Quantifying the contribution of moisture recycling to precipitation at Lake Taihu using isotope method

Yongbo Hu

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PART 1

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Outline

- 1 Introduction
- 2 Site and methods
- 3 Results and discussion
- 4 Conclusions

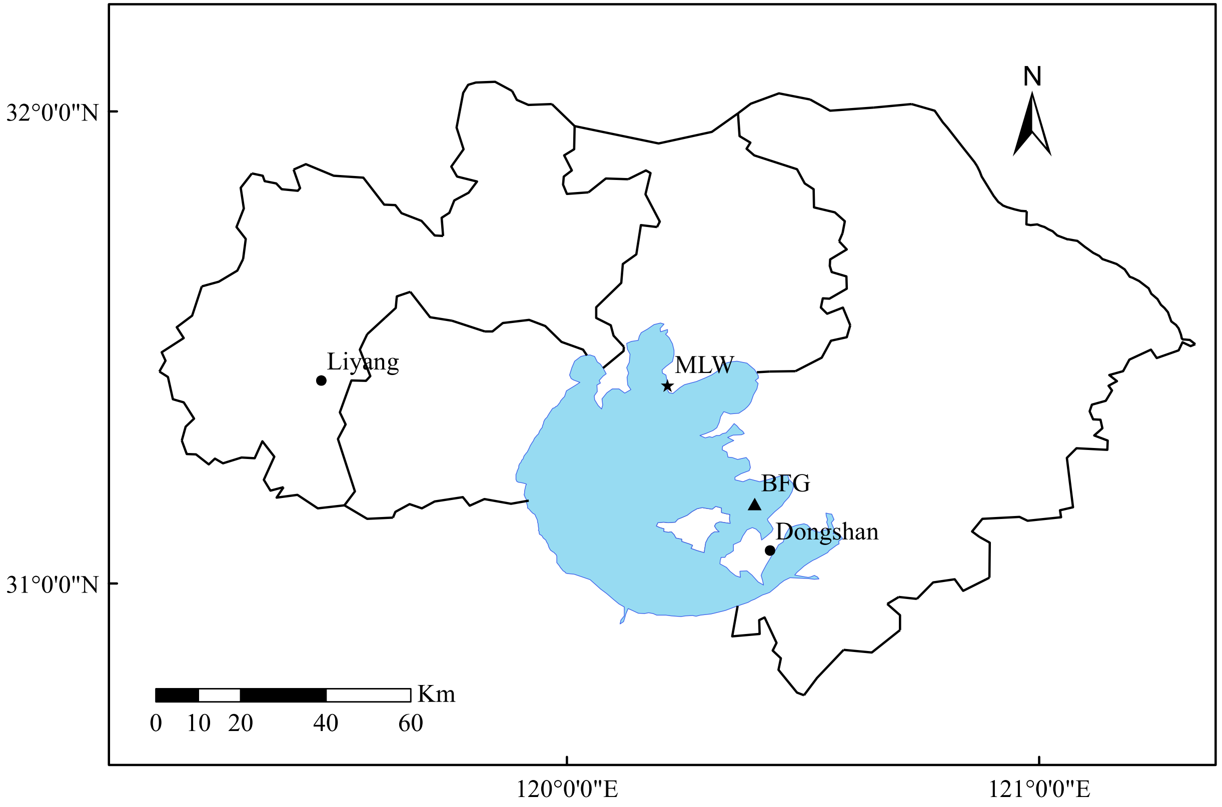
1 Introduction

- Precipitation plays an important role in the global and local water cycle. Horizontal transport (advection, ~90%) and local evapotranspiration (moisture recycling, ~10%) are two main sources of precipitation (Trenberth, 1999).

- However, at lake region, lake evaporation can increase the precipitation amount of the downwind sites, e.g. at the Great Lakes which up to 25% (Bryan et al., 2014; Corcoran et al., 2018). How about Lake Taihu?

2 Sites and methods

2.1 Sites



Liyang, Dongshan: Precipitation isotope
MLW: Water vapor isotope and meteorology
BFG: Lake surface temperature

2.2 Two-component mixing model

$$f = \frac{d_{\mathrm{down}} - d_{\mathrm{up}}}{d_{\mathrm{E}} - d_{\mathrm{up}}}$$

- *f*: contribution of the Lake Taihu evaporation
- d_{down}: *d*-excess of the precipitation at the cloud base of the downwind site (Liyang)
- d_{up}: *d*-excess of the precipitation at the cloud base of the upwind site (advected water vapor, Dongshan)
- d_{E}: *d*-excess of the Lake Taihu evaporation

3.3 Sub-cloud evaporation effect

The Rayleigh equilibrium slope a_{T} (Criss, 1999)

- Difference between the slope of the LWML and the Rayleigh equilibrium slope Δa can be used to test whether the precipitation isotope is influenced by the sub-cloud evaporation effect. (Salamalikis et al., 2016)

$$a_T = \frac{(\alpha_{\mathrm{eq}}^{\mathrm{D}} - 1)(\delta \mathrm{D}_{\mathrm{p}} + 1000)}{(\alpha_{\mathrm{eq}}^{18} - 1)(\delta^{18}\mathrm{O}_{\mathrm{p}} + 1000)}$$

- α_{eq}: the equilibrium fractionation factor of D/^{18}O for the liquid-vapor transition.

3.3 Sub-cloud evaporation effect

$$\Delta d = d_{\mathrm{cb}} - d_{\mathrm{gr}} = F_D - 8 \times F_O$$

Δd: d-excess difference between the cloud base (d_{cb}) and the ground (d_{gr}); F can be expressed as,

$$F = (1 - \frac{\gamma}{\alpha})(f^{\beta} - 1)$$

α is the equilibrium fractionation, f is the remaining fraction after evaporation. γ and β can be expressed as,

$$\gamma = \frac{\alpha h}{1 - \alpha (D/D')^{k}(1 - h)}$$

$$\beta = \frac{1 - \alpha (D/D')^{k}(1 - h)}{\alpha (D/D')^{k}(1 - h)}$$

h is relative humidity; D/D' is the ratio of the lighter and heavier isotope diffusion coefficient; k is an empirical constant, 0.58.

