

1 Arctic Sea Ice Loss Amplifies Local Evaporation Influence on Water Vapor Isotopes: 2 Insights from Cruise Observations

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7 **Abstract.** Rapid Arctic warming and sea ice retreat have increased atmospheric humidity, yet the relative contributions of
8 local evaporation and advected lower-latitude moisture remain poorly quantified. Here, we present high-resolution, ship-based
9 in situ measurements of near-surface water vapor isotopes across diverse Arctic sea ice regimes between 17 July and 23
10 September 2024. By integrating isotope fractionation models with multi-source meteorological data, we show that sea ice
11 changes act as a key modulator of Arctic water vapor isotopic variations. Under ice-covered conditions, water vapor isotopes
12 are strongly temperature-dependent, exhibiting depleted $\delta^{18}\text{O}$ and elevated d-excess due to **long-range transport with cloud**
13 **formation underway**. As sea ice retreats, kinetic fractionation from local evaporation becomes increasingly important,
14 particularly at temperatures above $\sim 5^\circ\text{C}$, generating enriched $\delta^{18}\text{O}$, elevated d-excess, and **a negative $\delta^{18}\text{O}$ –temperature**
15 **relationship**. A Bayesian isotope mixing model quantifies the resulting moisture source shift, showing local evaporation
16 contributions rise from 18.6 % (95 % CI: 11.1–40.9 %) in ice-covered regions to 43.8 % (95 % CI: 14.7–65.7 %) in ice-free
17 regions, despite advected moisture remaining predominant. These findings provide a process-based isotope framework for the
18 Arctic hydrological cycle, complementing conventional meteorological diagnostics and offering valuable observational
19 constraints for interpreting paleo-isotope archives.

20 1 Introduction

21 Stable water vapor isotopes ($\delta^{18}\text{O}$ and δD) and their derived deuterium excess (d-excess, defined as $\delta\text{D} - 8 \times \delta^{18}\text{O}$) are widely
22 used as tracers for investigating hydrological processes and identifying moisture sources (Gat, 1996). These isotopes
23 fractionate during phase transitions such as evaporation, condensation, and sublimation, encoding information about
24 environmental conditions both at the moisture source and along the atmospheric transport pathways (Dansgaard, 1964;
25 Galewsky et al., 2016; Bowen et al., 2019). As such, analysis of these isotopic parameters in modern water vapor and
26 paleoclimate archives provides key insights into moisture sources and climatic dynamics across both modern (Kurita, 2011;
27 Kopeck et al., 2016; Wang et al., 2023) and paleoclimate contexts (Klein and Welker, 2016; Opel et al., 2013; Porter et al.,
28 2019). This dual perspective is particularly critical in the Arctic, which serves not only as a modern laboratory of rapid
29 hydrological change but also as a primary archive of past climate preserved in its ice sheets.
30 Recent Arctic amplification and sea ice loss have moistened the Arctic atmosphere (Min et al., 2008; Bengtsson et al., 2011;
31 Bintanja and Selten, 2014), yet the relative roles of local evaporation and poleward moisture transport remain contested. Some

32 studies attribute these changes to enhanced local evaporation over newly melted ocean (Bintanja and Selten, 2014; Boisvert
33 and Stroeve, 2015; Kopec et al., 2016), while others emphasize strengthened poleward transport of lower-latitude moisture
34 (Bengtsson et al., 2011; Zhang et al., 2013; Zhong et al., 2018). Resolving this debate is crucial because water vapor plays a
35 central role in Arctic radiative and hydrological feedbacks, influencing cloud cover, precipitation, and surface warming. Stable
36 water vapor isotopes provide a process-based diagnostic for disentangling and quantifying the relative contributions of these
37 two sources. Such insights are essential not only for improving the representation of hydrological processes in climate models,
38 thereby enabling more reliable projections (Barras and Simmonds, 2009; Gao et al., 2011; Sturm et al., 2010; Xi, 2014), but
39 also for interpreting isotopic signals preserved in paleoclimate archives such as ice cores (Klein et al., 2016; Opel et al., 2013;
40 Porter et al., 2019).

41 However, the application of stable isotopes in the Arctic remains challenging due to conflicting interpretations of isotopic
42 signals, particularly regarding the role of local evaporation. One line of evidence associates high d-excess and low $\delta^{18}\text{O}$ over
43 open water with strong evaporation under cold, dry air masses, as observed in both water vapor (Kurita, 2011; Steen-Larsen
44 et al., 2013; Bailey et al., 2021) and precipitation (Mellat et al., 2021). In contrast, Klein et al. (2015) proposed that local
45 evaporation associated with sea ice loss leads to enriched $\delta^{18}\text{O}$ and suppressed d-excess in Arctic water vapor, relative to
46 advected lower-latitude moisture, a pattern that is also reflected in observed Arctic precipitation (Kopec et al., 2016; Song et
47 al., 2023). However, this interpretation is further complicated by recent evidence showing that low-latitude moisture can also
48 produce low vapor d-excess when influenced by synoptic-scale extratropical cyclones (Thurnherr and Aemisegger, 2022).
49 These divergent interpretations largely reflect the scarcity of spatially extensive, high-resolution vapor isotope observations
50 across the Arctic Ocean. Most existing studies are based on land-based observations, with limited coverage over the Arctic
51 Ocean, hindering efforts to disentangle complex isotope signals (Kurita, 2011; Kopec et al., 2016; Mellat et al., 2021). Klein
52 and Welker (2016) suggested the relative influence of local evaporation on water vapor isotopes may vary with sea ice extent
53 (SIE), proposing an anti-correlation between d-excess and SIE. However, subsequent studies have reported the opposite
54 relationship (Bonne et al., 2019). This persistent inconsistency underscores the need for broader, in-situ water vapor isotope
55 observations across different sea ice regimes.

56 To address this gap, we leverage a new set of high-resolution, in-situ water vapor isotope measurements obtained during a
57 comprehensive summer expedition across the Arctic Ocean aboard the RV *Xuelong 2*. This dataset provides extensive spatial
58 coverage across the Arctic, spanning sea ice regimes from near-continuous cover to open water. By integrating these isotopic
59 observations with meteorological data, and applying isotope-based diagnostics along with Lagrangian trajectory analysis, this
60 study aims to: (1) map the spatial patterns of summer Arctic vapor isotopes in relation to diverse sea ice regimes, (2) identify
61 the primary mechanisms behind isotope variability under distinct sea ice conditions, and (3) quantify the partitioning between
62 locally evaporated and advected moisture sources across different Arctic regions.

63 2 Data and Methods

64 2.1 Water Vapor Isotope Measurements and Calibration

65 Continuous, in situ measurements of near-surface water vapor isotopes and humidity were conducted using a Picarro L2130-i
66 cavity ring-down spectroscopy (CRDS) analyzer. The instrument was housed in the meteorological laboratory aboard RV
67 *Xuelong 2* during China's 14th Arctic Scientific Expedition. The data were obtained between 17 July and 23 September 2024,
68 during which RV *Xuelong 2* transited the Arctic Ocean from the Bering Strait through the Chukchi Sea and the central Arctic
69 Ocean to the Barents Sea, before returning to the Bering Strait via the East Siberian Sea (Fig. 1). Ambient air was drawn from
70 an inlet mounted on the balcony of the meteorological room, approximately 10 m above sea level. The inlet was equipped with
71 a hydrophobic polytetrafluoroethylene (PTFE) filter (Nanjing Liebo) to prevent particle contamination and icing during
72 sampling. The sample air was delivered to the analyzer through a 5-meter PTFE inlet tube. Most of the tubing was routed
73 inside the meteorological laboratory, where air conditioning maintained the temperature at ~20°C, preventing condensation
74 without additional heating.

75 All isotope values are reported in standard δ notation relative to the Vienna Standard Mean Ocean Water (VSMOW):

$$76 \quad \delta = \left(\frac{R_{\text{sample}}}{R_{\text{VSMOW}}} - 1 \right) \times 1000 \quad (1)$$

77 where R_{sample} and R_{VSMOW} are the isotope ratios ($^{18}\text{O}/^{16}\text{O}$ or $^2\text{H}/^1\text{H}$) of the sample and VSMOW, respectively. The
78 measurement precision was 0.25‰/0.08‰ for $\delta^{18}\text{O}$ and 1.6‰/0.5‰ for δD at 10 s/100 s averaging intervals, respectively. The
79 long-term analytical reproducibility, estimated from repeated measurements of internal laboratory standards over the study
80 period and derived from normalization residuals, was $\pm 0.06\text{‰}$ for $\delta^{18}\text{O}$ and $\pm 0.3\text{‰}$ for δD (1σ). The data were originally
81 collected at 1-minute intervals and have been resampled to hourly resolution for analysis. D-excess was calculated following
82 Dansgaard (1964) as:

$$83 \quad d\text{-excess} = \delta\text{D} - 8 \times \delta^{18}\text{O} \quad (2)$$

84 Because the isotope measurements from the Picarro analyzer are subject to ambient water vapor concentration, a humidity-
85 dependent correction was applied to ensure consistency across varying atmospheric conditions. Calibration was performed
86 using a Standard Delivery Module (SDM). To define the calibration gradient, we used the analyzer's internal humidity
87 measurements, which closely agree with independent sensors (Wang et al., 2023; Bonne et al., 2019), without relying on
88 external references. Following Liu et al. (2014), two VSMOW-standard water samples were analyzed at three flow rates (0.015,
89 0.04, 0.07 $\mu\text{L/s}$) to generate three humidity levels of 5000, 15000, and 25000 ppm, and each step was maintained for 20 minutes.
90 A full calibration cycle was performed approximately every 22 hours during atmospheric measurements. To minimize memory
91 effects, the first 10 minutes and final 2 minutes of each calibration step were excluded. A correction function was derived from
92 the valid calibration data to express isotope bias as a function of humidity and was applied to all samples to normalize their
93 values to a reference humidity of 20000 ppm:

$$\delta_{\text{measured}} - \delta_{\text{humidity calibration}} = f(\text{humidity}_{\text{measured}}) - f(20000) \quad (3)$$

$$f_{\delta^{18}\text{O}}(\text{humidity}) = 0.2976\ln(x) - 11.969 \quad (4)$$

$$f_{\delta\text{D}}(\text{humidity}) = 0.2843\ln(x) - 30.892 \quad (5)$$

Equations (4) and (5) describe the humidity-dependent isotope bias for $\delta^{18}\text{O}$ and δD , respectively, derived from the calibration measurements. Equation (3) was used to correct all isotope measurements to a reference humidity of 20000 ppm, thereby minimizing the influence of water vapor concentration on the measured isotope values.

