Minimal moisture models in convective penetration of a stably stratified layer

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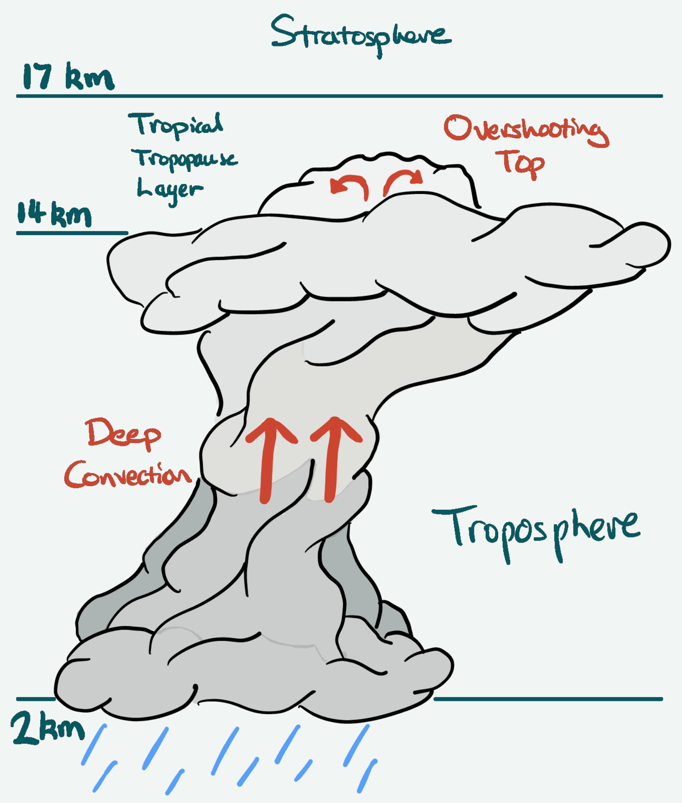
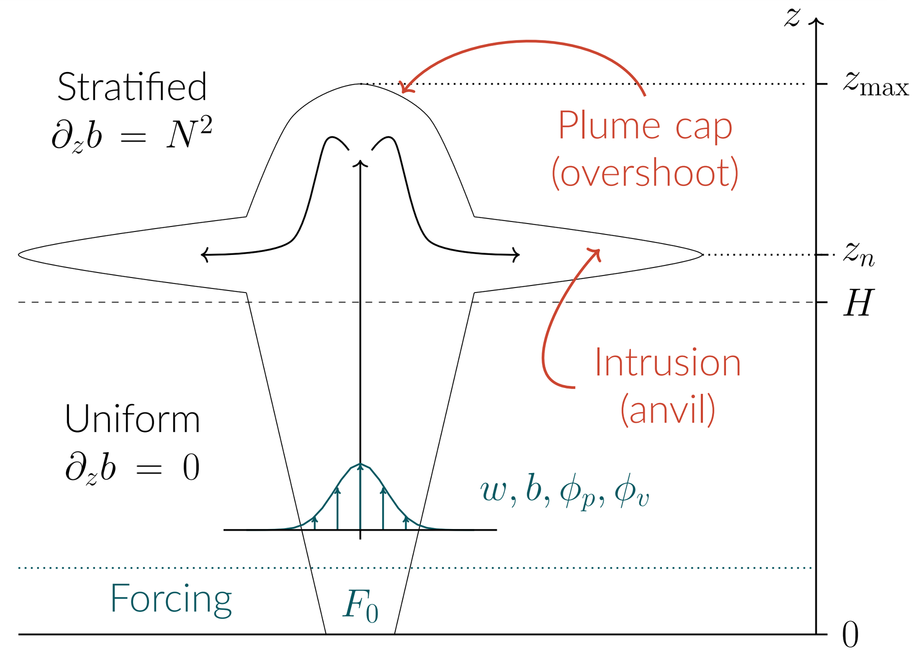
Background

- Consider a buoyant plume penetrating from an unstratified region into a strongly stably stratified layer.
- This is an idealised model of the tropical tropopause layer, where convective overshoots generated by strong thunderstorms can penetrate into the lower stratosphere.
- Net transport of water vapour by overshoots is potentially important (Fueglistaler et al., 2009), but simulation of entire thunderstorm complexes remains computationally expensive (Dauhut et al., 2018).
- Study the mechanisms leading to hydration of the stratified layer using a highly simplified moisture model, involving only three tracers.



