DIFF$$

$$\frac{dq_c}{dt} = COND - AUTC - ACCR$$

$$\frac{dq_p}{dt} = \frac{1}{\rho}\frac{\partial}{\partial z}(\rho q_p v_t) + AUTC + ACCR - DIFF$$

θ - potential temperature
q_v - *water vapor* mixing ratio
q_c - *cloud condensate (water or ice)* mixing ratio
q_p - *precipitation water (rain or snow)* mixing ratio
$COND$ - condensation rate (saturation adjustment)
$DIFF$ - diffusional growth rate
$AUTC$ - "autoconversion" rate: $q_c \rightarrow q_p$
$ACCR$ - accretion rate: $q_c, q_p \rightarrow q_p$

saturation: $q_{vs} = \alpha q_{vw} + (1-\alpha) q_{vi}$

cloud water: $q_w = \alpha q_c$; cloud ice: $q_i = (1-\alpha) q_c$
rain: $q_r = \alpha q_p$; snow: $q_s = (1-\alpha) q_p$
$DIFF = DIFF_r + DIFF_s$
$AUTC = AUTC_r + AUTC_s$
$ACCR = ACCR_r + ACCR_s$
$v_t = \alpha v_t(q_r) + (1-\alpha) v_t(q_s)$

Non-equilibrium approach

Lin et al. 1983

Rutledge and Hobbs 1984

BULK MODEL WITH ICE MICROPHYSICS:

- potential temperature θ:

$$\frac{d\theta}{dt} = \frac{L_v \theta_e}{c_p T_e} S_1 + \frac{L_s \theta_e}{c_p T_e} S_2 + \frac{L_f \theta_e}{c_p T_e} S_3$$

- water vapor mixing ratio q_v:

$$\frac{dq_v}{dt} = S_v$$

- cloud condensate variables q_c^i, $i = 1, N_c$ (typically, $N_c = 2$: cloud water, cloud ice):

$$\frac{dq_c^i}{dt} = S_c^i$$

- precipitating water variables q_p^i, $i = 1, N_p$: (typically, $N_p = 3$: rain, snow, graupel/hail):

$$\frac{dq_p^i}{dt} = \frac{1}{\rho} \frac{\partial}{\partial z} \left(\rho q_p^i v_t^i \right) + S_p^i$$

S – various sources/sinks due to phase changes

Rutledge and Hobbs, JAS 1984

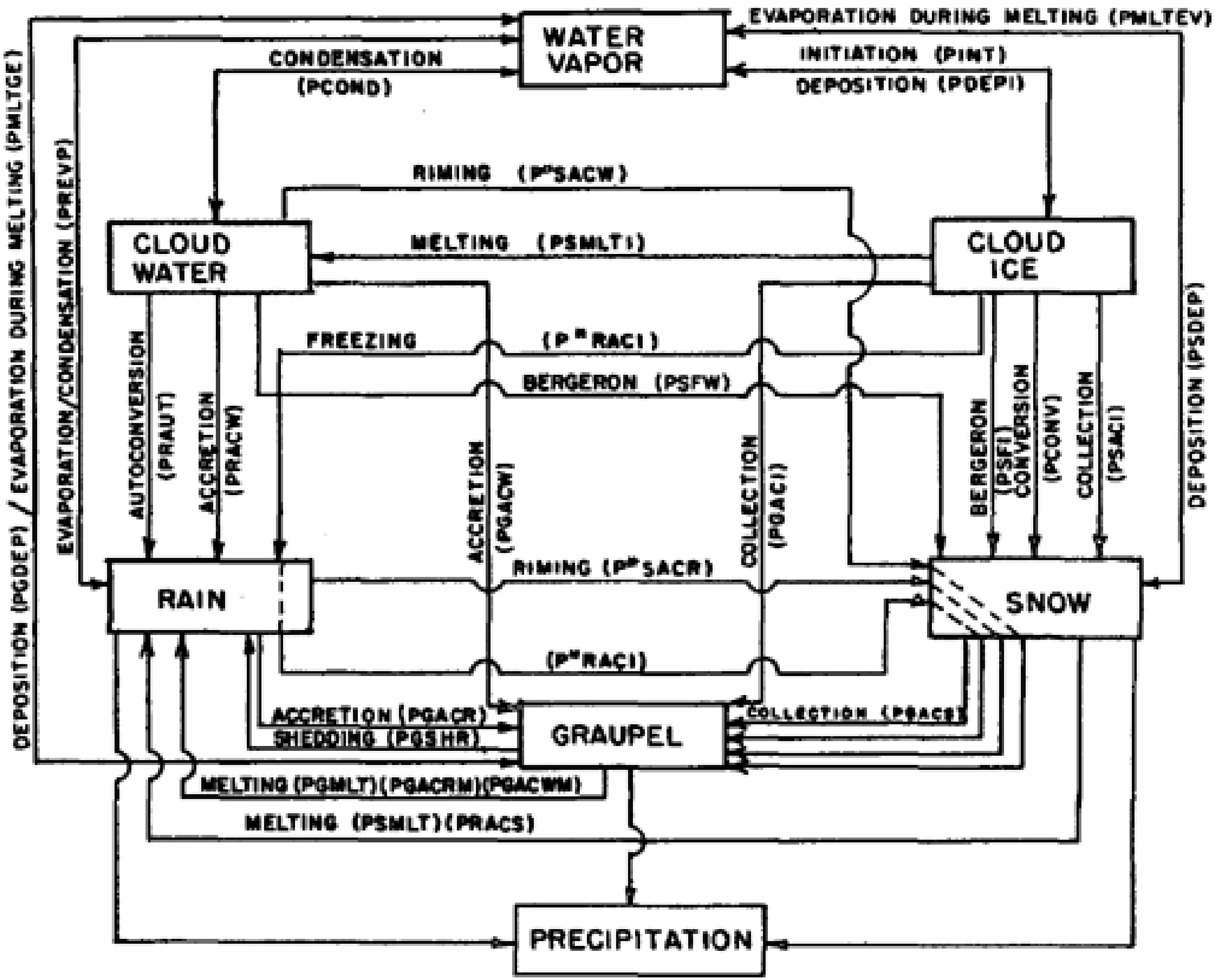


FIG. 1. Schematic depicting the cloud and precipitation processes included in the model for the study of narrow cold-frontal rainbands.