All vapor isotope measurements were then normalized to the VSMOW scale using two laboratory reference waters ($\delta^{18}\text{O} = -11.02\text{‰}$, $\delta\text{D} = -78.1\text{‰}$; $\delta^{18}\text{O} = -29.86\text{‰}$, $\delta\text{D} = -222.9\text{‰}$), whose isotope values match the expected range of water vapor isotopes in the Arctic. The certified uncertainties provided by the calibration laboratory were $\pm 0.021\text{‰}$ ($\delta^{18}\text{O}$) and $\pm 0.530\text{‰}$ (δD) for the first standard, and $\pm 0.006\text{‰}$ ($\delta^{18}\text{O}$) and $\pm 0.104\text{‰}$ (δD) for the second standard. Both standards were pre-corrected for humidity-dependent bias, and a linear calibration function was established based on the measured versus true values of the two standards and applied to all samples to complete the VSMOW normalization.

2.2 Meteorological and sea ice data

Meteorological data, including air temperature (T), relative humidity (RH), and air pressure, were obtained from the onboard weather station of the RV *Xuelong 2*, located in close proximity to the isotope inlet. Specific humidity (q) was calculated from the water vapor concentration measured by the Picarro instrument. All datasets were synchronized to the isotope measurement timestamps (UTC+8) and averaged to 1-minute resolution. Surface evaporation flux (E), representing the instantaneous vertical water vapor flux from the ocean surface, was obtained from the fifth-generation European Centre for Medium-Range Weather Forecasts reanalysis (ERA5; Hersbach et al. (2023)), available at 1-hour temporal and $0.5^\circ \times 0.5^\circ$ horizontal resolution.

The daily 4-km sea ice concentration (SIC) data were obtained from the National Snow and Ice Data Center (NSIDC), specifically from the MASIE-AMSR2 (MASAM-2) blended product (Fetterer, 2023). This product integrates data from the Multisensor Analyzed Sea Ice Extent (MASIE) and the Advanced Microwave Scanning Radiometer 2 (AMSR2). To assess the influences of sea ice on water vapor isotopes and moisture source partitioning, the Arctic was classified into three sea ice regimes using contemporaneous SIC data (Fig. 2):

- I. Ice-free Region ($\text{SIC} < 0.4$): Areas below satellite detection threshold (MASAM-2 $\text{SIC} = 0$)
- II. Sea Ice Region ($\text{SIC} > 0.85$): Predominantly ice-covered areas
- III. Transition Region ($0.4 \leq \text{SIC} \leq 0.85$): Intermediate ice conditions

2.3 Theoretical Isotope Modelling of Ocean Evaporation

We employed the MJ79 model (Merlivat and Jouzel, 1979) to estimate the isotopic composition of water vapor evaporated from the Arctic Ocean, following the formulation of Bonne et al. (2019):

$$R_{BL} = \frac{R_{SW}}{\alpha_{eq} \times (\alpha_k + RH(1 - \alpha_k))} , \quad (6)$$

where R_{BL} is the isotopic ratio of water vapor in the boundary layer, and R_{SW} is the isotopic ratio of surface seawater. RH is the relative humidity at the sea surface. α_{eq} and α_k are the equilibrium and kinetic fractionation coefficients, respectively. α_{eq} was calculated as a function of temperature, while α_k takes values of 1.0060 for $\delta^{18}\text{O}$ and 1.0053 for δD under smooth wind conditions, and 1.0035 for $\delta^{18}\text{O}$ and 1.0031 for δD under rough wind conditions (Bonne et al., 2019). Because surface seawater isotope values in the open Arctic Ocean are typically close to 0 ‰ (Namyatov et al., 2024, 2023), we prescribed $\delta^{18}\text{O}$ and δD of surface seawater as 0 ‰, from which R_{SW} was calculated and used as the model input.

2.4 Back-trajectory calculation

To identify moisture sources in the Arctic, we analyzed air mass trajectories using the Hybrid Single Particle Lagrangian Integrated Trajectory (HYSPLIT) model (Stein et al., 2015). The model was driven by $0.5^\circ \times 0.5^\circ$ Global Data Assimilation System (GDAS) meteorological fields from the National Oceanic and Atmospheric Administration (NOAA) Air Resources Laboratory.

Ten-day backward trajectories were computed hourly along the cruise track at five vertical levels (10, 20, 30, 40, and 50 m above the surface). This duration aligns with the typical residence time of atmospheric water vapor in polar regions (Gimeno et al., 2021; Wallace and Hobbs, 2006) and is long enough to capture moisture transport from lower latitudes. The selected vertical levels span the full range of the research vessel's measurements, enabling a more comprehensive representation of air mass histories. Building on these trajectory calculations, moisture-weighted source contributions from the Pacific and Atlantic sectors were quantified. All trajectories were batch-processed using the PySPLIT Python package (Warner, 2018).

2.5 Bayesian isotope mixing model

To quantify the contributions of different moisture sources to Arctic water vapor, we employed MixSIAR (Stock and Semmens, 2016a), an open-source Bayesian mixing model implemented in R. MixSIAR, developed from earlier MIXSIR and SIAR, has been widely applied in ecological and environmental studies. The model infers source contributions by representing each vapor isotope measurement as a mixture of multiple endmembers and exploring all feasible combinations of source proportions. Solutions that satisfy isotope mass balance within the uncertainty of both endmembers and observations are retained, forming the posterior distribution of source contributions. According to Stock et al. (2018), the model framework can be expressed as:

$$Y_j = \sum_k p_k u_{jk}^s , \quad (7)$$

where Y_j is the tracer value of the mixture for tracer j , p_k is the proportional contribution of source k , and u_{jk}^s is the mean tracer value of source k . Given the substantial isotopic variability in Arctic water sources and its propagation into the vapor mixture,

we evaluated three error structures (residual, process, multiplicative) and adopted the process error structure as most appropriate (Stock and Semmens, 2016b):

$$Y_{ij} \sim N\left(\sum_k p_k u_{jk}^s, \sum_k p_k^2 \omega_{jk}^{s2}\right) \quad (8)$$

where N represents the normal distribution and ω represents the source variance of the two stable isotope tracers. Posterior distributions were estimated using Markov Chain Monte Carlo (MCMC) sampling with the "Normal" run length setting (chain length = 100,000; burn-in = 50,000; thinning = 50; 3 chains), following the MixSIAR manual (Stock and Semmens, 2016a). Given the limited knowledge of moisture source partitioning in the Arctic, uninformative Dirichlet priors ($\alpha = 1$) were assigned to the source proportions, specifying equal prior probability for all potential source mixtures. Model convergence was assessed using the Gelman-Rubin diagnostic, with values below 1.05 indicating satisfactory convergence of the MCMC chains. To construct a representative lower-latitude moisture end-member, trajectory statistics were performed using a custom Python script. Source end-member data were input as means and standard deviations. Air trajectories from Pacific (160°E-160°W) and Atlantic (15°E-30°W) were integrated and weighted by the specific humidity of each backward trajectory at the time it crossed the Arctic boundary (66.5°N). The weight W_k for each channel k was calculated as:

$$W_k = \frac{\sum_{i=1}^{N_k} q_{i,cross}}{\sum_{all} q_{i,cross}} \quad (9)$$

where $q_{i,cross}$ represents the specific humidity of trajectory i at the moment of crossing 66.5°N. The composite isotopic mean δ_{mix} was then derived using Eq. (10), and the mixed standard deviation σ_{mix} was calculated following the Gaussian error propagation rule (Eq. 11):

$$\delta_{mix} = \sum_k (W_k * \delta_k) \quad (10)$$

$$\sigma_{mix} = \sqrt{\sum_k (W_k \cdot \sigma_k)^2} \quad (11)$$

A Rayleigh distillation correction was applied to the lower-latitude source end-member to account for isotopic depletion during poleward transport. The initial source region is defined as the North Pacific (50–60°N), with a saturation specific humidity of 6.6 g/kg (corresponding to ~8°C). Using the weighted approach described above, initial $\delta^{18}\text{O}$ and δD values of -17.68‰ and -134.4‰ were obtained. This air mass was modified according to a Rayleigh fractionation curve to a target humidity of 3.93 ± 1.84 g/kg, which represents the hourly mean saturation specific humidity across the Arctic domain resulting from rainout of lower-latitude water vapor along the trajectories. The calculated isotope values, representing the discrimination between the lower-latitude source and the Arctic background state due to Rayleigh fractionation, were incorporated into the MixSIAR model.

