

Ice-core d-excess reveals a shift to wind-driven control of oceanic evaporation in warm climates

Ajay Ajay

ajphy1@gmail.com

Department of Earth Sciences, Indian Institute of Science Education and Research Kolkata, 741246, West Bengal, India <https://orcid.org/0000-0001-5601-0767>

Alexandre Cauquoin

Institute of Industrial Science (IIS), The University of Tokyo <https://orcid.org/0000-0002-4620-4696>

Kei Yoshimura

The University of Tokyo <https://orcid.org/0000-0002-5761-1561>

Jeffrey Welker

University of Alaska Anchorage <https://orcid.org/0000-0002-3865-4822>

Prasanta Sanyal

Indian Institute of Science Education and Research Kolkata <https://orcid.org/0000-0002-9751-9743>

Article

Keywords:

Posted Date: September 9th, 2026

DOI: <https://doi.org/10.21203/rs.3.rs-10068906/v1>

License: © ⓘ This work is licensed under a Creative Commons Attribution 4.0 International License.

[Read Full License](#)

Additional Declarations: There is **NO** Competing Interest.

1 Ice-core *d-excess* reveals a shift to wind-driven control of
2 oceanic evaporation in warm climates

3

4 **Authors:** Ajay Ajay^{1,2}, Alexandre Cauquoin³, Kei Yoshimura³, Jeffrey M Welker^{2,4,5},
5 Prasanta Sanyal^{1,6}

6 **Affiliations:**

7 ¹ Department of Earth Sciences, Indian Institute of Science Education and Research
8 Kolkata, 741246, West Bengal, India

9 ² Ecology and Genetics Research Unit, University of Oulu, Finland

10 ³ Institute of Industrial Science, University of Tokyo, Kashiwa, Chiba, Japan

11 ⁴ Department of Biological Sciences, University of Alaska Anchorage, AK, USA

12 ⁵ UArctic, University of the Arctic, Rovaniemi, Finland

13 ⁶ Center for Climate and Environmental Studies, Indian Institute of Science Education
14 and Research Kolkata; 741246, West Bengal, India

15

16

17 **Key Points**

18

- 19 ● Recalibrated against modern ocean evaporation, ice-core deuterium excess yields
20 the first continuous 800,000-year record of oceanic evaporation.
- 21 ● The control on evaporation reverses with climate state: source temperature
22 dominates under glacial conditions, wind-driven transport in warm and transitional
23 periods.
- 24 ● Source temperature and wind jointly explain 81% of evaporation variance over
25 800 kyr, versus 55% for temperature alone, making circulation a first-order control
26 in warm climates.

27

28

29

30

31

32

33 **Abstract**

34 Deuterium-excess (*d-excess*) in Antarctic ice-cores has long been used to reconstruct
35 glacial–interglacial changes in moisture-source temperature (ΔT_{source}) or
36 relative-humidity. Yet temperature, relative-humidity, and wind-speed jointly and
37 inseparably influence *d-excess*, an ambiguity that has undermined single-variable
38 readings. Here we resolve it by reframing *d-excess* as a proxy for the oceanic evaporation
39 rate (E): all three controls drive both evaporation and its *d-excess* in the same direction.
40 We calibrate the modern spatial E–*d-excess* relationship against reanalysis and satellite
41 evaporation products ($R^2 = 0.83\text{--}0.97$) and apply it to the EPICA Dome C record to
42 produce the first continuous 800,000-year history of moisture-source evaporation.
43 Reconstructed E couples tightly with ΔT_{source} during cold glacial periods ($\leq -3.5\text{ }^{\circ}\text{C}$; $R^2 =$
44 0.57) but decouples through the transitional (-3.5 to $-1\text{ }^{\circ}\text{C}$; $R^2 = 0.10$) and warm (≥ -1
45 $^{\circ}\text{C}$; $R^2 = 0.23$) regimes, where wind, proxied by EDC dust-flux, becomes the dominant
46 control. Jointly, source-temperature and wind explain 81% of evaporation variance across
47 800 kyr, against 55% for temperature alone. The control on evaporation shifts from
48 thermodynamic to dynamic as climate warms, implying that the response of the water
49 cycle to future warming is governed by atmospheric-circulation as much as by
50 temperature.

51

52

53

54

55 Main

56 The ocean-atmosphere exchange of moisture represents one of Earth's most fundamental
57 processes, driving global energy transport, precipitation patterns, and climate variability
58 across all timescales^{1,2,3}. Despite its central role in Earth system processes, direct
59 quantification of past oceanic evaporation rates has remained elusive. Reconstructing past
60 oceanic evaporation requires constraining multiple boundary conditions such as sea
61 surface temperature (SST), air-sea humidity gradients (RH_{sea}), and boundary-layer
62 stability (wind-speed), which together govern moisture flux at the ocean-atmosphere
63 interface⁴.

64 Reasonable proxies exist for SST^{5,6,7} but the other two drivers are recorded only
65 indirectly. Wind could be inferred from ice-core dust record but also reflects source
66 aridity and sea-ice extent⁸. Humidity is recorded by deuterium excess (*d-excess*)^{9,10} but
67 also a function of SST and wind¹¹⁻¹⁸. Rather than assembling the flux from its parts, a
68 direct proxy of evaporation would benefit the scientific community.

69 The scientific narrative of *d-excess* in Antarctic ice cores reflects a forty-year evolution,
70 from reconstructing past relative humidity (RH)⁹ to process-based understanding of
71 moisture-source dynamics (latitudinal moisture-source shifts driven by orbital obliquity)
72¹⁹⁻²². Assuming insignificant change in the RH over the moisture source¹⁸, *d-excess* was
73 also inverted for past SST, often returning values incompatible with marine constraints
74²²⁻²³. Overall, *d-excess* responds to SST, humidity, and wind^{11,17,18} together and focusing
75 on only one variable is undetermined.

76 In this study, we demonstrate that *d-excess* could act as a direct proxy for oceanic
77 evaporation rates. This interpretation leverages the underlying physics: *d-excess* reflects
78 kinetic fractionation during evaporation at the air-sea interface, a process fundamentally
79 governed by evaporative flux. Critically, this framework is most robust for Antarctic ice
80 core records. Unlike low-latitude or northern hemispheric regions where primary

d-excess signals are modified by continental recycling and sub-cloud re-evaporation^{24,25}, these processes are physically suppressed by the extreme cold along high-latitude transport pathways in the Antarctic region²⁶. Furthermore, we focus on the EPICA Dome C (EDC) record, where the *d-excess* signal has been demonstrated to be exceptionally coherent and straightforward, maintaining a robust correlation with supersaturation sensitive logarithmic definitions of *d-excess* ($R \approx 0.80$) and source temperature ($R \approx 0.93$) without the complex, non-linear biases found at other East Antarctic sites^{20,22,27}.

By calibrating *d-excess* against modern evaporation rates over Dome C's primary moisture source regions (the subtropical Indian Ocean sector) from reanalysis products and applying this relationship to the 800-kyr EDC ice core record, we reconstruct the first continuous record of past oceanic evaporation spanning multiple glacial cycles. This record reveals a regime-dependent reversal in evaporation control, thermal dominance under glacial conditions, dynamic (wind) dominance during warm and transitional climates.

Results

Comparison of empirically derived oceanic vapor *d-excess* with GCM output

The magnitude of the empirically derived oceanic vapor *d-excess* depends on the specific linear regression used to convert SST and/or RH_{sea} into *d-excess*. Several such relationships exist in the literature, all derived from concurrent *in-situ* ship measurements but covering different oceanic sectors (Table-S1). We adopt the Uemura et al.¹⁴ relationship, because its calibration domain (35°S–65°S Southern Indian Ocean) most closely matches the established moisture-source region for the EPICA Dome C ice core²⁸. Application of any other regional formula would yield a *d-excess* field offset by a regionally dependent constant but preserving the same spatial pattern.

The empirically derived *d-excess* field shows strong spatial agreement with the WISOMIP isotope-enabled GCM ensemble across the global ocean (60°S–60°N; $R^2 =$

0.89, slope = 1.12; Fig. S1), confirming that the empirical formula captures the global oceanic isotopic pattern at field-realism level. While this GCM comparison serves as a field realism check rather than a primary physical validation, given known challenges in simulating *d-excess*^{29,30}, the exceptional spatial coherence provides high confidence in the pattern of our reference *d-excess* field, establishing its credibility before any comparison with evaporation products.

