Supplementary materials

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Supplementary Methods S1: Description of the isotope circulation model (ICM)

Supplementary Methods S1 provides the full description of the isotope circulation model (ICM), which largely follows Adachi and Yamanaka (2025).

S1.1 Model overview

The offline ICM used in this study simplifies the atmosphere into a single vertical layer and calculates variations in precipitation isotope ratios on the basis of the Rayleigh equation, using reanalysis data as meteorological forcing (Yoshimura et al., 2003). In this study, we employed a modified version of the model that incorporates isotopic variability of evaporating water vapor from the surface of the Earth and subcloud raindrop evaporation (Adachi and Yamanaka, 2025a; Fig. 1). The simulations were conducted continuously over 76 years, from September 1947 to March 2024. The time step (Δt) was set to 1 hour.

S1.2 Meteorological forcing

We used JRA-3Q (Kosaka et al., 2024) as the meteorological forcing because Adachi and Yamanaka (2025a) reported that it reproduced the day-to-day variability in the isotopic composition of precipitation more accurately in East Asia than other reanalysis datasets (JRA-55, ERA5, and MERRA-2). The spatial resolution is 1.25°×1.25°. The following parameters from the dataset were used as parameters for isotope balance: vertically integrated water vapor flux (Q); precipitable water (W); precipitation (P), which is divided into large-scale precipitation (Lsp) and convective precipitation (Cp); and evaporation (E), namely, evapotranspiration Additionally, temperature (T_a), surface temperature (T_s), and dew point temperature (T_d) were used as parameters for isotopic fractionation.

S1.3 Isotope mass balance and Rayleigh fractionation

The isotope mass balance at each grid cell (i.e., atmospheric column) before precipitation occurrence is calculated as follows:

$$\delta_{w(t+\Delta t)} W_{(t+\Delta t)} = \delta_{w(t)} W_{(t)} - \nabla \cdot \delta_w \vec{Q} \Delta t + \delta_{ET} E \Delta t \qquad \text{(S1)}$$

where δ_w and δ_{ET} are the isotope values of atmospheric water vapor and evapotranspiration flux, respectively. The initial δ_w is derived under the assumption that water vapor isotopes are in equilibrium with precipitation isotopes, as described by the latitude effect equation (Bowen and Wilkinson, 2002). The divergence of the water vapor flux (i.e., $\nabla \cdot \delta w \vec{Q}$) can be calculated as follows:

$$\nabla \cdot \delta_w \vec{Q} \Delta t = \frac{1}{R_c cos\varphi} \left(\frac{\partial \delta_w Q_\lambda}{\partial \lambda} + \frac{\partial \delta_w Q_\varphi cos\varphi}{\partial \varphi} \right) \qquad \text{(S2)}$$

where R_c is the radius of the Earth and where λ and φ are the longitude and latitude, respectively. The change in δ_w during the reduction in W due to precipitation is given by the following Rayleigh-type equation:

$$\delta_w = \left\{ \left(\frac{W}{W^*} \right)^{\alpha - 1} (1 + 10^{-3} \delta_w{}^*) - 1 \right\} \times 10^3 \qquad \text{(S3)}$$

where α represents the temperature-dependent equilibrium isotope fractionation factor (Majoube, 1971), and the asterisk indicates the values before precipitation occurs; that is, $W = W^* - P\Delta t$ and $\delta_w W = \delta_w{}^* W^* - \delta_{P0} P \Delta t$. Thus, the isotope ratio of the precipitation isotope (δ_{P0}) is given as follows:

$$\delta_{P0} = \frac{\delta_w{}^* - \delta_w f}{1 - f} \qquad \text{(S4)}$$

where f (=W/W^*) is the water vapor retention rate at each time step. Daily values of δ_P are obtained as flux-weighted averages.

