“Nimbostrophy”

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Nimbus (*latin*) – cloud

Strophe (from Greek στροφή, "turn, bend, twist")
is a concept in versification

Strophic - relating to, containing, or consisting of strophes

Loosely: applying order, as in poems containing *strophes*

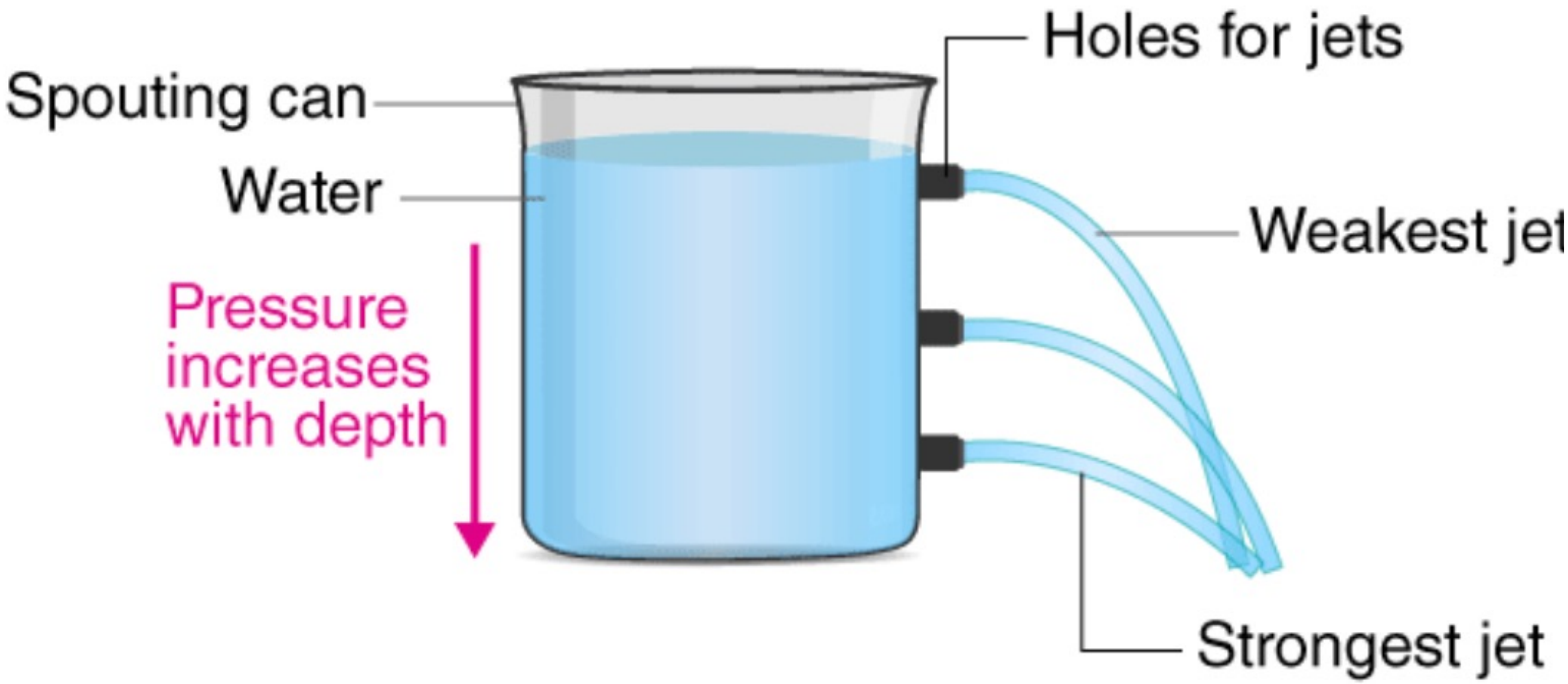
In fluid mechanics: *balanced flows*:

Hydrostatic flow ("hydrostrophy"?)

Cyclostrophic flow ("cyclostrophy")

Geostrophic flow ("geostrophy")

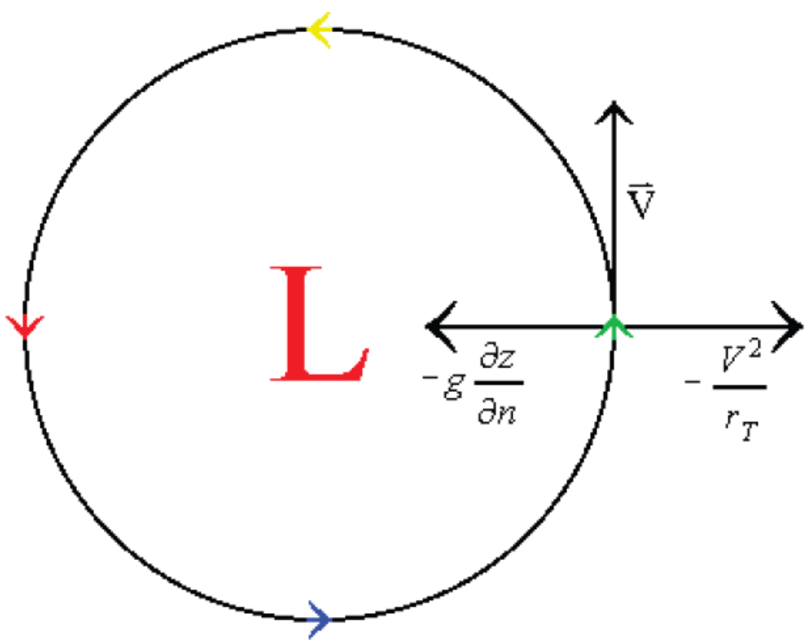
Hydrostatic balance: $\frac{dp}{dz} = -\rho z$



https://byjus.com/physics/hydrostatic-pressure/#:~:text=We%20know%20that%20matter%20in,or%20pressure%20of%20the%20liquid.

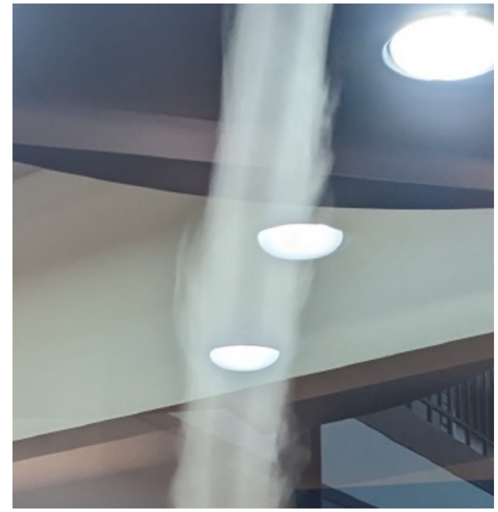
Cyclostrophic flow (“cyclostrophy”)

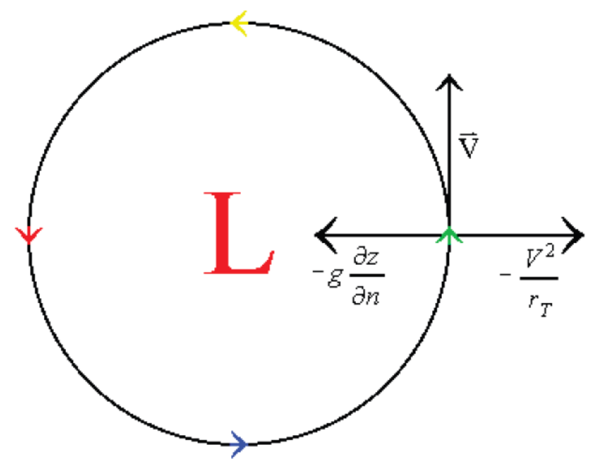
balance between pressure gradient and centrifugal forces