Quantification and validation of the link between d-excess and oceanic evaporation

The spatial distribution of mean annual *d-excess* closely mirrors that of the OAFlux-derived mean annual oceanic evaporation rate³¹ (compare Fig. 1A and 1B). Over the global ocean (60°S–60°N), annual-mean *d-excess* shows a robust and highly significant correlation with evaporation rates ($R^2=0.90$; Fig. 1C and Fig. 1E), with the relationship strongest over the EDC moisture source region where *d-excess* explains ~95% of evaporation variance (Table-S2; Fig. 1F), providing the empirical basis for the derived transfer function ($E = 0.266d^2 + 2.155d + 17.40$; refer to methods). The quadratic coefficient A derived from the ERA5-OAFlux calibration (0.266) falls within the $\pm 1\sigma$ spread of the GCM Multi-Model Ensemble (0.207 ± 0.120 ; Fig. S3, Supplementary Note 1 and Table-S4). The relationship is not merely spatial: at each grid point, the seasonal cycles of *d-excess* and evaporation co-vary, with both the coupling strength ($R^2 > 0.5$) and the regression sensitivity highest and most consistent over the EDC moisture source region (Fig. S4). This confirms a local, temporally resolved physical link rather than a spatial covariation artifact.

To confirm that the relationship between empirically derived *d-excess* and OAFlux-derived oceanic evaporation is not a mathematical artifact of shared reanalysis input (as both are derived from the reanalysis products), we validate the empirically derived *d-excess* against two purely satellite-derived evaporation products, J-OFURO³² and HOAPS³³, created entirely from satellite sensors without atmospheric reanalysis input. In the Southern Indian Ocean sector, the primary moisture source for the EDC

core²⁸, the correlation between *d-excess* and evaporation rate remains remarkably stable across independent retrieval methodologies ($R^2 > 0.75$; Table-S2). HOAPS exhibits higher instability compared to OAFlux and J-OFURO3, with noise concentrated in humid tropical regimes where passive microwave retrievals encounter elevated uncertainty^{33, 34}. In contrast to the mid-latitude regions, the *d-excess*–evaporation correlation weakens significantly in the tropics, where high boundary-layer humidity and convective moisture mixing decouple the surface evaporation signal from *d-excess*. (Fig. 1D; Supplementary Note 3).

Using concurrent in-situ, ship-based measurements of water-vapor *d-excess* over the Southern Indian Ocean¹⁴, we find that observed *d-excess* explains 83% of the variance in evaporative potential (Δq) at day-to-week timescales ($R^2 = 0.83$, $p < 0.0001$; Fig. S2C), significantly outperforming SST ($R^2 = 0.54$; Fig. S2A) and relative humidity ($R^2 = 0.65$; Fig. S2B) as individual predictors. This provides direct observational confirmation that *d-excess* tracks the thermodynamic driver of evaporation, independent of the reanalysis products used in our spatial calibration.

Past Southern Ocean Evaporation Intensity and temperature relationship

The Southern Hemisphere *d-excess*-evaporation midlatitude transfer function has been used to produce the continuous reconstruction of past oceanic evaporation rates from the 800-kyr EDC ice-core record (Fig. 2A). Crucially, the evaporation change varies non-linearly with the source temperature (Fig. 2C). The system transitions from a tightly constrained, SST-dominant linear regime at $\Delta T_{\text{source}} \leq -3.5^\circ\text{C}$ ($R^2 = 0.57$) through an intermediate plateau ($R^2 = 0.10$), ultimately reaching a warm-state divergence at $\Delta T_{\text{source}} \geq -1.0^\circ\text{C}$ ($R^2 = 0.23$). This regime-dependent structure is insensitive to transfer function calibration choice, remaining consistent across 15 independent scenarios spanning three flux products and five source region configurations (Table-S3). To confirm that the regime-dependent coupling structure is not an artifact of these specific threshold values, we recomputed $R^2(\Delta T_{\text{source}}, \Delta \text{Evaporation})$ within each regime across numerous threshold combinations spanning warm (-0.5 to -2.0°C) and cold (-2.5 to -4.5°C) values.

164 Cold-regime R^2 (0.47–0.58) consistently exceeds Warm-regime R^2 (0.15–0.33) in every
165 tested combination, and the intermediate-regime R^2 remains low throughout
166 (0.006–0.29), confirming that the coupling hierarchy is threshold-independent (Fig. S5).
167 Even after resampling to a uniform 1-kyr time step to remove ice-accumulation sampling
168 bias, the ranking at canonical thresholds is preserved (Cold-regime R^2 : 0.59,
169 intermediate-regime R^2 = 0.20, Warm-regime R^2 =0.27).

170 The decoupling between source temperature and evaporation rate is most clearly visible
171 at interglacial-to-glacial transitions, where peak source temperature does not align with
172 peak evaporation. For example, source temperature peaked at 122.8 ka while
173 reconstructed evaporation continued rising thereafter, reaching an absolute maximum of
174 +52% at 104.2 ka — a lag of 18.7 kyr (Fig. 2B and Fig. S6). Such decoupling suggests
175 that a secondary, non-thermal driver must be influencing evaporation.

176 **Wind speed as an independent driver of d -excess**

177 We identify this secondary driver as wind-driven transport. Establishing it, however, first
178 requires addressing a circularity concern: the coupling between reconstructed evaporation
179 (E) and ΔT_{source} is, by construction, vulnerable because E is derived from d -excess via our
180 transfer function, and ΔT_{source} is also partly derived from d -excess²⁰, so any correlation
181 between them could in principle be inflated by their shared d -excess origin rather than
182 reflecting genuine physical coupling. To test this, we asked whether the 800-kyr d -excess
183 signal contains variance that is independent of source temperature, using an external
184 proxy with no d -excess input — the EDC dust-flux record⁸.

185 The raw correlation between d -excess and dust flux is weak ($R^2 = 0.004$). However, once
186 the contribution of source temperature (reconstructed following Landais et al.²⁰) to
187 d -excess is removed, the residual variance correlates significantly with dust flux ($R^2 =$
188 0.37; Fig. 3). This correlation remains highly significant after accounting for the strong
189 serial autocorrelation of both records: an effective-sample-size t -test reduces the degrees
190 of freedom from 5785 to $N_{\text{eff}} = 491$ yet still yields $p < 10^{-15}$, and a moving-block
191 bootstrap places the 95% confidence interval for r at [0.49, 0.69] (a phase-randomization
192 surrogate test gives the same conclusion; Methods). This externally corroborated

independent variance shows that d-excess carries a genuine signal beyond source temperature — and therefore that the $E-\Delta T_{\text{source}}$ coupling reflects real physical structure, not their shared d-excess origin.

Thermodynamic-to-dynamic transition in evaporation control

Across the full 800-kyr record, moisture source temperature and dust flux together explain 81.0% of total evaporation variance ($R^2 = 0.81$), compared to 55.5% for temperature alone (Fig. S8). The contribution of each driver however, varies fundamentally across climatic regimes (Fig. 2E). Under cold glacial conditions ($\Delta T_{\text{source}} \leq -3.5^\circ\text{C}$), source temperature alone accounts for 57% of evaporation variance, and adding dust flux yields the smallest incremental gain of the three regimes (+26%), Fig. 2F). Under warmer interglacial conditions ($\Delta T_{\text{source}} \geq -1.0^\circ\text{C}$), dust flux becomes the dominant predictor; the combined model explains 68.7% of warm-regime variance, with dust contributing an incremental 46% beyond temperature alone. The dynamical channel is most informative precisely where the thermodynamic channel fails: in the transitional regime ($-3.5^\circ\text{C} < \Delta T_{\text{source}} < -1.0^\circ\text{C}$), where temperature alone explains only 10% of evaporation variance, adding dust flux raises the explained variance by 63%. Together, source temperature and dust flux explain approximately 83%, 73%, and 69% of evaporation variance in cold, transitional, and warm regimes, respectively (Fig. 2F). Four independent statistical frameworks, multiple linear regression, partial correlation analysis, variance inflation factors, and dominance analysis, confirm this regime-dependent reversal across all tests (Fig. S7). The temporal alignment of these signals is visible during the 123–104 ka window, when rising dust flux peaks at 103.3 ka, synchronous within chronological uncertainty with the reconstructed evaporation maximum (Fig. 2B), and the residual evaporation time series ($\varepsilon = \Delta E\% - f(\Delta T_{\text{source}})$) tracks the dust record across the full record.

