



Supplement of

Lake-Mountain synergy in convective clouds and precipitation over the Tibetan Plateau: a large-eddy simulation study

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The Froude number characterizes flow behavior near topography and is defined as:

$$Fr = \frac{u}{NH_0} \quad (1)$$

where u (unit: $\text{m}\cdot\text{s}^{-1}$) is the zonal wind speed, H_0 (unit: m) represents the maximum height of topography. N (s^{-1}) denotes the Brunt-Väisälä frequency, which describes the vertical oscillation frequency of an air parcel in a stably stratified atmosphere and serves as a key parameter linking moist convection to terrain-induced dynamical features. It is given by:

$$N = \sqrt{\frac{g}{\theta_v} \left(\frac{d\theta_v}{dz} \right)} \quad (2)$$

where θ_v (unit: K) is the virtual potential temperature and g (unit: $\text{m}\cdot\text{s}^{-2}$) is the gravity acceleration. It is worth noting that the Brunt-Väisälä frequency should be considered in three stages over the development of moist convection: (1) dry air conditions (original formulation), (2) unsaturated moist air (dry adiabatic processes), and (3) saturated moist air (moist adiabatic processes). The moist Brunt-Väisälä frequency is given by (Durran and Klemp, 1982; Liu and Gao, 2003):

$$N_m = \sqrt{\frac{g}{T} \left(\frac{dT}{dz} + \Gamma_m \right) \left(1 + \frac{Lq_s}{RT} \right) - \frac{g}{1+q_w} \frac{dq_w}{dz}} \quad (3)$$

Where the pseudo adiabatic lapse rate, Γ_m , is defined as:

$$\Gamma_m = \Gamma_d \frac{(1+q_w) \left(1 + \frac{Lq_s}{RT} \right)}{1 + \frac{c_{pv}q_s + c_wq_l + \frac{\epsilon L^2 q_s}{c_p R T^2} \left(1 + \frac{q_s}{\epsilon} \right)} \quad (4)$$

In Equations 3 and 4, T (unit: K) is the air temperature, L (unit: $\text{J}\cdot\text{kg}^{-1}$) is the latent heat of vaporization, q_s (unit: $\text{kg}\cdot\text{kg}^{-1}$) is the saturation mixing ratio of water vapor, R (unit: $\text{J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1}$) is the gas constant for dry air, q_w (unit: $\text{kg}\cdot\text{kg}^{-1}$) is the total mixing ratio of water substance, Γ_d (unit: $\text{K}\cdot\text{km}^{-1}$) is the dry-adiabatic lapse rate, c_{pv} (unit: $\text{J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1}$) is the specific heat capacity of water vapor, c_w (unit: $\text{J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1}$) is the specific heat capacity of liquid water, q_l (unit: $\text{kg}\cdot\text{kg}^{-1}$) is the liquid water mixing ratio, c_p (unit: $\text{J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1}$) is the specific heat capacity of dry air, and ϵ is the ratio of gas constants for dry air and water vapor. All constants are set to their respective values in the “Constants.F” module of the CM1 source code.

Equations 3 and 4 are based on the warm cloud assumption and account for the thermodynamic effects of water vapor condensation, liquid water, and air-water vapor mixtures. However, clouds over the TP exhibit distinct microphysical properties (Chang et al., 2019; Zhao et al., 2019). Mixed-phase clouds are more prevalent during the warm season over the TP, and the frequency of precipitation is primarily controlled by these clouds (Chang et al., 2019; Wang et al., 2021). To represent the transition from liquid to ice saturation conditions, a temperature-dependent weighting approach is adopted in mixed-phase clouds. Following standard cloud microphysics practice (Peters et al., 2022; Yau and Rogers,

1996), two reference temperatures were selected: $T_0 = 273.15$ K (0 °C) and $T_i = 233.15$ K (−40 °C), which were
 45 defined by (Pruppacher and Klett, 2010; Yau and Rogers, 1996). The mixed-phase weighting factor $\alpha(T)$, which
 represents the fractional contribution of ice, was prescribed as

$$\alpha(T) = \begin{cases} 0, & T \geq T_0 \\ \frac{T_0 - T}{T_0 - T_i}, & T_i < T < T_0 \\ 1, & T \leq T_i \end{cases} \quad (5)$$

This linear formulation ensures a smooth transition between liquid and ice regimes. Although smoother alternatives exist,
 the linear weighting has been shown to be effective and computationally efficient (Bolton, 1980; Durran and Klemp,
 50 1982; Emanuel, 1991; Peters et al., 2022).