Kiehl and Trenberth, BAMS 1997

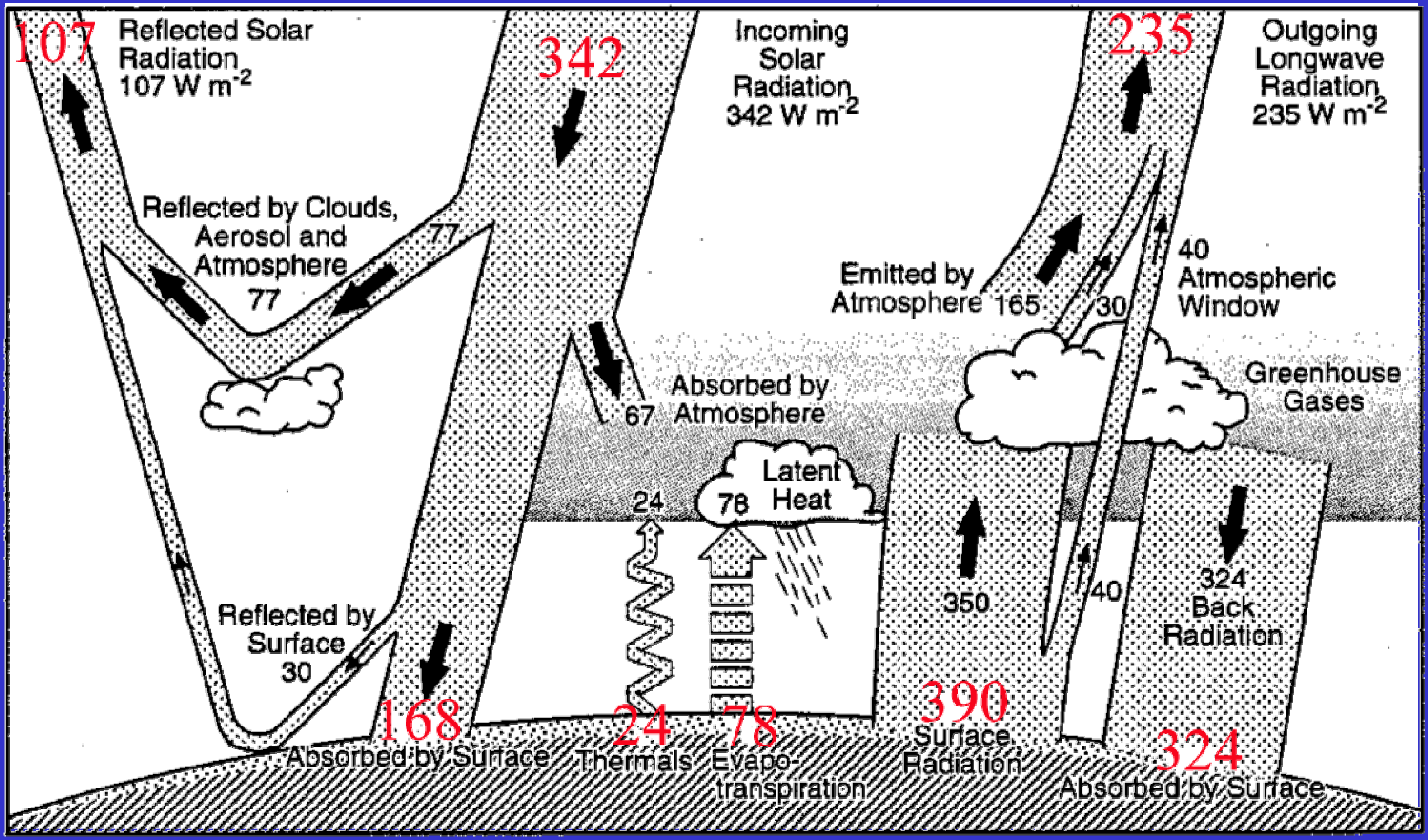
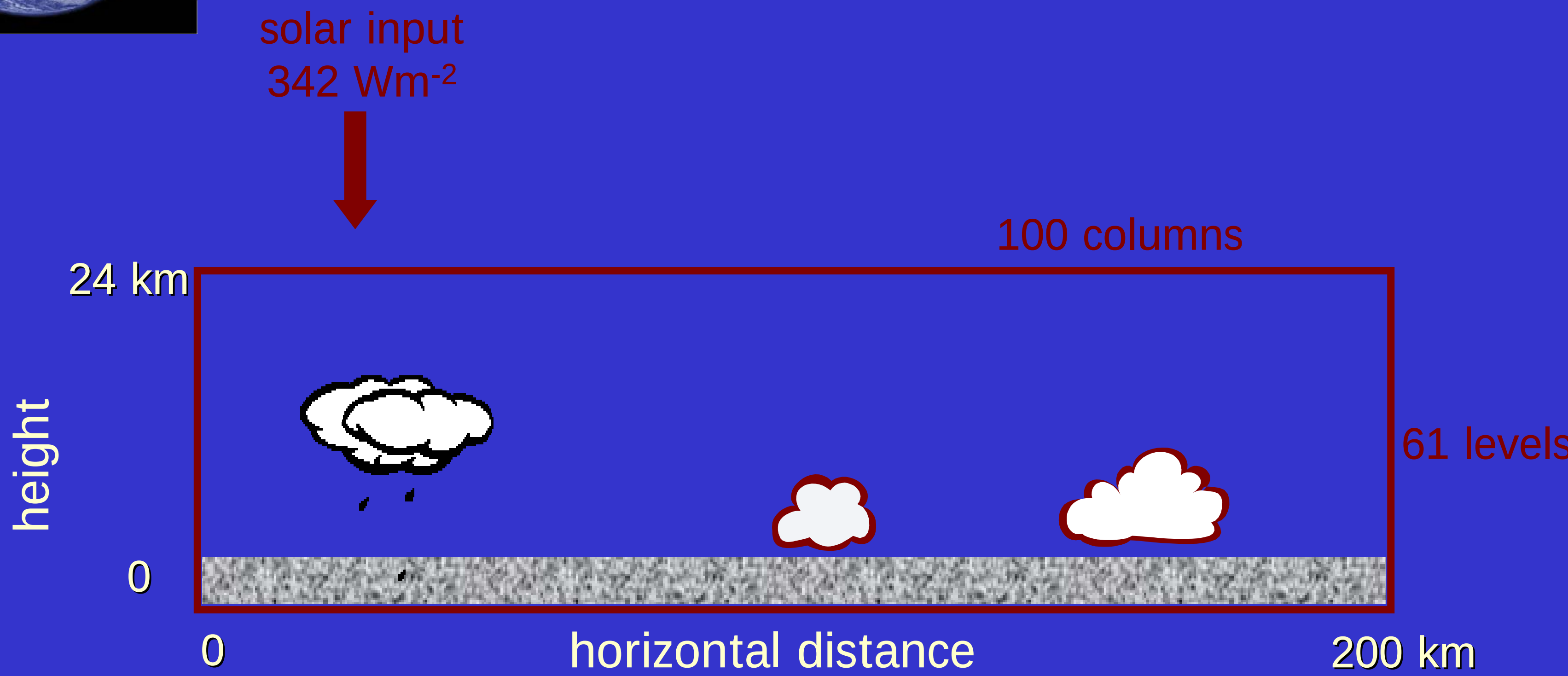


FIG. 7. The earth's annual global mean energy budget based on the present study. Units are W m^{-2}.