218 Discussion

219 Reframing *d-excess* as a physical proxy for evaporation

220 The bulk aerodynamic equation expresses evaporation as the product of wind speed and
221 the air–sea humidity gradient $(q_s - q_a)$ ^{4, 36} where q_s is the saturation specific humidity at
222 the sea surface temperature and q_a is the actual specific humidity of the near-surface air.
223 The Craig–Gordon expression governs *d-excess* through the normalised humidity deficit
224 $(1 - h)$, where $h = q_a/q_s$ is the relative humidity at the evaporation site^{37,38,39}. An increase
225 in this deficit raises both evaporation and the *d-excess* of the evaporated vapour, so
226 *d-excess* carries the signature of the thermodynamic side of evaporation. But the flux also
227 depends on wind speed, which this thermodynamic view leaves out. The *d-excess* can
228 therefore track the full evaporative flux only if wind, too, leaves an imprint on it. The
229 classical Merlivat–Jouzel framework suggests that *d-excess* decreases as wind speed
230 rises, as turbulent mixing replaces molecular diffusion in the transport of the
231 isotopologues^{15,40}. If this direct pathway dominated, stronger winds would drive *d-excess*
232 and evaporation apart rather than together. However, recent marine measurements find
233 this direct dependence to be weak or absent under realistic open-ocean conditions¹⁷.

234 Once the temperature signal is removed, the residual *d-excess* reflects variations in
235 relative humidity, which over glacial cycles are driven by wind. The dust record carries a
236 signature of source aridity, surface wind speed, and atmospheric circulation; the factor
237 common to both is wind. Their strong positive correlation therefore isolates this shared
238 wind signal and shows that *d-excess* rises with wind (Fig. 3). We suggest that *d-excess*
239 increases with wind speed when wind replaces moist surface air with drier air, widening
240 the humidity deficit $(1 - h)$. Over the mid-latitude Southern Ocean this occurs through
241 the cyclones of the westerly storm track: behind their cold fronts, cold dry air sweeps
242 equatorward over warmer water, drying the marine boundary layer and steepening the
243 air–sea humidity contrast^{41–44}. The widened deficit simultaneously enhances the kinetic
244 fractionation that generates *d-excess* and the bulk evaporative flux, so the two rise

245 together. Because the same winds mobilise and transport the dust, the dust flux indexes
246 this ventilation, the physical basis for the residual–dust correlation.

247 Our finding that *d-excess* rises with wind is not in tension with observations that *d-excess*
248 shows no direct wind dependence¹⁷: wind acts not by modulating the kinetic
249 fractionation, but by lowering the relative humidity to which *d-excess* responds.
250 Similarly, the empirically derived *d-excess* field carries no explicit wind-speed term (Fig
251 1); it still bears wind's imprint, because wind sets the relative humidity that *d-excess*
252 records. The *d-excess* can therefore be read as the isotopic shadow of the evaporative
253 flux, more faithful to its physics than a reading cast in sea surface temperature or relative
254 humidity (Fig. S2, Fig. 2D and Fig. 3B). Most importantly, evaporation is the variable
255 that matters and that can be tested: it governs moisture supply and latent heat, and can be
256 checked against bulk aerodynamic theory, model moisture budgets, and modern
257 observations, benchmarks unavailable to a source-condition reading of the same signal.

258 **Regime-dependent coupling between thermal and dynamical drivers**

259 Reading *d-excess* as the evaporative flux lets us ask not just what drives evaporation, but
260 which driver dominates, and when (Fig 2D-F). Because the flux responds to both
261 temperature, through the deficit, and wind, their balance need not be fixed, and across the
262 past 800,000 years it reverses between glacial and interglacial states.

263 During cold periods, low *d-excess* and reduced evaporation could reflect lower SST,
264 higher RH, or weaker winds. But the high dust flux of glacial periods signals vigorous
265 transport, ruling out weak winds as the primary cause. Vigorous winds with suppressed
266 evaporation point instead to a near-saturated boundary layer. When relative humidity is
267 near saturation⁹, the gradient $(1 - h)$ is insensitive to wind-driven advection (Fig. 2F). It
268 is set instead by temperature: a small rise in air temperature lifts the saturation vapour
269 pressure and opens the gradient. Source temperature therefore dominates through this
270 saturation-vapour-pressure mechanism, and evaporation tracks it with $R^2 = 0.57$.

271 The temperature–evaporation R^2 reaches its minimum, 0.10, in the intermediate regime
272 ($-3.5\text{ °C} < \Delta T_{\text{source}} < -1.0\text{ °C}$). It falls below both the cold and warm values, and the
273 incremental gain from adding dust peaks here ($+0.63\text{ }R^2$; Fig. 2F). The clearest
274 expression of this handoff is the lag between peak source temperature (122.8 ka) and
275 peak evaporation (104.2 ka). Such offsets appear at most interglacial-to-glacial transitions
276 (Fig. 2A). During the warming-to-cooling phase, source temperature declines while the
277 equator-to-pole temperature gradient steepens⁴⁶. This steepening invigorates atmospheric
278 transport, recorded in the rising dust flux⁴⁵, so dry-air advection over the source region
279 strengthens even as the source itself cools. Because evaporation here is wind-limited, it
280 keeps rising after the thermal peak.

281 During warm periods, the regulation of the gradient is fundamentally different. Although
282 dust flux is lower, wind exerts stronger control on evaporation than temperature does, by
283 acting as the rate-limiting step (Fig. 2F). What increases is the sensitivity of evaporation
284 to wind, not the magnitude of the wind itself. The warm atmosphere holds more moisture,
285 so the boundary layer saturates readily. Evaporation is then limited by how fast wind
286 advects this vapour away and re-establishes the gradient. Source temperature here nears
287 its saturation-vapour-pressure ceiling, where extra warming yields diminishing returns.
288 Wind-driven transport of unsaturated air becomes rate-limiting, and the coupling weakens
289 to $R^2 = 0.23$. This is a dynamically modulated regime, where evaporation can intensify
290 while source temperature stagnates. It is consistent with observations of stalled or
291 declining ocean evaporation despite rising SST³.

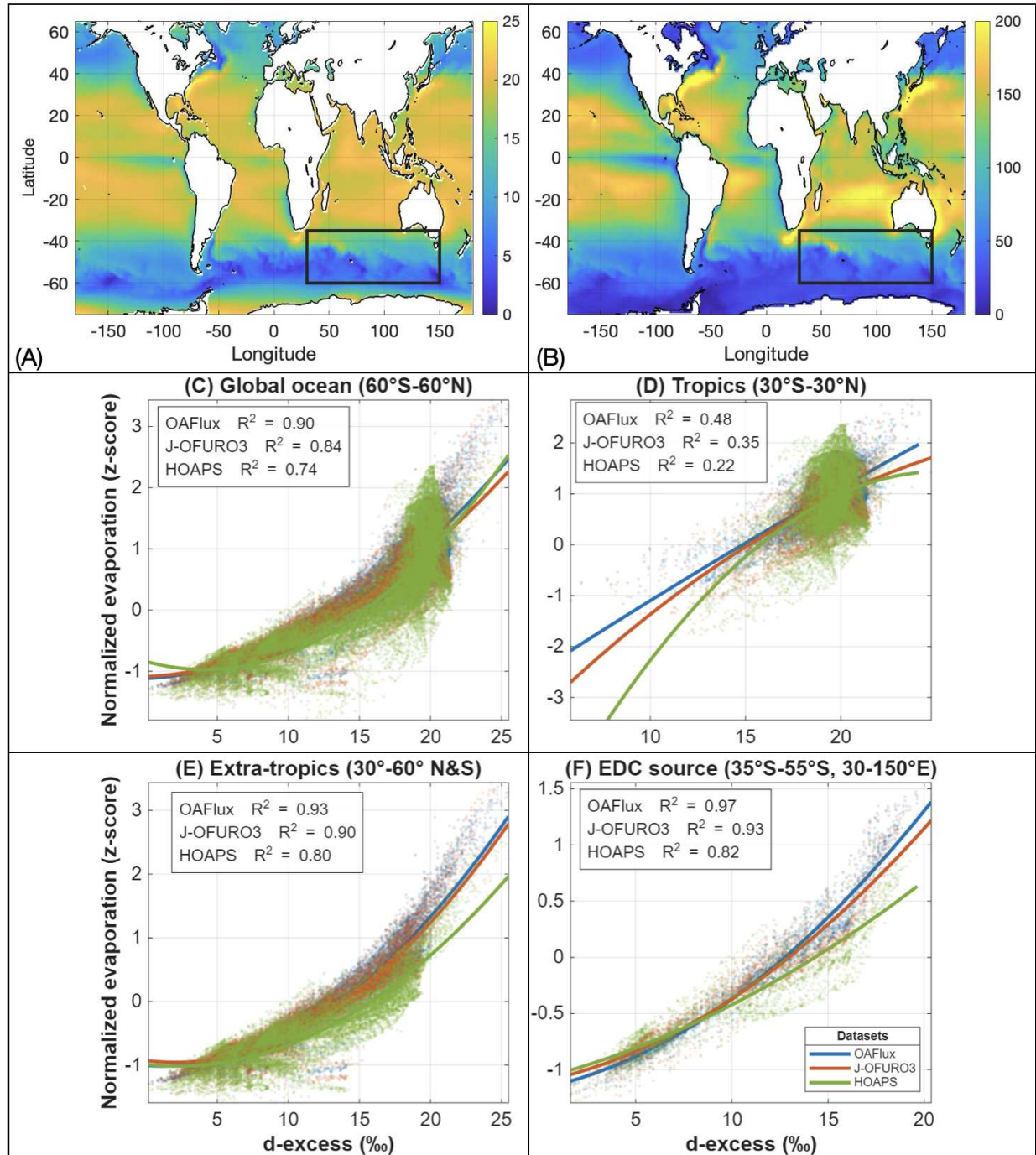
292 Limitations

293 Because E and ΔT_{source} derive from the same *d-excess* record, we rely on the external
294 dust-flux proxy⁸ to isolate a temperature-independent wind signal; this constrains but
295 does not fully remove their shared origin, and dust itself integrates source aridity and
296 sea-ice extent alongside wind, so the dynamic driver is corroborated rather than directly
297 measured. The regime-dependent structure of our results is robust across calibration
298 scenarios, statistical frameworks, and resampling, whereas the absolute amplitude carries

