The Impact of Aerosol on Cloud Water: A Heuristic Perspective

F. Hoffmann[1], F. Glassmeier[2], and G. Feingold[3]

[1] Ludwig Maximilian University, Germany; [2] Delft University of Technology, Netherlands; [3] Chemical Sciences Laboratory, NOAA, USA

Why are Cloud Water Adjustments Important?

The change in short-wave cloud albedo A due to a change in the aerosol concentration N can be expressed as

$$\frac{\mathrm{d}\ln(A)}{\mathrm{d}\ln(N)} = \frac{1-A}{3}\left[1+\frac{5\,\mathrm{d}\ln(L)}{2\,\mathrm{d}\ln(N)}\right].$$

The first term depicts the increase in A due to more cloud droplets scattering radiation more efficiently at higher N (the Twomey effect). The second term describes so-called cloud water adjustments by $\mathrm{d}\ln(L)\,/\,\mathrm{d}\ln(N) \equiv m$, where L is the vertically integrated cloud water content. An $m > 0$ is usually associated with precipitation suppression due to smaller cloud droplets (the Albrecht effect), while $m < 0$ is considered a result of increased entrainment that evaporates the cloud more readily at higher N (sedimentation- or evaporation-entrainment feedbacks).

The magnitude and sign of m are uncertain: Estimates vary from 0.1 to 0.4 for precipitation suppression at low N, and – 0.2 to – 0.4 for entrainment feedbacks at high N. To better understand the behavior of m and its variability, **we derive a reduced-order model that describes the dynamics of L and N.**

The Reduced-Order Model

To predict the development of L, we consider losses by precipitation and the charge/discharge to the thermodynamic carrying capacity of the system as

$$\frac{\mathrm{d}L}{\mathrm{d}t} = -\frac{L}{\tau_\mathrm{p}} + \frac{L_{\infty,\mathrm{h}}(N)-L}{\tau_\mathrm{t}}, \tag{1}$$

where τ_p is a precipitation timescale, τ_t a thermodynamic timescale, and the thermodynamic carrying capacity is

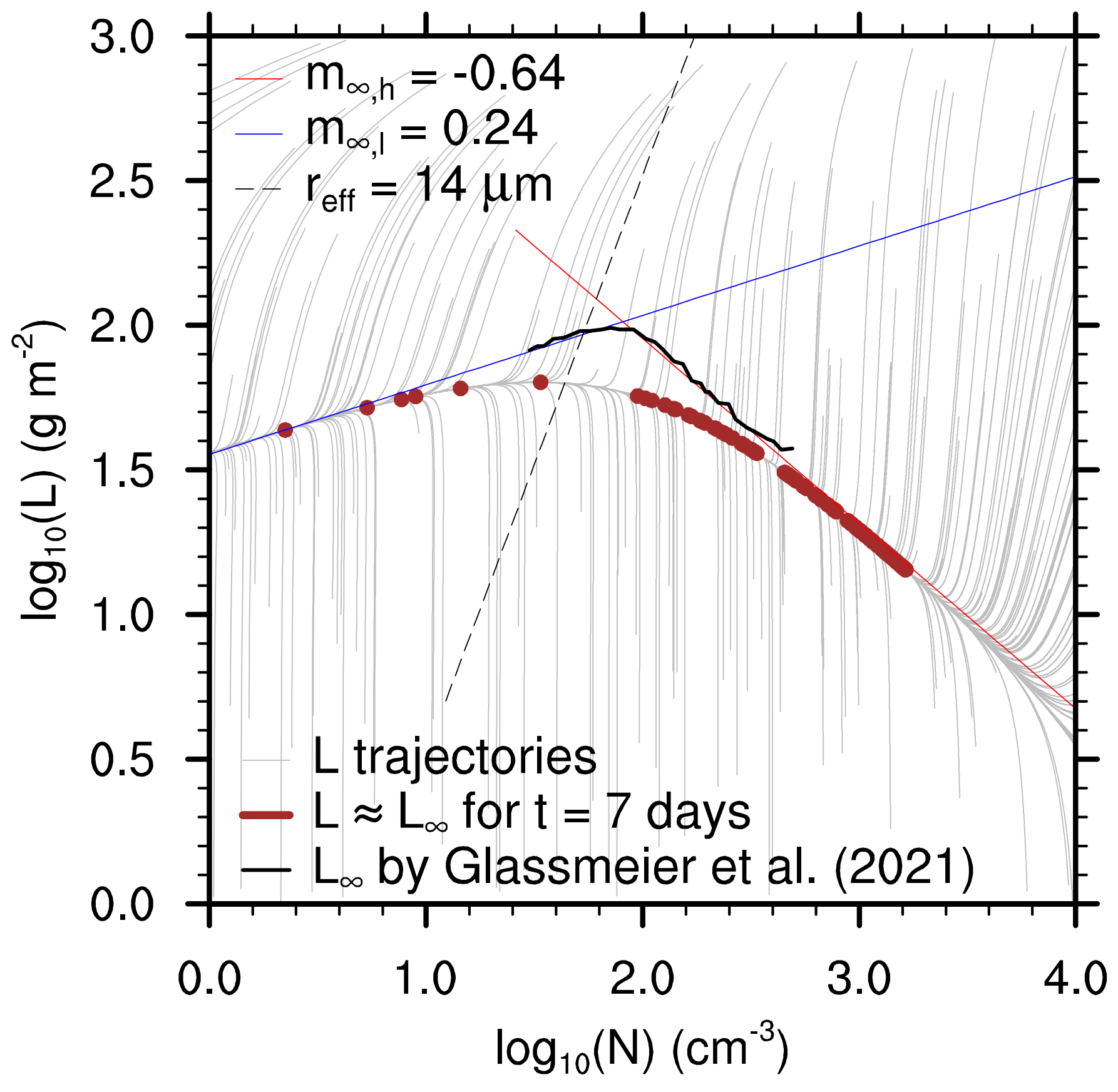
$$L_{\infty,\mathrm{h}} = L_0\left(\frac{N}{N_0}\right)^{m_{\infty,\mathrm{h}}},$$