The local evaporation end-member was derived using hourly air temperature and relative humidity from shipboard observations in ice-free region as inputs for the MJ79 model (see section 2.3), implemented via a Python script. The model

181 was run in batch mode to calculate the isotopic composition of vapor due to local evaporation. The resulting means and standard
182 deviations were used to define the local evaporation endmember in the MixSIAR model.

183 To assess the sensitivity of the inferred source contributions to endmember variability and prior assumptions, sensitivity
184 analyses were conducted. For endmember selection, the isotopic compositions of both local and remote source endmembers
185 were perturbed by $\pm 20\%$, corresponding to approximately $\pm 1\sigma$ of the observed variability during the campaign, and 1,000
186 randomized endmember sets were generated and propagated through the MixSIAR framework. For the Bayesian mixing model
187 configuration, alternative Dirichlet priors were tested, including an uninformative prior ($\alpha = (1,1)$), a weakly informative prior
188 ($\alpha = (2,2)$), and biased priors favoring each endmember ($\alpha = (3,1)$ and $\alpha = (1,3)$).

189 **3 Results**

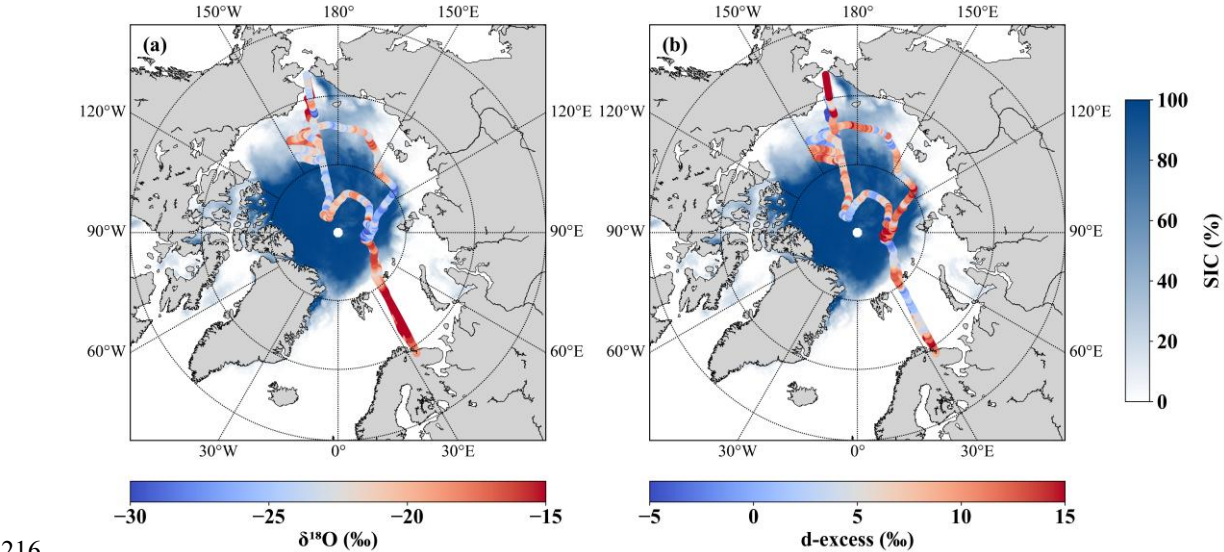
190 **3.1 Water Vapor Isotope Variability Along the Cruise Track**

191 Figure 1 shows the evolution of near-surface water vapor isotopes and sea ice coverage along the trans-Arctic RV *Xuelong 2*
192 expedition track spanning the Chukchi Sea, central Arctic Basin, Barents Sea, and Laptev–East Siberian Seas, suggesting co-
193 variability between them. $\delta^{18}\text{O}$ exhibited a pronounced spatial gradient along the cruise track, with the most enriched values
194 (-10.8‰) in the ice-free Barents Sea, gradually becoming depleted poleward to the most depleted values (-34.19‰) in the
195 heavily ice-covered central Arctic ($\text{SIC} > 90\%$), a pattern that inversely mirrors the gradient in sea ice coverage (Fig. 1a). This
196 anti-phase variation between water vapor $\delta^{18}\text{O}$ and sea ice coverage is further confirmed by the $\delta^{18}\text{O}$ distributions across
197 different sea-ice coverage regions (Fig. 2a), which reveal a systematic $\delta^{18}\text{O}$ enrichment from the Sea Ice Region (median =
198 -23.97‰), through the Transition Region (-20.54‰), to the Ice-free Region (-18.83‰). These results suggest that sea ice
199 changes may contribute to variations in water vapor isotope composition across the Arctic.

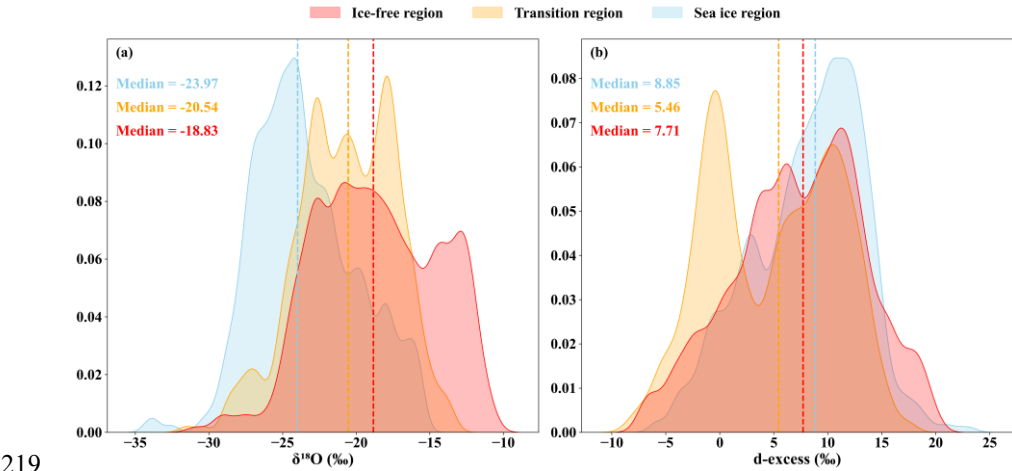
200 In contrast, the second-order parameter d-excess exhibits a more complex spatial pattern, with elevated values primarily
201 observed both near coastal regions and in some localized zones of the ice-covered central basin (Fig. 1b). The highest values
202 (23.31‰) were recorded in the central basin under steady sea ice cover, while the lowest (-8.35‰) occurred in the Chukchi
203 Sea amid fluctuating sea ice (SIC varied from 40% to 85%). When grouped by SIC regimes, both the Ice-free and Sea Ice
204 Regions sustained relatively higher d-excess values (median = 7.71‰ and 8.85‰ , respectively), whereas the Transition
205 Region yielded the lowest values (median = 5.46‰) (Fig. 2b). This different d-excess pattern across sea ice regimes points to
206 the influence of sea ice changes on Arctic water vapor isotopic composition, likely through shifts in moisture sources and
207 kinetic fractionation processes.

208 To examine how sea ice changes modulate Arctic water vapor isotopes, we show the variations of $\delta^{18}\text{O}$ and d-excess in relation
209 to temperature, humidity, and local evaporation along the cruise track (Fig. 3). A comparison across the three sea ice regimes
210 reveals that $\delta^{18}\text{O}$ and d-excess exhibited pronounced, anti-phase fluctuations (Fig. 3b). Depleted $\delta^{18}\text{O}$ generally coincided with
211 colder, drier air in the ice-covered central Arctic, while enriched values occurred in warm, moist sectors like the Barents

212 Sea (Fig. 3d). The episodes of high d-excess were closely associated with conditions of reduced relative humidity and enhanced
 213 evaporation, particularly in the Ice-free Region (Fig. 3c), highlighting the enhanced influence of kinetic fractionation over
 214 open water. These co-variations demonstrate that the sea ice regimes set the boundary conditions for the atmospheric
 215 controls—from equilibrium (temperature-driven) to kinetic (evaporation-driven)—on Arctic water vapor isotopes.