Tornado

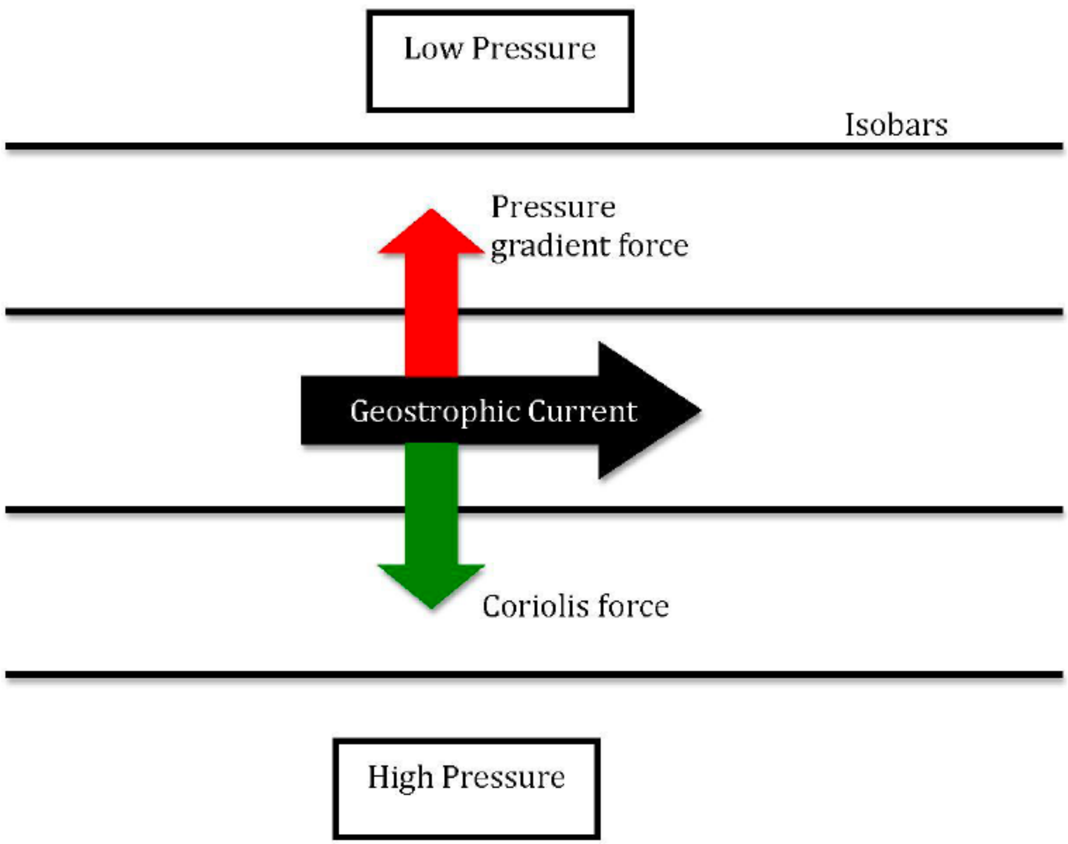

wild

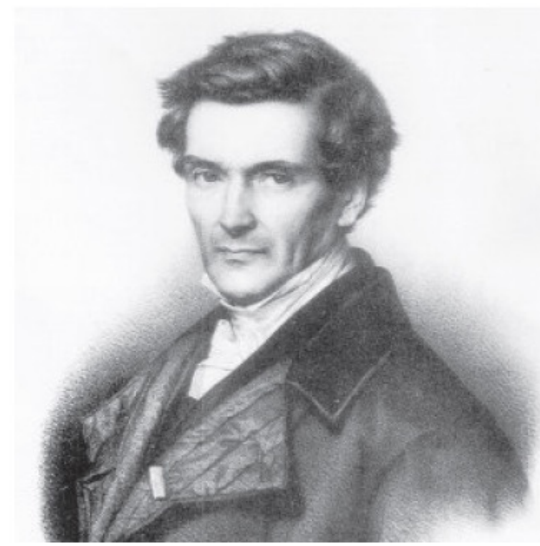



L
$\vec{V}$
$-g\,\frac{\partial z}{\partial n}$
$-\frac{V^2}{r_T}$

Geostrophic flow ("geostrophy")

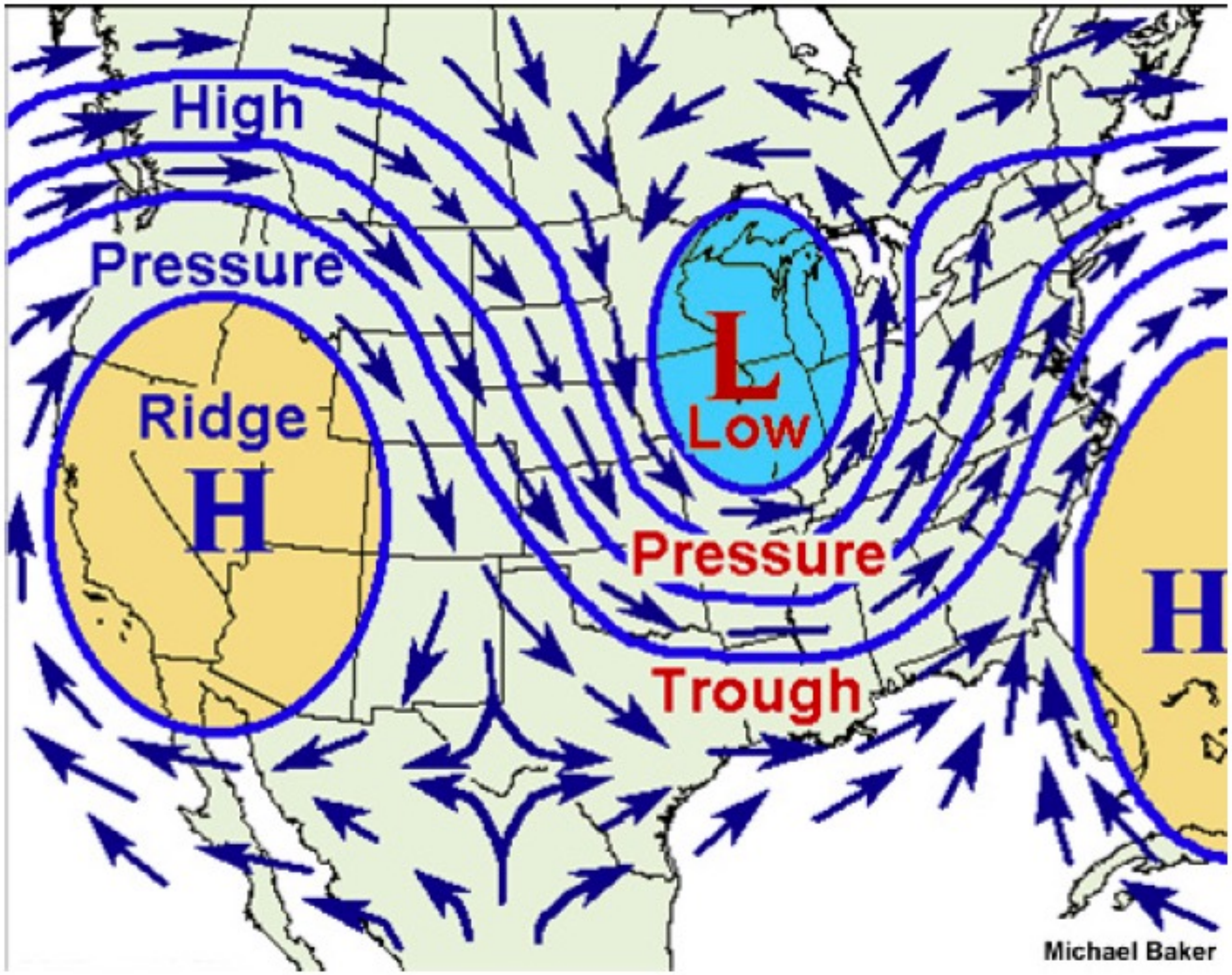
balance between pressure gradient and Coriolis forces





Gaspard-Gustave de Coriolis
1792-1843

Geostrophic flow ("geostrophy")



Examples of balanced flows:

	Antitriptic flow	Geostrophic flow	Cyclostrophic flow	Inertial flow	Gradient flow	Ekman flow
curvature	N	N	Y	Y	Y	N
friction	Y	N	N	N	N	Y
pressure	Y	Y	Y	N	Y	Y
Coriolis	N	Y	N	Y	Y	Y

https://en.wikipedia.org/wiki/Balanced_flow

What about clouds?

Not necessarily about cloud dynamics – we know that the small-scale air flow within a cloud is affected by buoyancy and nonhydrostatic pressure gradient.

Rather, "nimbostrophy" is about understanding basic *thermodynamic* balances that clouds follow, especially warm clouds, that is, clouds with no complications of ice processes.

Warm clouds develop through condensation of water vapor to form and grow cloud droplets.

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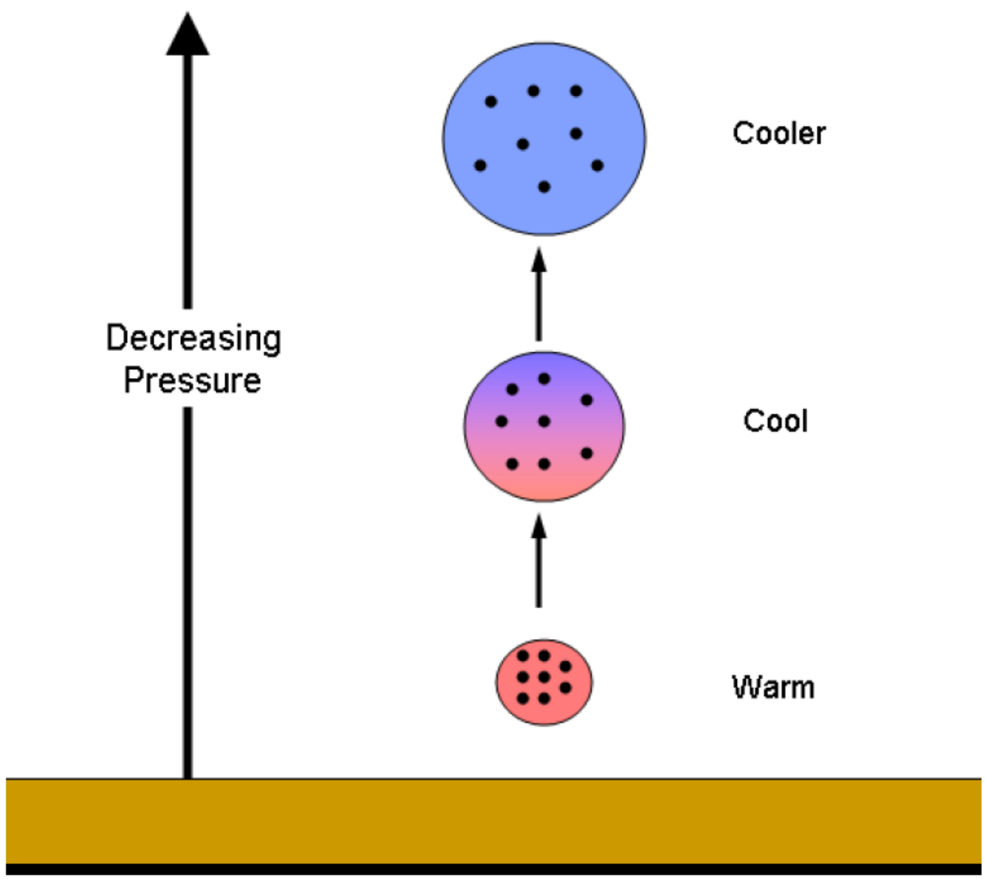
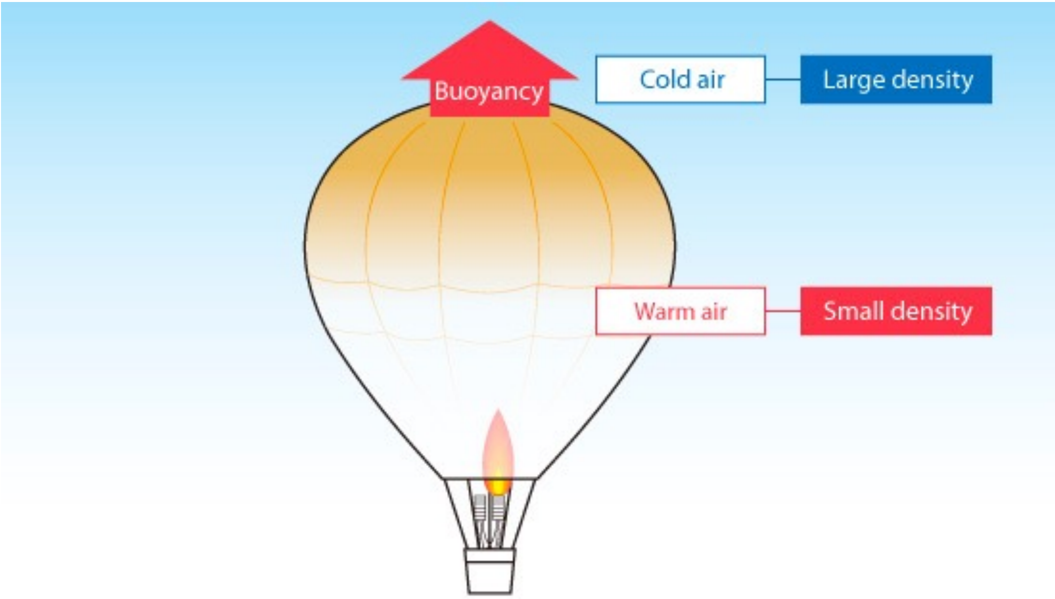
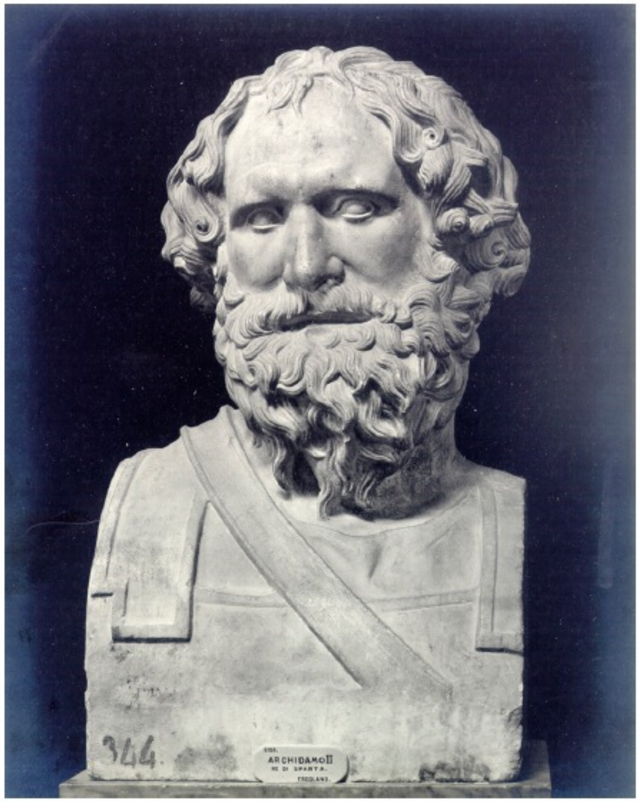