S1.4 Surface evaporation processes

The isotope values of evaporation flux (i.e., evaporating water vapor; δ_E) from the surface of the Earth have been conventionally calculated using the Craig–Gordon model (Craig and Gordon, 1965), which is expressed as follows:

$$\delta_E = \frac{(1/\alpha)\delta_s - h\delta_w - \varepsilon}{1 - h + 10^{-3}\Delta\varepsilon} \tag{S5}$$

where δ_s is the isotope value of sea water (δ_{sea}) or shallow soil water (δ_{ssw}) and h is the relative humidity normalized by T_s. ε and $\Delta\varepsilon$ represent equilibrium and kinetic enrichment, respectively, which are expressed as follows:

$$\varepsilon = \left(1 - \frac{1}{\alpha}\right) \cdot 10^3 + \Delta\varepsilon, \qquad \Delta\varepsilon = (1 - h)\frac{\rho_M}{\rho} n[(D_v/D_{vi}) - 1] \cdot 10^3 \tag{S6}$$

where ρ is the total resistance to water vapor transport, ρ_M is the transport resistance within the diffusion layer, n is a coefficient, and (D_V/D_{vi}) is the ratio of the molecular diffusivities in air (Merlivat, 1978). Under the closure assumption (i.e., $\delta_w = \delta_E$), δ_E can also be given as follows (Merlivat and Jouzel, 1979):

$$\delta_E = \frac{(1/\alpha)\delta_s - \varepsilon}{1 + 10^{-3}\Delta\varepsilon} \tag{S7}$$

A semiclosure assumption is applied, where atmospheric water vapor is a mixture of surface-evaporated vapor and advected vapor, with the closure ratio (clr) varying according to relative humidity, expressed as the following equation:

$$\delta_E = clr \cdot {}^{cl}\delta_E + (1 - clr) \cdot {}^{cg}\delta_E \tag{S8}$$

$$clr = \frac{1}{1 + \{7 \cdot (1 - h)\}^{14}} \tag{S9}$$

where ${}^{cl}\delta_E$ and ${}^{cg}\delta_E$ are the δ_E values calculated via Equations (S7) and (S5), respectively.

We assume that $\delta_s = \delta_{sea} = 0$ for sea grid cells and that $\delta_s = \delta_{ssw}$ for land grid cells. In our ICM, the land is vertically divided into three layers, each of which contains shallow soil water, deep soil water, and ground water. It is assumed that soil surface evaporation (E_s) occurs from shallow soil water, whereas transpiration (Tr) occurs from deep soil water without isotopic fractionation (White, 1989); that is, the isotope ratio of transpiration flux (δ_{Tr}) equals that of deep soil water (δ_{dsw}). Under these assumptions, the isotope mass balance for shallow soil water and deep soil water is expressed as follows:

$$\delta_{ssw_{(t+\Delta t)}} c_{ssw} = \delta_{ssw_{(t)}} c_{ssw} + \delta_P P\Delta t - \delta_E E_s \Delta t - \delta_A (P - E_s)\Delta t \tag{S10}$$

$$\delta_{dsw_{(t+\Delta t)}} c_{dsw} = \delta_{dsw_{(t)}} c_{dsw} + \delta_A (P - E_s)\Delta t - \delta_{Tr} Tr \Delta t - \delta_B (P - E_s - Tr)\Delta t \tag{S11}$$

where δ_A is determined by the sign of $(P - E_s)$. If $(P - E_s)$ is positive, δ_A is equal to $\delta_{ssw_{(t)}}$; whereas if $(P - E_s)$ is negative, δ_A corresponds to $\delta_{dsw_{(t)}}$. Similarly, δ_B is determined by the sign of $(P - E_s - Tr)$. If $(P - E_s - Tr)$ is positive, δ_B is represented by $\delta_{dsw_{(t)}}$; if it is negative, it is represented by δ_{gw}. For simplicity, the capacities of shallow soil water (c_{ssw}) and deep soil water (c_{dsw}) are assumed to be constant (50 mm). Adachi and Yamanaka (2025a) identified this setting as optimal on the basis of East Asian tests, and we adopt it here for consistency. We employed a global precipitation isotope map (Bowen et al., 2005) with initial values of δ_{ssw}. Similarly, at the beginning of each month, δ_{dsw} was reset to the long-term monthly average from the map to avoid isotopic drift, and δ_{gw} was set to the long-term annual average. The fraction of transpiration to evaporation (F_T) is given by the following relationship for mixed forests (Wei et al., 2017):