(Stewart, 1975)

3.4 Craig-Gordon model

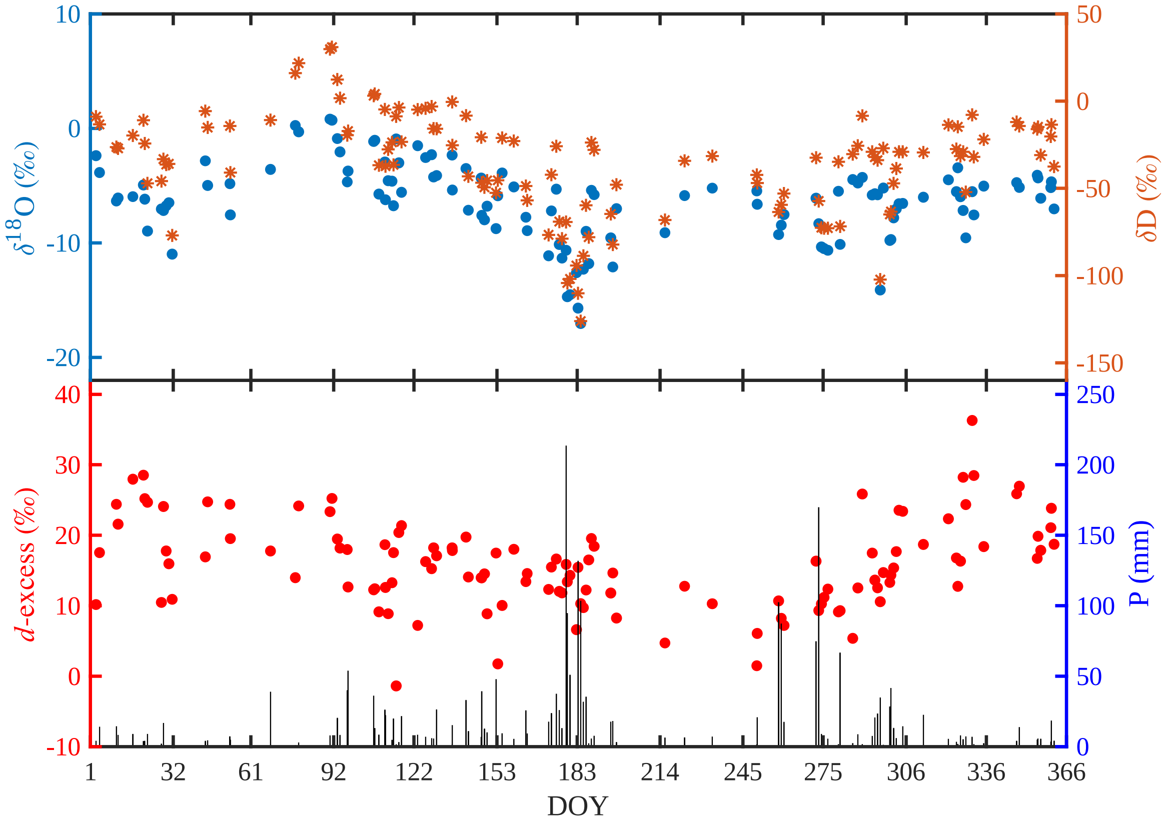
$$\delta_{\mathrm{E}} = \frac{\alpha_{\mathrm{eq}}^{-1}\delta_{\mathrm{L}} - h\delta_{\mathrm{v}} - \varepsilon_{\mathrm{eq}} - (1-h)\varepsilon_{\mathrm{k}}}{1 - h + 0.001(1-h)\varepsilon_{\mathrm{k}}}$$

- α_{eq}: equilibrium fractionation factor;
- δ_{L}: isotopic composition of the lake surface water;
- h: relative humidity in reference to lake surface temperature;
- δ_{V}: isotopic composition of the atmospheric water vapor;
- $\varepsilon_{\mathrm{eq}} = 10^3(1 - 1 / \alpha_{\mathrm{eq}})$: equilibrium fractionation factor in delta notation (‰);
- ε_{k}: kinetic fractionation factor in delta notation (‰).