Buoyancy-driven flows: gravity plus density difference



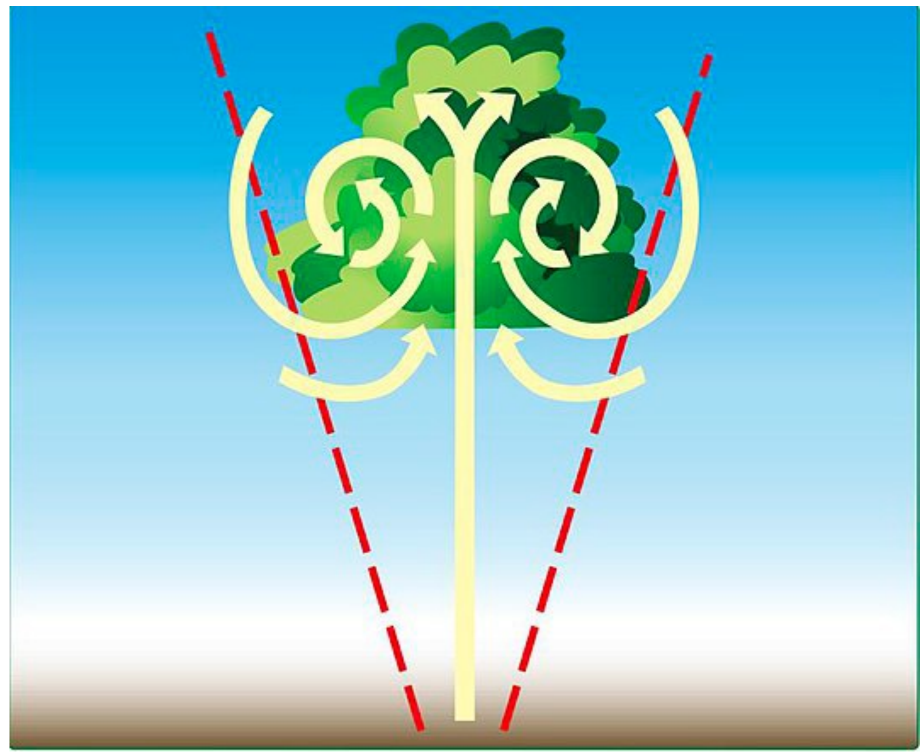


Archimedes, c. 287 BCE – c. 212 BCE

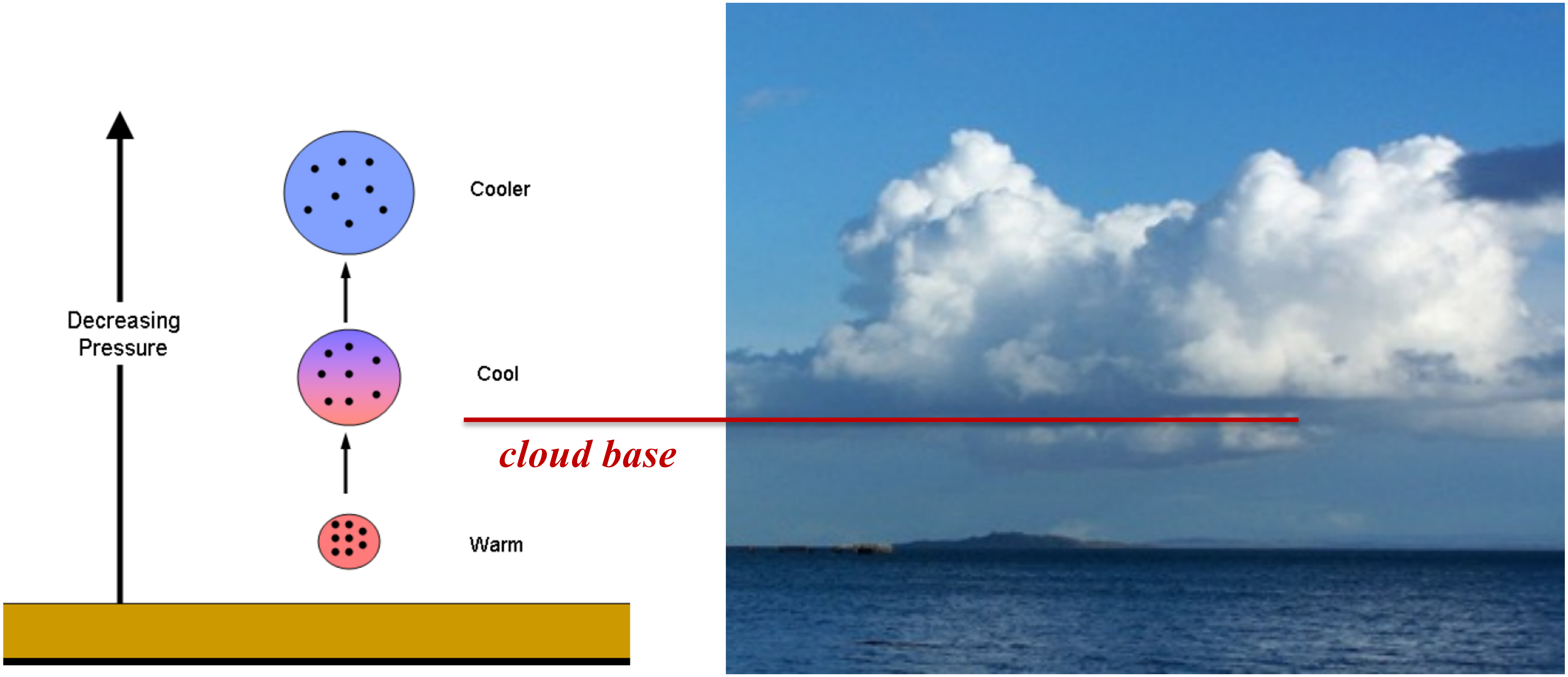
Buoyancy-driven flows: gravity plus density difference

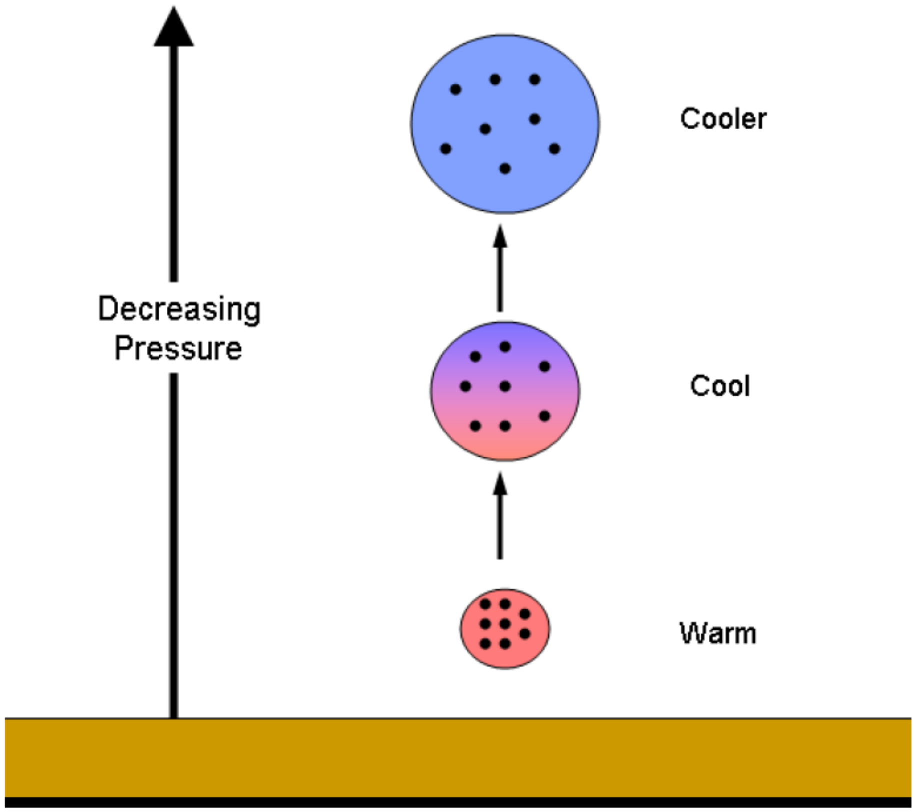
Small cloud

Idealized thermal



https://en.wikipedia.org/wiki/Thermal


Decreasing
Pressure
Cooler
Cool
cloud base
Warm



A simple model of an adiabatic parcel: a small volume of air rising in the atmosphere, decreasing its temperature, and eventually reaching water saturation. Subsequent rise leads to cloud formation.