which is determined by a prescribed $m_{\infty,\mathrm{h}}$. Following Baker and Charlson (1990), changes in N are described by

$$\frac{\mathrm{d}N}{\mathrm{d}t} = c_2 N\left(-\frac{L}{\tau_\mathrm{p}}\right) - c_3 N^2 + S_\mathrm{N}, \tag{2}$$

depicting precipitation scavenging, Brownian coagulation, and potential sources (e.g., sea spray).

The steady-state values of m show a good agreement with the large-eddy simulation (LES) results of Glassmeier et al. (2021).



Cloud Water Adjustments at Low and High N

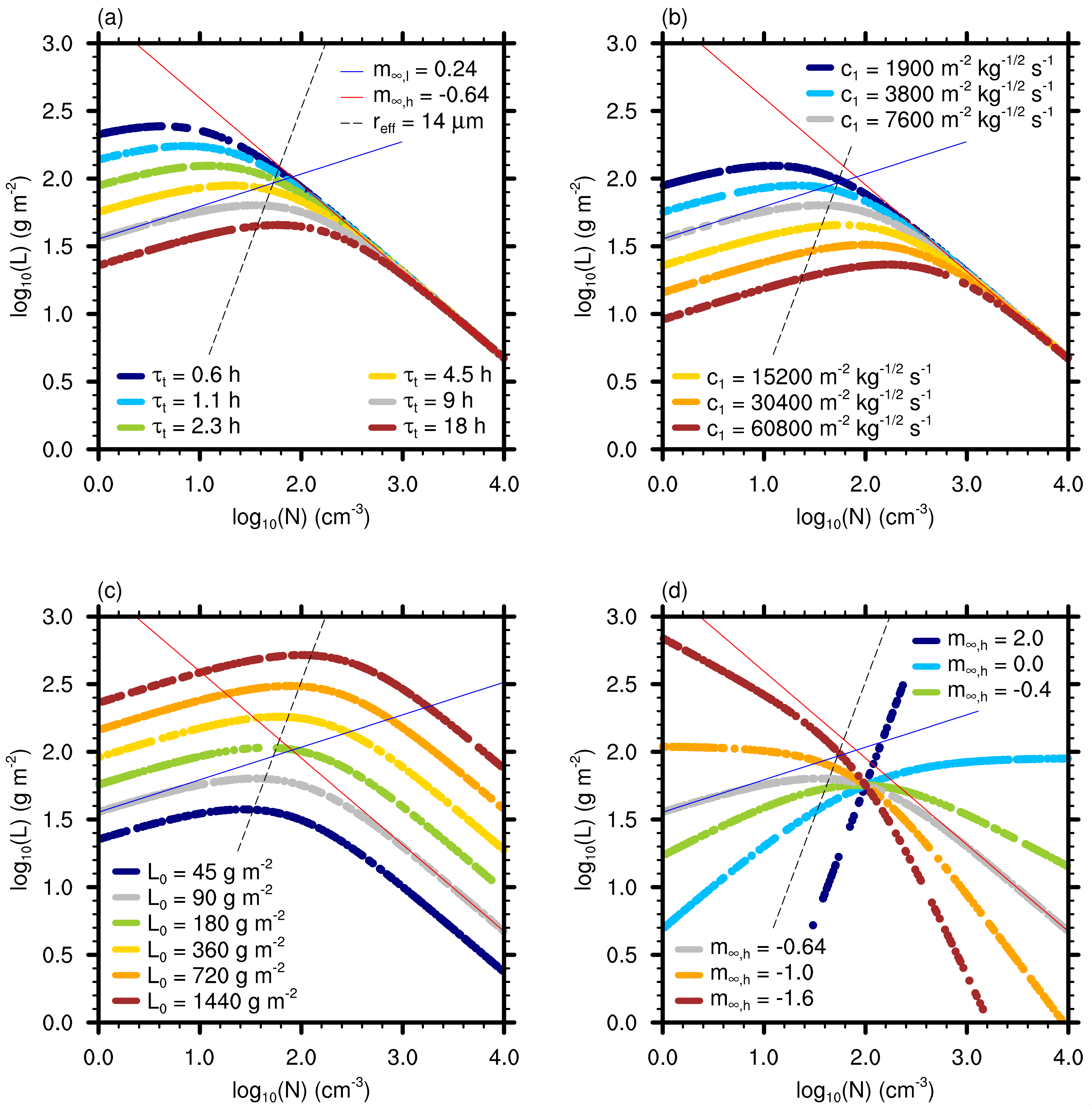
Panels (a) to (c) show the effect of the thermodynamic timescale τ_t, the precipitation timescale $\tau_\mathrm{p} \propto c_1$, and the thermodynamic carrying capacity $L_{\infty,\mathrm{h}} \propto L_0$ on the steady-state distribution of L. **While L varies, the steady-state m_∞ for low and high N are not affected.**

In Panel (d), the impact of the parameter $m_{\infty,\mathrm{h}}$ is assessed. **The m_∞ for high N changes as prescribed, but also the m_∞ for low N adapts in a systematic manner.** What is the relation between the m_∞ for low and high N?