Idealised numerical simulations

- Large eddy simulations of Boussinesq equations in a doubly periodic domain. Plume with source buoyancy flux F_0 generated at bottom of uniform layer. Stably stratified layer above has constant vertical buoyancy gradient N^2.
- Plume carries a passive tracer with concentration $\phi_p(\boldsymbol{x}, t)$. The buoyancy is $b(\boldsymbol{x}, t)$, related to potential temperature perturbation $\delta\theta = \frac{\theta_0}{g} b$ from reference value $\theta_0 = 300\,\mathrm{K}$.



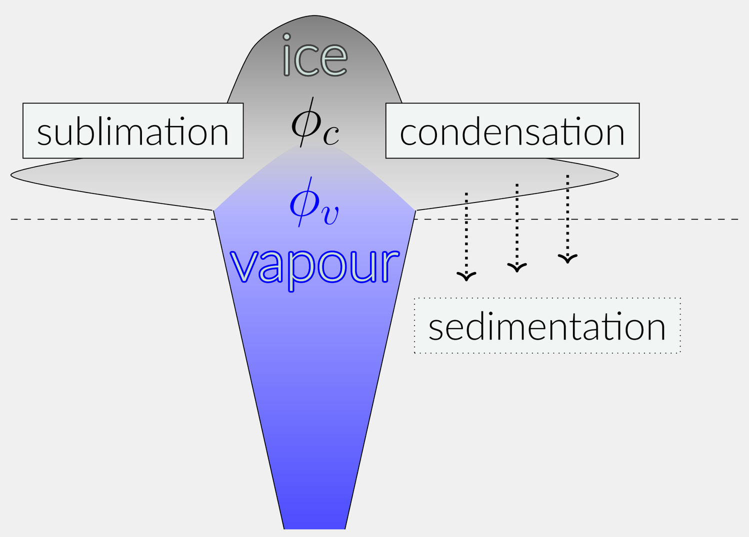
Minimal moisture model

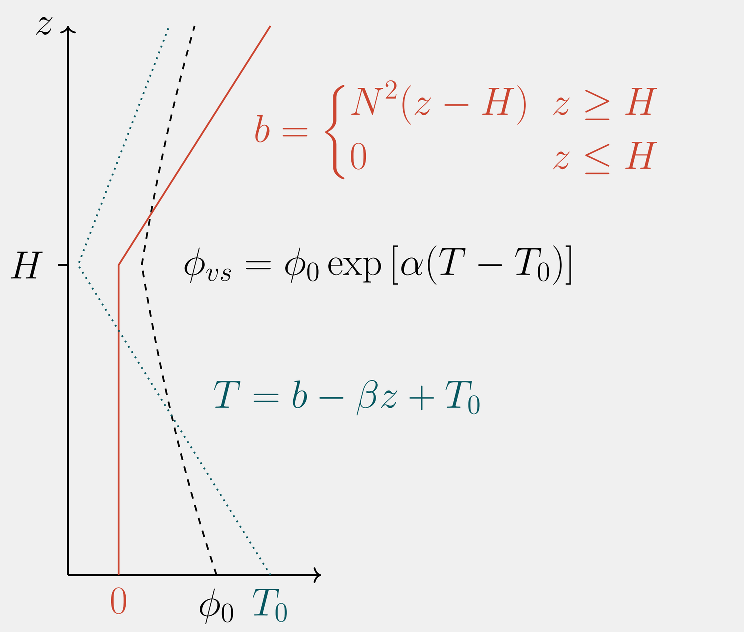
- Plume carries two moist species: water vapour concentration $\phi_v(\boldsymbol{x}, t)$ and ice condensate concentration $\phi_c(\boldsymbol{x}, t)$. Water vapour ϕ_v and ϕ_p have identical forcing, ϕ_c unforced.
- Model three moist processes: condensation of water vapour into ice, sublimation of ice into water vapour, and sedimentation of ice at a constant velocity w_s.