Radiative-convective quasi-equilibrium mimicking planetary energy budget using a cloud-resolving model



Surface temperature = 15° C
Surface relative humidity = 80%
Surface albedo = 0.15

Numerical model:

- Dynamics: 2D super-parameterization model (Grabowski 2001) with simple bulk microphysics (warm-rain plus ice; Grabowski 1998)
- Radiation: NCAR's Community Climate System Model (CCSM) (Kiehl et al 1994) in the Independent Column Approximation (ICA) mode
- 100 columns (Δx=2km) and 61 levels (stretched; 12 levels below 2 km; top at 24 km)

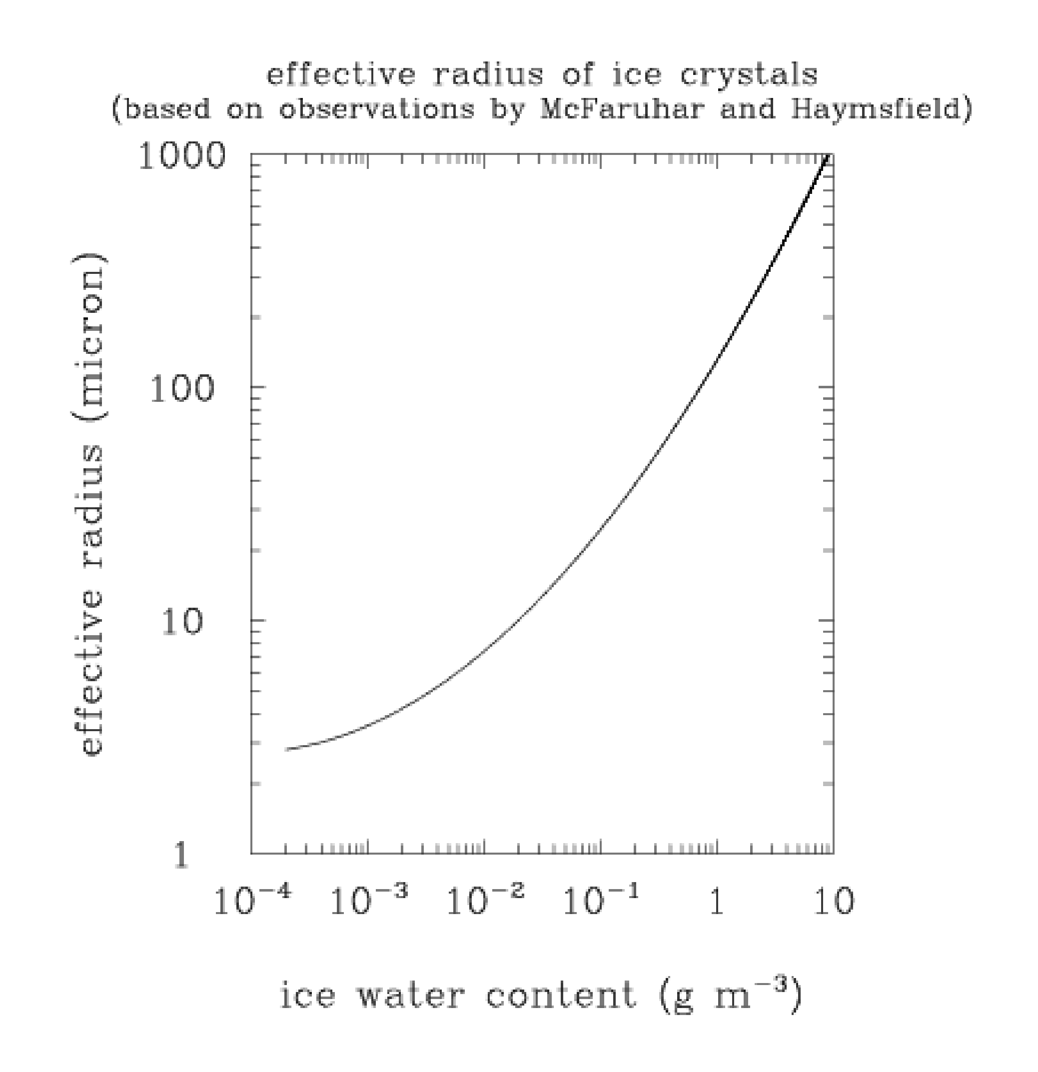
How do dynamic and radiation transfer models communicate?

- Dynamics provides profiles of temperature, moisture, and cloud properties in each model column (condensate mixing ratios, optical properties of cloud particles - effective radii[1] of cloud droplets and ice crystals in particular)

- Radiation calculates radiative fluxes based on solar input, surface characteristics, and properties within the column; divergence of solar and longwave fluxes are fed to the dynamics

[1] – effective radius is the ratio between the third and the second moments of droplet/ice crystal size distribution

Effective radius for ice particles...



This formula is used in all simulations (i.e., no indirect impact on ice particles).

Effective radius for water droplets...

$r_{eff} \sim r_v$ (e.g., Martin et al. JAS 1994)