an uncertainty of $\sim\pm 30\%$ (Supplementary Note 1) that scales magnitude without affecting timing or the driver hierarchy. The framework is calibrated on modern open-ocean evaporation and is least secure at glacial maxima, where sea-ice-sourced sublimated vapour^{17,47,56} and enhanced supersaturation⁴⁸ can perturb the *d-excess*–evaporation link and EDC may approach the behavior of Dome F²²; glacial minima are therefore lower-confidence than interglacial and transitional values^{49,50}.

Conclusion

Deuterium excess, long read as an ambiguous tracer of source conditions, emerges here as a physically grounded, quantitative measure of oceanic evaporation. By calibrating it against modern evaporation and applying it to the EPICA Dome C record, we reconstruct evaporation at an Antarctic moisture source continuously across the past 800,000 years, the first such record of the first step of the global water cycle. The control on evaporation is not fixed but shifts with climate state: thermodynamic, set by source temperature, under cold glacial conditions, and dynamic, set by wind-driven ventilation of the marine boundary layer, during warm and transitional climates. Past oceanic evaporation therefore cannot be inferred from sea surface temperature or the humidity gradient alone. Atmospheric circulation is a first-order control on the strength of the hydrological cycle, and it dominates in precisely the warm states most relevant to a warming climate. Cast in physical units, the reconstruction can now be tested directly against the theory and models of the ocean-atmosphere water cycle.

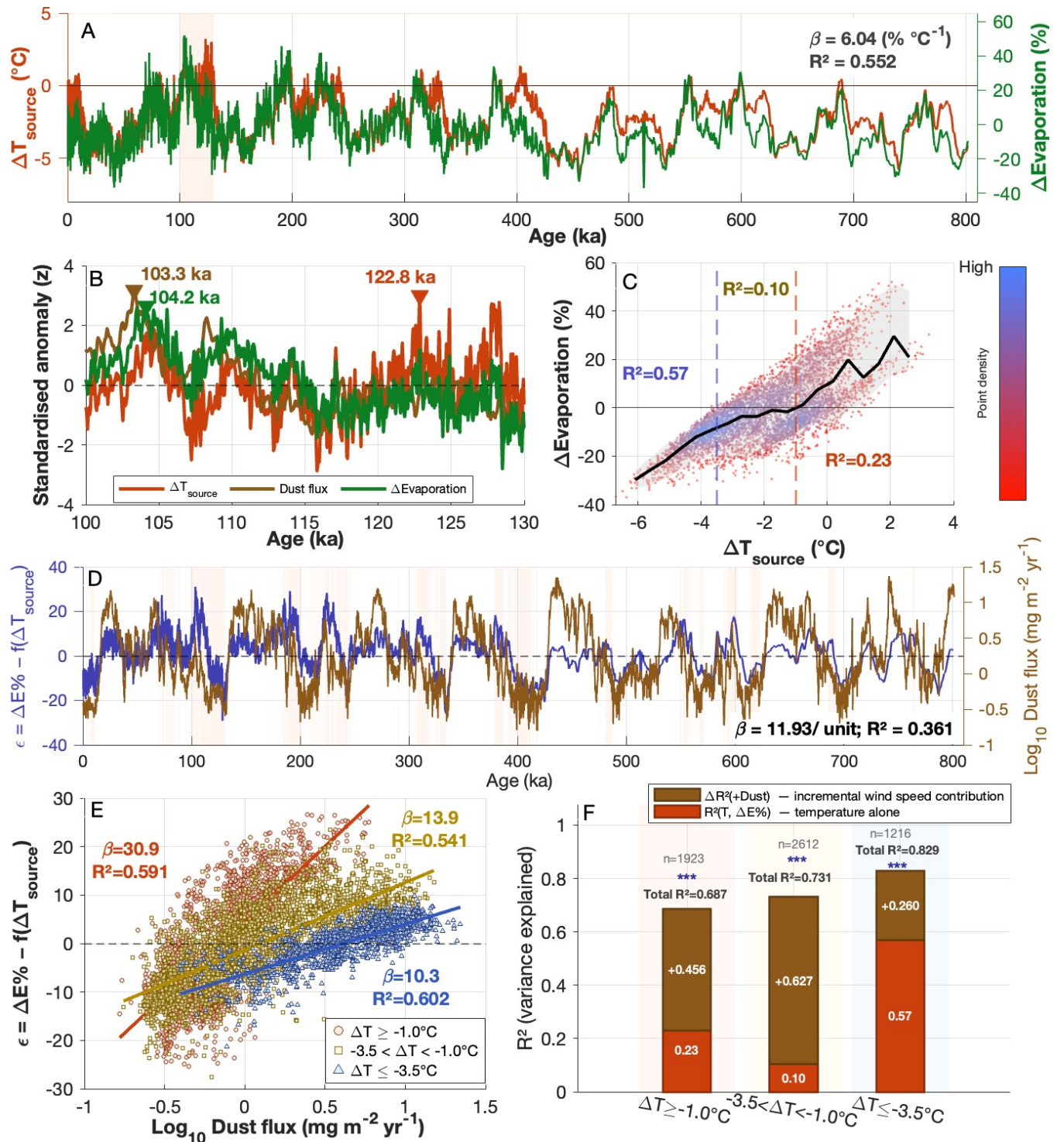


320

Fig. 1: Spatial and statistical validation of water vapor *d-excess* as a proxy for oceanic evaporation. (A) Reference water vapor *d-excess* field (‰) calculated using the Southern Indian Ocean-specific empirical relationship of Uemura et al. (2008) applied to ERA5 reanalysis data. (B) Climatological annual mean oceanic evaporation rate (cm/yr) derived from the OAFlux product (1980–2020). Scatter plots illustrating the correlation between water vapor *d-excess* and normalized oceanic evaporation rates across four

327 spatial domains: **(C)** global ocean (60N–60S), **(D)** tropics (30N–30S), **(E)** combined
328 extra-tropics (30–60 N&S), and **(F)** the integrated EDC moisture source region (35S–55S,
329 30E–150E). To ensure cross-comparability between independent retrieval methodologies,
330 evaporation data from OAFlux (hybrid), J-OFURO3 (satellite-LHF), and HOAPS
331 (satellite-EVP) were Z-score normalized. Solid curves denote second-order polynomial
332 regressions. Inset boxes provide the coefficient of determination (R^2).