$$\frac{D\phi_v}{Dt} = \kappa\nabla^2\phi_v - \frac{\phi_v - \phi_{vs}}{\tau_m}\mathcal{H}(\max\{\phi_c, \phi_v - \phi_{vs}\})$$

$$\frac{D\phi_c}{Dt} - w_s\frac{\partial\phi_c}{\partial z} = \kappa\nabla^2\phi_c + \frac{\phi_v - \phi_{vs}}{\tau_m}\mathcal{H}(\max\{\phi_c, \phi_v - \phi_{vs}\})$$

- Vapour condenses into ice when its concentration exceeds the saturation vapour concentration ϕ_{vs} which we assume to be a function of temperature, T, alone.



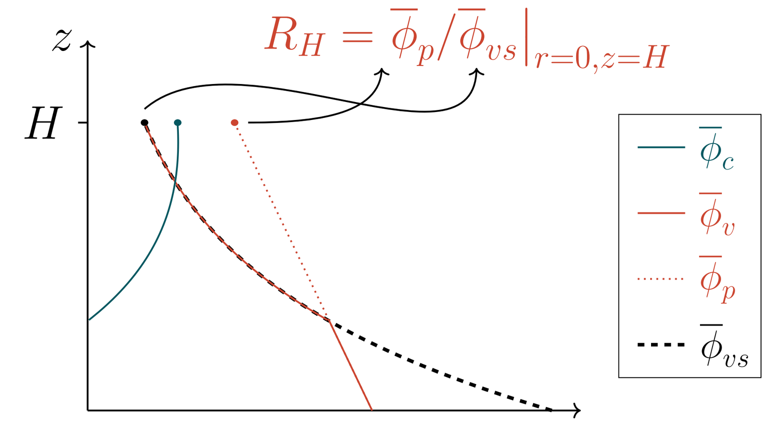


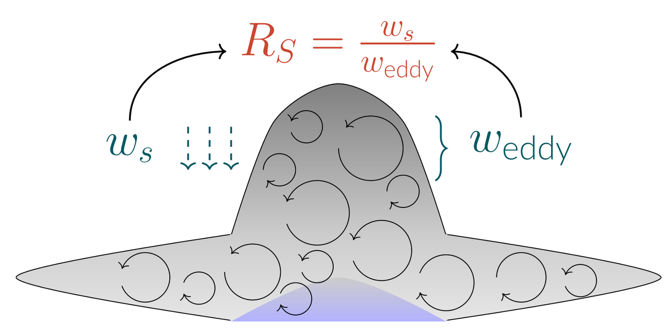
Model assumptions

- Boussinesq: deviations from a neutral profile are small, penetration depth small compared to the density scale height.
- Under Boussinesq assumption, b acts as proxy for θ and height z for pressure p. Then, temperature $T(p, \theta) \equiv T(b, z)$ depends on buoyancy b and height z alone, with lapse rate $-\partial T/\partial z = \beta$.
- Fast autoconversion/accretion: water vapour condenses directly into ice crystals, which rapidly grow to a size that will precipitate (Grabowski, 1998).
- Fast condensation/sublimation: moist processes occur on a timescale τ_m much shorter than the dynamical timescale (Vallis et al., 2019).

Characterising regimes

- The passive tracer ϕ_p acts as the conserved vapour, representing the vapour concentration if moist processes are 'switched off'; the forcing is the same but ϕ_p cannot condense/sublime.