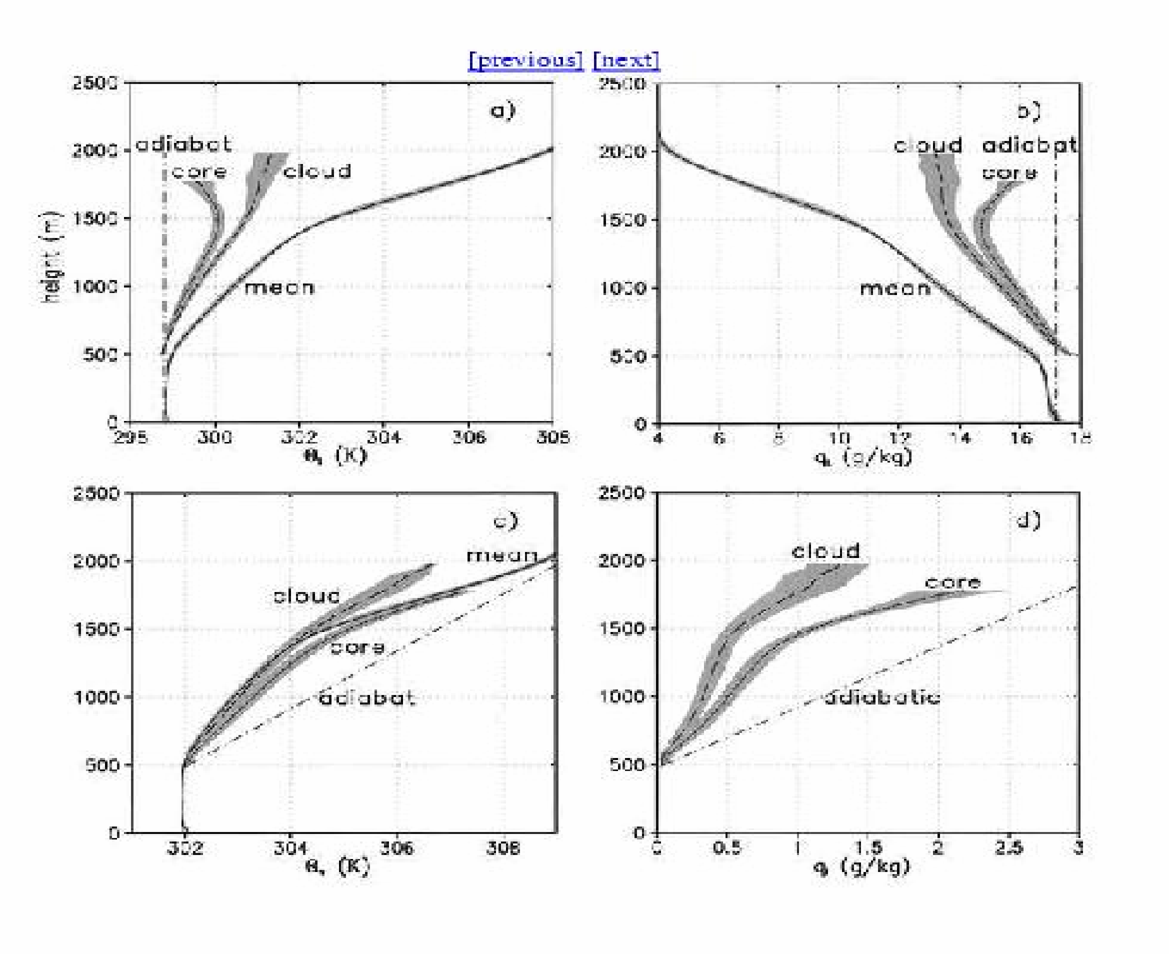
$\longrightarrow r_v \equiv \langle r^3 \rangle^{1/3}$ – mean volume radius;

$\longrightarrow$ cloud water $q_c \sim N r_v^3$

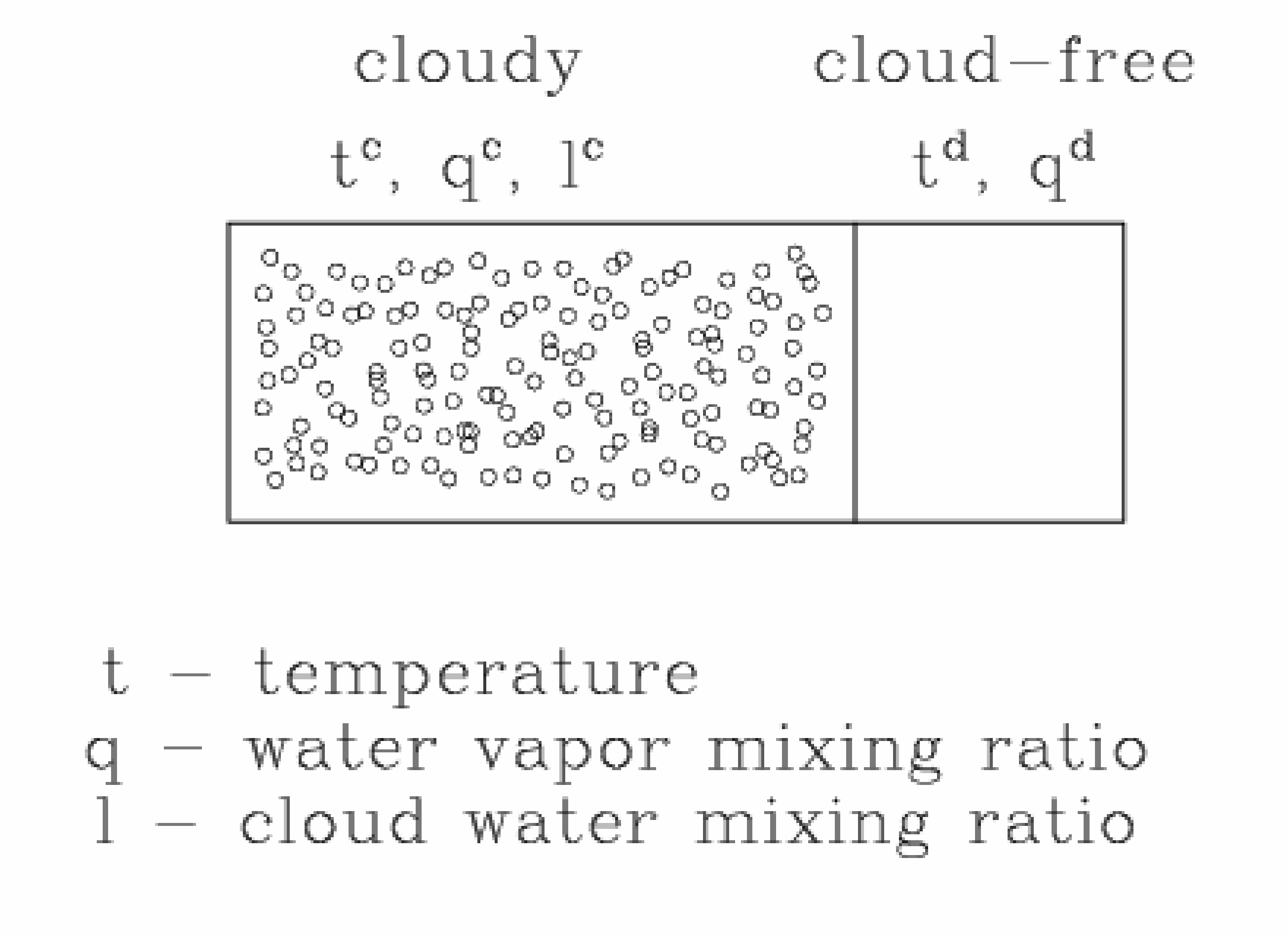
bulk model provides q_c*; how to estimate* r_v*?*

Shallow convective clouds are strongly diluted by entrainment

Siebesma et al. JAS 2003



Bulk mixing between cloudy and cloud-free air (adiabatic, isobaric)



What is wrong with this picture?