216
 217 **Figure 1. Spatial distributions of water vapor $\delta^{18}\text{O}$ (a) and d-excess (b) along the cruise track. Colored dots indicate measured**
 218 **isotopic values (‰) and shading shows the mean sea ice concentration (SIC) during the observation period.**



219
 220 **Figure 2. Probability density distributions of $\delta^{18}\text{O}$ (a) and d-excess (b) in the sea ice Ice-free (red), Transition (orange), and Sea Ice**
 221 **regions. Median values for each region are indicated.**

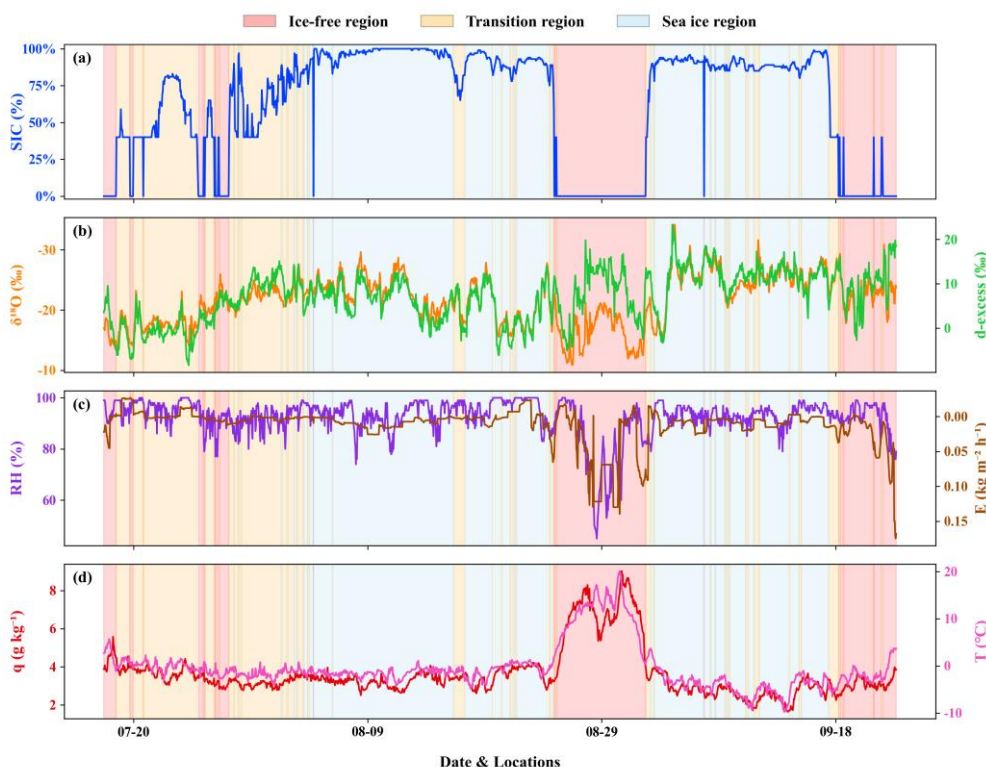


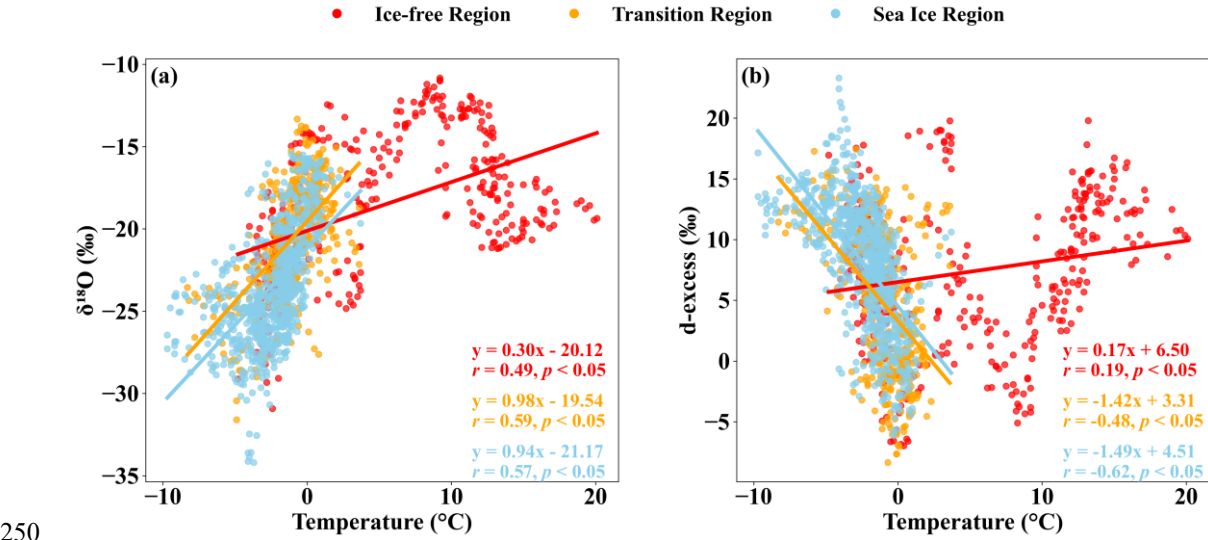
Figure 3. Temporal and spatial evolution of observed water vapor isotopes, SIC, and meteorological parameters along the cruise routes. (a) SIC (%); (b) $\delta^{18}\text{O}$ (‰) and d-excess (‰); (c) Relative humidity (RH; %) and Evaporation (E ; $\text{kg m}^{-2} \text{h}^{-1}$); (d) Specific humidity (q ; g kg^{-1}) and air temperature (T ; $^{\circ}\text{C}$). Background colors indicate the three different sea ice regimes.

3.2 Isotopic Responses to Temperature and Humidity across Arctic Sea Ice Regimes

To investigate isotopic responses to temperature and humidity, we first examined the relationship between $\delta^{18}\text{O}$ and air temperature in the three distinct sea ice regions. As expected, $\delta^{18}\text{O}$ values exhibit significant positive correlations with temperature in all regions (Fig. 4a), consistent with the classical “temperature effect” (Dansgaard, 1964). However, this relationship weakens notably in the Ice-free Region and reverses when temperatures exceed 5°C , during which $\delta^{18}\text{O}$ becomes negatively correlated with temperature, indicating a departure from equilibrium fractionation and the emergence of dominant kinetic processes under warm, ice-free conditions.

In contrast, d-excess shows a more variable response to temperature. Significant negative correlations are observed in the Sea Ice and Transition Regions ($r = -0.62$ and -0.49 , respectively; $p < 0.05$; Fig. 4b), whereas the Ice-free Region exhibits a weak positive correlation ($r = 0.19$; $p < 0.05$; Fig. 4b). In the Ice-free Region, the correlation shifts from negative to positive when temperature is above 5°C , suggesting a shift in the dominant controls on isotopic variability. Although d-excess is theoretically not expected to directly depend on temperature (Shao et al., 2021; Xiang et al., 2022), previous studies have shown that a

238 positive correlation can emerge as a signature of oceanic evaporation (Uemura et al., 2008). This suggests that the correlation
 239 between d-excess and temperature may serve as a diagnostic of the influence of local evaporation on Arctic water vapor
 240 (Brunello et al., 2023). Collectively, these patterns indicate that fractionation mechanisms undergo fundamental shifts between
 241 ice-covered and open-water environments, with an enhanced role of kinetic effects over the ice-free ocean.
 242 Relative humidity (RH) serves as a diagnostic of atmospheric sub-saturation, which controls the magnitude of kinetic isotopic
 243 fractionation during evaporation (Bonne et al., 2019). Across all sea ice regimes, d-excess exhibits a consistent negative
 244 correlation with RH (Fig. 5a), with the strongest correlation in the Ice-free Region ($r = -0.52$, $p < 0.05$), followed by the
 245 Transition ($r = -0.37$) and Sea Ice Regions ($r = -0.33$). A similar pattern is observed for the relationship between d-excess and
 246 surface evaporation (Fig. 5b), with the strongest positive correlation in the Ice-free Region ($r = 0.51$) and weaker ones in the
 247 Transition ($r = 0.40$) and Sea Ice ($r = 0.24$) Regions. These results indicate that as sea ice retreats, local evaporation increasingly
 248 governs the isotopic composition of Arctic water vapor, consistent with the elevated d-excess observed in ice-free areas, while
 249 its influence remains relatively weak in ice-covered regions.



251 **Figure 4. Relationships of water vapor isotopes with air temperature across three sea ice regimes. (a) $\delta^{18}\text{O}$ versus temperature; (b)**
 252 **d-excess versus temperature. Dots and linear regression lines are color-coded by regime: ice-free (red), transition (yellow), and sea**
 253 **ice (blue). Regression statistics are displayed in corresponding colors.**

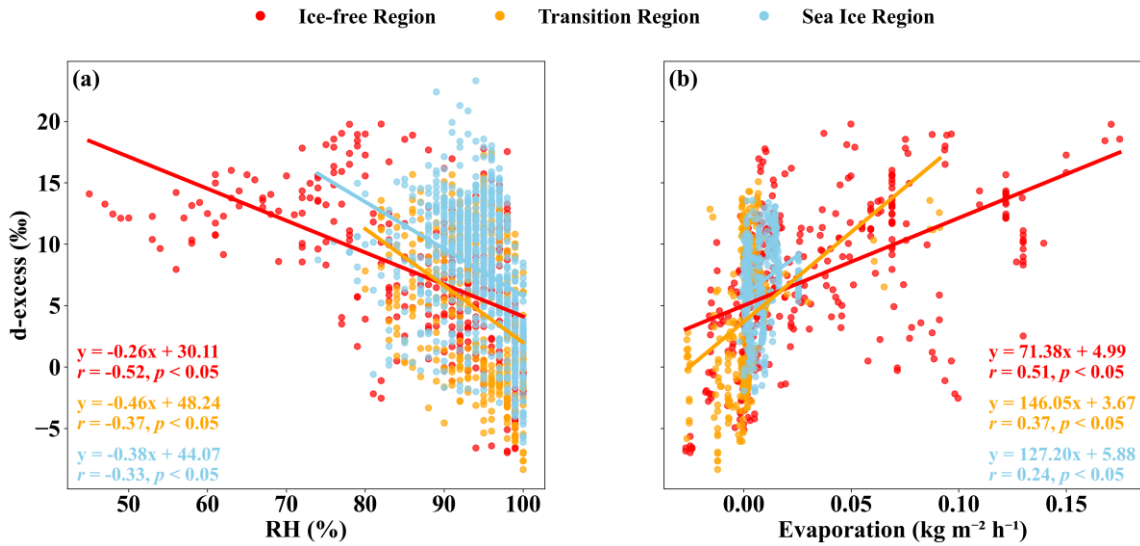


Figure 5. Relationships of d-excess with relative humidity and evaporation across three sea ice regimes. (a) d-excess versus relative humidity (RH); (b) d-excess versus evaporation flux. Dots and linear regression lines are color-coded by regime: ice-free (red), transition (yellow), and sea ice (blue). Regression statistics are displayed in corresponding colors.