The weighting was then applied to interpolate between liquid- and ice-based thermodynamic properties. The effective
 saturation vapor pressure is calculated as

$$e_s^* = (1 - \alpha)e_{sw}(T) + \alpha e_{si}(T) \quad (6)$$

where $e_{sw}(T)$ and $e_{si}(T)$ are the saturation vapor pressures over water and ice, respectively. For consistency, the
 55 Murphy and Koop (2005) formulations are used to compute these reference values. To ensure physical consistency and
 numerical stability, supersaturation is prevented in the model by limiting the saturation vapor pressure e_s^* to be less than
 the ambient pressure. This approach is commonly employed in cloud microphysics models to avoid unphysical
 supersaturation and to stabilize the numerical solution.

Similarly, the latent heat was interpolated as

$$60 \quad L_s^* = (1 - \alpha)L_v(T) + \alpha L_s(T) \quad (7)$$

where $L_v(T)$ is the latent heat of vaporization and $L_s(T)$ is the latent heat of sublimation.

The corresponding saturation mixing ratio was then obtained as

$$q_s^* = \varepsilon \frac{e_s^*}{p - e_s^*} \quad (8)$$

where p denotes the ambient pressure. This formulation ensures that both thermodynamic stability and moist processes
 65 are represented consistently across the liquid, mixed-phase, and ice regimes.

Using the effective saturation mixing ratio (q_s^*) and the mixed-phase latent heat (L_s^*), the pseudo-adiabatic lapse rate can
 be written as

$$\Gamma_m = \Gamma_d \frac{(1 + q_t) \left(1 + \frac{L_s^* q_s^*}{RT} \right)}{1 + \frac{c_p v q_s^* + c_w q_l + c_i q_i}{c_p} + \frac{\varepsilon (L_s^*)^2 q_s^*}{c_p R T^2} \left(1 + \frac{q_s^*}{\varepsilon} \right)} \quad (9)$$

where q_t is the total hydrometeor mixing ratio, q_s^* is the effective saturation mixing ratio, and L_s^* is the mixed-phase
 70 latent heat. Substituting this mixed-phase lapse rate into the moist stability formulation gives the modified moist Brunt-
 Väisälä frequency (N_m),

$$N_m = \sqrt{\frac{g}{T} \left(\frac{dT}{dz} + \Gamma_m \right) \left(1 + \frac{L_s^* q_s^*}{RT} \right) - \frac{g}{1 + q_t} \frac{dq_t}{dz}} \quad (10)$$

which accounts for both thermal stratification and the vertical distribution of total water under liquid, mixed-phase, and ice saturation conditions.

75 Using the modified mixed-phase moist Brunt-Väisälä frequency, we define the moist flow-stability index as

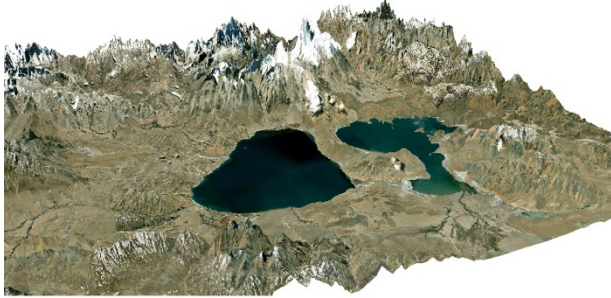
$$I_{mfs} = \frac{u}{N_m H_0}, \quad (11)$$

where u is the cross-mountain wind speed and H_0 is the maximum mountain height. In this study, u and N_m vary spatially, whereas H_0 is retained as a fixed terrain scale. The spatial variability of I_{mfs} therefore reflects changes in the local cross-mountain flow strength relative to mixed-phase moist stability.