$$F_T = 0.52 LAI^{0.26} \tag{S12}$$

where *LAI* is the leaf area index. For this estimation, a global monthly LAI dataset (Mao and Yan, 2019) was employed. The isotopic composition of bulk evapotranspiration (δ_{ET}) was then obtained via Eq. S13:

$$\delta_{ET} = \delta_E (1 - F_T) + \delta_{Tr} F_T \tag{S13}$$

S1.5 Post condensation processes

The isotope ratio of precipitation is altered by postcondensation processes. To address this, the altered isotope values of precipitation were calculated using the following nonequilibrium Rayleigh equation (Gonfiantini, 1986):

$$\delta_P = \left\{ \left(10^{-3}\delta_{P0} - \frac{A}{B} \right) f_r^{\,B} + \frac{A}{B} \right\} \times 10^3 \tag{S14}$$

$$A = \frac{h \cdot 10^{-3}\delta_a + 10^{-3}\Delta\varepsilon + 1 - (1/\alpha)}{1 - h + 10^{-3}\Delta\varepsilon}, B = \frac{h \cdot 10^{-3}\delta_a - \left(1 - (1/\alpha)\right)}{1 - h + 10^{-3}\Delta\varepsilon} \tag{S15}$$

where δ_P represents the isotope values of precipitation after the postcondensation process, δ_{P0} represents the initial isotope values of falling raindrops and f_r represents the retention rate of the raindrops. These equations consider raindrop evaporation as well as isotopic exchange between raindrops and ambient water vapor under the cloud base. It was assumed that f_r differs depending on the type of precipitation. The precipitation types were classified into large-scale condensation precipitation and convective precipitation, with retention rates of 0.950 for large-scale condensation (f_{r_Lsp}) and 0.667 for convective precipitation (f_{r_Cp}). These values were set with reference to satellite-based estimates by Worden et al. (2007), who reported average subcloud evaporation fractions of approximately 20% (and less than 50%) in tropical convective systems and negligible evaporation in high-latitude stratiform precipitation. Water vapor is lost from raindrops because evaporation returns to the atmosphere. While some GCMs use the time-dependent formulation of Stewart (1975), the ICM simplifies the atmospheric column into a single vertical layer. To represent raindrop evaporation under such vertically average conditions, we employed the simpler nonequilibrium Rayleigh equation.

S1.6 Temporal aggregation of model outputs

We converted the daily model outputs to monthly values ($\delta_{m,y}$) using precipitation-weighted means (Eq. S16). Climatological monthly means (δ_{clim}) were defined as the arithmetic average of the monthly values for each calendar month from January 1948 to December 2023 (N = 76) (Eq. S17). We also computed the precipitation-weighted annual means for each year ($\delta_{ann,y}$) and then averaged them for 1948–2023 to obtain the long-term annual mean ($\overline{\delta_{ann}}$) (Eqs. S18 and S19).

$$\delta_{m,y} = \frac{\sum_{d=1}^{D_{m,y}} P_{m,y,d} \cdot \delta_{m,y,d}}{\sum_{d=1}^{D_{m,y}} P_{m,y,d}} \quad \text{(S16)}$$

$$\delta_{clim} = \frac{1}{N}\sum_{y=1}^{N} \delta_{m,y} \quad \text{(S17)}$$

$$\delta_{ann,y} = \frac{\sum_{d=1}^{D_{y}} P_{y,d} \cdot \delta_{y,d}}{\sum_{d=1}^{D_{y}} P_{y,d}} \quad \text{(S18)}$$

$$\overline{\delta_{ann}} = \frac{1}{N}\sum_{y=1}^{N} \delta_{ann,y} \quad \text{(S19)}$$

where m is the calendar month, y is the year, d is the number of days within (m, y), $D_{m,y}$ is the number of days in month m of year y, $P_{m,y,d}$ is the daily precipitation, $\delta_{m,y,d}$ is the daily isotope value, D_y is the number of days in year y, and N = 76 (January 1948–December 2023).