Homogeneous mixing:

all droplets are exposed to the same conditions
during mixing

Extremely inhomogeneous mixing:

some droplets evaporate completely, the rest
does not change size at all

Simulations of decaying moist turbulence (final stages of cloud entrainment)

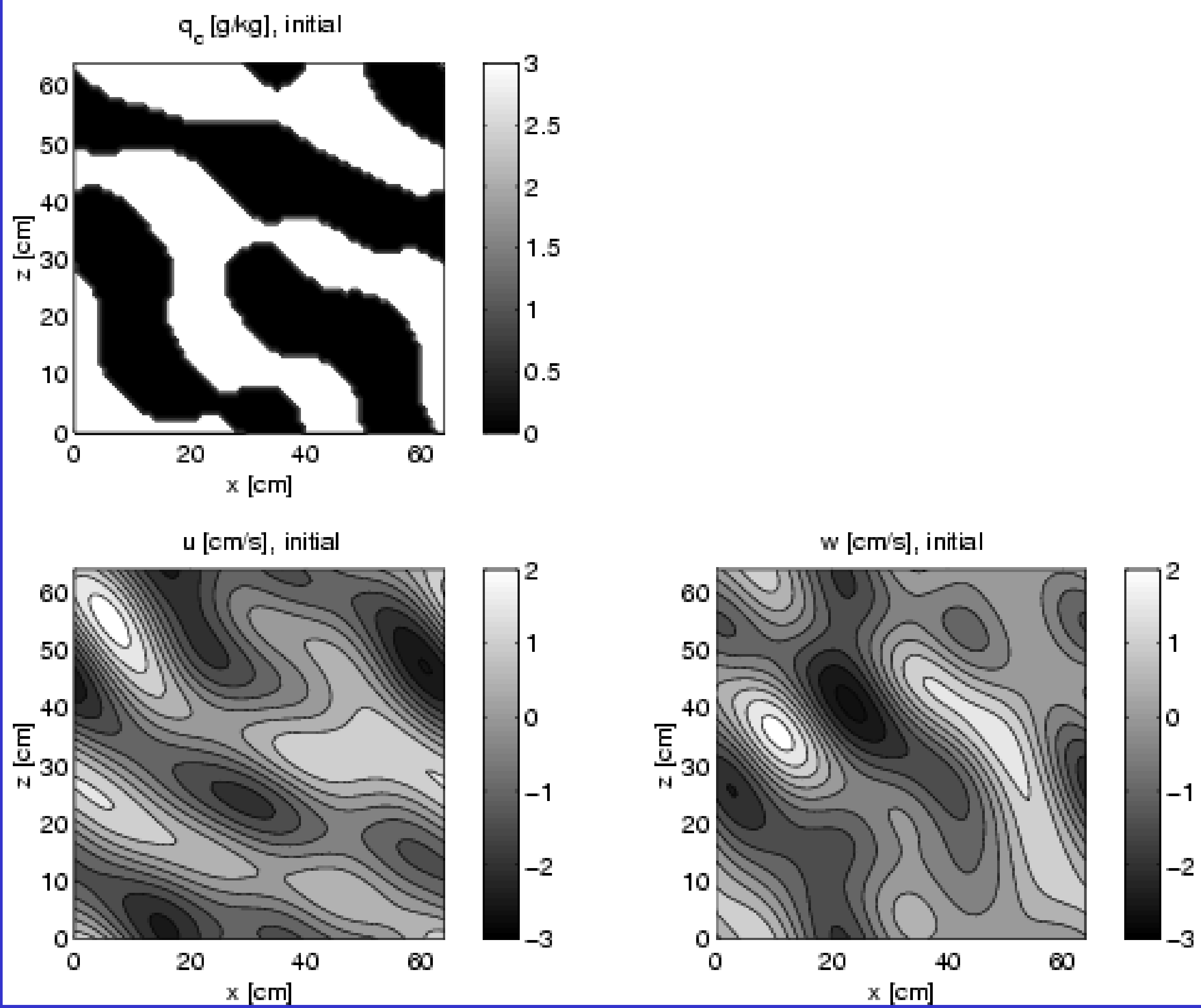
Andrejczuk et al., JAS, 2004

T=293 K (both cloud and clear air)

RH=65%

q_c=3.2 g/kg

(filaments neutrally buoyant)



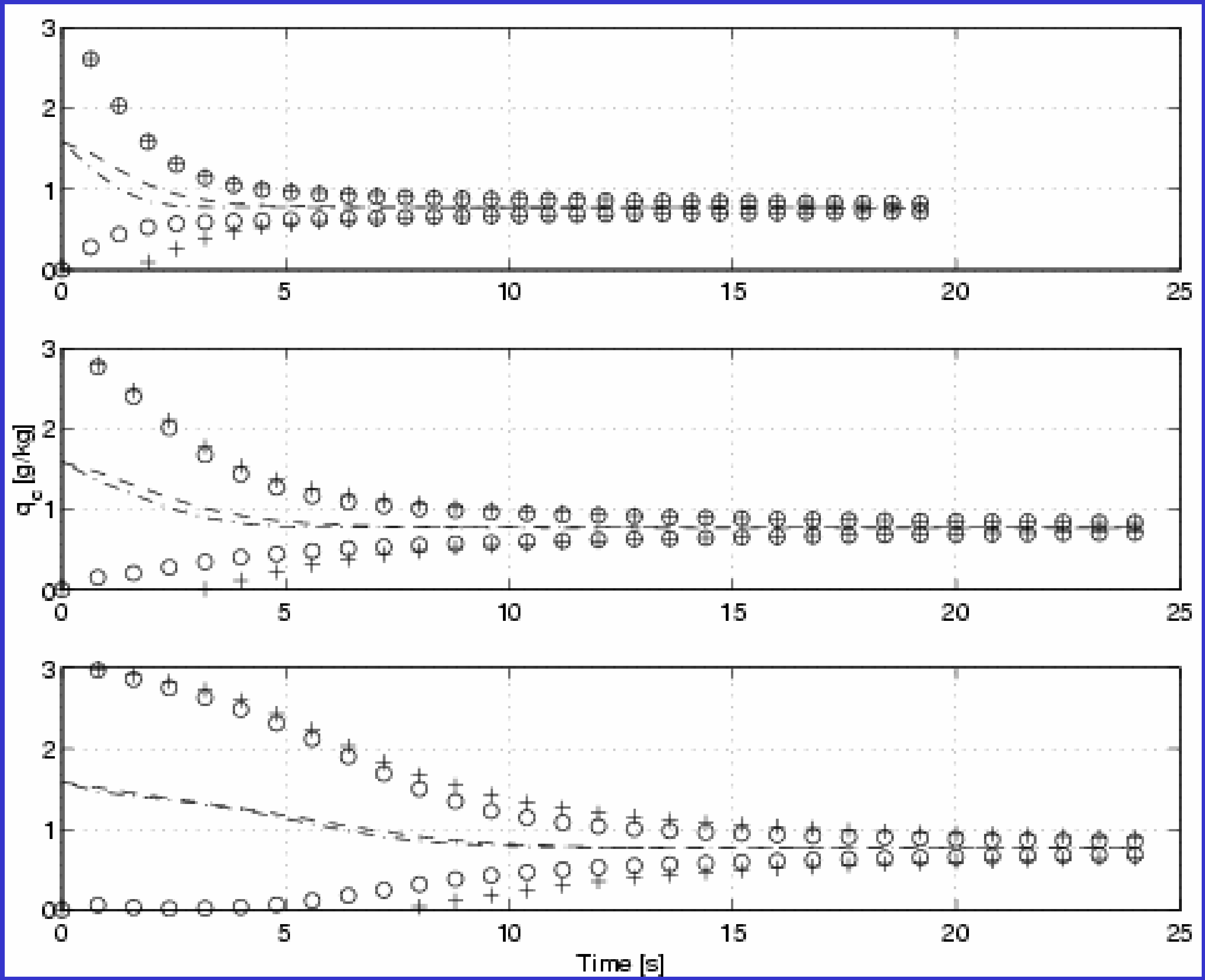
Velocity scale for low TKE (x10 for high TKE)