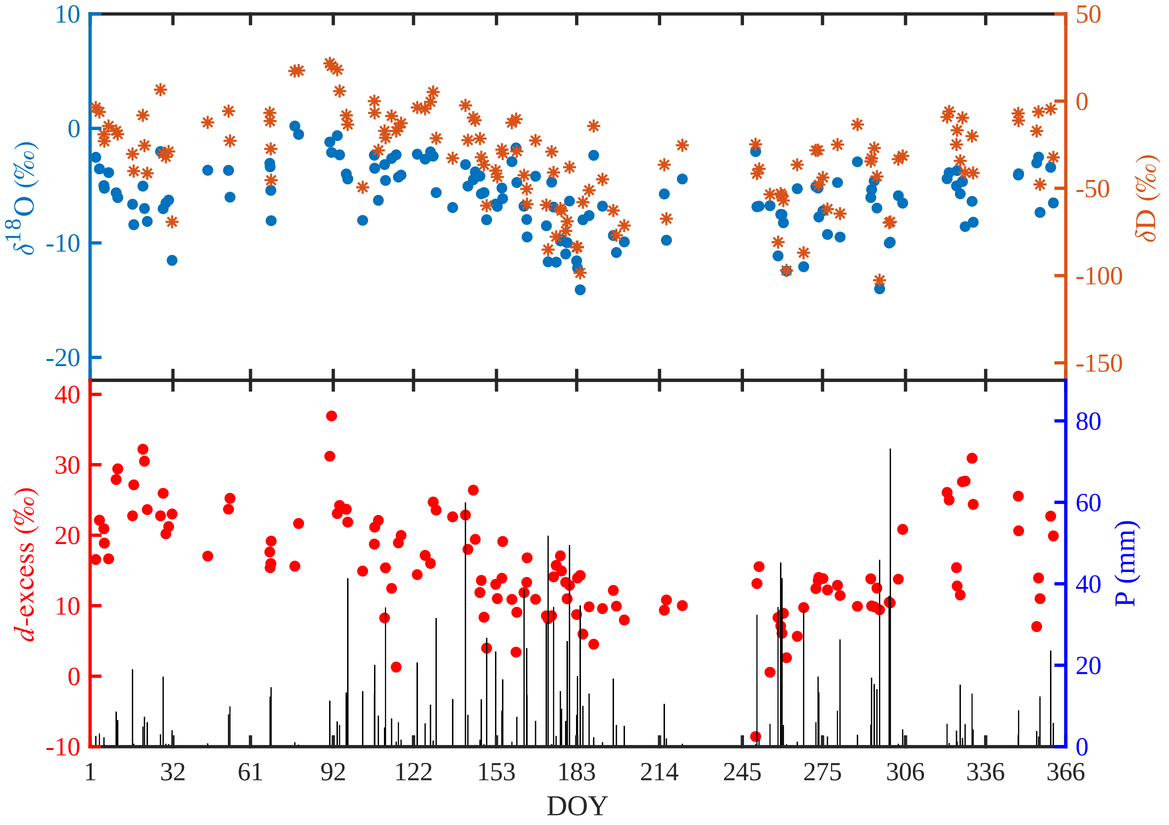
3 Results and discussion

3.1 Basic information of precipitation isotope

3.1.1 Temporal variation of precipitation isotope

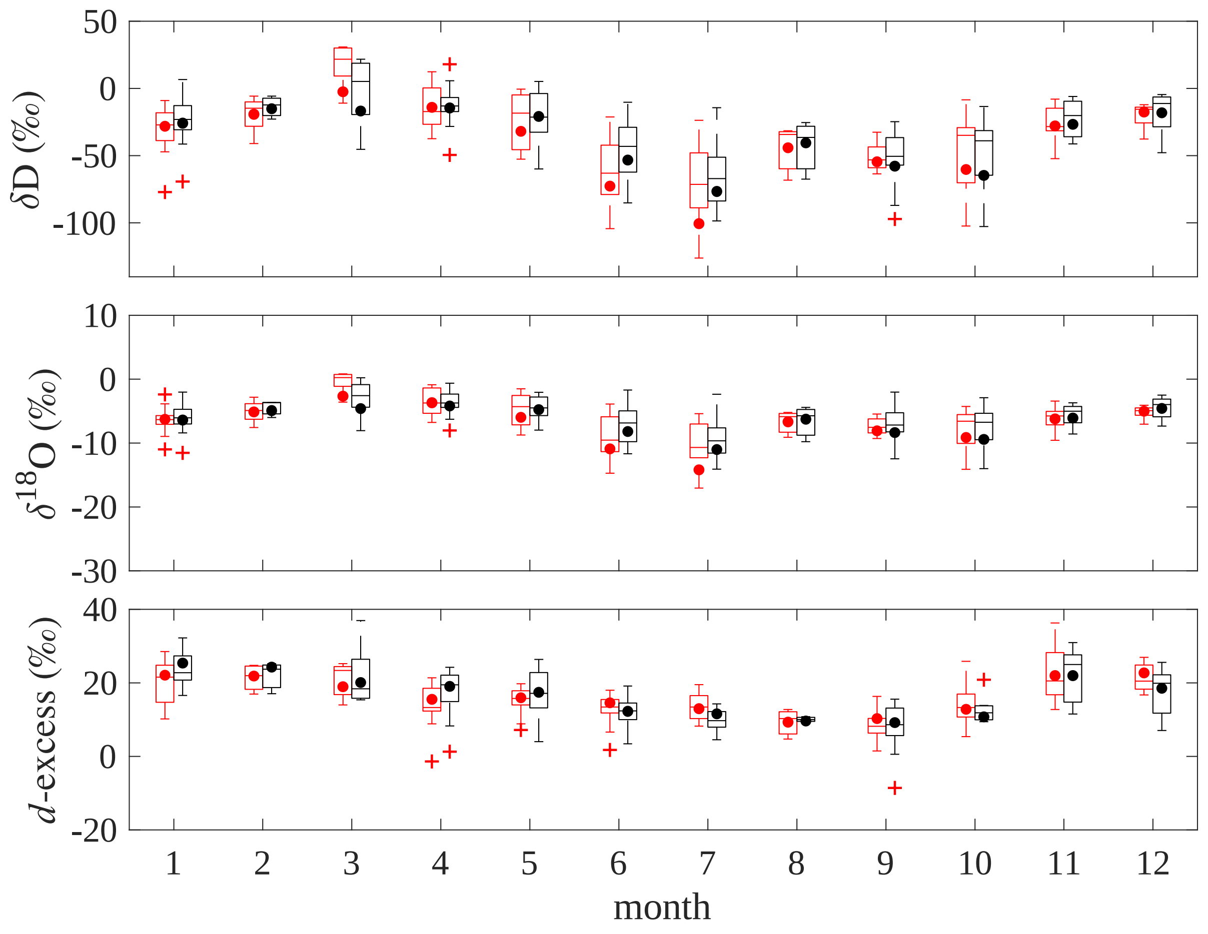


Liyang



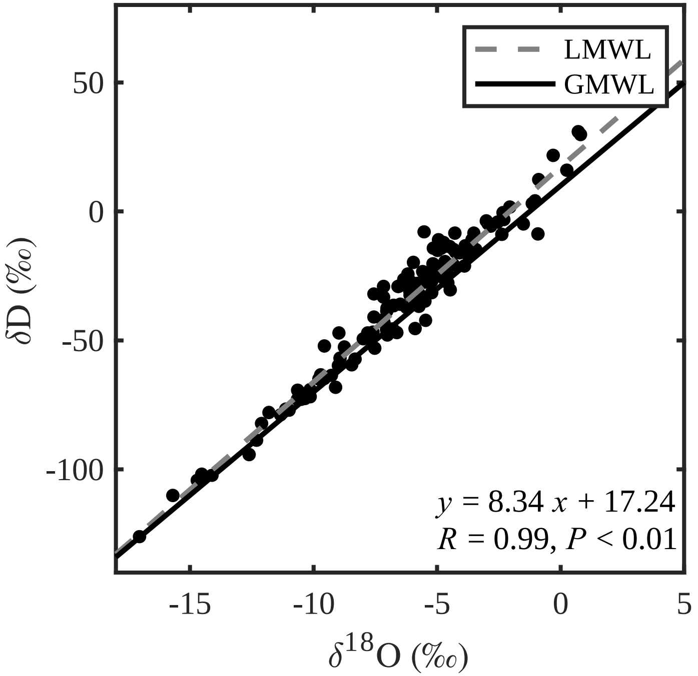
Dongshan

3.1.2 Spatial variation

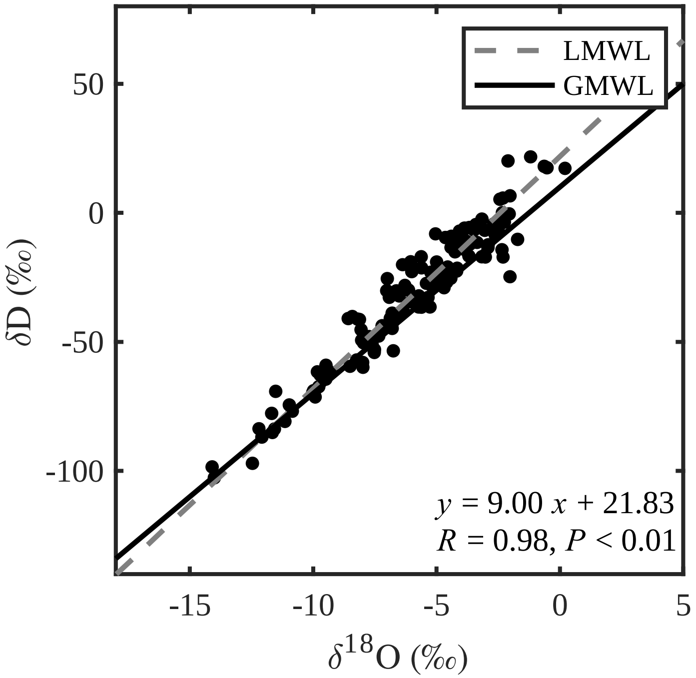


Red box: Liyang
Black box: Dongshan

3.1.3 Local meteoric water line



Liyang



Dongshan

3.2 Sub-cloud evaporation

Comparison between the slope of LWML and the Rayleigh equilibrium slope a_T during the summer monsoon period