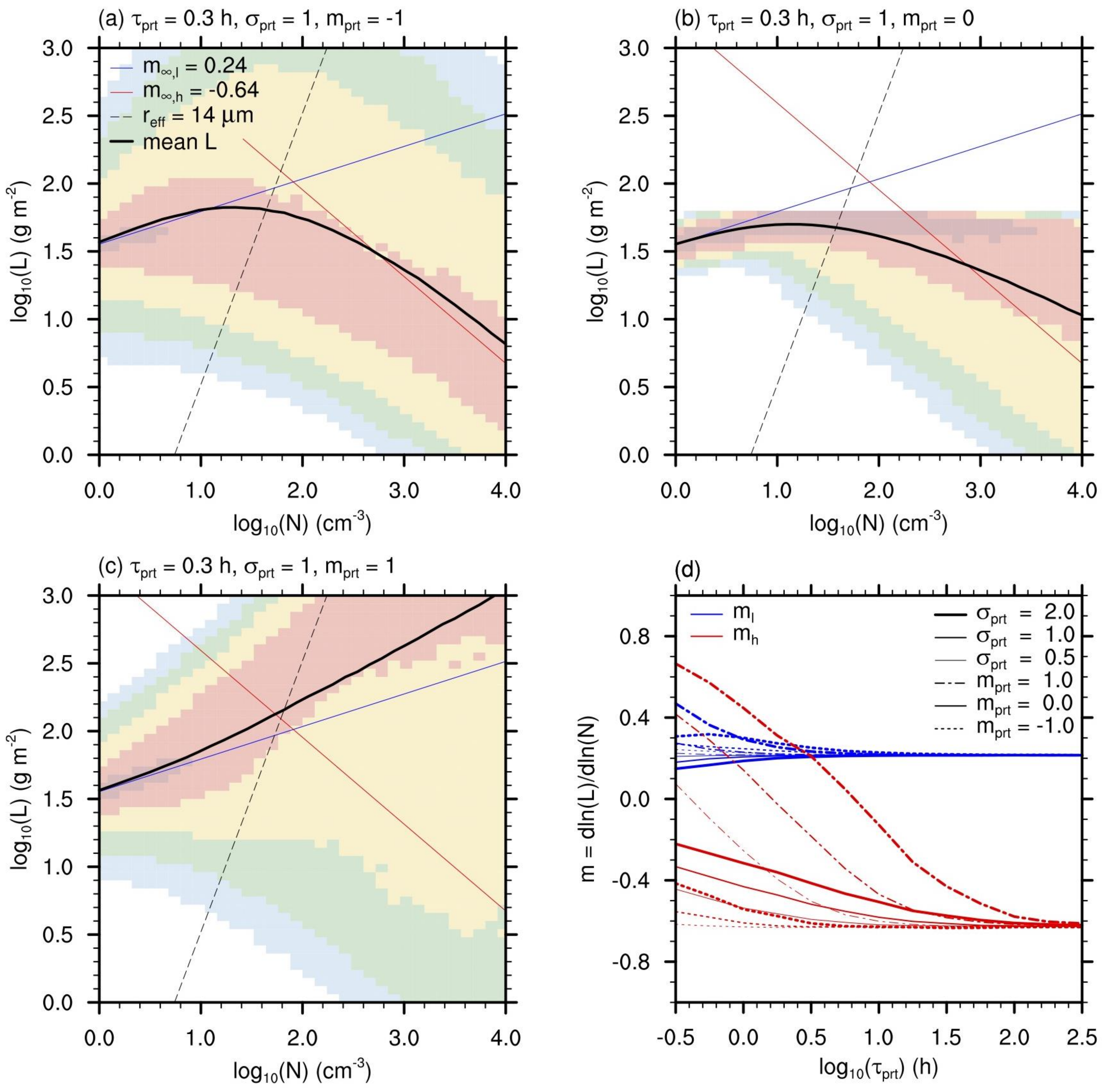
The steady-state solution of (1) relates these slopes as

$$m_{\infty,\mathrm{l}} = \frac{2}{3}\left(m_{\infty,\mathrm{h}}+1\right). \tag{3}$$

Because estimates for $m_{\infty,\mathrm{l}}$ are between 0.1 to 0.4, (3) requires $m_{\infty,\mathrm{h}}$ to be between – 0.9 and – 0.4, which is much more negative than our observational evidence (– 0.2 to – 0.4). **One explanation is that $m_{\infty,h}$ is a function of N, and strengthens (gets more negative) for smaller N (see Yaosheng Chen's poster).**



The Impact of Perturbations



N perturbations exist at various temporal and spatial scales, covering highly localized aerosol emissions such as ship tracks to phenomena on regional scales like volcanic eruptions. At the same time, these perturbations can exhibit correlations with L. **How do perturbations affect m_∞?**

We extended (1) and (2) to stochastic differential equations to assess the effect of perturbations. Perturbations are applied by prescribing a timescale (τ_prt), relative magnitude (σ_prt), and a correlation between L and N (m_prt).

Perturbations tend to increase (weaken) the negative slope at high N. This is due to the larger range of potential positive L perturbations that require more time to be offset by the thermodynamic discharge to $L_{\infty,\mathrm{h}}$. Interestingly, perturbations barely affect the slope at low N. This is because of the comparably short precipitation timescale τ_p that dominates L dynamics at lower N.

Perturbations in L and N are another explanation why cloud water adjustments at high N are weaker than expected from (3).