3.3 Moisture Sources and Fractionation Processes across Arctic Sea Ice Regimes

To disentangle the influences of equilibrium and kinetic fractionation on Arctic water vapor isotopes, we employ q - δ diagrams as a diagnostic tool (Fig. 6a). Within a Rayleigh framework, $\delta^{18}\text{O}$ decreases systematically as vapor is progressively depleted through condensation during long-range transport, whereas d-excess remains relatively constant (Worden et al., 2007; Galewsky and Samuels-Crow, 2015; Noone, 2012). In the q - δ diagram, the idealized Rayleigh process forms a reference curve along which data distributions reflect the dominant influence of equilibrium fractionation. Data points deviating from this trajectory indicate additional physical processes: values above the curve often reflect admixture with other air masses (Wang et al., 2023), those below the mixing line reflect interactions between air masses of contrasting humidity (Galewsky and Samuels-Crow, 2015), and points below the Rayleigh curve typically signify local evaporation, such as from rain or open water (Worden et al., 2007). To specifically diagnose kinetic effects, we also examine a q -d-excess diagram (Fig. 6b), accounting for realistic polar conditions with Rayleigh curves modified for ice supersaturation ($\text{RH}_i = 100\text{--}110\%$). Under these conditions, kinetic fractionation suppresses d-excess below equilibrium predictions (Jouzel and Merlivat, 1984; Jensen and Pfister, 2005). Within this framework, points falling below the equilibrium curve primarily reflect d-excess suppression by ice-supersaturated cloud formation, whereas points above it indicate contributions from evaporation or sublimation (Samuels-Crow et al., 2014; Kopec et al., 2019). Collectively, these dual diagrams allow us to distinguish the roles of **rainout during long-range transport**, air-mass mixing, ice-phase microphysics, and local evaporation in shaping Arctic vapor isotopes across the sea ice regimes.

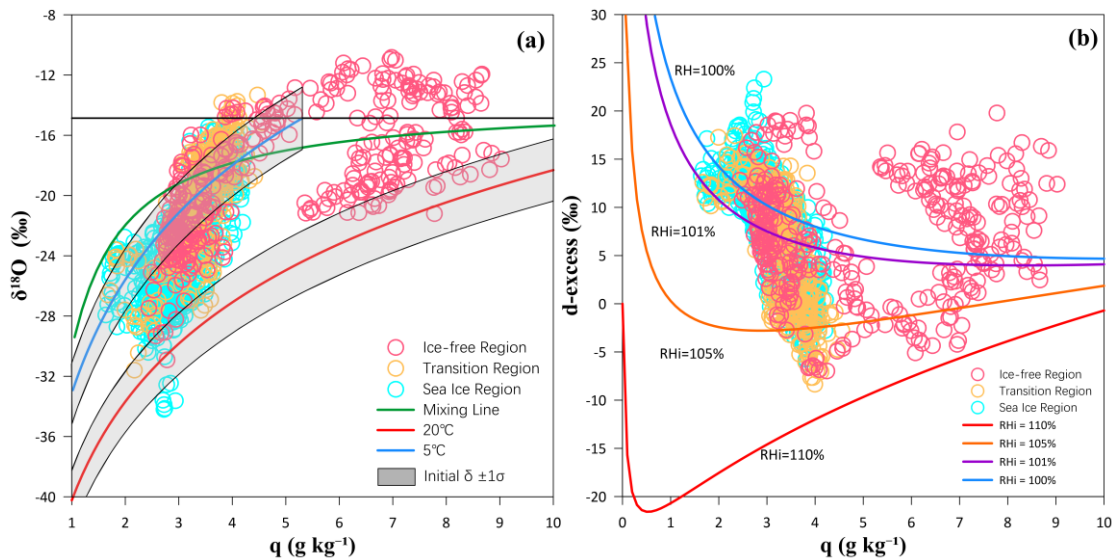


Figure 6. Water vapor $\delta^{18}\text{O}$ (a) and d-excess (b) versus specific humidity (q). Data points are color-coded by sea ice regime, theoretical Rayleigh distillation curves and mixing/supersaturation curves are shown as colored lines (see legends). In (a), Rayleigh curves were initialized using isotopic compositions representative of mid-latitude conditions ($40\text{--}60^\circ\text{N}$; $\delta^{18}\text{O} = -14.86\text{‰}$). Shaded areas indicate the range of theoretical Rayleigh curves associated with ± 1 standard deviation variations in the initial $\delta^{18}\text{O}$ values. Initial specific humidity was calculated assuming saturation at the corresponding source temperature ($q_0 = 14.3$ and 5.31 g kg^{-1}). In (b), colored curves represent supersaturation scenarios initialized with d-excess = 5.01‰ under the same fixed initial specific humidity ($q_0 = 14.3 \text{ g kg}^{-1}$).

In the Sea Ice and Transition Regions, q – $\delta^{18}\text{O}$ observations cluster near or between the 5°C and 20°C Rayleigh curves (Fig. 6a), consistent with vapor originating from advection and progressive dehydration during long-range transport. The q –d-excess relationship in these regions is located around the Rayleigh curve (Fig. 6b), supporting the interpretation that Rayleigh distillation, rather than local kinetic fractionation, primarily governs water vapor isotopic variation in these regions. This also explains the contrasting temperature correlations observed in Fig. 4: a positive correlation for $\delta^{18}\text{O}$ but a negative one for d-excess, reflecting the advection of warm, humid air from lower-latitude oceanic regions toward the Arctic. However, the decline in d-excess is steeper than predicted by the equilibrium Rayleigh curve, with many points falling below it (Fig. 6b). This deviation indicates additional kinetic fractionation, most likely from cloud formation under ice-supersaturated conditions, where the higher diffusivity of HDO compared to H_2^{16}O leads to stronger fractionation of deuterium, suppressing d-excess in surrounding vapor (Clark and Fritz, 2013; Jouzel and Merlivat, 1984). Our calculations show that RH with respect to ice (RH_i) exceeding 105% —common in polar regions—can reduce d-excess by over 10% . This supersaturation-induced decline in water vapor d-excess has also been observed during dew deposition (Thurnherr and Aemisegger, 2022), indicating that supersaturated conditions drive d-excess reduction regardless of the phase transition pathway. Conversely, a limited number of points above the Rayleigh curve likely reflect contributions from sublimation of snow or sea ice (Kopec et al., 2019). Overall, the isotopic composition of water vapor in the Sea Ice and Transition Regions is primarily governed by Rayleigh fractionation during advective transport, while d-excess provides a distinct, sensitive record of local kinetic processes at the

ice-atmosphere interface, such as supersaturation and sublimation. However, our interpretation of d-excess variations over sea ice is largely dependent on Rayleigh fractionation curves, with contributions from sublimation and processes under supersaturated conditions remaining poorly constrained.

In contrast, the Ice-free Region shows clear deviations from the theoretical Rayleigh distillation curves. Observations show a broad scatter in both q - $\delta^{18}\text{O}$ and q -d-excess diagrams, indicating the interplay between advected and locally evaporated vapor under ice-free conditions. At lower specific humidity ($< 5 \text{ g/kg}$), isotopic patterns resemble those of ice-covered regions (Fig. 6), suggesting a persistent influence of advective moisture. However, as humidity rises ($> 5 \text{ g/kg}$)—which coincides with surface air temperatures exceeding $\sim 5^\circ\text{C}$ —systematic deviations from Rayleigh behavior become evident, reflecting an increased influence of local processes. In the q - $\delta^{18}\text{O}$ diagram, two distinct patterns emerge: one cluster is enriched well above the mean mid-latitude $\delta^{18}\text{O}$ value, indicating input from warm, isotopically heavy sources; the other falls below the 5°C Rayleigh curve, where its scattered, non-logarithmic distribution—though still bounded by the 20°C curve and the mixing line—points to strong kinetic fractionation. This divergence is systematically mirrored in the q -d-excess diagram (Fig. 6b), where values split both below and above the equilibrium curve. The co-occurrence of $\delta^{18}\text{O}$ enrichment and elevated d-excess in this high-humidity regime supports the interpretation that kinetic fractionation during local evaporation is an important driver of Arctic water vapor isotopic composition under ice-free conditions.

In summary, Arctic water vapor isotopes exhibited contrasting characteristics between ice-covered and ice-free regions. In the Sea Ice and Transition Regions, isotope variations were broadly consistent with Rayleigh distillation during long-range water vapor transport, with possible modulation by kinetic effects associated with ice-related processes such as supersaturation and sublimation. In the Ice-free Region, however, the observed isotopic behavior deviated from the classical Rayleigh framework, reflecting an enhanced influence of local evaporation and mixing with advected moisture over open water. Together, these results suggest that Arctic water vapor isotopic variability differs across sea ice conditions.