site	a_{LWML}	a_T
Dongshan	8.05	8.64
Liyang	7.90	8.52

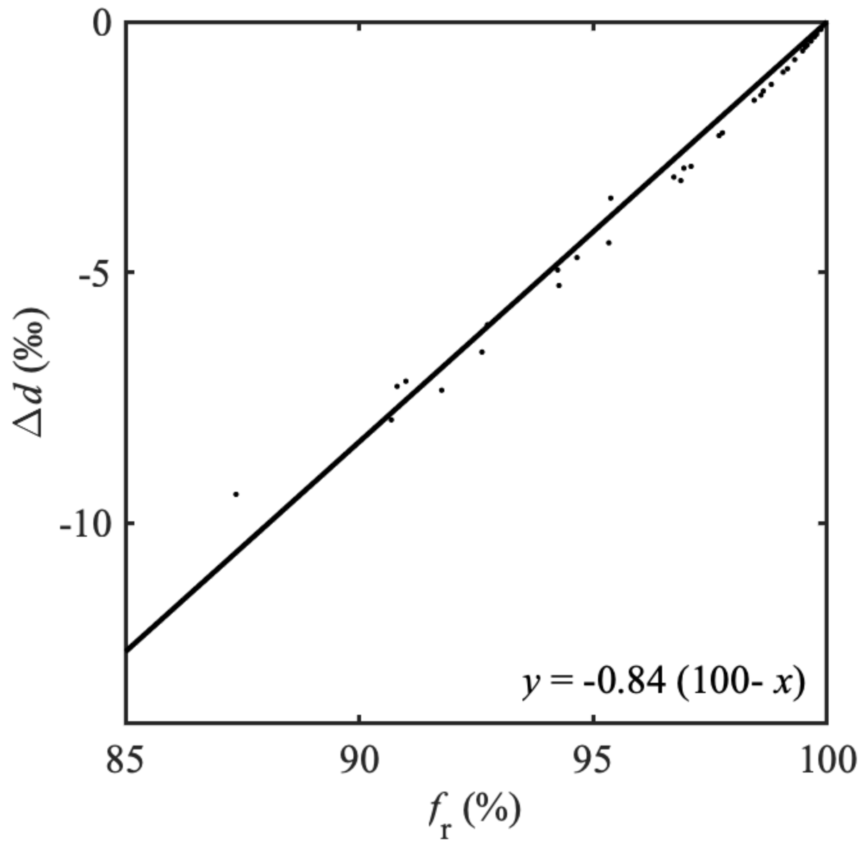
$a_T > a_{LWML}$, which means that the precipitation will be influenced by the sub-cloud evaporation effect.

Comparison between amount-weighted *d*-excess before and after sub-cloud evaporation calibration

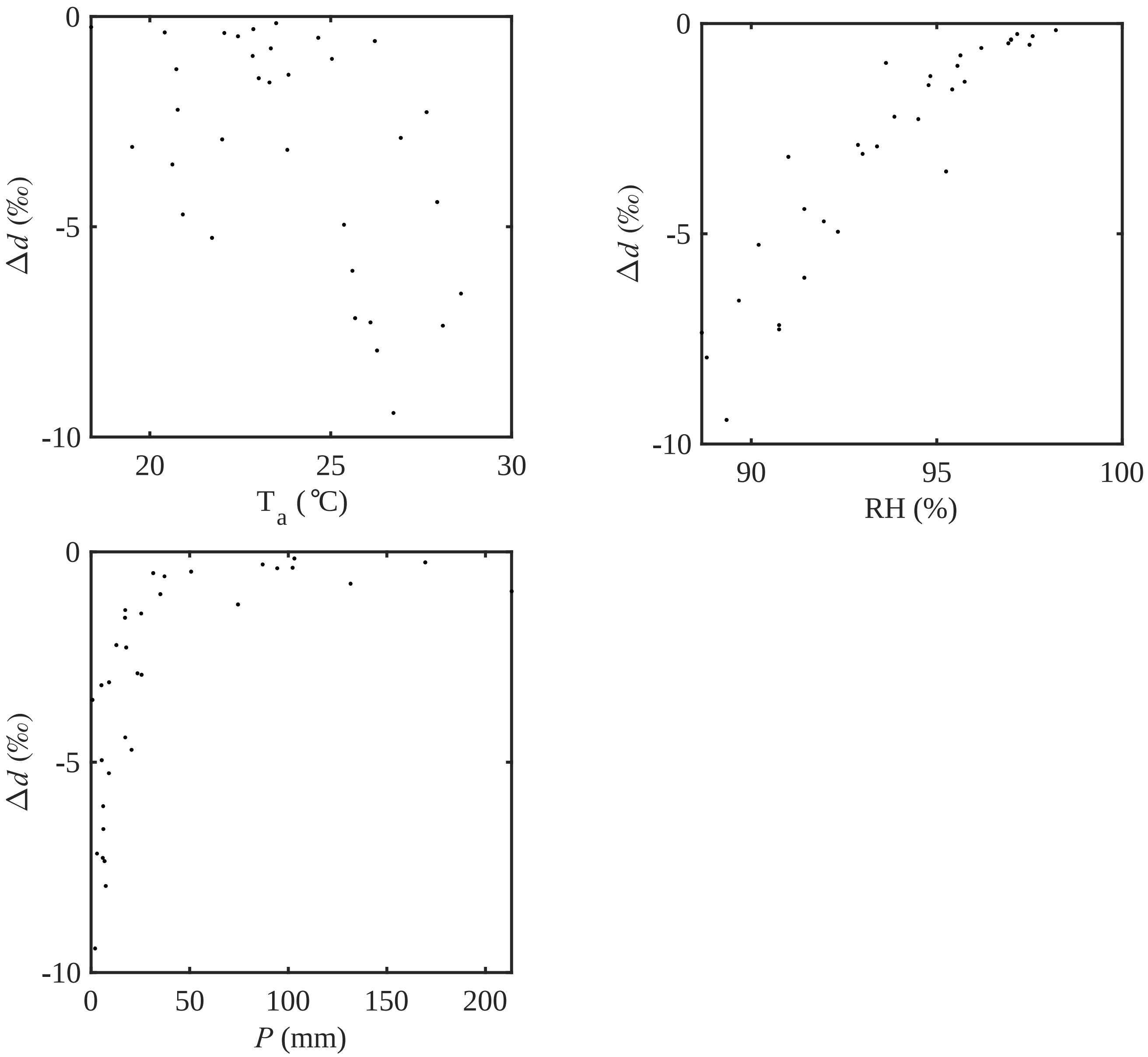
		Dongshan	Liyang
Jun. – Sep.	Before SCEC	11.06	12.61
	After SCEC	12.62	13.73
Jun. – Jul.	Before SCEC	12.08	13.93
	After SCEC	13.47	15.13
Aug. – Sep.	Before SCEC	9.02	10.24
	After SCEC	11.06	11.21