Evolution of the cloud water

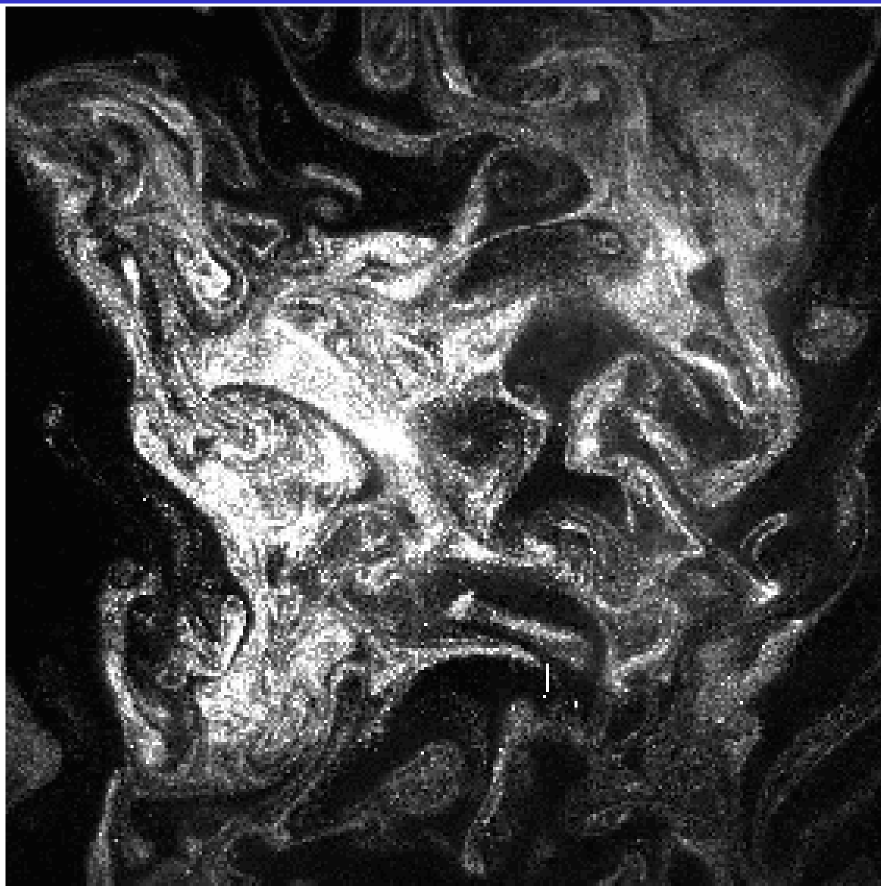
High TKE

Moderate TKE

Low TKE



Laboratory experiment



60 cm

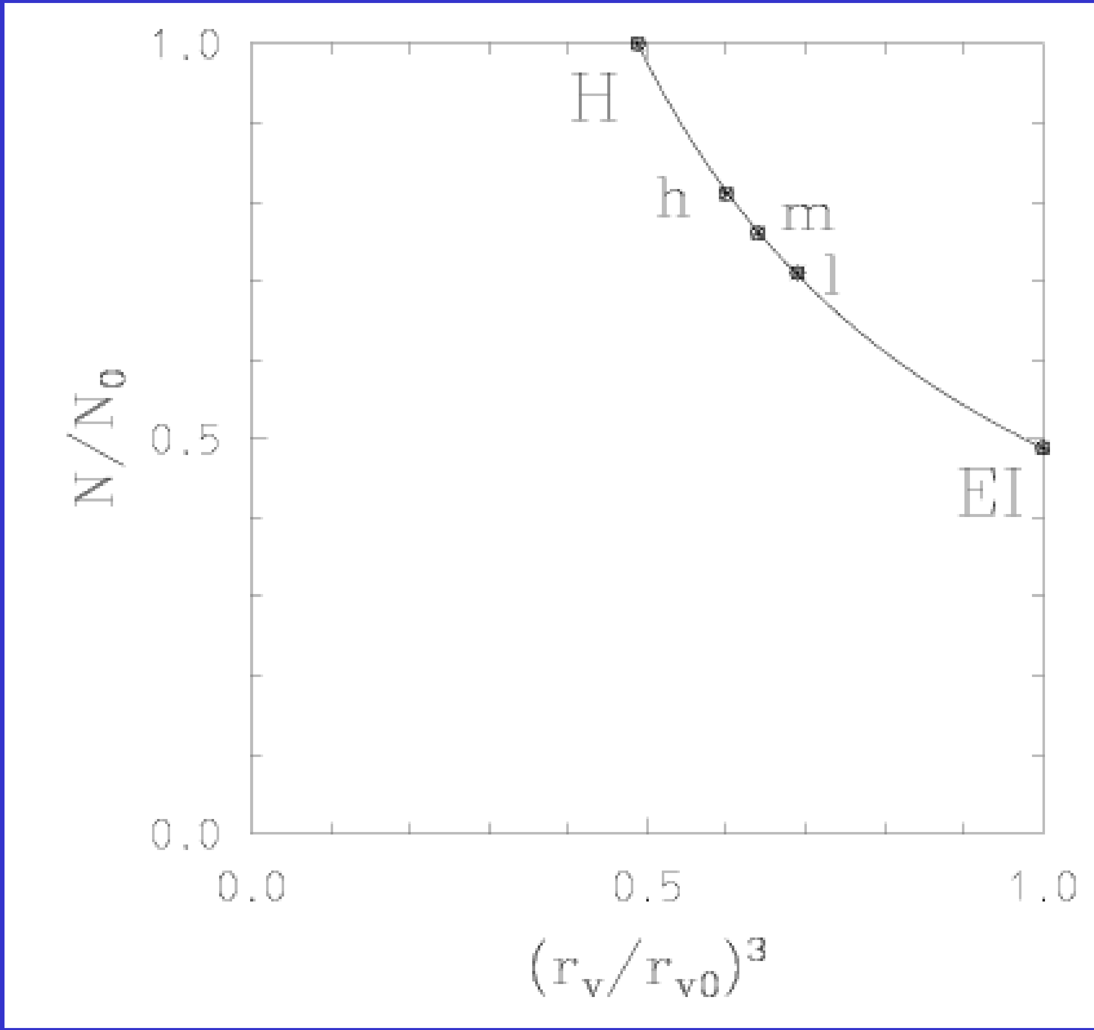
Model simulation

60 cm

Microphysical transformations during cloud-clear air mixing (Andrejczuk et al. JAS 2004)

HOMOGENEOUS MIXING

r-N diagram



EXTREMELY INHOMOGENEOUS MIXING

prescribe number of droplets:

$$r_v^3 \sim q_c/N$$

HOMOGENEOUS MIXING

prescribe size of droplets:

$$r_v = r_{ad}(z)$$

EXTREMELY INHOMOGENEOUS MIXING

1st and 2nd indirect effects of water clouds in a cloud-resolving model:

	"pristine"	"polluted"
cloud droplet concentration (cm^{-3})	100	1,000

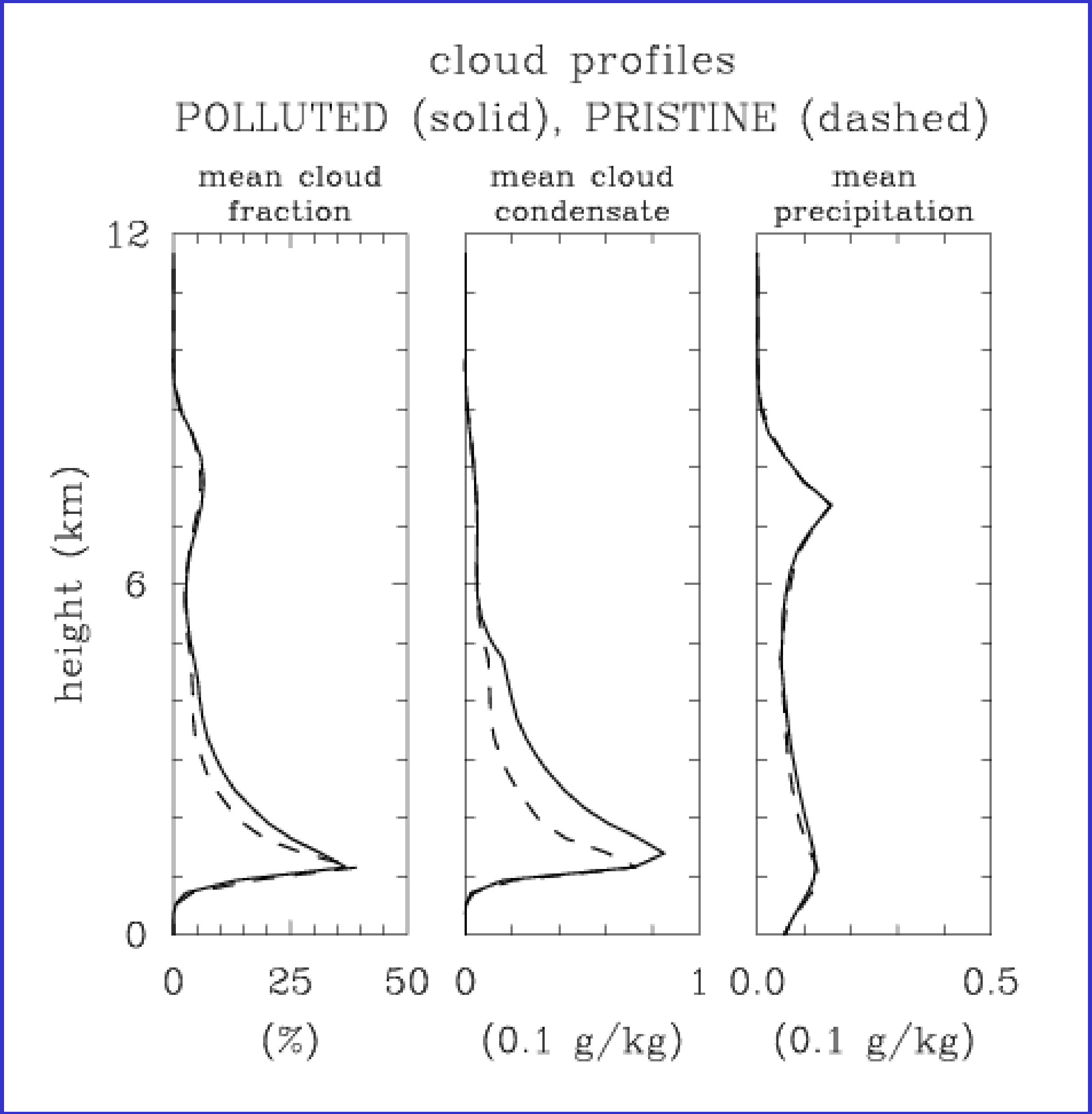
1st effect: a simple parameterization of the effective radius

2nd effect: Berry's parameterization of the conversion from cloud water to rain

Model simulations:

- 100 per cc versus 1,000 per cc
- Mixing:
  - homogeneous (h)
  - extremely inhomogeneous (ei)
  - between "h" and "ei" (the same changes in N and r^3 (rN)

Each simulation run for 120 days, results averaged over the last 60 days



Homogeneous mixing (droplet concentration the same everywhere…)