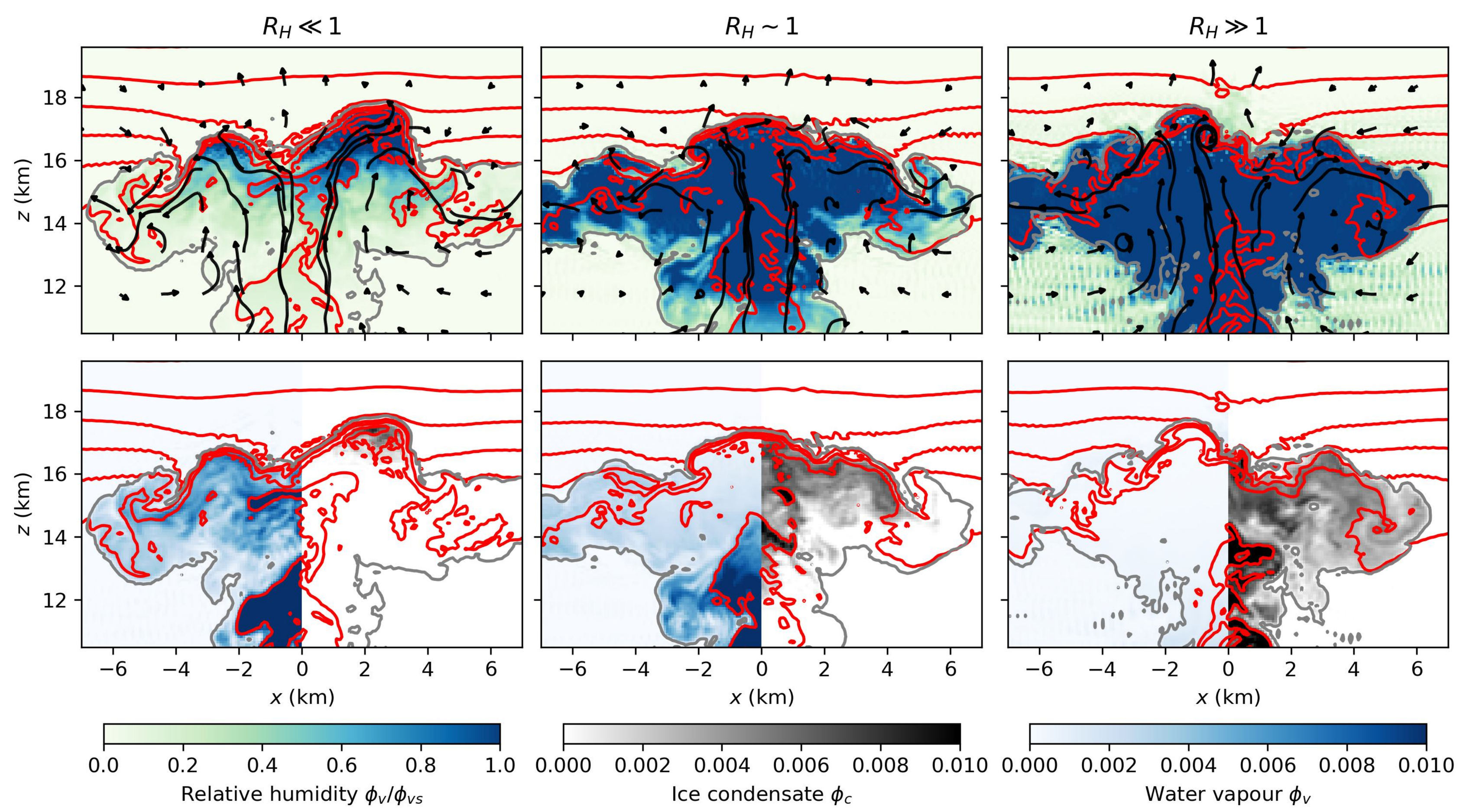




- Importance of saturation quantified by R_H, the relative humidity *in the absence of condensation.*
- $R_H \ll 1$: vapour saturation not reached at penetration. $R_H \gg 1$: vapour condensed into ice before penetration.
- Sedimentation strength R_S: ratio of sedimentation velocity w_s to typical vertical turbulent eddy velocity w_{eddy}.
- $R_S \ll 1$: eddies keep ice in suspension. $R_S \gg 1$: ice sediments out.