333



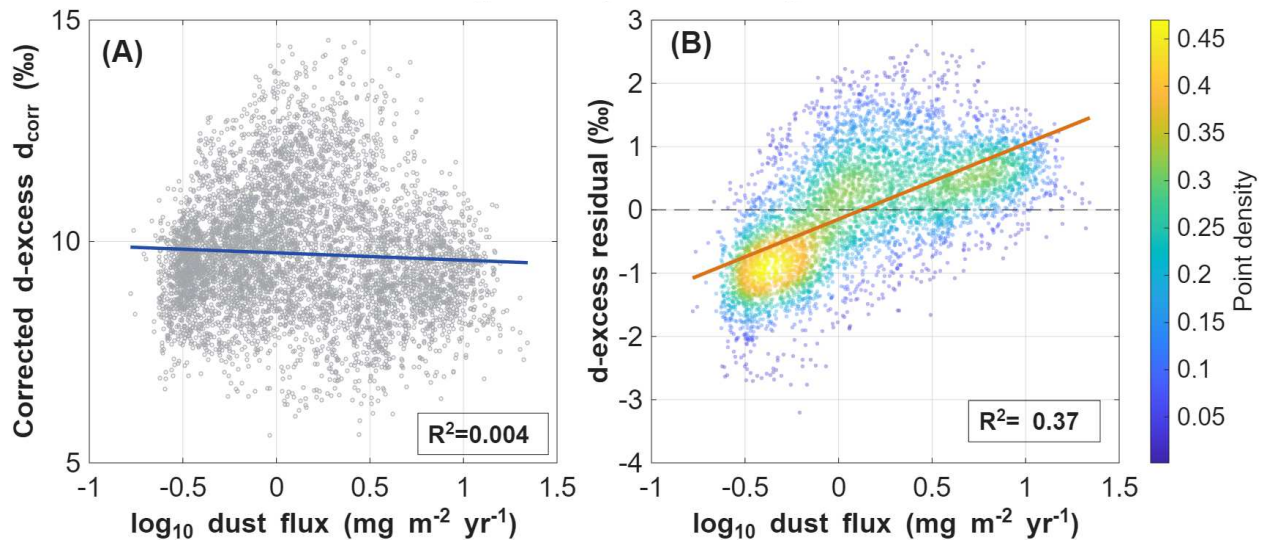
334

335 **Fig 2. Temporal variation in Evaporation intensity across 800,000 years and its**
 336 **relationship with temperature and wind-speed. (A) Reconstructed evaporation intensity**
 337 **($\Delta\text{E}\%$, green) and source temperature anomaly (ΔT_{source} , orange) from EPICA Dome C.**
 338 **Dashed lines mark thermal regime boundaries (-1.0°C and -3.5°C). Pink shading**

denotes the MIS 5e inception window. **(B)** Standardised z-scores of ΔT_{source} (orange), dust flux (brown; Lambert et al.⁸), and $\Delta \text{Evaporation}$ (green) over 100–130 ka. Peak ages are labelled; purple arrow marks the 18.7 kyr lag between thermal and hydrological maxima. **(C)** Phase space of ΔT_{source} versus $\Delta \text{Evaporation}$ coloured by point density. Black line = median evaporation trend; grey envelope = 90% confidence spread. R^2 values show temperature-evaporation coupling strength per regime. **(D)** Temperature-independent evaporation residuals ($\varepsilon = \Delta E\% - f(\Delta T_{\text{source}})$, blue) and log dust flux (brown) across 800 kyr. Pink shading marks warm intervals ($\Delta T_{\text{source}} \geq -1.0^\circ\text{C}$). **(E)** Dust flux versus the temperature-independent evaporation residual ε by thermal regime; annotated R^2 is the fraction of *residual* variance explained by dust and β its sensitivity (slope). **(F)** Partition of *total* evaporation variance within each regime into temperature alone ($R^2(T, \Delta E\%)$) and the incremental variance added by dust (ΔR^2). All dust-residual correlations significant at $p < 0.001$.

352

353



354

Fig. 3. Unmasking of the 800-kyr d-excess wind signal. (A) Raw corrected d-excess (d_{corr}) shows essentially no relationship with $\log_{10}$ dust flux ($R^2 = 0.004$); large-scale source-temperature variation masks the mechanical signal in the raw record. (B) After the source-temperature component is removed, the d-excess residual (coloured by point density) correlates with $\log_{10}$ dust flux ($R^2 = 0.37$), isolating the wind-driven signal carried by d-excess. Reading through the transfer function, this same

temperature-independent signal corresponds to the evaporation residual whose regime-dependent coupling to dust is shown in Fig. 2E.

Methods

Modern Gridded Datasets and *d-excess* Estimation

The *d-excess* value of oceanic surface vapor is estimated using the empirical relationship of Uemura et al.¹⁴ ($d-excess = -0.42 \times RH_{sea} + 0.45 \times T_{SST} + 37.9$). The T_{SST} is directly retrieved from ERA5, and RH_{sea} has been calculated using the Magnus formula as follows,

$$RH_{sea} = RH_{air} \times \exp \left[17.625 \left(\frac{T_{air}}{243.04 + T_{air}} - \frac{T_{SST}}{243.04 + T_{SST}} \right) \right] \times \frac{P - 0.232 \times \exp \left(\frac{17.625 \times T_{SST}}{243.04 + T_{SST}} \right)}{P - 0.232 \times \exp \left(\frac{17.625 \times T_{air}}{243.04 + T_{air}} \right)}$$

where RH_{air} , T_{2m} , T_{SST} , and P are 2-meter relative-humidity, 2-meter air temperature, sea surface temperature, and surface atmospheric pressure, respectively. All atmospheric variables were obtained from ERA5 at monthly resolution⁵¹ and annually averaged to produce annual-mean *d-excess* fields. We applied the Uemura et al.¹⁴ relationship as it is calibrated specifically to the 35°S–60°S Southern Indian Ocean, the primary moisture source for EPICA Dome C ice core sites. While alternative regional calibrations exist, all are linear functions of SST and RH with similar physical sensitivities, differing only in absolute scaling (Table S1). As a supplementary analysis, we compared ERA5-derived *d-excess* fields against outputs from seven isotope-enabled general circulation models (GCMs) from the Water Isotope Model Intercomparison Project (WISOMIP)^{29,30}.

Validation against Satellite-Derived Evaporation Products

The link between the *d-excess* value of the ocean vapor and the oceanic evaporation rate has been studied using flux data from the OAFlux³¹ (Objectively Analyzed air-sea

Fluxes; monthly evaporation estimates derived from optimal synthesis of satellite observations and atmospheric reanalysis), HOAPS³³ (Hamburg Ocean Atmosphere Parameters and Fluxes from Satellite Data; evaporation estimates from passive microwave observations) and J-OFURO3³² (Japanese Ocean Flux Data Sets with Remote Sensing Observations; monthly latent heat flux estimates derived purely from satellite sensors without reanalysis input). Critically, J-OFURO3 and HOAPS are derived purely from satellite observations without atmospheric reanalysis input unlike OAFlux, providing independent validation of the *d-excess*-evaporation relationship. All datasets were regridded to 1°×1° grid using bilinear interpolation before analysis.

Physical Parameterization and Statistical Framework

To establish a non-circular, empirical link between *d-excess* and the moisture-source evaporation potential, we conducted a validation study using an independent observational dataset from the Uemura et al.,¹⁴. This region corresponds to the primary moisture source for the EPICA Dome C (EDC) ice core. The dataset provides concurrent in-situ measurements of T_{SST} , T_{2m} , RH_{sea} , and atmospheric pressure, alongside laboratory-measured vapor *d-excess* values. We parameterized the thermodynamic "speed" of evaporation as the specific humidity gradient (Δq) between the sea surface and the overlying atmosphere. This calculation followed a bulk aerodynamic formulation. Saturation Vapor Pressure (e_s): Computed using the Magnus-Tetens formula⁵²:

$$e_{s(T)} = 6.112 \exp[17.67 \times T / (T + 243.5)]$$

where T is temperature in °C. This was determined for both the sea surface $e_s(T_{SST})$ and the atmosphere $e_s(T_{air})$. The vapor pressures were converted to specific humidity (g/Kg) at ocean surface (q_s) and atmosphere (q_a),

$$q_a = \frac{0.622 \times e_{s(T_{air})} \times RH/100}{P - (0.378 \times e_{s(T_{air})} \times RH/100)} \times 1000$$

408

$$q_s = \frac{0.622 \times 0.98 \times e_{s,(T_{SST})}}{P - (0.378 \times 0.98 \times e_{s,(T_{SST})})} \times 1000$$

409 To account for the reduction in vapor pressure over seawater, saturation values at the
 410 surface were adjusted by a salinity factor of 0.98. The evaporative potential,
 411 thermodynamic gradient was defined as $\Delta q = q_s - q_a$. To evaluate the predictive power of
 412 various proxies, we compared the variance explained (R^2) by Δq , SST, and RH. Statistical
 413 superiority was tested using Steiger's Z-test.

414 Derivation of the Evaporation Transfer Function

415

416 To reconstruct past hydrological intensity, we established an empirical transfer function
 417 between *d-excess* and oceanic evaporation from OAFLux data. Based on our modern
 418 calibration in the Southern Indian Ocean, we utilized a second-order polynomial
 419 regression:

420

$$E = 0.266 \times d^2 + 2.155 \times d + 17.40$$

421 where E represents the absolute evaporation rate and d is the corrected *d-excess*. To
 422 isolate the relative change in the mechanical intensity of the evaporative engine, all
 423 reconstructed values were converted to percentage anomalies relative to the modern (0
 424 ka) baseline:

425

$$\Delta E (\%) = \left(\frac{E - E_{modern}}{E_{modern}} \right) \times 100$$

426 E_{modern} is computed by applying the same transfer function to the modern (0 ka) *d-excess*
 427 value retrieved from the top of the EDC record.