This can be described by a simple set of equations describing changes of the temperature, water vapor density, and the mass of condensed (liquid) cloud water.

BULK MODEL OF CONDENSATION:

$$\frac{D\theta}{Dt} = \frac{L_v \theta}{c_p T} C_d$$

$$\frac{Dq_v}{Dt} = -C_d$$

$$\frac{Dq_c}{Dt} = \quad C_d$$

$$\frac{D}{Dt} \equiv \frac{\partial}{\partial t} + \mathbf{u} \cdot \nabla$$

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 ($\approx \theta_e/T_e$, i.e., environmental hydrostatic pressure)

Potential temperature equation:

$$\frac{D\theta}{Dt} = \frac{L_v \theta}{c_p T} C_d$$

Temperature equation:

$$\frac{DT}{Dt} = -\frac{g}{c_p} w + \frac{L_v}{c_p} C_d$$

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$$C_d \sim N_c \, \frac{dm}{dt}$$

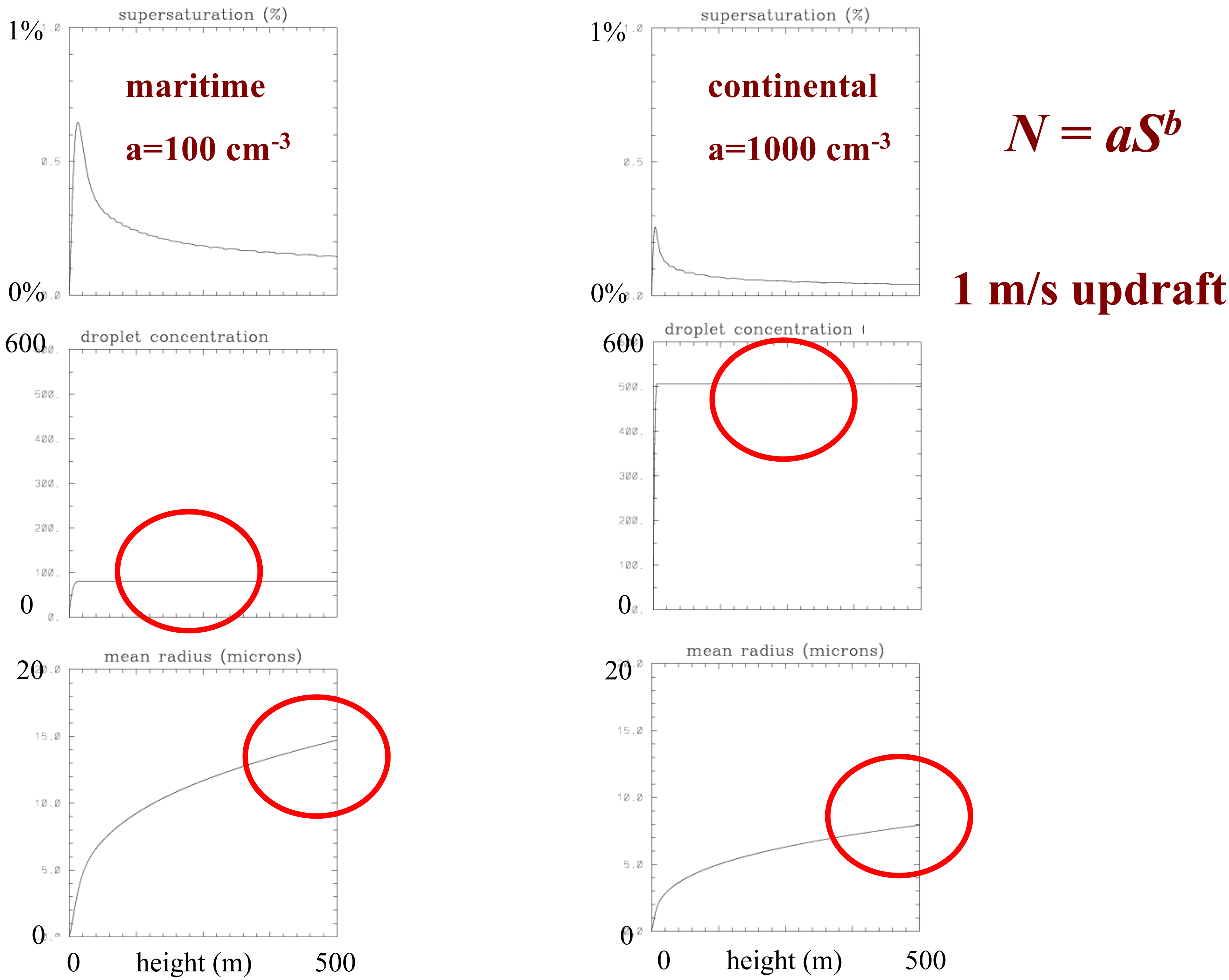
N_c - droplet concentration, $\frac{dm}{dt}$ - droplet mass growth rate

$$\frac{dr}{dt} = A \, \frac{S}{r} \,, \quad A = A(p, T) \,, \quad S - \text{supersaturation}$$

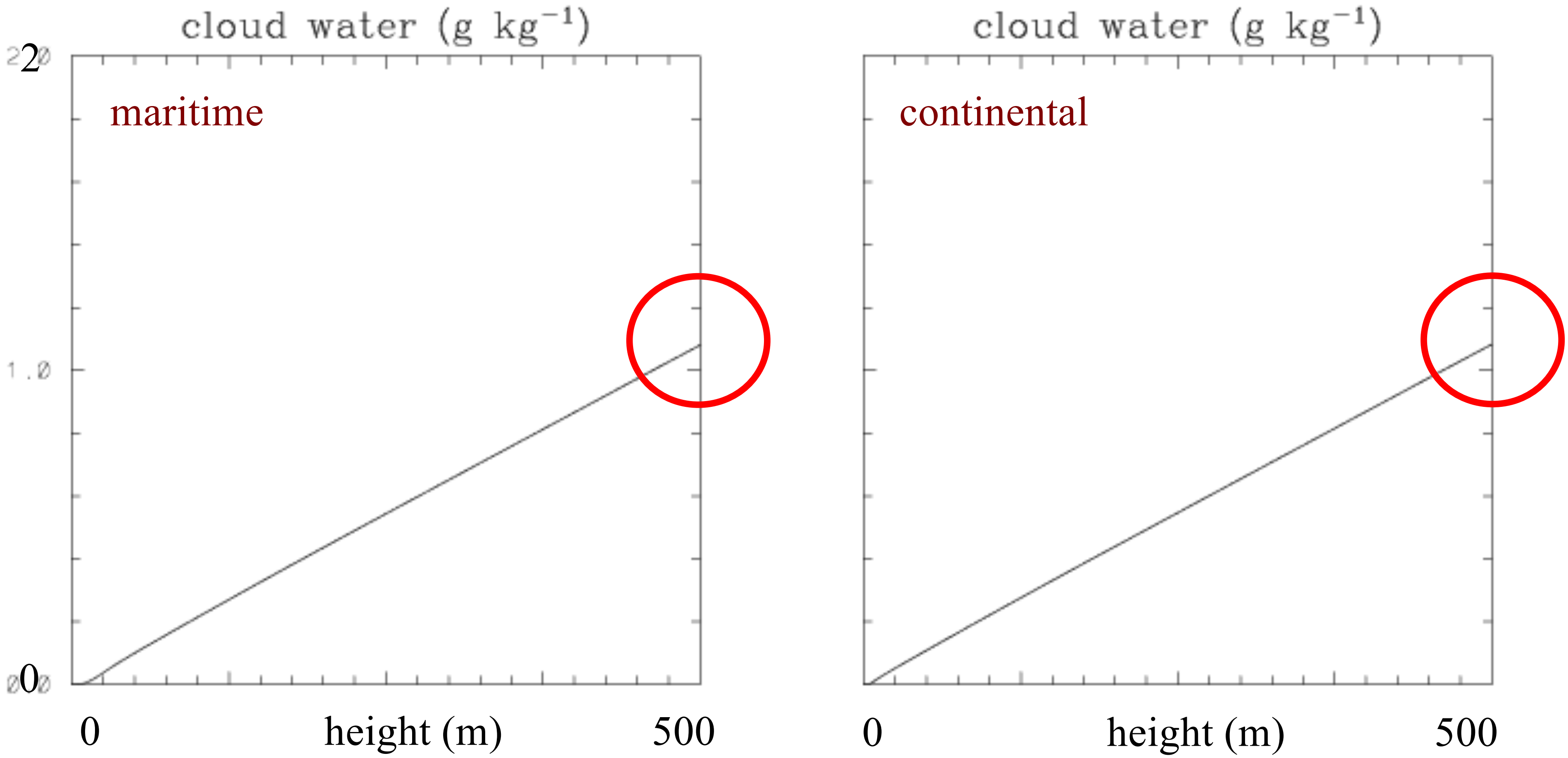
$$S = \frac{q_v}{q_{vs}} - 1$$

$$\boxed{C_d \sim N_c \; r \; S}$$

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



supersaturation (%)
1%
0%
maritime
a=100 cm-3
continental
a=1000 cm-3
$N = aS^b$
1 m/s updraft
droplet concentration
600
0
mean radius (microns)
20
0
0
height (m)
500


cloud water (g kg^{-1})
2
1.0
0
maritime
0
height (m)
500
cloud water (g kg^{-1})
continental
0
height (m)
500

Since the finite supersaturation has such a small impact on the amount of condensed water (and thus on the temperature change), can we simply assume that supersaturation vanishes?