	PRISTINE 100 per cc			POLLUTED 1,000 per cc			KT97
	h	rN	ei	h	rN	ei	
TOA albedo	0.35	0.31	0.29	0.41	0.38	0.34	0.31
OLR (W m^{-2})	242	242	243	240	241	242	235
radiative cooling of atmosphere (W m^{-2})	-101	-100	-100	-101	-100	-99	-102
solar energy absorbed at surface (W m^{-2})	163	176	182	141	153	164	168
surface net longwave flux (W m^{-2})	72	73	73	70	71	73	66
surface sensible heat flux (W m^{-2})	20	20	20	19	19	19	24
surface latent heat flux (W m^{-2})	73	73	73	75	74	74	78
surface energy budget (W m^{-2})	**-2 (7)**	**10 (5)**	**17 (5)**	**-23 (9)**	**-12 (9)**	**-2 (7)**	**0**

1st indirect effect

1st indirect effect

2nd indirect effect

Table 2: Energy fluxes averaged over 60-day period (days 61-120) for various simulations assuming PRISTINE and POLLUTED cloud conditions. Columns marked "h", "ei", and "rN" show results from simulations where, respectively, homogeneous, extremely inhomogeneous, and intermediate mixing scenarios were assumed to prescribe the effective radius of cloud droplets. Values in brackets in the surface energy budget show standard deviations for the 60-day averaging period. Estimates of global mean energy budgets from Kiehl and Trenberth (1997) are shown in KT97 column.

Key results:

1. Water and energy fluxes surprisingly similar to Kiehl &Trenberth (1997)!

2. Insignificant effects on longwave fluxes, surface sensible and latent heat fluxes (no 2nd indirect effect)!

3. Significant impact on solar radiation: changes in "planetary" albedo and surface energy budget:

	Clean 100 per cc			Polluted 1000 per cc		
Surface energy	h	rN	ei	h	rN	ei
budget (W/m^2)	-2	10	17	-23	-12	-2

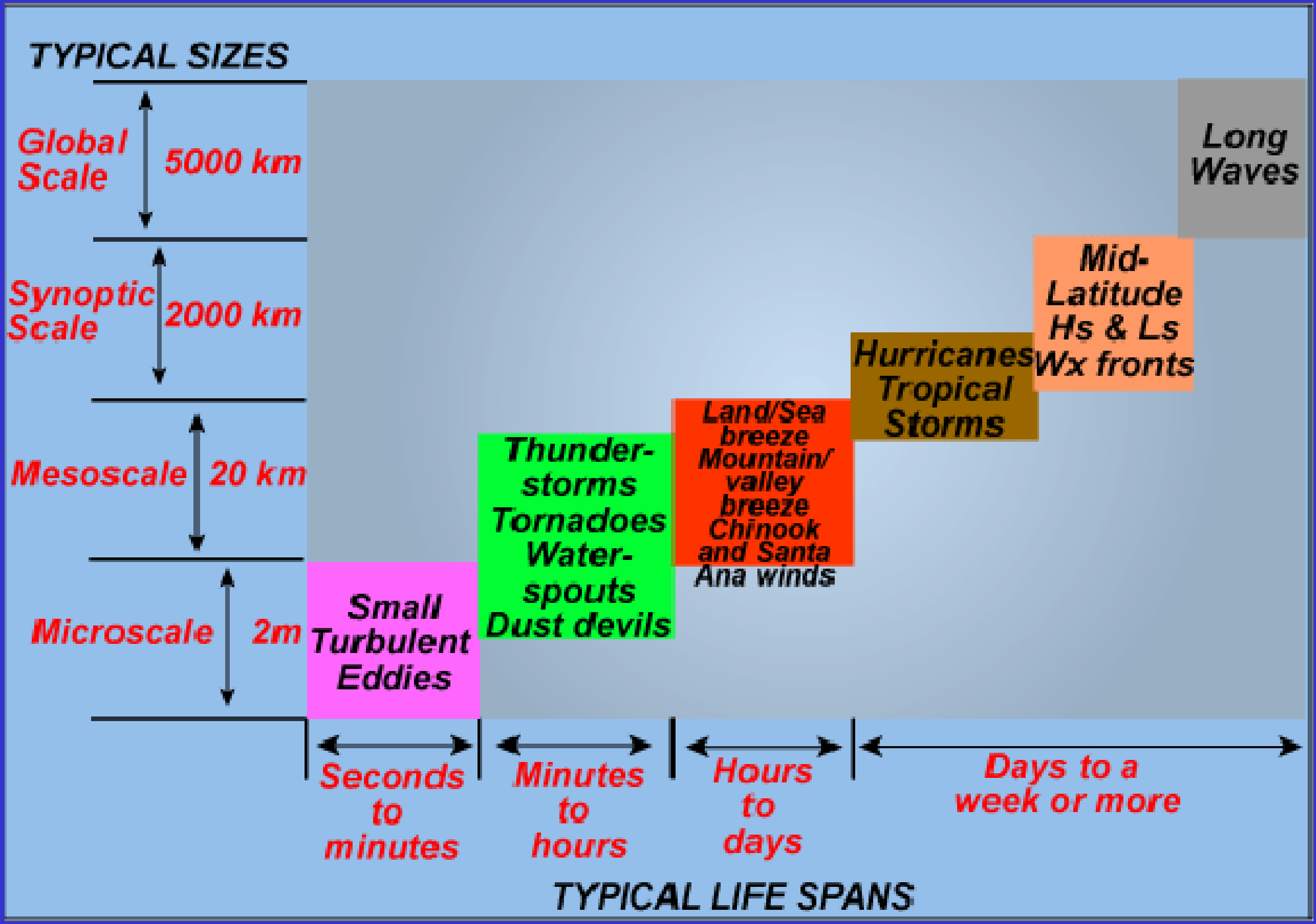
4. The 1st indirect effect (Twomey effect) can be offset by assumptions about mixing!

But...

- dynamic model is 2D,
- microphysics is extremely simple,
- clouds are under-resolved (especially shallow convection),
- radiative transfer is 2D…

- Assumptions about spatial variability of effective radius in warm convective clouds seem to impact the surface energy budget at a level comparable to the 1st indirect effect itself.
- This conclusion comes from low-spatial resolution 2D model and two-stream independent column radiation transfer model. LES-type simulations applying 3D radiative transfer are needed.
- Combination of in-situ aircraft and ground-based remote sensing observations, together with laboratory studies and numerical modeling should provide sufficient data to develop useful parameterizations of microphysical properties of shallow convective clouds.

Why is it so hard to simulate the Earth climate system?



Because some of the key processes are even not on this diagram….

Conclusions:

Cloud microphysics is one of key elements of climate and climate change.

Cloud microphysics is tightly coupled to cloud dynamics; super-parameterization was proposed to provide a better framework for the cloud microphysics in the clouds-in-climate problem.

Lower tropospheric clouds are especially important because of their strong impact on solar radiation reaching the surface and small impact on longwave radiation. Their microphysical properties are critical.