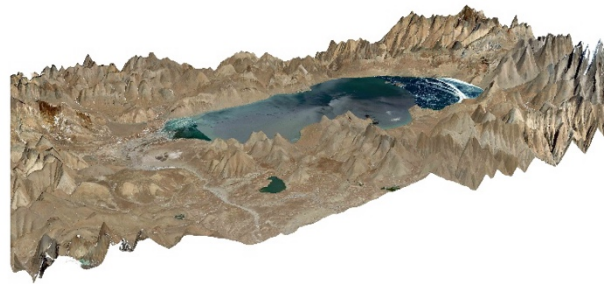
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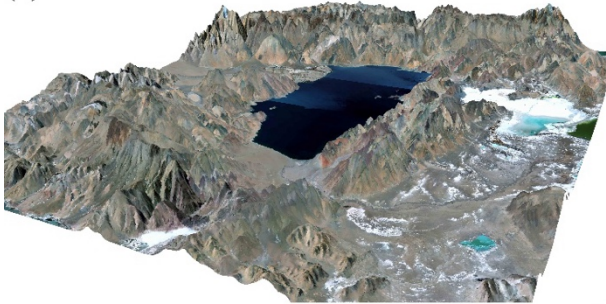
(a) Mapam Yumco



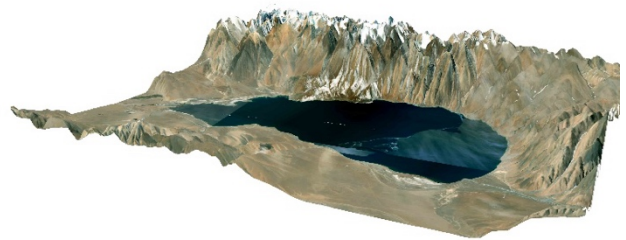
(b) Pumo Yumco



(c) Taro Co



(d) Xuru Co



(e) Zharin Nam Co

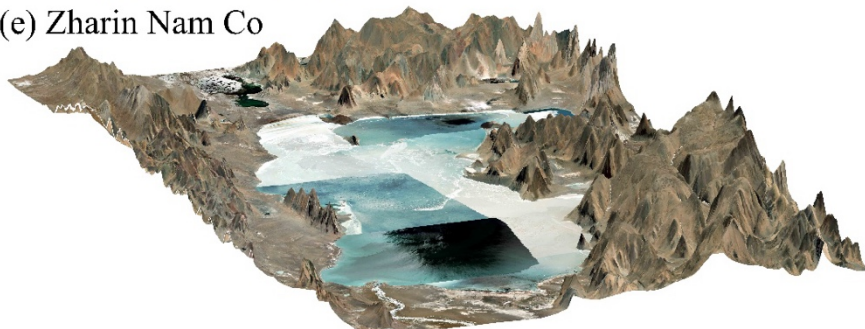


Figure S1. Map of a typical lake-mountain configuration over the TP, derived from Landsat imagery and SRTM topographic dataset.