428 While the primary transfer function was derived using the empirical relationship between
 429 in-situ *d-excess* and the OAFLux product, we performed a comprehensive sensitivity
 430 analysis to ensure the robustness of our reconstruction. This involved cross-checking the
 431 *d-excess*–evaporation relationship against two independent satellite-only products
 432 (J-OFURO3 and HOAPS) across five distinct moisture source sectors (Table S2).

EDC Ice Core Processing and Isotopic Corrections and moisture source temperature estimation

We applied *d-excess*-evaporation calibration to reconstruct past oceanic evaporation rates from the EPICA Dome C (EDC) ice core. The EDC isotopic records span the past 800,000 years on the AICC2012 chronology⁵³ and provide exceptional temporal resolution and analytical precision^{23,54}. Following Landais et al.²⁰, we applied a correction for past changes in mean ocean isotopic composition to derive corrected *d-excess* values. We adopted the 500-year smoothing window for our primary analysis, as this timescale effectively preserves orbital-to-millennial signals while filtering sub-millennial noise. We intentionally utilized the corrected *d-excess* signal without additional site-temperature compensation to avoid the mathematical circularity inherent in traditional inversion methods; a detailed physical justification for this approach and the preservation of the kinetic source signal during transport is provided in Supplementary Note 2.

Source-temperature anomalies (ΔT_{source}) were reconstructed from the corrected isotopic records following the co-isotopic inversion of Landais et al.²⁰. Raw δD and $\delta^{18}O$ from EDC were placed on the AICC2012 chronology and corrected for changes in mean ocean isotopic composition, and corrected *d-excess* was computed as $d_{\text{corr}} = \delta D_{\text{corr}} - 8 \cdot \delta^{18}O_{\text{corr}}$. Both records were expressed as anomalies relative to the modern (0 ka) value at the top of the core ($\Delta \delta D_{\text{corr}}$, Δd_{corr}), and source temperature was obtained as a linear combination of the two, $\Delta T_{\text{source}} = 0.06 \cdot \Delta \delta D_{\text{corr}} + 0.93 \cdot \Delta d_{\text{corr}}$, with coefficients taken from the isotopic-model inversion of Landais et al.²⁰.

To characterize evaporation changes across different climate regimes, we classified the 800-kyr record into three climate states based on moisture source temperature anomaly:

- Warm regime : $\Delta T_{\text{source}} \geq -1.0^{\circ}\text{C}$ (modern-like conditions)
- Intermediate: $-3.5^{\circ}\text{C} < \Delta T_{\text{source}} < -1.0^{\circ}\text{C}$ (Transition)
- Cold Regime : $\Delta T_{\text{source}} \leq -3.5^{\circ}\text{C}$ (full glacial conditions)

460 The result of the study is insensitive to the choice of the threshold to define the regimes
461 (Fig.S5).

462 Because ΔT_{source} is thus derived predominantly from *d-excess*, the same signal used to
463 reconstruct evaporation, the resulting $E-\Delta T_{\text{source}}$ coupling is tested against the external
464 dust-flux record to exclude a shared-origin artefact.

465 **Temperature-independent residuals**

466 To isolate the component of each record not explained by source temperature, we
467 computed two residual series. For the isotopic analysis (Fig. 3), corrected *d-excess*
468 (dcorr) was regressed linearly on ΔT_{source} and the *d-excess* residual taken as dcorr minus
469 this fit. For the evaporation analysis (Fig. 2D, E), the same procedure was applied to the
470 reconstruction, yielding $\varepsilon = \Delta E\% - f(\Delta T_{\text{source}})$, where f is the least-squares linear fit of
471 $\Delta E\%$ to ΔT_{source} . Because ΔT_{source} is itself derived from the isotopic data, these residuals
472 retain only the variance orthogonal to the source-temperature signal; their correlation
473 with the independent EDC dust-flux record therefore provides an external, non-circular
474 test for a temperature-independent (wind) control.

475 **Significance testing under autocorrelation**

476 Because both the *d-excess* residual and the dust-flux record are strongly autocorrelated,
477 the nominal sample size ($N = 5785$) overstates the independent information available and
478 would yield a spuriously small p-value if used directly. We therefore replaced N with an
479 effective sample size following Bretherton et al.⁵⁷. $N_{\text{eff}} = N(1 - a_x a_y)/(1 + a_x a_y)$,
480 where a_x and a_y are the lag-1 autocorrelation coefficients of the two series. The
481 correction discounts the sample size by the degree to which the two records jointly
482 persist: when either series is uncorrelated in time the factor approaches unity and $N_{\text{eff}} \approx$
483 N , whereas when both are highly persistent it approaches zero and N_{eff} falls well below
484 N , because spurious agreement arises only when runs of similar values coincide in both
485 series. For our records ($a_x a_y \approx 0.84$) this reduced the effective sample size to $N_{\text{eff}} =$
486 491, and the two-sided t-test on the correlation, evaluated at $N_{\text{eff}} - 2$ degrees of freedom,

487 remained significant at $p < 10^{-15}$. Because this estimate assumes a first-order
488 autoregressive (red-noise) structure summarised by the lag-1 coefficient alone, we
489 additionally confirmed the result with a moving-block bootstrap and an Ebisuzaki
490 phase-randomization test⁵⁸, which preserve the full autocorrelation and power spectrum,
491 respectively.

492 Attribution of evaporation variance to thermal and dynamical drivers

493 To quantify the relative contributions of source temperature and wind-driven transport to
494 reconstructed evaporation, we fitted nested multiple linear regression models with $\Delta E\%$
495 as the response and two predictors: ΔT_{source} and the base-10 logarithm of EDC dust flux
496 (Lambert et al.⁸), the latter Savitzky–Golay-smoothed and interpolated onto the EDC
497 isotope timescale. A temperature-only model ($\Delta E\% = \beta_0 + \beta_1 \Delta T_{\text{source}}$) was compared
498 with a combined model ($\Delta E\% = \beta_0 + \beta_1 \Delta T_{\text{source}} + \beta_2 \log_{10}[\text{Dust}]$); the incremental
499 variance explained (ΔR^2) and its significance were assessed with a partial F-test. Both
500 models were fitted across the full 800-kyr record and independently within each thermal
501 regime (Fig. 2E, F).

502 To confirm that the inferred driver hierarchy was not an artefact of any single approach,
503 we examined it with four complementary diagnostics (Fig. S7). First, standardised
504 regression coefficients, computed on z-scored predictors and response, placed the
505 temperature and dust effects on a common scale for direct comparison of effect size.
506 Second, partial correlations isolated each predictor's association with evaporation while
507 controlling for the other — $r(\text{Dust}, \Delta E\% \mid \Delta T_{\text{source}})$ and $r(\Delta T_{\text{source}}, \Delta E\% \mid \text{Dust})$ — with
508 significance from a two-sided t-test. Third, variance inflation factors, $\text{VIF} = 1/(1 -$
509 $r^2(\Delta T_{\text{source}}, \text{Dust}))$, tested for collinearity between the two predictors. Fourth, dominance
510 analysis ranked the predictors by their average marginal contribution to R^2 across all
511 possible model subsets. Model adequacy was verified from the residual distributions and
512 the residual-versus-predicted structure (Fig. S8).

513 Use of large language models

514 During the preparation of this work, the authors used a large language model (Claude,
515 Anthropic) to assist with drafting and debugging the MATLAB analysis and
516 figure-generation code, and with editing and improving the clarity of the manuscript text.
517 All study design, data analysis, interpretation, and conclusions are the authors' own. The
518 authors reviewed and verified all model-assisted outputs and took full responsibility for
519 the content of the manuscript.

520 Data Availability

521 The 800-kyr EPICA Dome C ice core isotopic records and the AICC2012 chronology
522 used for this reconstruction were obtained from the National Oceanic and Atmospheric
523 Administration (NOAA) Paleoclimatology Database, with mean ocean isotopic
524 corrections applied following Landais et al.²⁰. The EPICA Dome C dust-flux record,
525 used here as the external wind proxy, is that of Lambert et al.⁸ and is available from the
526 PANGAEA database (<https://doi.pangaea.de/10.1594/PANGAEA.778213>). Modern
527 d-excess fields were derived from ERA5 monthly reanalysis — sea-surface temperature,
528 2 m air temperature and 2 m relative humidity — distributed by the Copernicus Climate
529 Data Store (<https://cds.climate.copernicus.eu>). Oceanic evaporation products were
530 obtained from OAFlux (<https://oaflux.whoi.edu/data-access/>), J-OFURO3³² and
531 HOAPS³³. The concurrent in-situ, ship-based measurements of water-vapour d-excess
532 used for the transfer-function calibration and observational validation are tabulated in
533 Uemura et al.¹⁴. Isotope-enabled simulations from the WISOMIP general circulation
534 models^{29,30} were obtained from the NASA GISS data portal
535 (<https://portal.nccs.nasa.gov/datashare/giss-publish/pub/paleoclimate/wisomip/>).