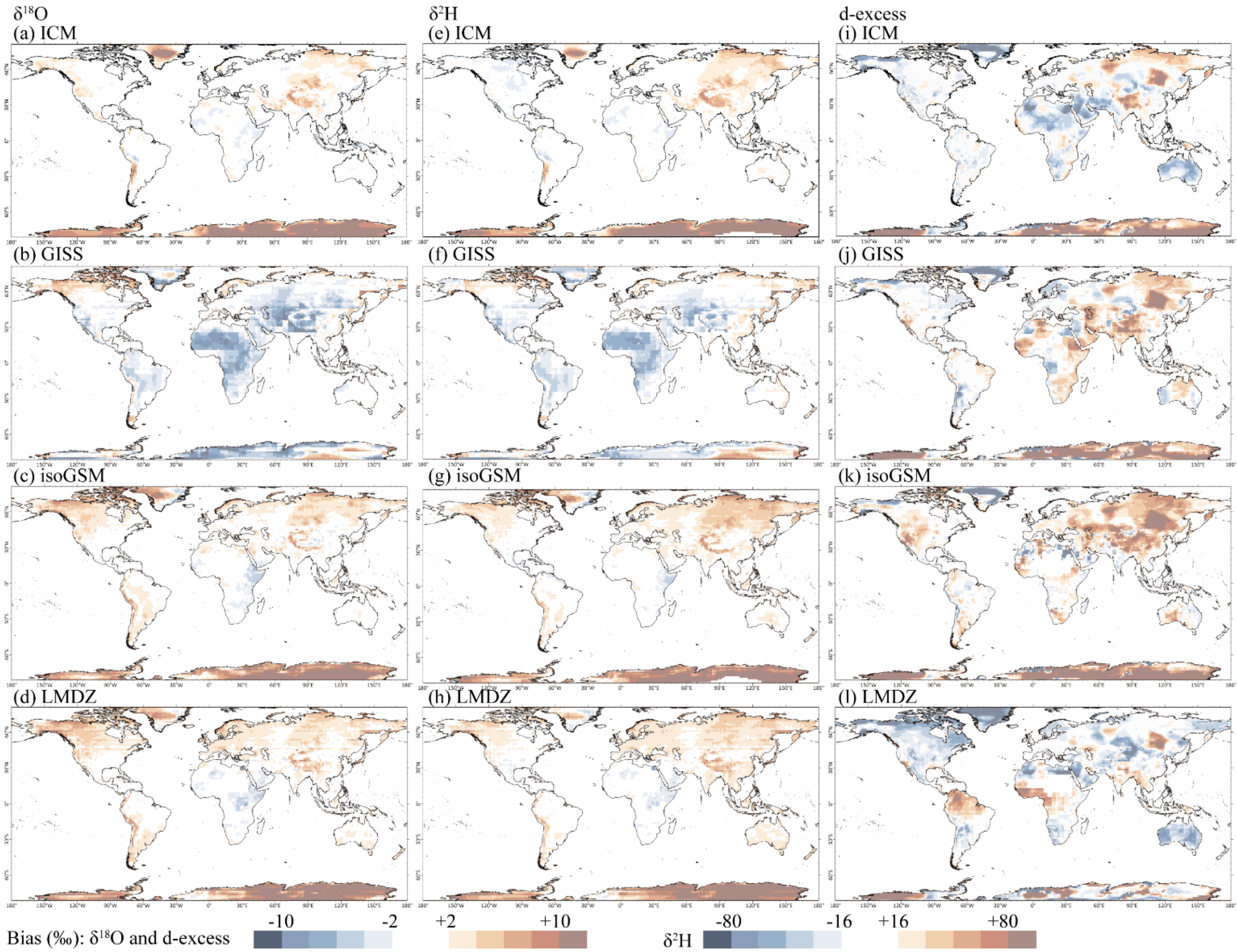

Figure S1: Annual-mean bias maps relative to the observation-based global isoscape of Bowen et al. (2005). Columns show $\delta^{18}O$ (a–d), $\delta^{2}H$ (e–h), and d-excess (i–l); rows show the ICM (a, e, and i) and nudged isotope GCMs (Risi et al., 2012): GISS (b, f, and j), isoGSM (c, g, and k), and LMDZ (d, h, and l). Bias is defined as the model annual mean − observation-based annual mean (Bowen et al., 2005) at each grid cell.