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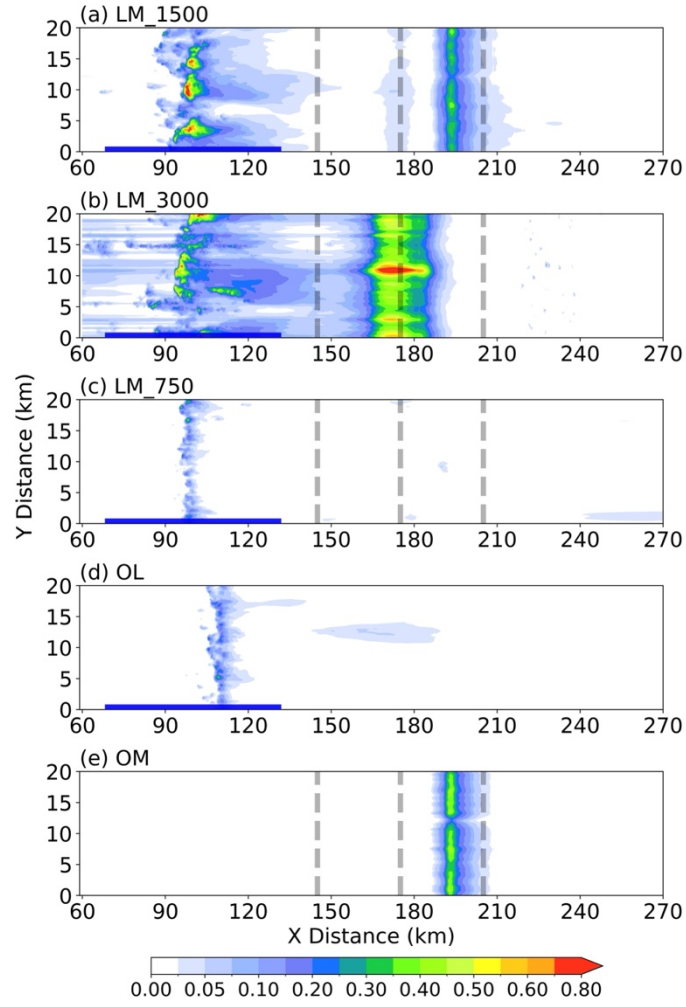


Figure S2. (a) Time-mean precipitation rate (mm h^{-1}) of 20 h in LM_1500. The thick blue line indicates the lake area ($x=70\text{-}130$ km), and the gray dashed lines mark the windward foothill ($x=145$ km), mountain summit ($x=175$ km), and leeward foothill ($x=205$ km).