3.4 Evaluating the Influence of Local Evaporation on Ice-free Region Water Vapor Isotopes

To test whether local evaporation can account for the observed deviations from Rayleigh behavior, particularly elevated d-excess under warm and humid conditions, we applied the MJ79 marine boundary layer evaporation model (Merlivat and Jouzel, 1979; Bonne et al., 2019). The model was driven by observed surface temperature and relative humidity and accounts for both equilibrium fractionation, controlled by temperature, and kinetic fractionation, modulated by relative humidity. Following Bonne et al. (2019), we adopted kinetic fractionation factors representative of the rough-wind regime ($> 7 \text{ m s}^{-1}$) for all simulations.

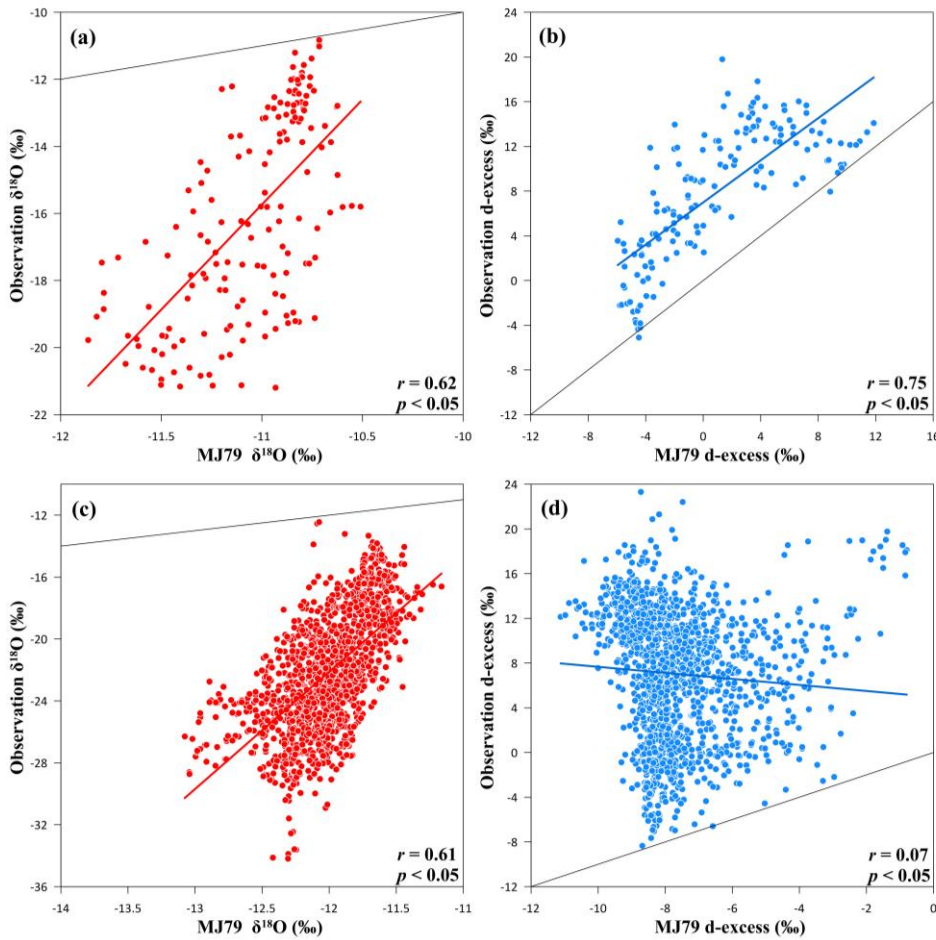


Figure 7. Observed versus MJ79-simulated water vapor isotopes ($\delta^{18}\text{O}$ and d-excess) across sea ice regions under different temperature regimes: (a, b) $>5^\circ\text{C}$; (c, d) $<5^\circ\text{C}$. The black line denotes the 1:1 line ($y = x$).

The first set of simulations were restricted to samples with temperatures exceeding 5°C , corresponding to ice-free, high temperature-low relative humidity conditions where local evaporation is expected to dominate. Under these settings, the model reproduces the observed isotopic variability reasonably well, with correlations of 0.62 for $\delta^{18}\text{O}$ and 0.75 for d-excess (Fig. 7a, b). This agreement supports the interpretation that kinetic fractionation during local evaporation contributes substantially to the isotopic composition of Arctic water vapor in the Ice-free Region. However, the model substantially overestimates the observed values, consistent with a previous study by Bonne et al. (2019). This overestimation largely reflects the closure assumption inherent in the MJ79 model, which treats local evaporation as the sole moisture source and neglects atmospheric mixing.

In contrast, when the temperature is below 5°C , although the simulation shows a reasonable correlation with the observed $\delta^{18}\text{O}$ ($r = 0.61$), it fails to reproduce the d-excess variability ($r = 0.07$) (Fig. 7c, d). This failure of the model to represent d-excess variability suggests that, under cold conditions, the moisture budget is dominated by non-local sources rather than local

341 evaporation. This further supports our interpretation that advective moisture transport dominates isotopic variations in ice-
342 covered regions.

343 **3.5 Moisture Source Attribution from Lagrangian Trajectory Analysis**

344 To test our isotope-based interpretation of moisture sources, we computed backward trajectories along the cruise track using
345 the HYSPLIT model (Fig. 8 and Figs. S2-4). The analysis reveals a clear dichotomy in air mass origins aligned with sea ice
346 regimes. In the Sea Ice and Transition Regions, trajectories show air masses undergoing rapid cooling and dehydration during
347 poleward transport, as reflected by declining specific humidity and rising relative humidity. Under these cold, saturated
348 conditions, evaporation is suppressed, allowing Rayleigh distillation of advected moisture to emerge as the characteristic
349 process, consistent with the observed alignment to Rayleigh curves (Fig. 6). Moreover, the frequent recirculation within the
350 Arctic interior points to prolonged moisture residence under stable, cold conditions. This extended residency favors non-
351 equilibrium ice-related processes, such as crystal formation and dew deposition under supersaturated conditions and
352 sublimation, providing a coherent explanation for the distinctive d-excess signals observed in these regions (Fig. 6b).

353 In the Ice-free Region, however, trajectories reveal two distinct source pathways, explaining its complex isotopic signals. Most
354 originate from lower latitudes, carrying warm air northward across the Barents Sea. Elevated temperatures in the Barents Sea
355 create a large saturation vapor deficit (via the Clausius–Clapeyron relationship), which maintains low relative humidity despite
356 high specific humidity. These conditions promote strong evaporation over the ice-free ocean, contributing to the d-excess
357 enrichment and reversed δ -temperature relationship observed in this region (Fig. 4) and confirming that local evaporation is a
358 major driver of isotopic variability under ice-free conditions. A smaller subset of air masses recirculates within the cold,
359 saturated Arctic interior. Their limited interaction with open water suppresses evaporation, preserving an advective signature
360 similar to that of ice-covered regions. Together, these dual air mass origins—one characterized by strong local evaporation
361 and the other by advected moisture—explain the coexistence of both isotopic end-members observed in the q - $\delta^{18}\text{O}$ and q - d -
362 excess relationships.