This is the idea behind bulk methodology to cloud modeling referred to as the "saturation adjustment", i.e., maintaining $S=0$.

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This is the idea behind bulk methodology to cloud modeling referred to as the "saturation adjustment", i.e., maintaining S=0.

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$$\frac{Dq_v}{Dt} = -C_d$$

$$\frac{Dq_c}{Dt} = C_d$$

C_d is defined such that cloud is always at saturation:

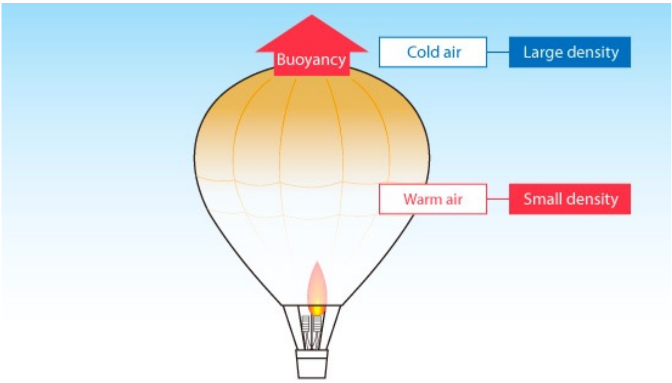
$$q_c = 0 \quad \text{if} \quad q_v < q_{vs}$$

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

$$q_{vs}(p,T) = 0.622 \frac{e_s(T)}{p - e_s(T)}$$

Parcel buoyancy: *density temperature* or *density potential temperature*

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



$$T_d = T \frac{1 + q_v/\epsilon}{1 + q_v + q_c}$$

T - air temperature

q_v - water vapor mixing ratio ($\sim 10^{-2}$)

q_c - condensed water mixing ratio ($\sim 10^{-3}$)

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

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

$$\boxed{T_d \approx T\left(1 + 0.61 q_v - q_c\right)}$$

Density potential temperature θ_d:

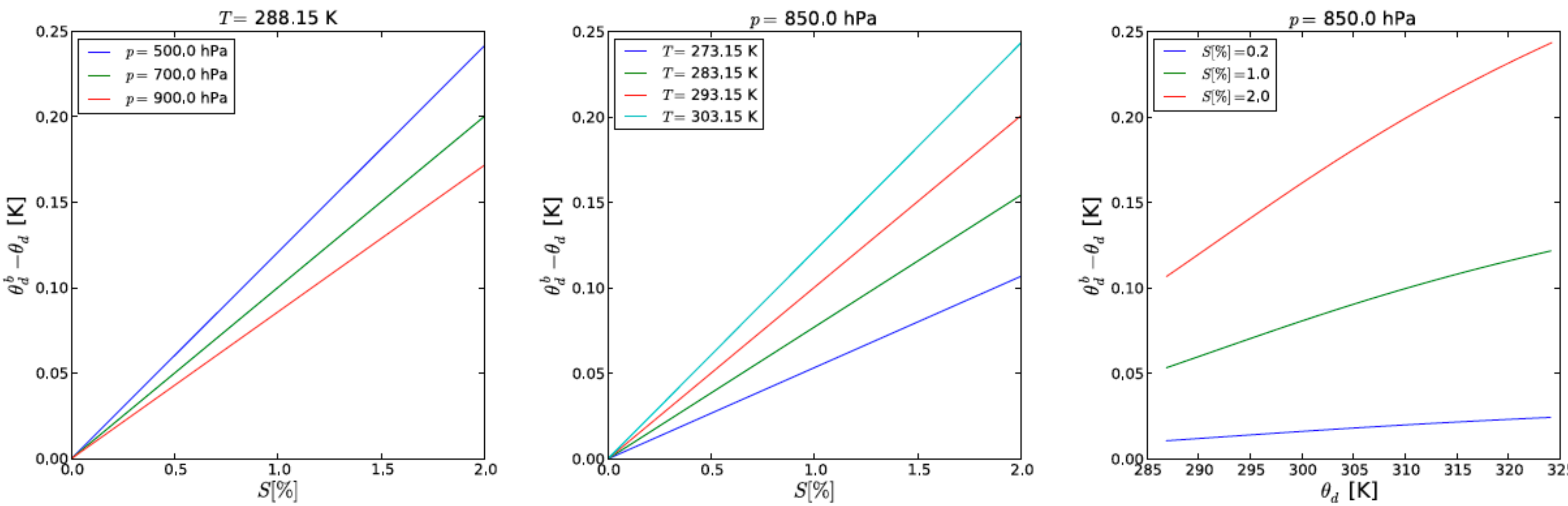
$$\boxed{\theta_d \approx \theta\left(1 + 0.61 q_v - q_c\right)}$$

$$\theta_d \approx \theta \left(1 + 0.61 q_v - q_c\right)$$

Does the final supersaturation (S, typically a fraction of 1%) affect cloud buoyancy?

Compare density potential temperature with finite S, and corresponding density potential temperature with S=0, the so-called *saturation adjustment bulk* temperature):

$$\theta_d \ , \ \theta_d^b$$



saturation adjustment (S=0) provides slightly more cloud buoyancy….

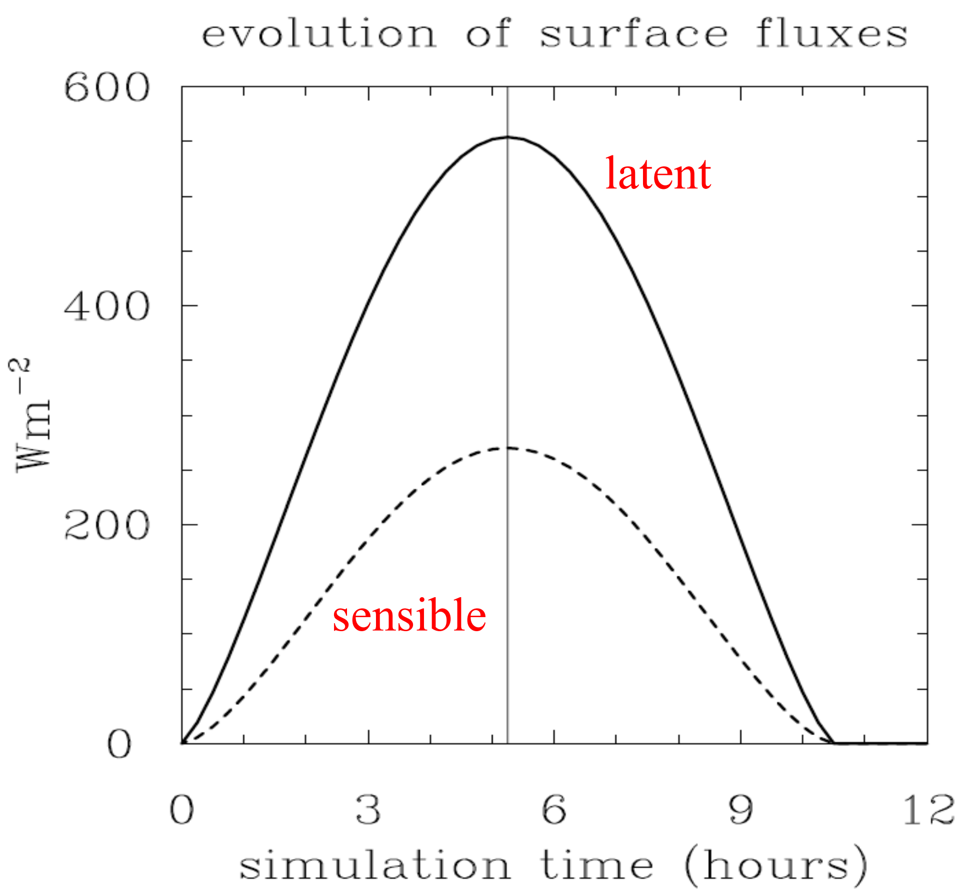
Grabowski and Jarecka *JAS* 2016

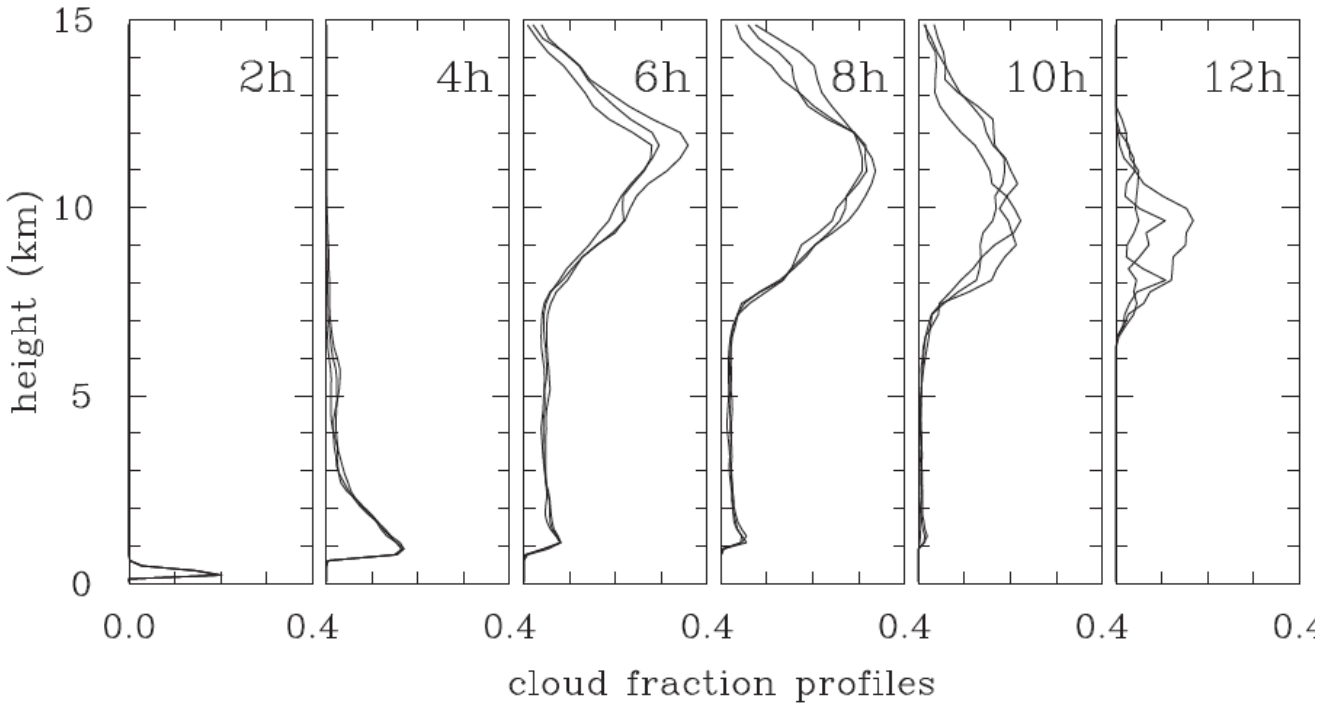
Does the finite supersaturation affects cloud dynamics?