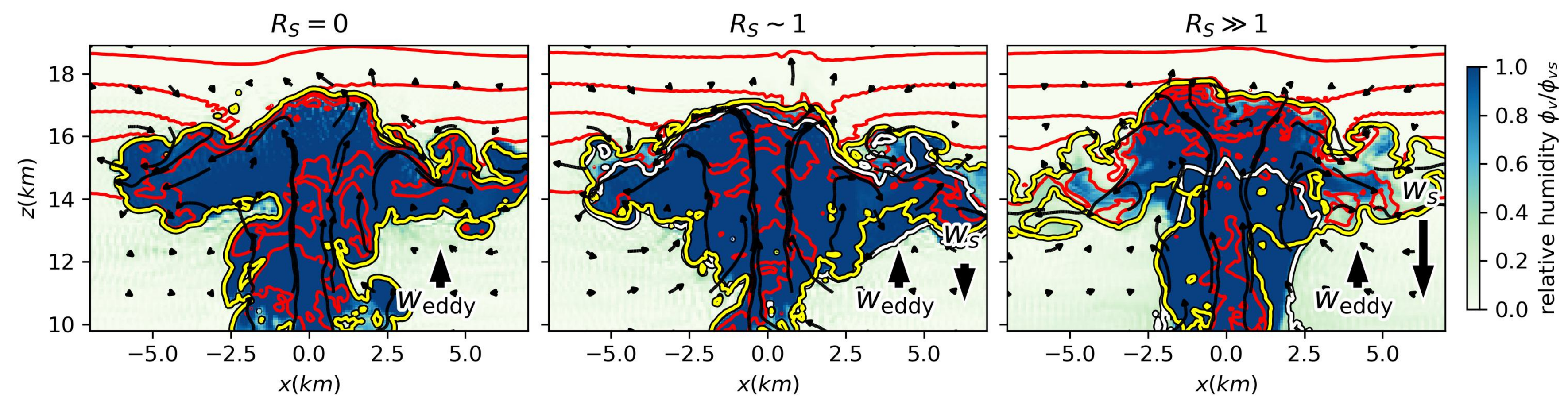
Vapour transport in $\mathrm{R_H}$ regimes

- Vary ϕ_0 in ϕ_{vs}, keeping tracer forcing fixed, so total water is fixed. Choose $\phi_0 = 0.2, 2, 15$ for $R_H \ll 1, \sim 1$ and $\gg 1$ regimes.
- No sedimentation $\implies$ total water conserved, i.e. $\phi_p = \phi_v + \phi_c$. Contour of ϕ_p marking plume edge shown in gray. Contours of θ every $20\,\mathrm{K}$ from $310\,\mathrm{K}$ in red.
- R_H controls partitioning of total water into vapour and ice. When $R_H \gg 1$, plume and immediate surroundings are saturated. Hence ice is abundant, which is mixed into warmer environment & sublimated into vapour. Saturation reached quickly since ϕ_{vs} small.