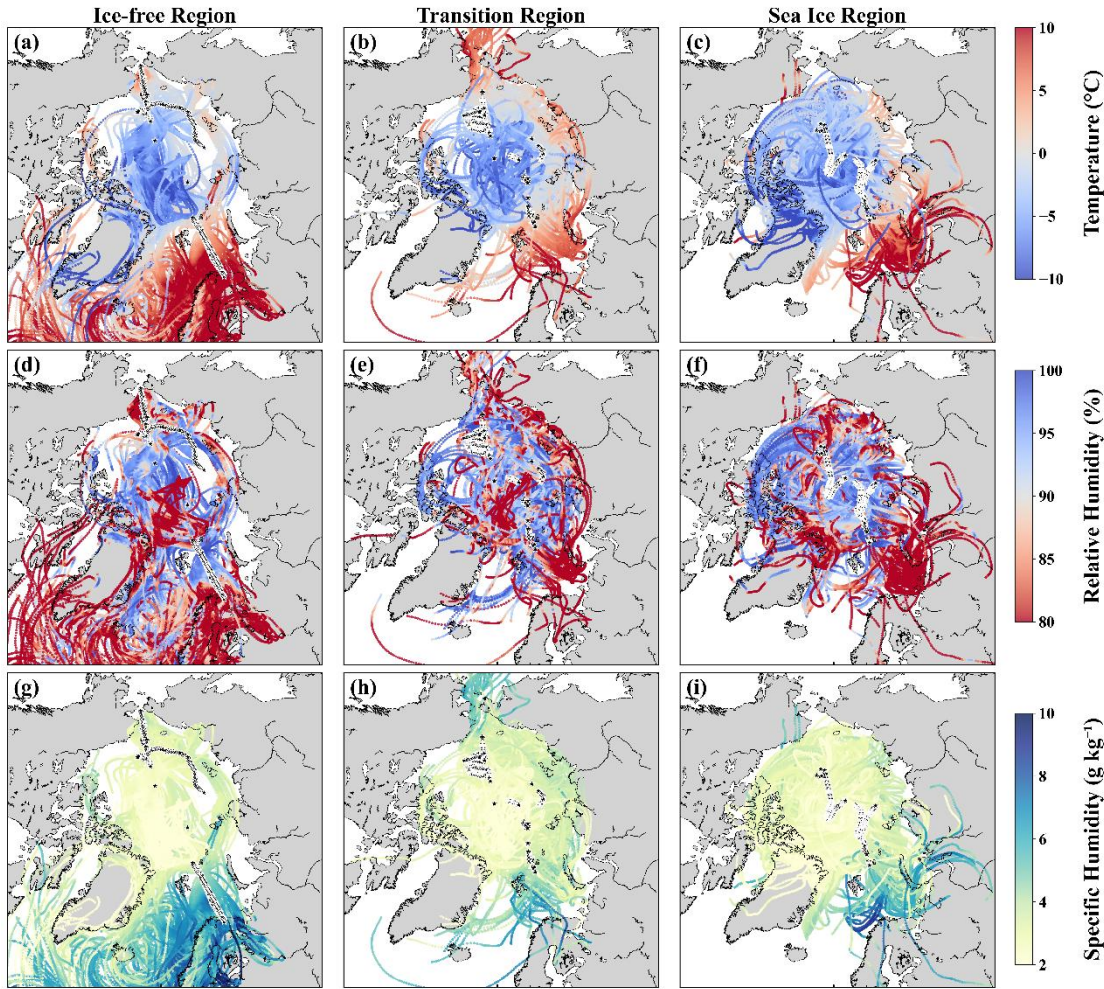
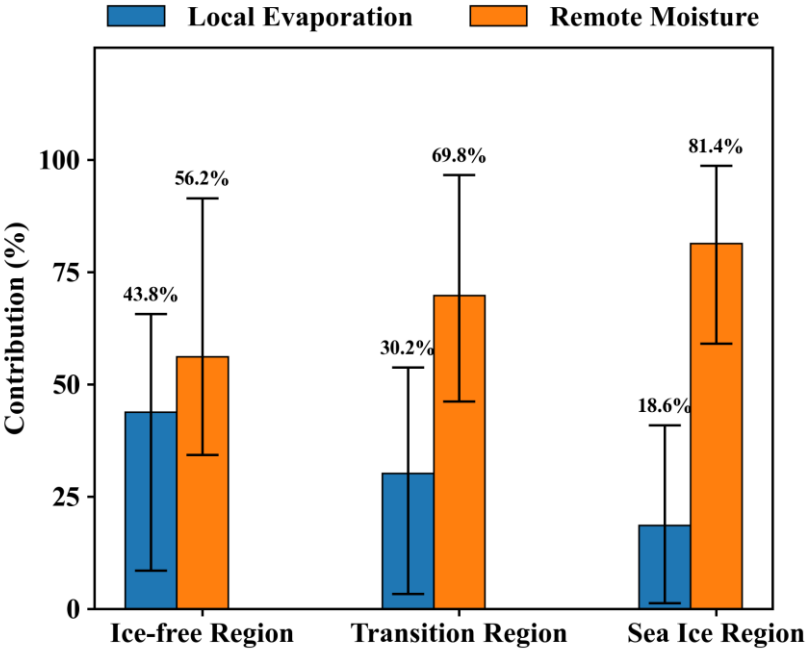


Figure 8. Ten-day HYSPLIT back trajectories from 10 m and associated meteorology by sea ice regime: ice-free (left), transition (middle), and sea ice (right). Rows show (a–c) air temperature (°C), (d–f) relative humidity (%), and (g–i) specific humidity (g kg⁻¹). black stars denote start points.

3.6 Relative Contributions of Moisture Sources Across Arctic Sea Ice Regimes

To quantify the relative contributions of locally evaporated and advected moisture to Arctic boundary layer water vapor across different sea ice regimes, we apply the Bayesian isotope mixing model MixSIAR (Stock et al., 2018) to $\delta^{18}\text{O}$ and δD measurements. Locally evaporated moisture is derived from Arctic open-water evaporation, with $\delta^{18}\text{O} = -11.77 \pm 0.59 \text{ ‰}$ and $\delta\text{D} = -96.0 \pm 7.7 \text{ ‰}$, estimated using the MJ79 model (see Data and Methods). Advected lower-latitude moisture, as identified by HYSPLIT trajectories (Fig. 8), primarily originates from the North Atlantic and the North Pacific, with isotopic endmembers taken from Benetti et al. (2017) near 50–60°N for the Atlantic ($\delta^{18}\text{O} = -18.25 \pm 4.79 \text{ ‰}$, $\delta\text{D} = -138.5 \pm 34.8 \text{ ‰}$) and from our observations over 50–60°N for the Pacific ($\delta^{18}\text{O} = -15.54 \pm 1.67 \text{ ‰}$, $\delta\text{D} = -118.9 \pm 9.0 \text{ ‰}$). A weighted average

375 based on trajectory specific humidity (Atlantic = 79.07 % and Pacific = 20.93 %; see Data and Methods) is used to represent
 376 a single lower-latitude source ($\delta^{18}\text{O} = -17.68 \pm 3.8 \text{ ‰}$, $\delta\text{D} = -134.4 \pm 27.5 \text{ ‰}$). Calibrated vapor isotope measurements from
 377 each sea ice regime serve as the mixture input. A process error structure is applied to account for spatial variability and source
 378 uncertainty (see Data and Methods), and MixSIAR is run independently for each regime to estimate the probability
 379 distributions of source contributions.
 380 MixSIAR results reveal a clear spatial gradient in moisture source contributions across sea ice regimes (Fig. 9). Local
 381 evaporation contributes most in the Ice-free Region, with a mean of 43.8 % (95 % CI: 14.7–65.7 %), decreasing to 30.2 %
 382 (95 % CI: 13.8–53.8 %) in the Transition Region and 18.6 % (95 % CI: 11.1–40.9 %) in the Sea Ice Region. Although the
 383 absolute contribution of local evaporation is modest, it increases progressively with decreasing sea ice cover. Across all
 384 regimes, advected lower-latitude moisture remains the dominant source, accounting for 56.2–81.4 % of boundary-layer vapor.
 385 Sensitivity analyses further indicate that these source attribution results are not sensitive to reasonable variations in endmember
 386 definitions or prior assumptions (Figs. S5–S6). These findings indicate that, while long-range transport governs the Arctic
 387 vapor budget, local evaporation contributes a notable fraction in ice-free areas.



388
 389 **Figure 9. Contributions of local and low-latitude remote moisture estimated using a Bayesian isotope mixing model. Blue and orange**
 390 **bars indicate the proportions of local and low-latitude sources, respectively. Error bars denote the 95% credible intervals (2.5–**
 391 **97.5%), and the values above each bar represent the mean contributions.**

392 **4 Discussion and Conclusions**

393 **4.1 Water Vapor Isotopic Responses to Distinct Sea Ice Regimes**

394 Our sea-ice-based isotope framework helps reconcile previously conflicting interpretations of Arctic vapor d-excess by
395 showing that, although d-excess has traditionally been treated predominantly as a source signature (Dansgaard, 1964), it is also
396 highly sensitive to humidity and local evaporation conditions associated with different sea-ice regimes. Since the first polar
397 vapor isotope measurements reported unexpectedly higher d-excess in Arctic-sourced moisture compared with lower latitudes
398 (Kurita, 2011), subsequent studies have alternately documented both high (Kopec et al., 2019) and low (Klein et al., 2015;
399 Brunello et al., 2023) values under similar surface conditions. These apparently conflicting results likely reflect sampling under
400 contrasting sea-ice and humidity conditions associated with different kinetic fractionation environments, rather than intrinsic
401 differences in moisture sources.

402 Previous studies have reported conflicting relationships between polar vapor d-excess and sea ice. Bonne et al. (2019) found
403 a positive correlation with sea ice, whereas Klein and Welker (2016) reported a negative correlation. This divergence likely
404 stems from differences in the study domains, which sample distinct sea-ice regimes. Bonne et al. (2019) primarily analyzed
405 regions with sea-ice coverage greater than zero, corresponding to the transition-to-ice-covered regime in our study, where d-
406 excess increases with SIC (Fig. S1a). In contrast, Klein and Welker (2016) focused largely on the low-ice Bering Strait, an
407 environment analogous to our ice-free-to-transition region, in which the negative correlation reflects enhanced local
408 evaporation over open water (Fig. S1b). By spanning multiple sea-ice regimes, our observations during the cruise period reveal
409 that d-excess does not exhibit a simple monotonic relationship with sea ice, but instead shows a bimodal distribution, with
410 elevated values in both ice-covered and ice-free regions. This pattern reflects contrasting controlling mechanisms between the
411 two regimes: the ice-covered region is dominated by rainout during long-range moisture transport, with additional modification
412 by sublimation, whereas the ice-free region is governed primarily by local evaporation dynamics.

413 Our analyses are based on a classification of observations into sea-ice regimes, which provides an analytical framework for
414 organizing the data. While these regime-dependent differences highlight distinct isotopic behaviors, the sea-ice-based
415 classification should be viewed as an analytical framework rather than a strict representation of discrete physical boundaries.
416 Some overlap between categories, particularly between the transition and ice-covered regimes, is therefore expected due to the
417 continuous nature of environmental variability along the cruise track. Results should thus be interpreted in terms of
418 environmental gradients rather than strictly separated regimes. Future studies could further explore isotope variability using
419 continuous variables such as SST and humidity to complement this regime-based perspective.