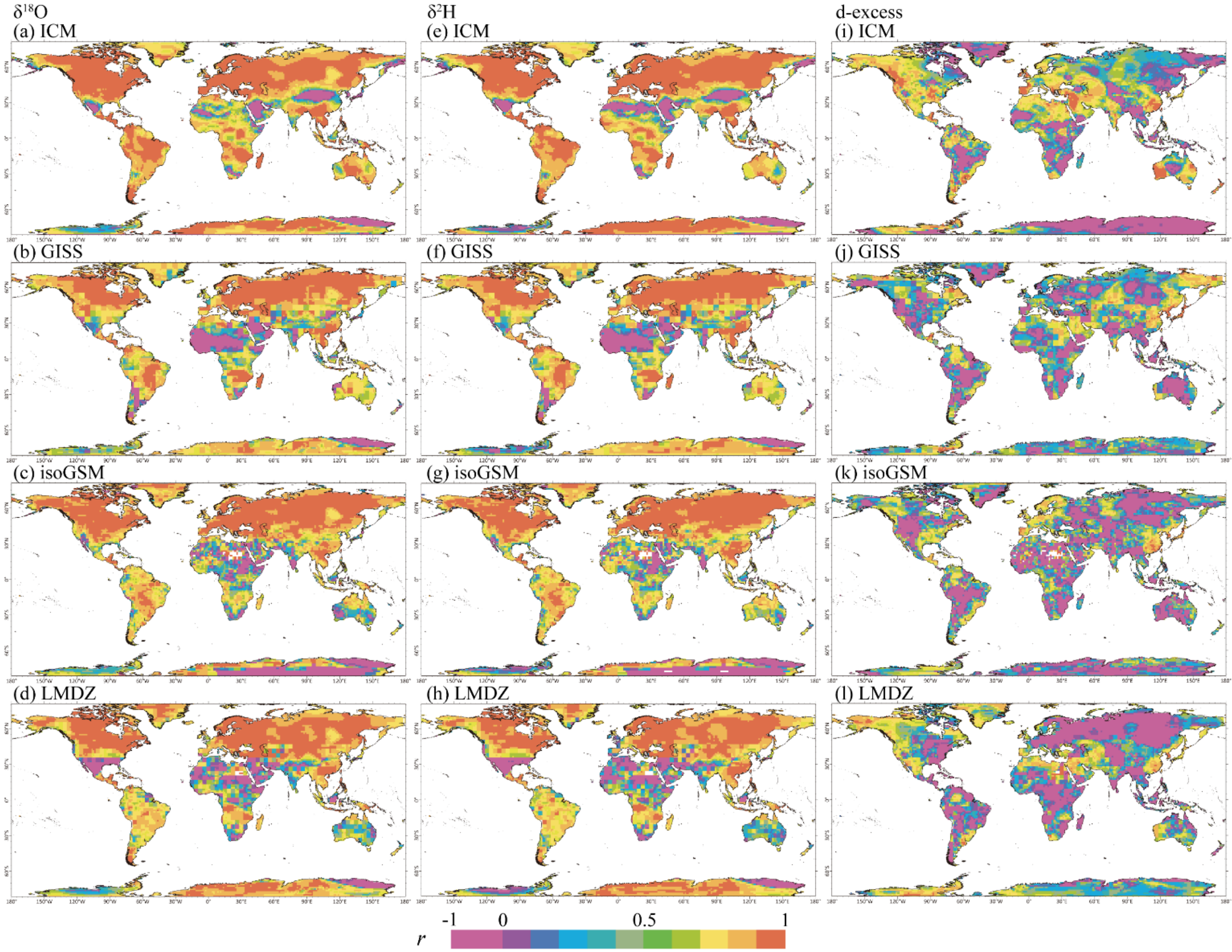

Figure S2: Pointwise correlation (per grid cell) with the observation-based global isoscape (Bowen et al., 2005). Columns show $\delta^{18}O$, $\delta^{2}H$, and d-excess; rows show ICM and nudged isotope GCMs (GISS, isoGSM, LMDZ; Risi et al., 2012). At each grid cell, the Pearson *r* is computed between the modeled climatological monthly cycle and the corresponding Bowen monthly cycle.