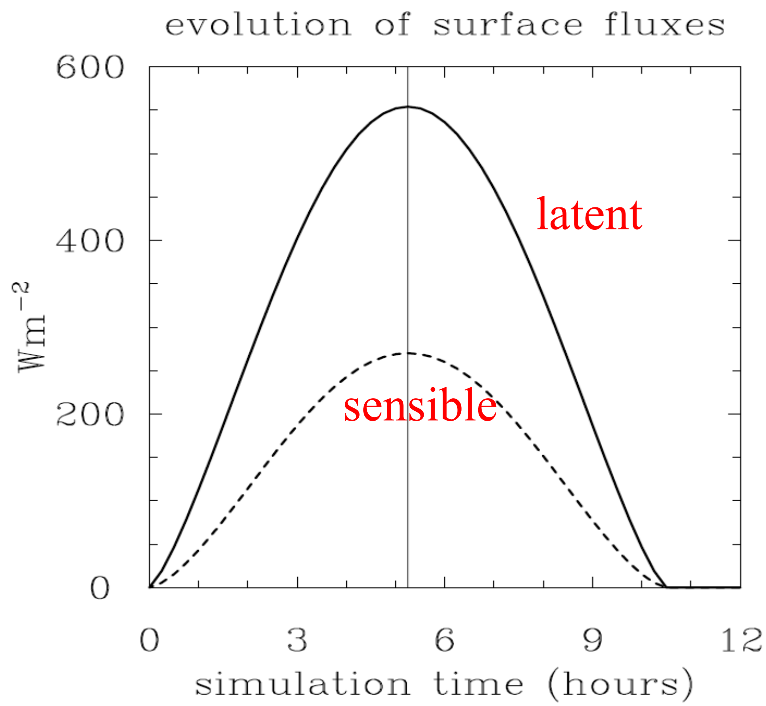
Q. J. R. Meteorol. Soc. (2006), **132**, pp. 317–344 doi: 10.1256/qj.04.147

Daytime convective development over land: A model intercomparison based on LBA observations

By W. W. GRABOWSKI[1]*, P. BECHTOLD[2], A. CHENG[3], R. FORBES[4], C. HALLIWELL[4], M. KHAIROUTDINOV[5], S. LANG[6], T. NASUNO[7], J. PETCH[8], W.-K. TAO[6], R. WONG[8], X. WU[9] and K.-M. XU[3]







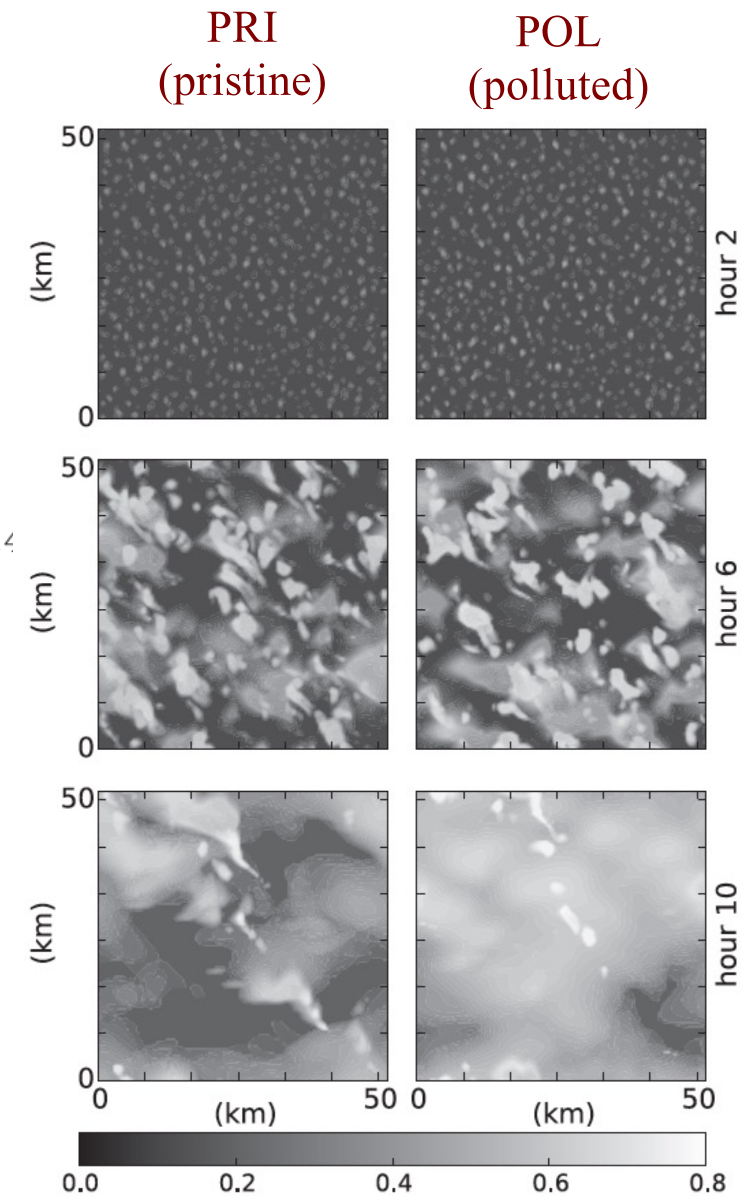
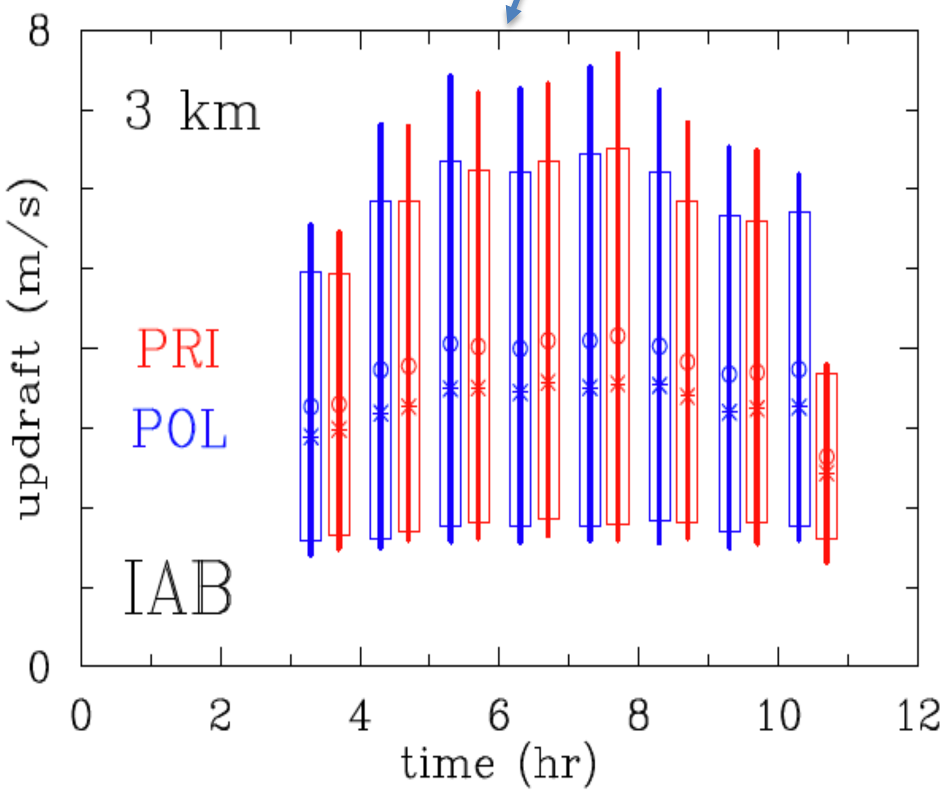
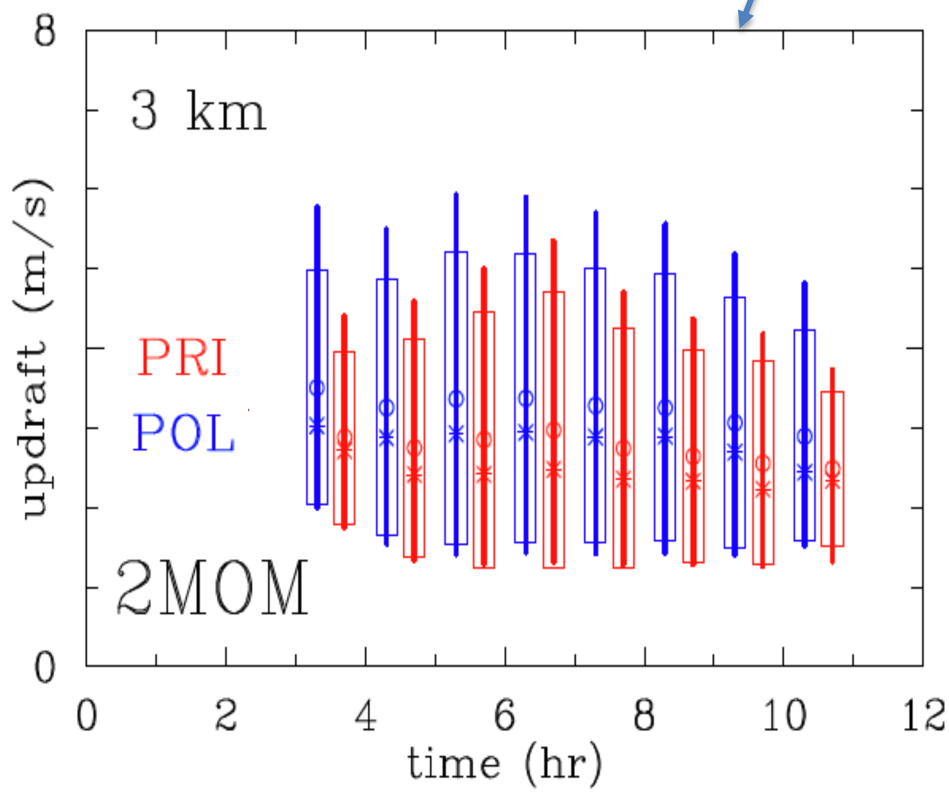


FIG. 2. Albedo at hours (top)–(bottom) 2, 6, and 10 for two simulations from (left) D-PRI and (right) D-POL ensembles.