SCEC: sub-cloud evaporation calibration

Relationship between remaining fraction of raindrops (f_r) and Δd ($d_{cb} - d_{gr}$)



Relationship between Δd ($d_{cb} - d_{gr}$) and T_a, RH, and P



3.3 Moisture recycling

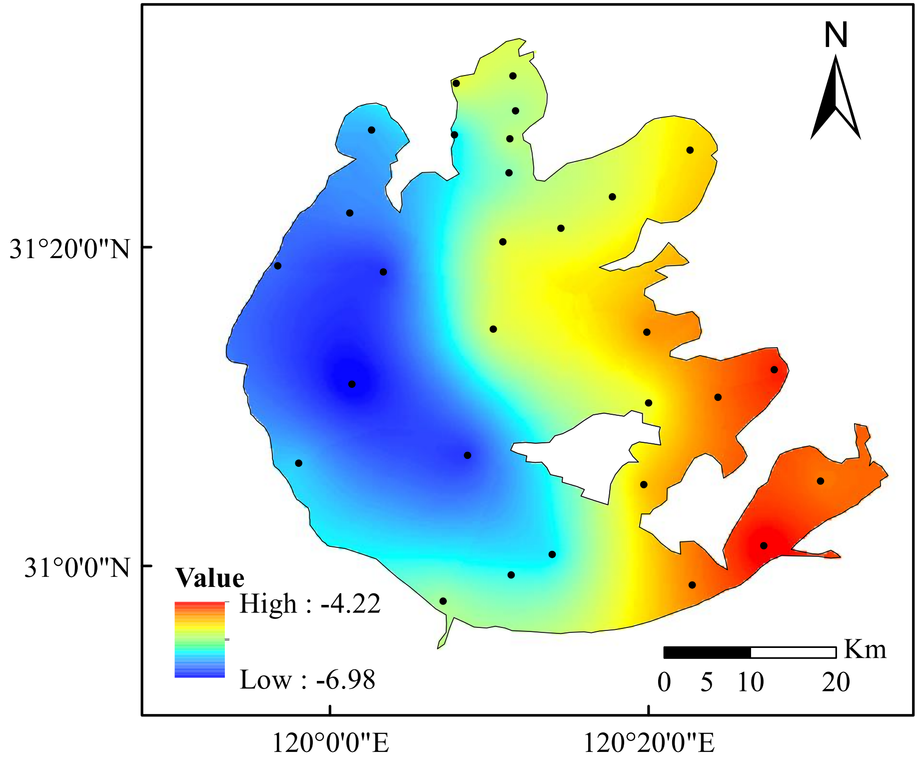
$$f = \frac{d_{\mathrm{down}} - d_{\mathrm{up}}}{d_{\mathrm{E}} - d_{\mathrm{up}}}$$

	Dongshan	Liyang	d_E	f
Jun. – Sep.	12.62	13.73	33.17	5.40
Jun. – Jul.	13.47	15.13	25.56	13.73
Aug. – Sep.	11.06	11.21	40.61	0.51

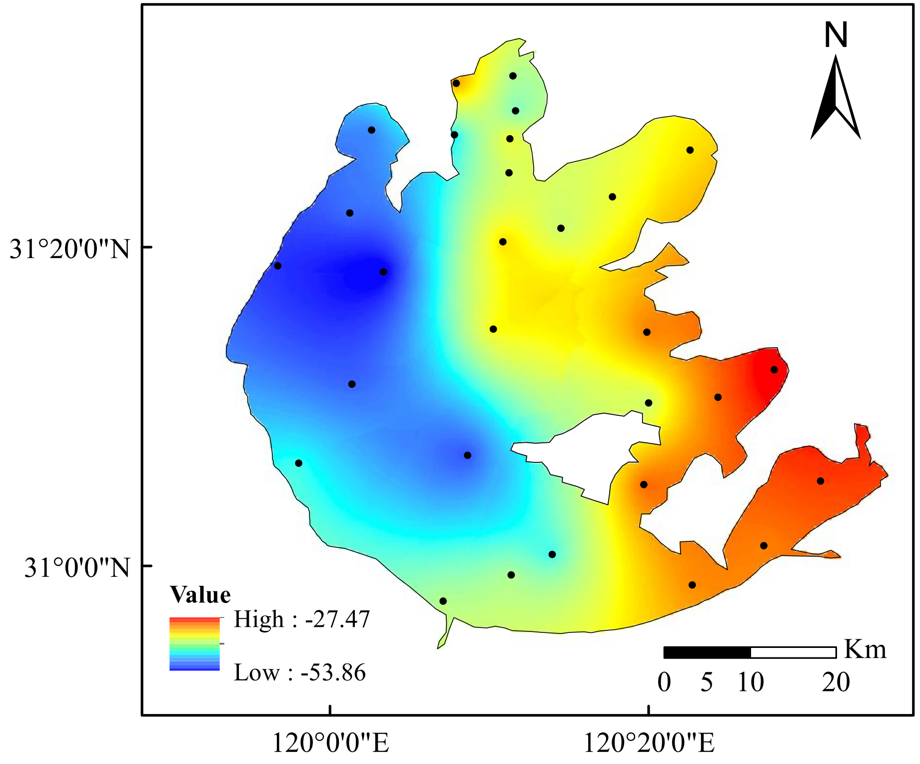
Difference in the fraction of moisture recycling between the Jun. – Jul. and Aug.– Sep. maybe because the rain type (convective/stratiform)