536 Code Availability

537 The MATLAB code used to process the records, perform the calibration, reconstruct past
538 evaporation, and generate all figures is openly available at Zenodo
539 (<https://10.5281/zenodo.20729400>), under the [CC-BY-4.0 / MIT] License.

540 Funding statement.

541 The authors declare no relevant funding for this work.

542 Acknowledgements

543 We acknowledge the EPICA community for the EPICA Dome C ice-core records, and the
544 providers of the ERA5 (Copernicus Climate Data Store), OAFlux, J-OFURO3, HOAPS,
545 and WISOMIP datasets used in this study, all of which are publicly available.

546 References:

- 547 1. Schmitt, R. W. The ocean's role in climate. *Oceanography* **31**, 32–40 (2018).
- 548 2. Findell, K. *et al.* *Rising Temperatures Increase Importance of Oceanic Evaporation*
549 *as a Source for Continental Precipitation*.
550 <https://doi.org/10.5194/egusphere-egu2020-10401> (2020).
- 551 3. Ma, N., Zhang, Y. & Yang, Y. Recent decline in global ocean evaporation due to
552 wind stilling. *Geophysical Research Letters* **52**, (2025).
- 553 4. Fairall, C. W., Bradley, E. F., Hare, J. E., Grachev, A. A. & Edson, J. B. Bulk
554 Parameterization of Air–Sea Fluxes: Updates and Verification for the COARE Algorithm.
555 *Journal of Climate* **16**, 571–591 (2003).
- 556 5. Elderfield, H. & Ganssen, G. Past temperature and $\delta^{18}\text{O}$ of surface ocean waters
557 inferred from foraminiferal Mg/Ca ratios. *Nature* **405**, 442–445 (2000).
- 558 6. Ho, S. L. *et al.* Sea surface temperature variability in the Pacific sector of the
559 Southern Ocean over the past 700 kyr. *Paleoceanography* **27**, (2012).
- 560 7. Chandler, D. & Langebroek, P. Southern Ocean sea surface temperature synthesis:
561 Part 2. Penultimate glacial and last interglacial. *Quaternary Science Reviews* **271**, 107190
562 (2021).
- 563 8. Lambert, F., Bigler, M., Steffensen, J. P., Hutterli, M. & Fischer, H. Centennial
564 mineral dust variability in high-resolution ice core data from Dome C, Antarctica.
565 *Climate of the Past* **8**, 609–623 (2012).
- 566 9. Jouzel, J., Merlivat, L. & Lorius, C. Deuterium excess in an East Antarctic ice
567 core suggests higher relative humidity at the oceanic surface during the last glacial
568 maximum. *Nature* **299**, 688–691 (1982).
- 569 10. Pfahl, S. & Sodemann, H. What controls deuterium excess in global precipitation?
570 *Climate of the Past* **10**, 771–781 (2014).
- 571 11. Benetti, M. *et al.* Deuterium excess in marine water vapor: Dependency on

relative humidity and surface wind speed during evaporation. *Journal of Geophysical Research: Atmospheres* **119**, 584–593 (2014).

12. Galewsky, J. & Rybecky, M. Environmental controls on water vapor deuterium excess in the coastal boundary layer: An information theory perspective. *Journal of Geophysical Research: Atmospheres* **130**, (2025).

13. Pfahl, S. & Wernli, H. Air parcel trajectory analysis of stable isotopes in water vapor in the eastern Mediterranean. *Journal of Geophysical Research: Atmospheres* **113**, (2008).

14. Uemura, R., Matsui, Y., Yoshimura, K., Motoyama, H. & Yoshida, N. Evidence of deuterium excess in water vapor as an indicator of ocean surface conditions. *Journal of Geophysical Research: Atmospheres* **113**, (2008).

15. Merlivat, L. & Jouzel, J. Global climatic interpretation of the deuterium-oxygen relationship for precipitation. *Journal of Geophysical Research: Oceans* **84**, 5029–5033 (1979).

16. Thurnherr, I. *et al.* Meridional and vertical variations of the water vapour isotopic composition in the marine boundary layer over the Atlantic and Southern Ocean. *Atmospheric Chemistry and Physics* **20**, 5811–5835 (2020).

17. Bonne, J.-L. *et al.* Resolving the controls of water vapour isotopes in the Atlantic sector. *Nature Communications* **10**, (2019).

18. Werner, M. *et al.* Glacial–interglacial changes in H₂¹⁸O, HDO and deuterium excess – results from the fully coupled ECHAM5/MPI-OM Earth system model. *Geoscientific Model Development* **9**, 647–670 (2016).

19. Vimeux, F., Masson, V., Jouzel, J., Stievenard, M. & Petit, J. R. Glacial–interglacial changes in ocean surface conditions in the Southern Hemisphere. *Nature* **398**, 410–413 (1999).

20. Landais, A. *et al.* Interglacial Antarctic–Southern Ocean climate decoupling due to moisture source area shifts. *Nature Geoscience* **14**, 918–923 (2021).

21. Vimeux, F. *et al.* A 420,000 year deuterium excess record from East Antarctica: Information on past changes in the origin of precipitation at Vostok. *Journal of Geophysical Research: Atmospheres* **106**, 31863–31873 (2001).

22. Uemura, R. *et al.* Ranges of moisture-source temperature estimated from Antarctic ice cores stable isotope records over glacial–interglacial cycles. *Climate of the Past* **8**, 1109–1125 (2012).

23. Stenni, B. *et al.* The deuterium excess records of EPICA Dome C and Dronning Maud Land ice cores (East Antarctica). *Quaternary Science Reviews* **29**, 146–159 (2010).

24. Xia, Z. Quantifying the fingerprint of oceanic moisture source conditions in deuterium and ¹⁷O excess parameters of precipitation. *Geophysical Research Letters* **50**,

609 (2023).

610 25. Xia, Z., Welker, J. M. & Winnick, M. J. The seasonality of deuterium excess in
611 non-polar precipitation. *Global Biogeochemical Cycles* **36**, (2022).

612 26. Risi, C., Bony, S., Vimeux, F. & Jouzel, J. Water-stable isotopes in the LMDZ4
613 general circulation model: Model evaluation for present-day and past climates and
614 applications to climatic interpretations of tropical isotopic records. *Journal of*
615 *Geophysical Research: Atmospheres* **115**, (2010).

616 27. Gao, Q., Sime, L. C., McLaren, A. J. & Werner, M. *Moisture Source Controls on*
617 *Water Isotopes in Antarctic Precipitation -Insights from Innovative Water Tracers in*
618 *ECHAM6-Wiso*. <https://doi.org/10.22541/essoar.173386109.93218804/v1> (2024).

619 28. Sodemann, H. & Stohl, A. Asymmetries in the moisture origin of Antarctic
620 precipitation. *Geophysical Research Letters* **36**, (2009).

621 29. Bong, H. *et al.* *Water Isotope Model Intercomparison Project (WisoMIP):*
622 *Present-Day Climate*. <https://doi.org/10.22541/essoar.175330911.17791287/v1> (2025).

623 30. Bong, H. *et al.* Process-Based intercomparison of water isotope-enabled models
624 and reanalysis nudging effects. *Journal of Geophysical Research: Atmospheres* **129**,
625 (2024).

626 31. Yu, L. & Weller, R. A. Objectively analyzed air–sea heat fluxes for the global
627 ice-free oceans (1981–2005). *Bulletin of the American Meteorological Society* **88**,
628 527–540 (2007).

629 32. Tomita, H., Hihara, T., Kako, S., Kubota, M. & Kutsuwada, K. An introduction to
630 J-OFURO3, a third-generation Japanese ocean flux data set using remote-sensing
631 observations. *Journal of Oceanography* **75**, 171–194 (2018).

632 33. Andersson, A. *et al.* Evaluation of HOAPS-3 ocean surface freshwater flux
633 components. *Journal of Applied Meteorology and Climatology* **50**, 379–398 (2011).

634 34. Brunke, M. A., Wang, Z., Zeng, X., Bosilovich, M. & Shie, C.-L. An Assessment
635 of the Uncertainties in Ocean Surface Turbulent Fluxes in 11 Reanalysis,
636 Satellite-Derived, and Combined Global Datasets. *Journal of Climate* **24**, 5469–5493
637 (2011).