Does the finite supersaturation affects cloud dynamics?

Comparing updraft statistics at 3 km for cloud field simulations (shallow to deep convection transition) applying saturation adjustment (i.e., S=0) with similar simulations with S prediction.





Hour-by-hour statistics of convective updrafts:

circle – mean, star – median
box – standard deviation
line – 10 to 90 percentile

Grabowski and Morrison *JAS* 2021

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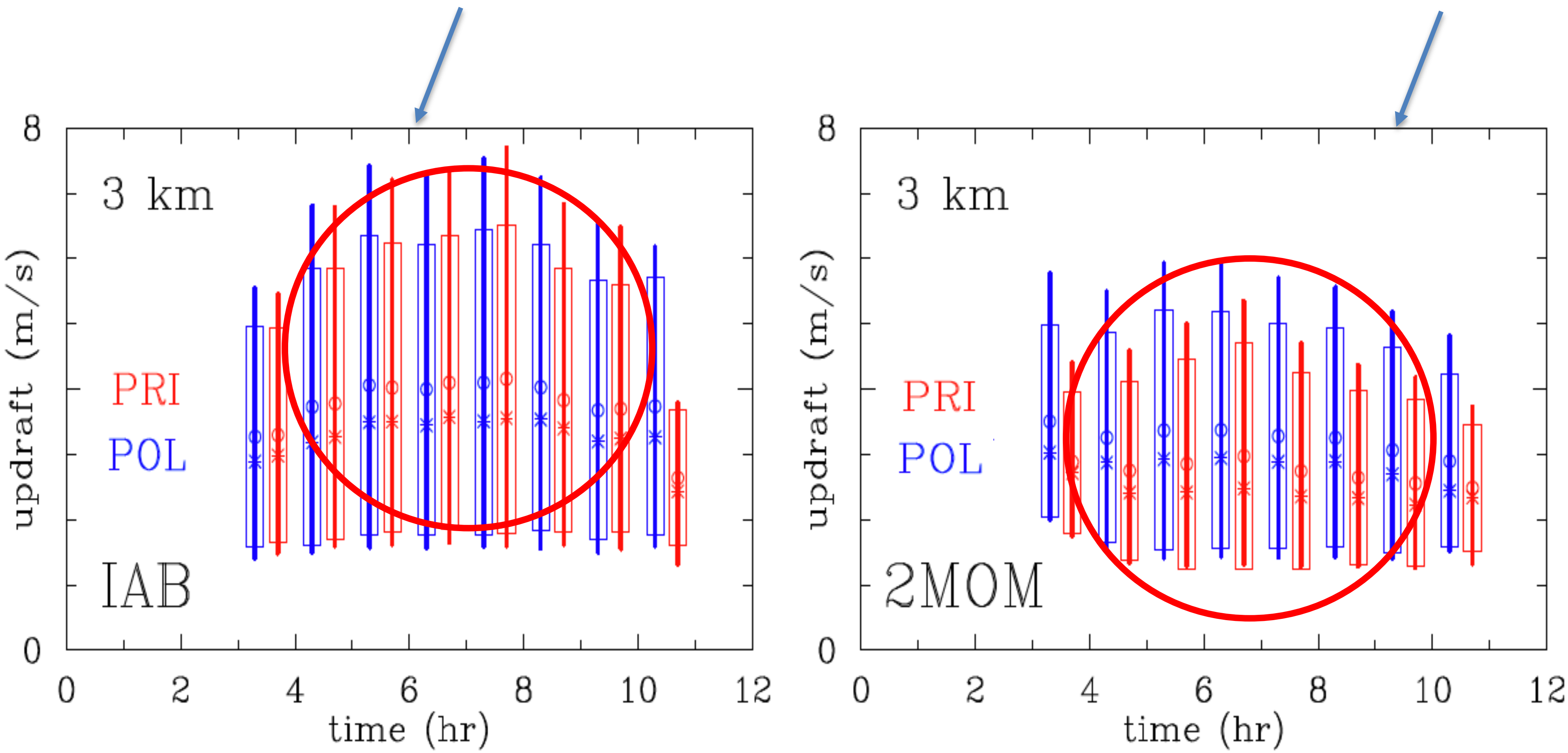

S=0 provides noticeably more buoyancy as shown by the updraft statistics.

Grabowski and Morrison *JAS* 2021

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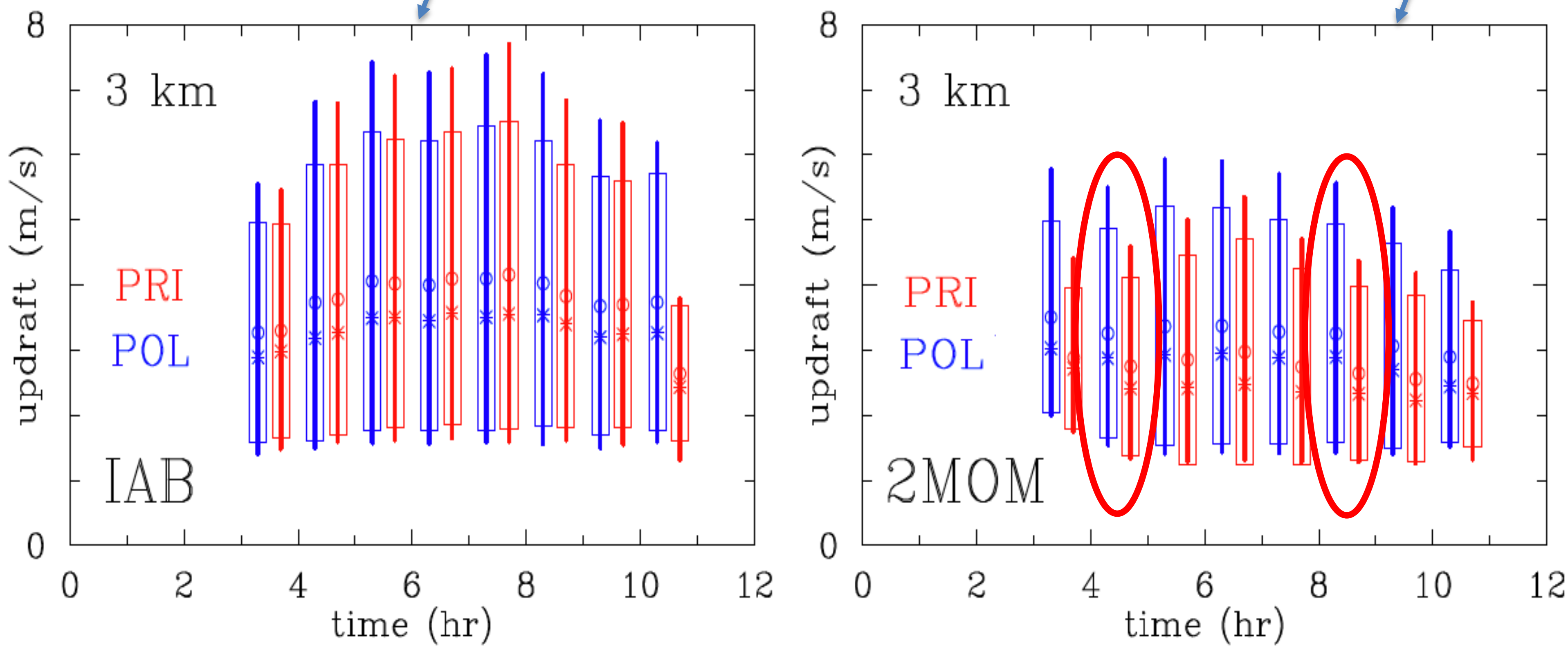

small differences between pristine (high S) and polluted (low S) provides noticeably more buoyancy

Grabowski and Morrison *JAS* 2021

Is there anything better (*more physical*) than saturation adjustment?