Uncertainty in the moisture recycling

Spatial variation of the $\delta^{18}O$ and δD at Lake Taihu



$\delta^{18}O$



δD

Compare with the Great Lakes

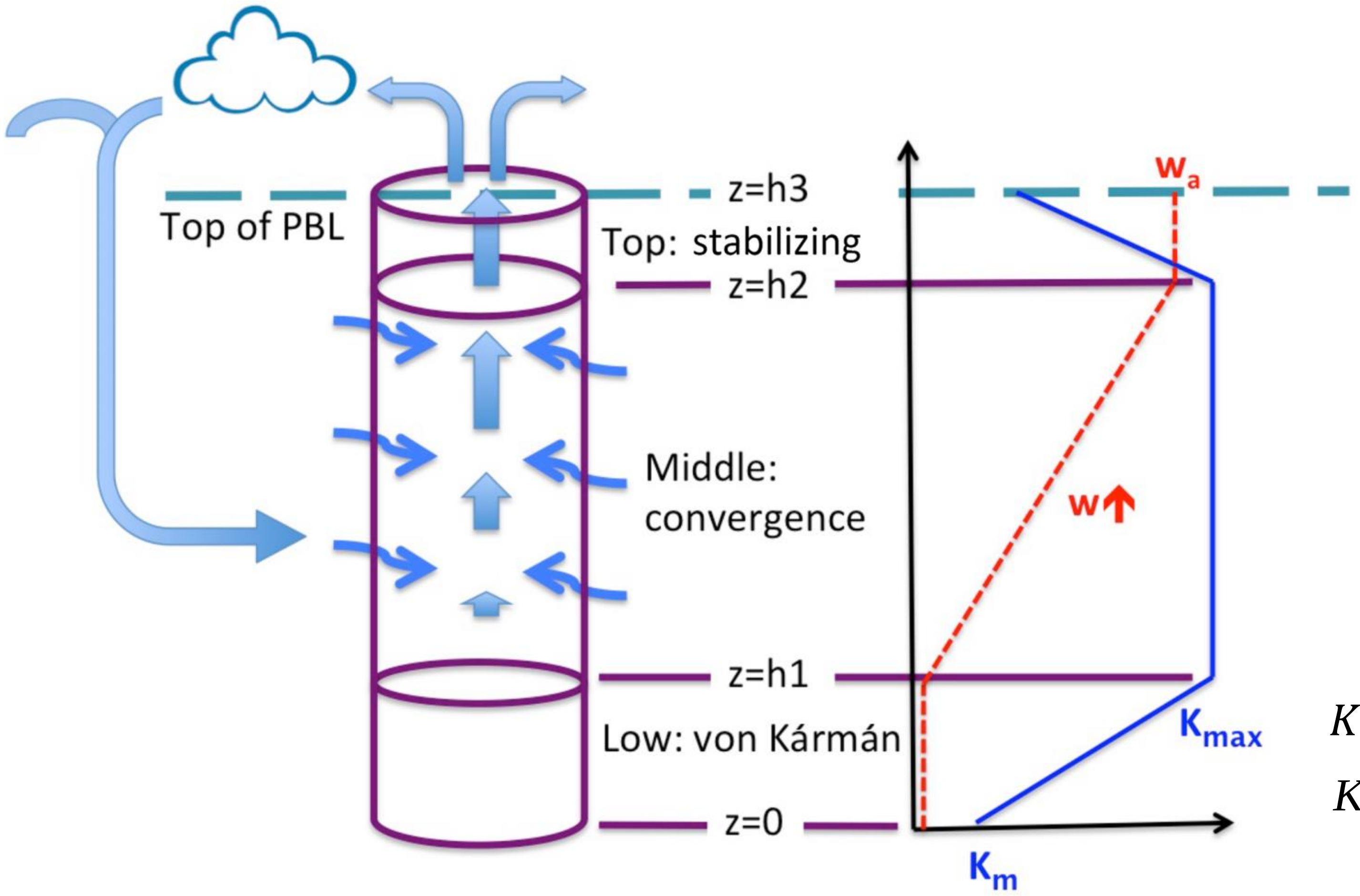
Region	Season	fraction	Literature
The Great Lakes	Summer	5.7% -9.5%	Gat et al., 1994
Lake Michigan	Summer	9%-16%	Machavaram and Krishmurthy, 1995
The Great Lakes	Whole year	4%-17%	Corcoran et al., 2018
The Great Lakes	Summer	16.8±2.8%	Xiao et al., 2018
Lake Taihu	Summer monsoon	5.4%	This study

4 Conclusion

- Precipitation isotope was depleted in summer monsoon period and enriched in winter monsoon period.
- 1% increasing of evaporated raindrop caused 0.84 decreasing of the *d*-excess in precipitation. During the summer monsoon period, relative humidity and precipitation amount were the main factors that controlling the sub-cloud evaporation effect.
- The fraction of the moisture recycling was 5.40%, 13.73% and 0.51% during the summer monsoon period, Jun. – Jul., and Aug. – Sep., respectively. Difference in the fraction of moisture recycling between the Jun. – Jul. and Aug.– Sep. may be because of the rain types. Spatial variation of the $\delta^{18}O$ and δD may cause some uncertainty in the fraction of moisture recycling.