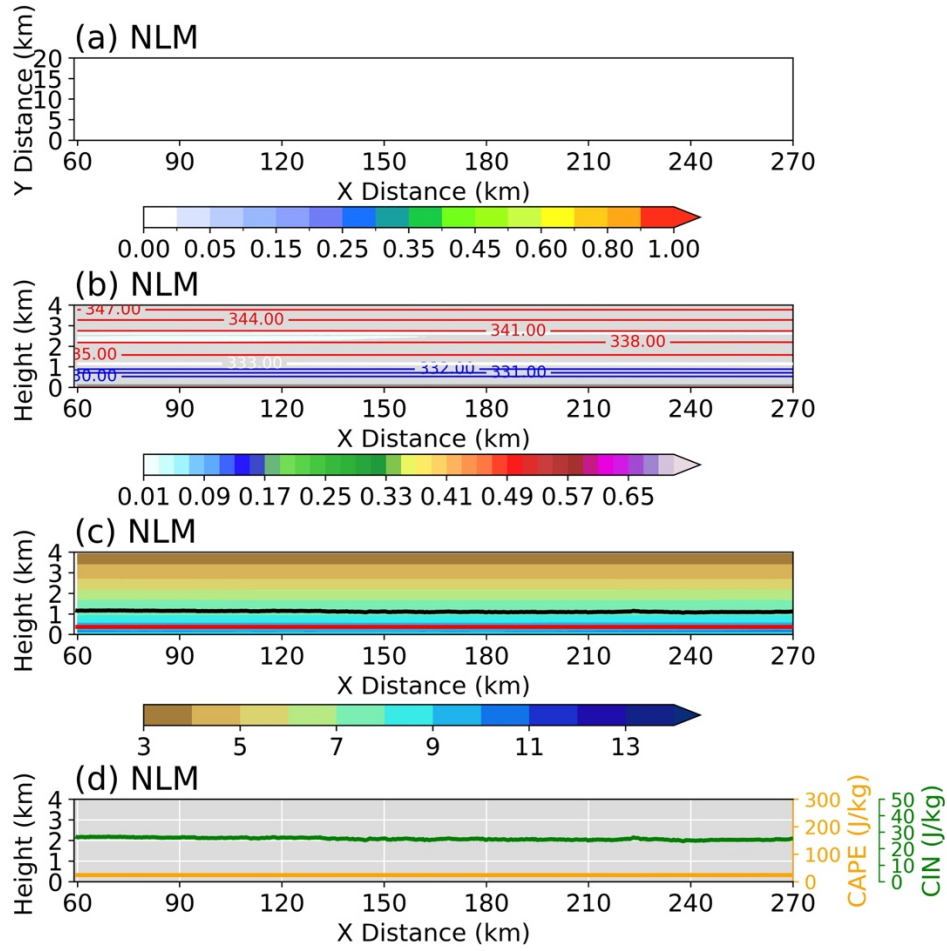


Figure S3. (a) Time-mean precipitation rate (mm h^{-1}) during 7-20 h in NLM (No lake & mountain experiment). (b) Vertical cross section of the 7-20 h time-mean, total hydrometeor mixing ratio (shading, g kg^{-1}) in NLM. Potential temperature (blue-to-red contours, K) and vertical velocity (black contours, only positive values, m s^{-1}) are superimposed. The vertical velocity field is smoothed using a 20×20 -point uniform filter. (c) Vertical cross section of the 7-20 h time-mean, water vapor mixing ratio (shading, g kg^{-1}) in NLM. (d) Surface-based CAPE (orange line, orange right axis) and CIN (green line, green right axis) along the cross section. The LCL (red lines) and LFC (black lines) are shown in panel (c).

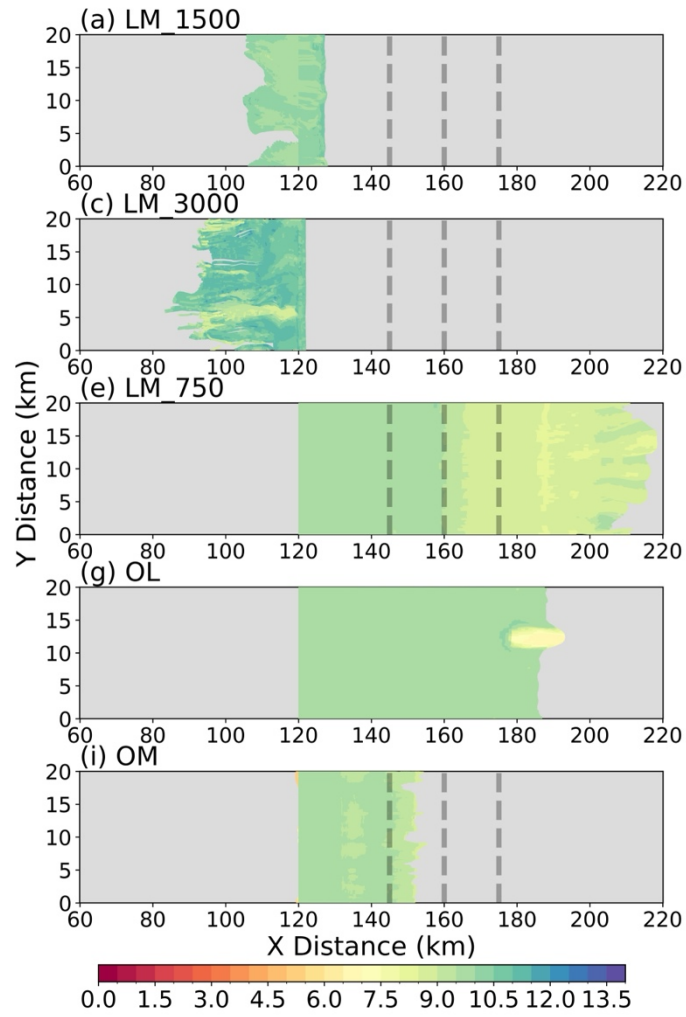


Figure S4. Forward trajectories of water vapor (g kg^{-1}) tracer particles released at $x = 120$ km and $z = 50$ m in

experiments with lake presence, including (a) LM_1500, (b) LM_3000, (c) LM_750, (d) OL and (e) OM. All parcels were released at 7 h of the simulation and integrated forward for 13 h, until 20 h. All trajectories were computed 20 point $\times$ 20 point $\times$ 2 point uniformly filter.