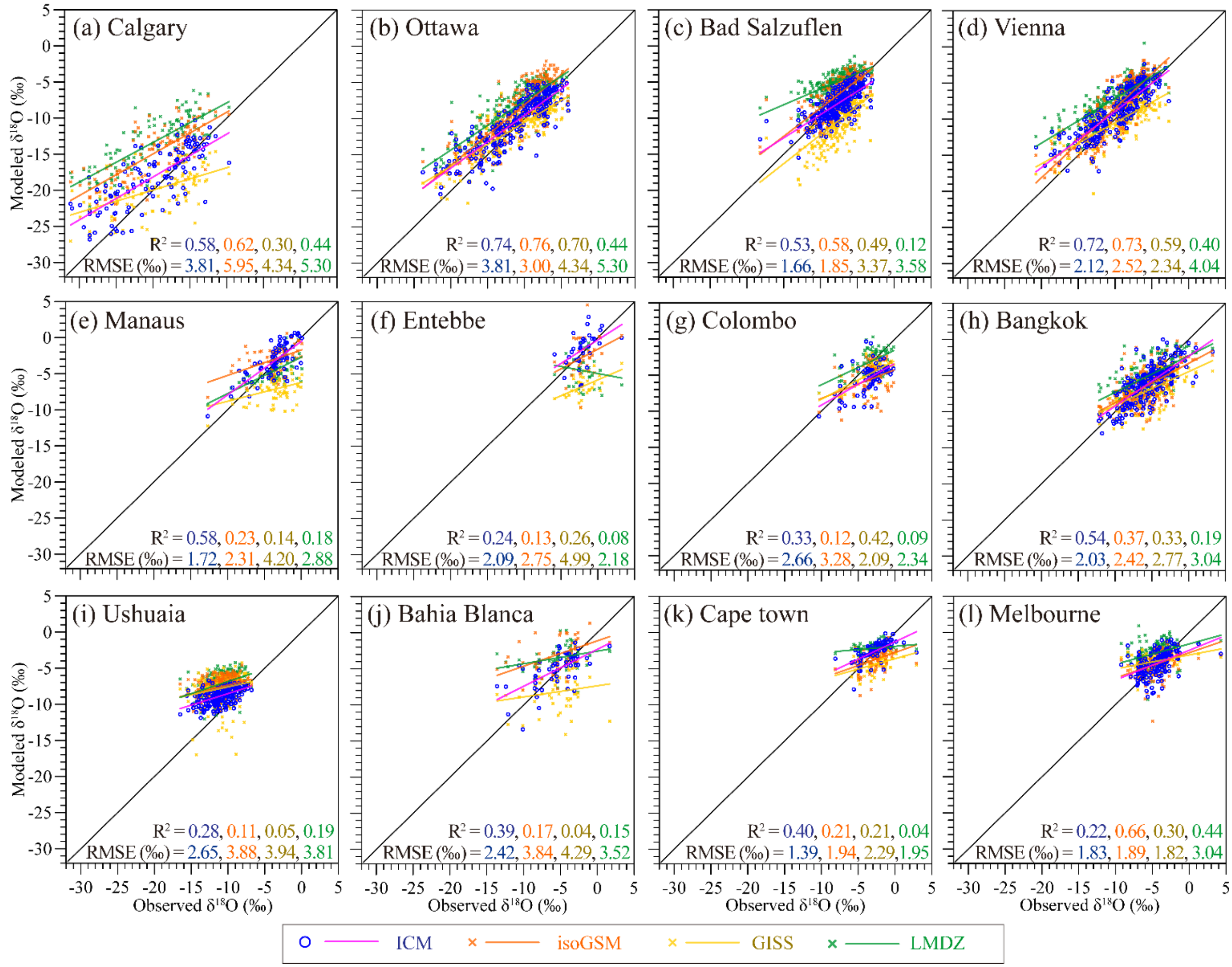

Figure S3: Monthly model–observation comparisons of δ^{18}O at representative GNIP stations. Panels (a–l) correspond to the stations labeled. Symbols/colors denote models: ICM (blue circles; annually bias-corrected model values), isoGSM (orange crosses), GISS (yellow crosses), and LMDZ (green crosses). The black line is the 1:1 line; the colored lines are the least-squares fits for each model. The numbers report R^2 and RMSE (‰) in the corresponding colors.