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$$\frac{D\theta}{Dt} = \frac{L_v \theta}{c_p T} C_d \qquad S = \frac{q_v}{q_{vs}} - 1$$

$$\frac{Dq_v}{Dt} = -C_d$$

$$C_d \sim N_c \; r \; S$$

$$\frac{Dq_c}{Dt} = \quad C_d$$

τ - phase relaxation time

$$\frac{dS}{dt} = A_1 w - \frac{S}{\tau}$$

$A_1 \sim 10^{-4}$ m^{-1}

TABLE 1. Time constant characterizing supersaturation. (Values of $\tau = 1/(a_2 I)$ s for $p = 771$ mb, $T = 4.3°C$)

Radius (μm)	Droplet concentration (cm^{-3}) 100	300	500	1000
2	14.1	4.7	2.8	1.4
3	8.7	2.9	1.7	0.87
5	4.9	1.6	0.98	0.49
10	2.3	0.77	0.46	0.23

Politovich and Cooper *JAS* 1988

$$\frac{dS}{dt} = A_1 w - \frac{S}{\tau} \qquad \tau \sim 1 \text{ sec}$$

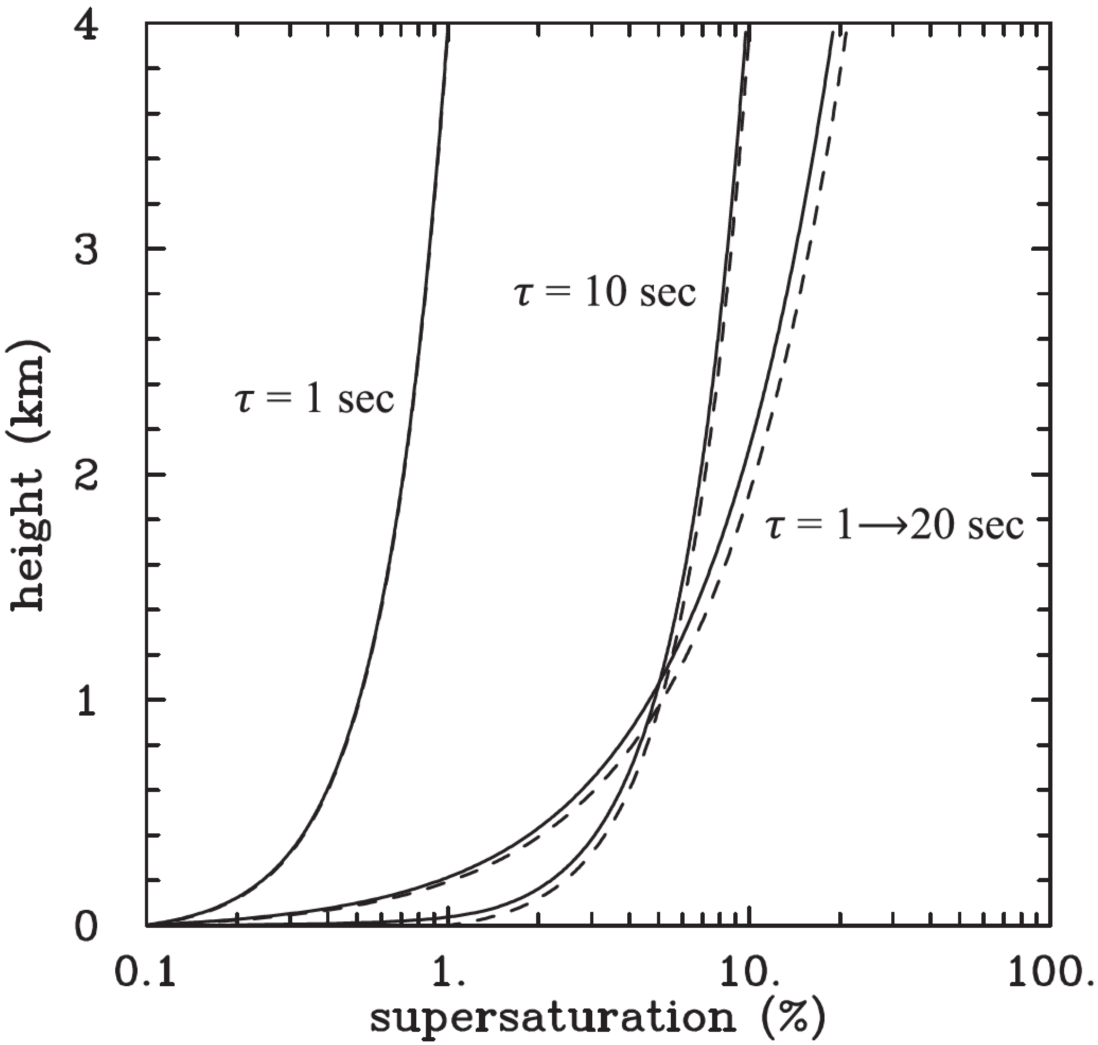
If vertical velocity varies on time scales larger than τ then one may assume:

$$\frac{dS}{dt} = 0; \quad \text{if so, then } \boxed{S = A_1 \text{ w } \tau}$$

This is referred to as the *quasi-equilibrium supersaturation*, S_{qe}

$$\frac{dS}{dt} = A_1 w - \frac{S}{\tau}$$

Updraft w increases from 2 m/s at z=0 to 20 m/s to 4 km



solid lines – solution of S equation

dashed lines – S_{qe} assuming w at that height

$$S_{qe} = A_1 w \tau$$

Grabowski and Morrison *JAS* 2021

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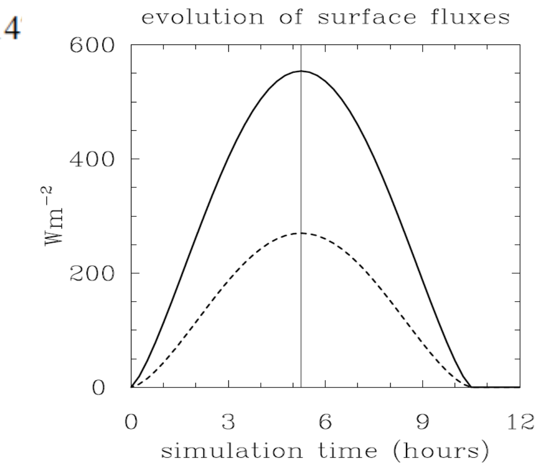

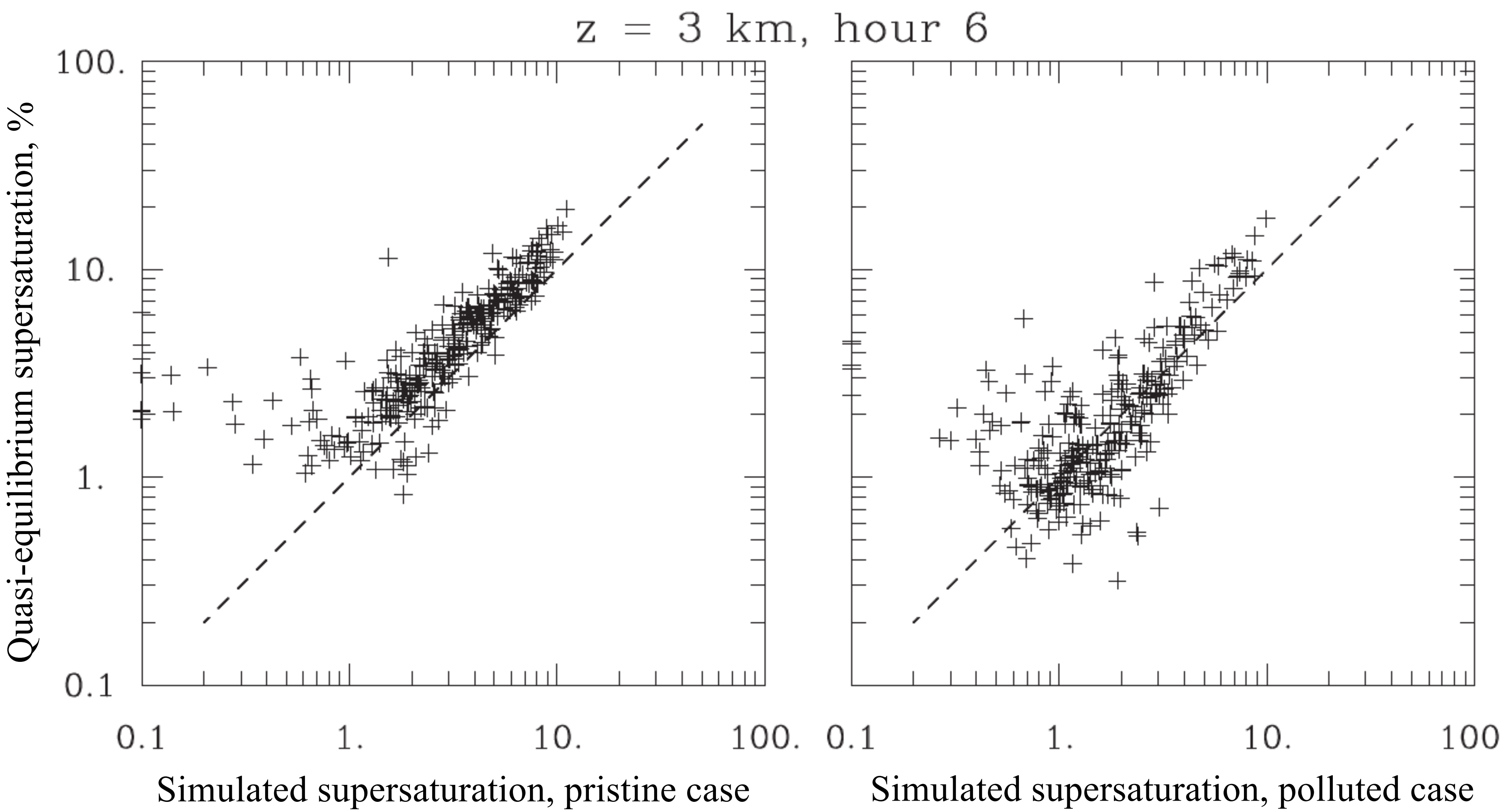

Grabowski and Morrison *JAS* 2021

Summary:

"Nimbostrophy" refers to thermodynamic balances that clouds tend to follow.

Since in-cloud supersaturations are typically small, assuming S=0 provides a powerful modeling methodology, used in early cloud models (stating in 1960ies) and still used today in many practical applications due to its simplicity and numerical stability. However, as shown in Grabowski and Jarecka (2016) and not discussed here, *saturation adjustment* is problematic when modeling entrainment and cloud dilution because cloud water is assumed to evaporate instantaneously.

Since phase relaxation time is typically of the order of 1 sec, a more physically sound approach is to assume that the in-cloud supersaturations is close to the *quasi-equilibrium supersaturation* S_{qe}. However, when the phase relaxation time is large (i.e., CCN activation and cloud droplet formation, cloudy volumes with extremely low droplet concentrations due to washout by rain, entrainment, etc.) S_{qe} provides poor estimate of the in-cloud supersaturation.