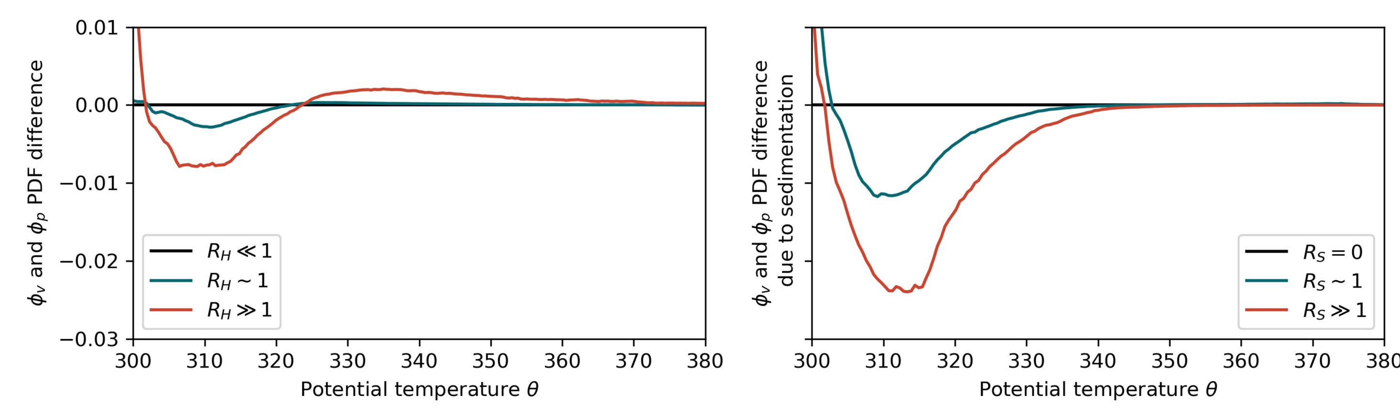
Effect of sedimentation

- Vary w_s to yield regimes $R_S = 0, 1, \gg 1$. $R_H \sim 1$ fixed.
- When $R_S > 0$, conserved vapour ϕ_p (white contours) no longer matches $\phi_v + \phi_c$ (yellow contours) – note increasing discrepancy as R_S increases.



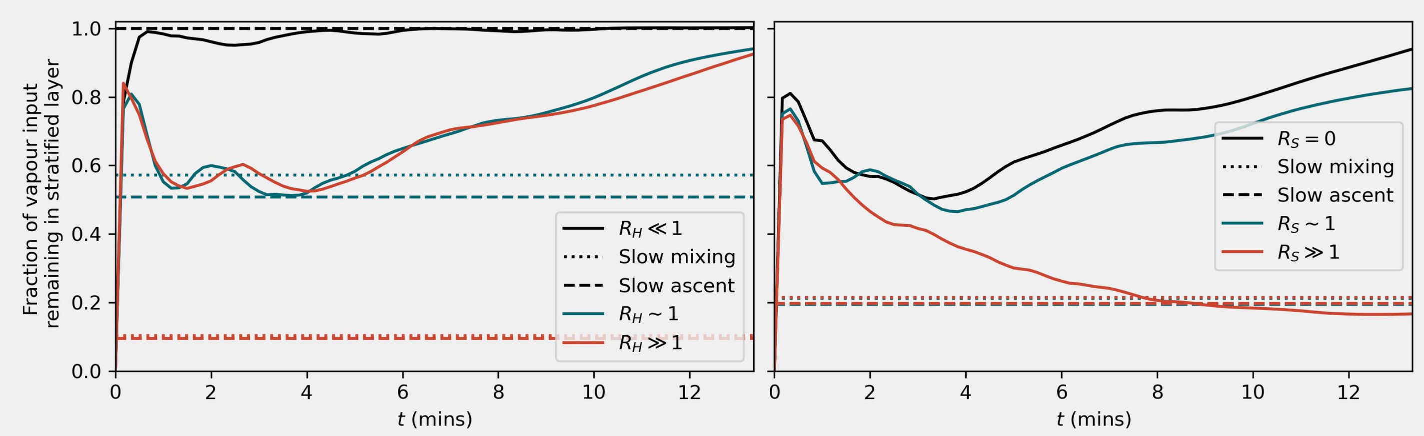
Comparing hydration properties

- Assess hydration of stratified layer via difference between PDFs of ϕ_v and ϕ_p with respect to θ – indicates changes in the vapour distribution due to moist processes.
- When varying R_S (with R_H fixed), subtract the $R_S = 0$ result to isolate changes due to sedimentation.
- Overall, less vapour is present when $R_H \gg 1$, but lies at greater potential temperatures.
- Sedimentation reduces availability of ice to sublimate into vapour, over a large range of θ. Hence reducing the hydrating effect seen when $R_H \gg 1$.



Why simulations?

- Consider proportion of vapour arriving in stratified layer which is retained after moist processes and dynamics equilibriate.
- Compare with semi-quantitative arguments: slow ascent – ϕ_v is limited by minimum in T at $z = H$ (cold point) – and slow mixing – ϕ_v is set by buoyancy distribution at penetration.
- Simulations demonstrate enhanced hydration of the stratified layer when $w_s = 0$ (left). Strong sedimentation reduces this effect (right).



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