PART 2

Low Layer model



$K(z) = K_m + bh$

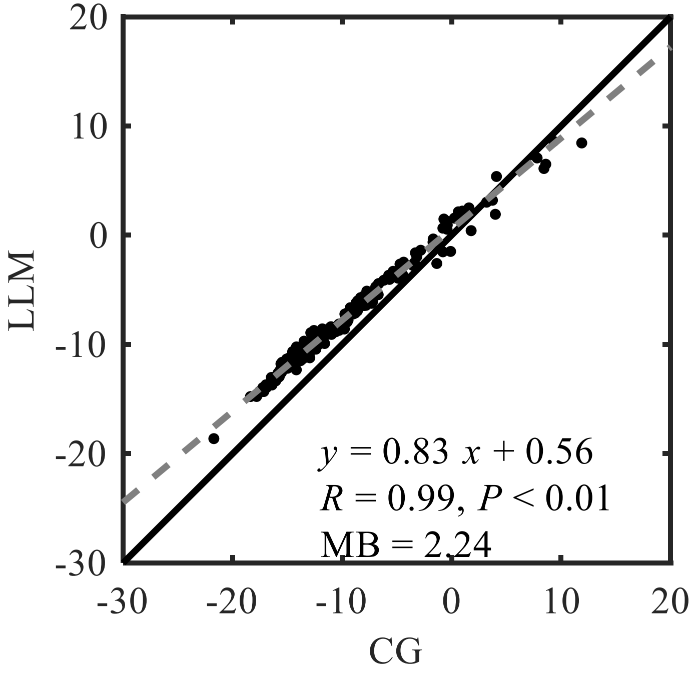
K: diffusion coefficient

$$F = -\rho * \frac{(K_{max}-K_m)(C_1-C_0)}{h_1 \ln[K_{max}/K_m]} = -\rho * b * \frac{(C_1-C_0)}{\ln[K_{max}/K_m]} = -\rho * b * \frac{(C(3.5)-C_0)}{\ln[(K_m+3.5b)/K_m]}$$

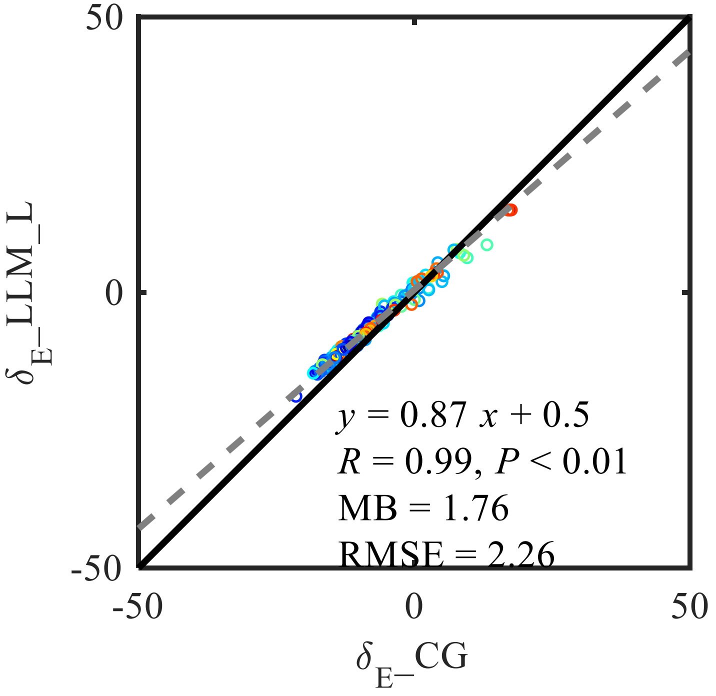
$$R_E = \frac{F_{18}}{F_{16}}$$

(Feng et al., 2019)

Comparison between $\boldsymbol{\delta}_E{}^{18}O$ from CG and LLM



b from the solver

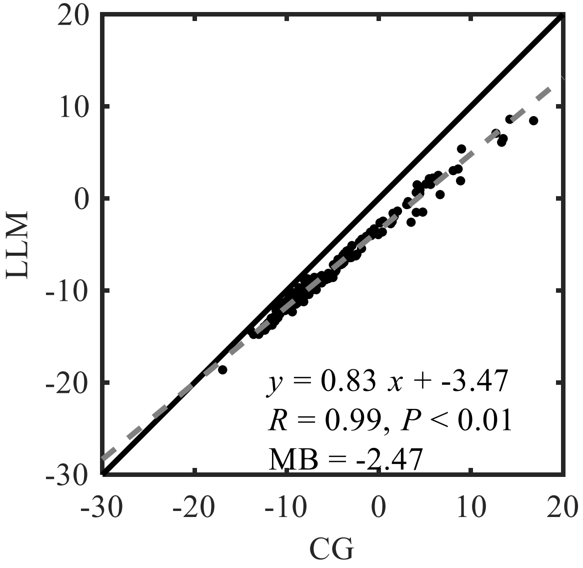


Let E_LLM=E_EC to get b

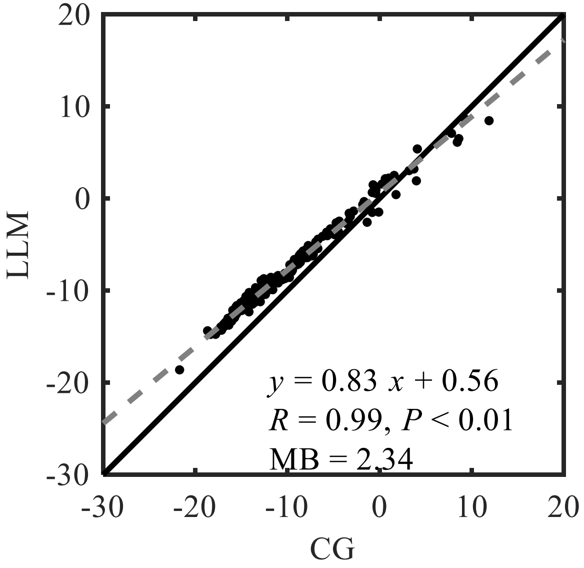
$$\varepsilon_k = \theta(1-h) \times n(1-\frac{D_i}{D}) \times 1000$$