420 **4.2 Relationship between isotopes and local meteorological conditions**

421 A negative d-excess–RH relationship is widely observed at both regional (Uemura et al., 2008) and global scales (Pfahl and
422 Sodemann, 2014) and has been explained mechanistically by the sign and magnitude of surface fluxes induced by large-scale
423 advection of air masses (Aemisegger and Sjolte, 2018; Thurnherr et al., 2021). In the Arctic, however, the relationship remains

poorly constrained due to the paucity of isotopic measurements. Our cross-Arctic water vapor isotope observations reveal an overall negative relationship between d-excess and RH, but with a progressively weaker correlation from the ice-free region to the sea-ice-covered region, indicating a diminishing influence of local evaporation as sea ice increases (Fig. 5). The slopes of the d-excess–RH relationship range from -0.46‰ to -0.26‰ across the different Arctic regimes and are comparable to values reported at lower latitudes (e.g., -0.4‰ ; Thurnherr et al. (2020)), highlighting the important role of local evaporation in shaping near-surface water-vapor d-excess. However, it should be noted that the relative humidity used in this study is calculated with respect to the in situ air temperature rather than normalized to the sea surface temperature (SST). Since relative humidity with respect to SST is the physically more relevant for isotopic fractionation during ocean evaporation (Aemisegger and Sjolte, 2018), the use of air temperature-based RH may partly weaken the observed correlations and affect the estimated regression slopes. Nevertheless, because air temperature and SST co-vary closely over the Arctic (Klein and Welker, 2016), the inter-regional contrasts in the d-excess–RH relationship identified here remain valid.

In contrast, our observations reveal pronounced differences in isotope–temperature relationships between the ice-free and ice-covered regions. The $\delta^{18}\text{O}$ –temperature correlation weakens from the ice-covered region toward the ice-free region and becomes clearly negative when temperatures exceed $5\text{ }^{\circ}\text{C}$ in the ice-free region (Fig. 4a). This interpretation is further supported by the d-excess–temperature relationships, which show weak correlation in the ice-free region but strong one in the ice-covered region (Fig. 4b). These patterns are consistent with previous observations over open water (Bonne et al., 2019; Klein and Welker, 2016; Sodemann et al., 2024) and in the ice-covered region (Brunello et al., 2023; Brunello et al., 2024), respectively. However, Brunello et al. (2024) and Sodemann et al. (2024) attributed the positive $\delta^{18}\text{O}$ –temperature and negative d-excess–temperature relationships to warm air intrusions, based on observed positive RH–temperature coupling. By contrast, our observations show a clear negative correlation between RH and temperature, indicating that the observed isotope variability is more consistent with local processes under contrasting sea-ice conditions rather than synoptic-scale moisture advection (Fig. 3). This difference likely reflects the contrasting observational frameworks, with our short-duration transect sampling being less sensitive to episodic warm air intrusions.

4.3 Quantifying the Shift in Moisture Sources

Stable water isotopes provide a quantitative framework to partition moisture source contributions across the study domain. The contribution from local evaporation increases from 18.6 % in the Sea Ice Region to 43.8 % in the Ice-free Region, demonstrating enhanced local recycling despite persistent advective dominance. This gradient aligns with observed surface moisture flux trends (Boisvert and Stroeve, 2015) and contextualizes previous observationally or model-based estimates ranging from 8 % (Domínguez et al., 2018) to over 35 % (Zhong et al., 2018). By providing a novel quantification based on vapor isotopes, this study highlights the capacity of in-situ isotopic monitoring to resolve complex Arctic hydrological processes.

Although we quantify the relative contributions of remotely advected and local evaporated moistures across different ice regimes, there are at least two caveats that should be noted. First, uncertainties arise from the determination of the end-members.

457 In particular, the remote end-member is constrained by limited low-latitude sampling and by the use of boundary humidity
458 rather than a full Lagrangian moisture budget, which may bias the estimated contribution of remote sources by neglecting
459 moisture turnover along transport pathways. In addition, although the MJ79 model serves as a standard local-evaporation end-
460 member, its closure assumption neglects atmospheric mixing and synoptic processes (Bonne et al., 2019). Consequently, the
461 simulated isotopic composition should be interpreted as a theoretical local-evaporation end-member rather than an exact
462 representation of atmospheric conditions. Second, the Bayesian mixing model relies on the combined use of $\delta^{18}\text{O}$ and δD ,
463 which are strongly correlated. This high correlation reduces the effective dimensionality of the isotopic constraints, limiting
464 the ability of the model to robustly distinguish between the two end members and introducing additional uncertainty. Despite
465 these caveats, our results suggest that enhanced local moisture contributions are associated with reduced sea-ice conditions,
466 consistent with the hypothesized feedback whereby sea-ice retreat promotes a more localized Arctic water cycle, as recorded
467 in both water-vapor (Bailey et al., 2021) and precipitation (Kopeck et al., 2016) isotopes. These results provide in-situ isotopic
468 observational constraints for evaluating near-surface humidity, precipitation recycling, and the fidelity of regional climate
469 models.

470 **4.4 Implications for Arctic Hydroclimate Research**

471 Beyond contemporary implications, our findings necessitate refinements in paleoclimate interpretations from polar ice-core
472 records. The canonical $\delta^{18}\text{O}$ –temperature relationship, long used as a paleo-thermometer, is modulated by sea ice extent: under
473 extensive ice cover, strong positive correlations reflect temperature-driven advective distillation, whereas in Transition and
474 Ice-free Regions, the relationship weakens or reverses, indicating that $\delta^{18}\text{O}$ integrates local evaporation and humidity signals
475 alongside temperature. Enriched $\delta^{18}\text{O}$ with elevated d-excess during warm, ice-reduced periods (e.g., interglacial) reflects
476 reorganized local moisture sources rather than temperature alone. We therefore advocate interpreting $\delta^{18}\text{O}$ –d-excess co-
477 variation within the context of different sea-ice regimes to help disentangle past temperature and sea ice signals, thereby linking
478 modern process understanding with paleo-isotope interpretation.

479 While our study proposes a sea ice–mediated isotopic framework, several research frontiers warrant further investigation.
480 Extending observations across seasons and years could help assess the climatological robustness of the proposed framework,
481 while expanding spatial coverage and incorporating additional tracers (e.g., ^{17}O -excess) may provide further constraints on the
482 underlying mechanisms. Integrating this framework with isotope-enabled climate models could facilitate evaluation of model
483 performance across both past and projected climates, while applications to ice-core records may offer additional insights into
484 the relative influences of temperature and sea-ice variability. Collectively, these efforts could advance our understanding of
485 Arctic isotope processes and hydroclimate responses under ongoing climate change.

486 In summary, this study demonstrates that Arctic near-surface water vapor isotopes exhibit contrasting characteristics between
487 ice-covered and ice-free conditions, reflecting differences in moisture sources and fractionation processes. **Under extensive
488 sea ice cover, the isotopic composition is primarily controlled by rainout during long-range moisture transport, resulting in a**

489 strong $\delta^{18}\text{O}$ –temperature correlation and elevated d-excess in colder regions. As sea ice retreats, local evaporation over open
490 water contributes more, imprinting a kinetic fingerprint of enriched $\delta^{18}\text{O}$ and elevated d-excess under warm, dry conditions.
491 This shift breaks down the canonical temperature effect, resulting in a negative $\delta^{18}\text{O}$ –temperature relationship. Quantifying
492 this regime shift, our Bayesian mixing model constrained by in-situ isotopes reveals that local evaporation contributions rise
493 from 18.6% in the Sea Ice Region to 43.8% in the Ice-free Region—demonstrating a 2.4-fold enhancement of local recycling
494 despite persistent advective dominance. These results highlight the important influence of contrasting sea-ice conditions on
495 Arctic isotopic variability across the ice–open water transition, and provide observational constraints for both contemporary
496 hydroclimate variability and paleoclimate interpretation.

497 **Data and code availability**

498 The ERA5 data are publicly available at: <https://cds.climate.copernicus.eu/datasets> (last access: September 2025). The SIC
499 data were adopted from the National Snow and Ice Data Center (NSIDC) at: <https://nsidc.org/data/g10005/versions/2> (last
500 access: September 2025). The Global Data Assimilation System (GDAS) meteorological fields were obtained from NOAA's
501 Air Resources Laboratory (<ftp://arlftp.arlhq.noaa.gov/pub/archives/gdas1/> (last access: September 2025)). The water vapor
502 isotope dataset has been submitted to the Polar Research Institute of China data repository (<https://www.chinare.org.cn/>).

503 **Author contributions**

504 Yuankun Zhang and Zhongfang Liu designed the research. Yuankun Zhang, Dongsheng Li, Zhiqing Li and Hebin Shao
505 performed the investigation, including the deployment and calibration of the ship-based instruments. Yuankun Zhang,
506 Zhongfang Liu, and Dongsheng Li performed the analysis. All authors contributed to the discussion of the results and the final
507 article. Yuankun Zhang drafted the paper with contributions from all co-authors. Zhongfang Liu checked and modified the
508 paper.

509 **Competing interests**

510 The authors declare that they have no conflict of interest.

511 **Acknowledgments**

512 We gratefully acknowledge the open-source contributions of Stock and Semmens for the MixSIAR R package, NOAA for the
513 HYSPLIT model, and Daniel Warner for the development and sharing of the pysplit package. Furthermore, we extend our
514 sincere gratitude to all members of the 14th Chinese National Arctic Research Expedition, especially the crew and scientists

515 aboard the RV *Xuelong 2*, for providing an exceptional operational environment and foundational support for the
516 measurements conducted in this study.

517 **Financial support**

518 This work is supported by the National Natural Science Foundation of China (Grant 42025602).

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