Reference:

- 120 Bolton, D.: The computation of equivalent potential temperature, *Mon. Weather Rev.*, 108, 1046–1053, [https://doi.org/10.1175/1520-0493\(1980\)108%3C1046:TCOEPT%3E2.0.CO;2](https://doi.org/10.1175/1520-0493(1980)108%3C1046:TCOEPT%3E2.0.CO;2), 1980.
- Chang, Y., Guo, X., Tang, J., and Lu, G.: Aircraft measurement campaign on summer cloud microphysical properties over the tibetan plateau, *Sci. Rep.*, 9, 4912, <https://doi.org/10.1038/s41598-019-41514-5>, 2019.
- Durran, D. R. and Klemp, J. B.: On the effects of moisture on the brunt-väisälä frequency, *J. Atmos. Sci.*, 39, 2152–2158, [https://doi.org/10.1175/1520-0469\(1982\)039%3C2152:OTEOMO%3E2.0.CO;2](https://doi.org/10.1175/1520-0469(1982)039%3C2152:OTEOMO%3E2.0.CO;2), 1982.
- 125 Emanuel, K. A.: A scheme for representing cumulus convection in large-scale models, *J. Atmos. Sci.*, 48, 2313–2329, [https://doi.org/10.1175/1520-0469\(1991\)048%3C2313:ASFRCC%3E2.0.CO;2](https://doi.org/10.1175/1520-0469(1991)048%3C2313:ASFRCC%3E2.0.CO;2), 1991.
- Liu, D. and Gao, S.: THE BRUNT-VAISALA FREQUENCY IN A SATURATED ATMOSPHERE AND THE REVISED EQUIVALENT POTENTIAL TEMPERATURE., *Acta Meteorol. Sin.*, (3), 379–383, <https://doi.org/10.11676/qxxb2003.037>, 2003.
- 130 Murphy, D. M. and Koop, T.: Review of the vapour pressures of ice and supercooled water for atmospheric applications, *Q. J. R. Meteorol. Soc. J. Atmos. Sci. Appl. Meteorol. Phys. Oceanogr.*, 131, 1539–1565, 2005.
- Peters, J. M., Mulholland, J. P., and Chavas, D. R.: Generalized lapse rate formulas for use in entraining CAPE calculations, *J. Atmos. Sci.*, 79, 815–836, <https://doi.org/10.1175/JAS-D-21-0118.1>, 2022.
- 135 Pruppacher, H. and Klett, J.: *Microphysics of clouds and precipitation*. Springer, 2010.
- Wang, J., Jian, B., Wang, G., Zhao, Y., Li, Y., Letu, H., Zhang, M., and Li, J.: Climatology of cloud phase, cloud radiative effects and precipitation properties over the tibetan plateau, *Remote Sens.*, 13, 363, <https://doi.org/10.3390/rs13030363>, 2021.
- Yau, M. K. and Rogers, R. R.: *A short course in cloud physics*, Elsevier, 1996.
- 140 Zhao, P., Li, Y., Guo, X., Xu, X., Liu, Y., Tang, S., Xiao, W., Shi, C., Ma, Y., Yu, X., Liu, H., Jia, L., Chen, Y., Liu, Y., Li, J., Luo, D., Cao, Y., Zheng, X., Chen, J., Xiao, A., Yuan, F., Chen, D., Pang, Y., Hu, Z., Zhang, S., Dong, L., Hu, J., Han, S., and Zhou, X.: The tibetan plateau surface-atmosphere coupling system and its weather and climate effects: The third tibetan plateau atmospheric science experiment, *J. Meteorol. Res.*, 33, 375–399, <https://doi.org/10.1007/s13351-019-8602-3>, 2019.