Testing the n value

n=0.1

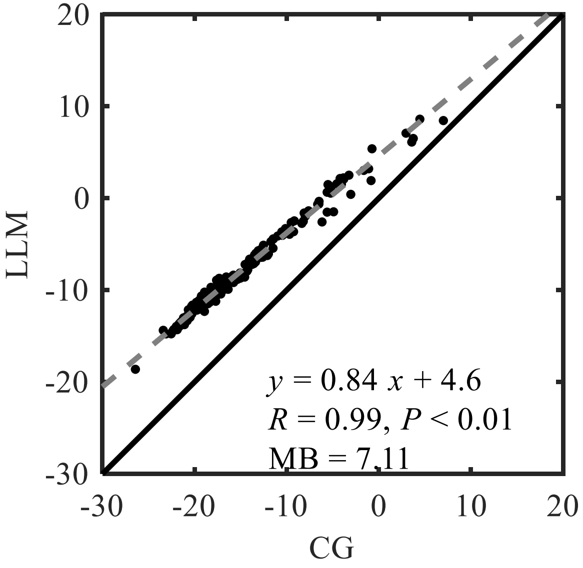


n=0.3

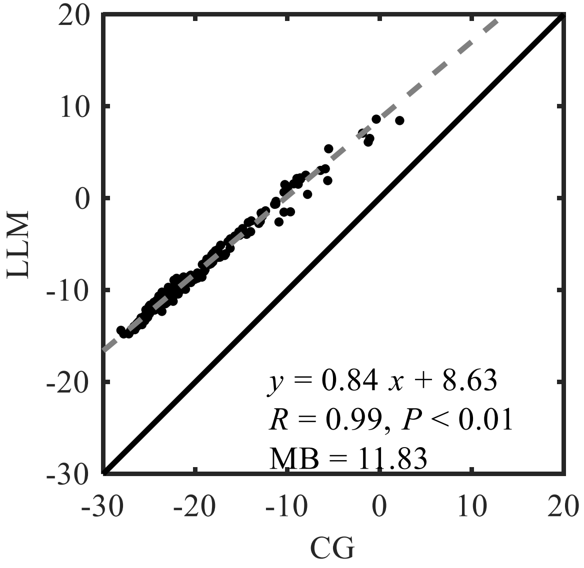


The mean offset is sensitive to n, but not the slope.

n=0.5



n=0.7



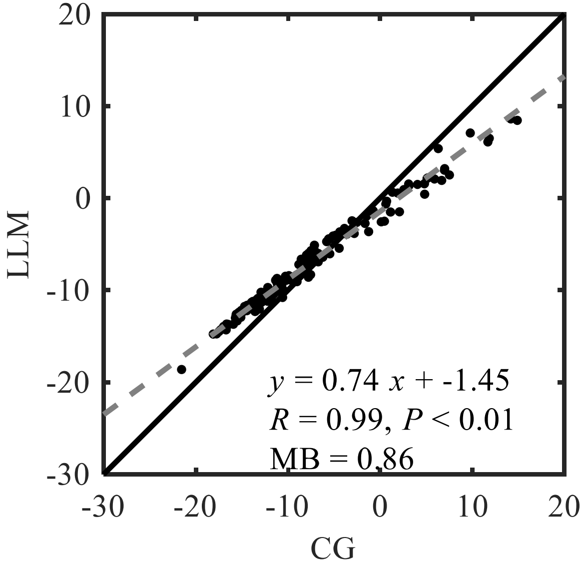
$$e_k = q(1-h)\times n(1-\frac{D_i}{D})\times 1000$$

n=0.3
Di/D=0.9723
θ=0.88

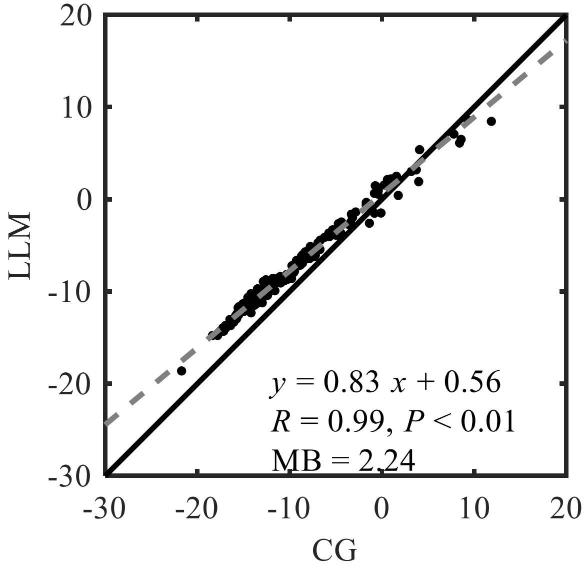
Testing relative humidity calculation.

Different c values are used to calculate relative humidity

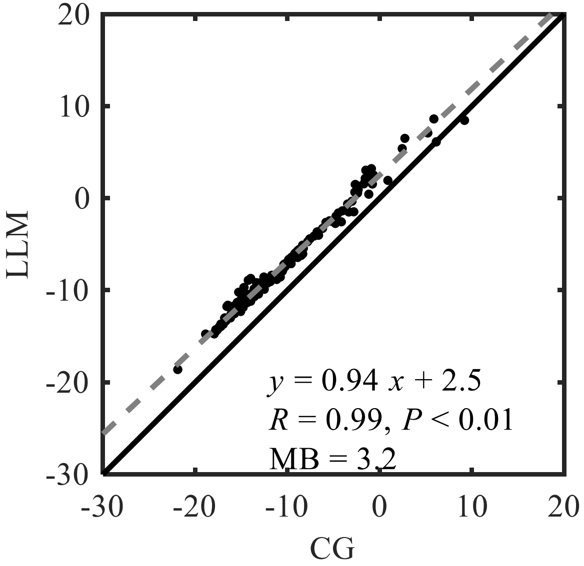
c is the lower height observation



c is the higher height observation



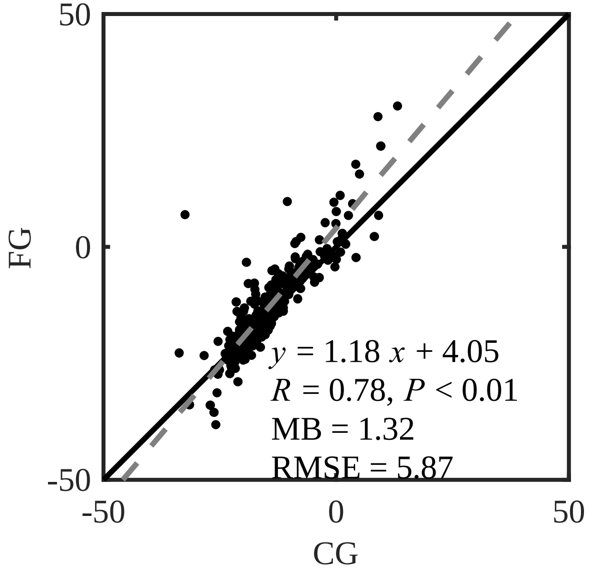
c at the higher height minus 500 ppm



$$h = \frac{c \times p * 10^{-6}}{e(T_s)}$$

c: water vapor concentration (ppm)
p: air pressure (hPa)
e: saturated water pressure (hPa)
Ts: lake surface temperature

Does it mean that the observation height should be higher?



Hypothesis

Measurement error and propagation is large for the FG calculation because the gradients are smaller in the atmosphere (1.1 m and 3.5 m) than between the atmosphere and near surface (like the LLM calculation and CG model). In other words, the signal to noise ratio is low.

Thanks for your attention!