638 35. Dietrich, L. J. *et al.* Snow-Atmosphere humidity exchange at the ice sheet surface
639 alters annual mean climate signals in ice core records. *Geophysical Research Letters* **50**,
640 (2023).

641 36. Liu, W. T., Katsaros, K. B. & Businger, J. A. Bulk Parameterization of Air-Sea
642 Exchanges of Heat and Water Vapor Including the Molecular Constraints at the Interface.
643 *Journal of the Atmospheric Sciences* **36**, 1722–1735 (1979).

644 37. Craig, H. & Gordon, L. I. Deuterium and oxygen 18 variations in the ocean and
645 the marine atmosphere. *Proc. Stable Isotopes in Oceanographic Studies and*

646 *Paleotemperatures* (1965).

647 38. Bui, T. Q., Do, N. T., Vu, T. T., Nguyen, D. T. & Trinh, D. A. Calculation of
648 Kinetic Fractionation of Open-Water Evaporation Over an Upland Reservoir. *Earth and*
649 *Space Science* **11**, (2024).

650 39. Horita, J., Rozanski, K. & Cohen, S. Isotope effects in the evaporation of water: a
651 status report of the Craig–Gordon model. *Isotopes in Environmental and Health Studies*
652 **44**, 23–49 (2008).

653 40. Hellmann, R. & Harvey, A. H. First-Principles diffusivity ratios for kinetic isotope
654 fractionation of water in air. *Geophysical Research Letters* **47**, (2020).

655 41. Noone, D. The influence of midlatitude and tropical overturning circulation on the
656 isotopic composition of atmospheric water vapor and Antarctic precipitation. *Journal of*
657 *Geophysical Research: Atmospheres* **113**, (2008).

658 42. Aemisegger, F. & Sjolte, J. A climatology of strong large-scale ocean evaporation
659 events. Part II: Relevance for the deuterium excess signature of the evaporation flux.
660 *Journal of Climate* **31**, 7313–7336 (2018).

661 43. Aemisegger, F. & Papritz, L. A Climatology of Strong Large-Scale Ocean
662 Evaporation Events. Part I: Identification, Global Distribution, and Associated Climate
663 Conditions. *Journal of Climate* **31**, 7287–7312 (2018).

664 44. Thurnherr, I. et al. The role of air–sea fluxes for the water vapour isotope signals
665 in the cold and warm sectors of extratropical cyclones over the Southern Ocean. *Weather*
666 *and Climate Dynamics* **2**, 331–357 (2021). doi:10.5194/wcd-2-331-2021

667 45. McGee, D., Broecker, W. S. & Winckler, G. Gustiness: The driver of glacial
668 dustiness? *Quaternary Science Reviews* **29**, 2340–2350 (2010).

669 46. MARGO Project Members , Constraints on the magnitude and patterns of ocean
670 cooling at the Last Glacial Maximum. *Nature Geoscience* **2**, 127–132 (2009).

671 47. Zhang, Y., Liu, Z., Li, D., Li, Z. & Shao, H. *Arctic Sea Ice Loss Amplifies Local*
672 *Evaporation Influence on Water Vapor Isotopes: Insights from Cruise Observations.*
673 <https://doi.org/10.5194/egusphere-egu26-3379> (2026).

674 48. Galewsky, J. Constraining Supersaturation and Transport Processes in a South
675 American Cold-Air Outbreak Using Stable Isotopologues of Water Vapor. *Journal of the*
676 *Atmospheric Sciences* **72**, 2055–2069 (2015).

677 49. Lewis, S. C., LeGrande, A. N., Kelley, M. & Schmidt, G. A. Modeling insights
678 into deuterium excess as an indicator of water vapor source conditions. *Journal of*
679 *Geophysical Research: Atmospheres* **118**, 850–865 (2013) doi:10.1002/jgrd.50053.

680 50. Jouzel, J. et al. Water isotopes as tools to document oceanic sources of
681 precipitation. *Water Resources Research* **49**, 7469–7486 (2013).

682 51. Hersbach, H. et al. The ERA5 global reanalysis. *Quarterly Journal of the Royal*

683 *Meteorological Society* **146**, 1999–2049 (2020).

684 52. Murray, F. W. On the Computation of Saturation Vapor Pressure. *Journal of*
685 *Applied Meteorology* **6**, 203–204 (1967).

686 53. Veres, D. *et al.* The Antarctic ice core chronology (AICC2012): An optimized
687 multi-parameter and multi-site dating approach for the last 120 thousand years. *Climate*
688 *of the Past* **9**, 1733–1748 (2013).

689 54. Jouzel, J. *et al.* Orbital and millennial Antarctic climate variability over the past
690 800,000 years. *Science* (2007) **317**, 793–796.

691 55. Worden, J., Noone, D. & Bowman, K. Importance of rain evaporation and
692 continental convection in the tropical water cycle. *Nature* **445**, 528–532 (2007).

693 56. Klein, E. S. & Welker, J. M. Influence of sea ice on ocean water vapor isotopes
694 and Greenland ice core records. *Geophysical Research Letters* **43**, 12,475–12,483 (2016).
695 doi:10.1002/2016GL071748

696 57. Bretherton, C. S., Widmann, M., Dymnikov, V. P., Wallace, J. M. & Bladé, I. The
697 effective number of spatial degrees of freedom of a time-varying field. *Journal of*
698 *Climate* **12**, 1990–2009 (1999).
699 doi:10.1175/1520-0442(1999)012<1990:TENOSD>2.0.CO;2

700 58. Ebisuzaki, W. A method to estimate the statistical significance of a correlation
701 when the data are serially correlated. *Journal of Climate* **10**, 2147–2153 (1997).
702 doi:10.1175/1520-0442(1997)010<2147:AMTETS>2.0.CO;2

703

704

705

706

707

708

709

710

711

712

713

Ice-core *d-excess* reveals a shift to wind-driven control of oceanic evaporation in warm climates

Authors: Ajay Ajay^{1,2}, Alexandre Cauquoin³, Kei Yoshimura³, Jeffrey M Welker^{2,4,5},
Prasanta Sanyal^{1,6}

Affiliations:

¹ Department of Earth Sciences, Indian Institute of Science Education and Research
Kolkata, 741246, West Bengal, India

² Ecology and Genetics Research Unit, University of Oulu, Finland

³ Institute of Industrial Science, University of Tokyo, Kashiwa, Chiba, Japan

⁴ Department of Biological Sciences, University of Alaska Anchorage, AK, USA

⁵ UArctic, University of the Arctic, Rovaniemi, Finland

⁶ Center for Climate and Environmental Studies, Indian Institute of Science Education
and Research Kolkata; 741246, West Bengal, India

This Supplementary Information presents the extended validation, methodological justification, and robustness analyses that support the main-text conclusions but are too detailed to include there. Supplementary Texts S1–S3 justify the central analytical choices — the physical robustness of the transfer-function curvature against an independent GCM ensemble (S1), the use of corrected *d-excess* without supersaturation correction at EPICA Dome C (S2), and why tropical *d-excess*–evaporation decoupling does not affect the high-latitude EDC source (S3). Supplementary Tables S1–S4 compile the published *d-excess* relationships and document the sensitivity of the calibration and reconstruction across regions, flux products, and model frameworks. Supplementary Figures S1–S8 provide the spatial, seasonal, observational, and statistical validation of the proxy and show that the regime-dependent reversal in evaporation control is robust to calibration choice, regime thresholds, smoothing window, and statistical method.

Supplementary Note 1. Validation of the transfer function curvature against the WISOMIP GCM ensemble

To test whether the quadratic transfer function $E = 0.266d^2 + 2.155d + 17.40$ reflects genuine atmospheric physics or is an artifact of the ERA5+OAFlux calibration, we derived independent transfer functions from all seven models participating in the Water Isotope Model Intercomparison Project (WISOMIP^{29,30}) over the EDC moisture source region (35°S–60°S, 30°E–150°E). For each model, annual mean evaporation (E , mm day⁻¹) and the d -excess of the evaporative flux (d , ‰) were extracted from ocean grid cells using a 0.25° land mask, and a quadratic transfer function of the form $E = Ad^2 + Bd + C$ was fitted independently. All seven GCMs reproduce a positive, monotonically increasing relationship between d -excess and evaporation, with R^2 values ranging from 0.73 to 0.93 (Table S4).

The curvature coefficient A , the only parameter invariant under translation of the d -excess axis, and therefore the physically meaningful one, ranges from 0.073 to 0.357 across the ensemble (multi-model ensemble mean: 0.204 ± 0.120). The ERA5-calibrated value ($A = 0.266$) falls within this 1σ spread, confirming that the curvature of the d -excess–evaporation relationship is consistent between observations and models and is not a calibration artifact.

The linear coefficient B and intercept C differ more substantially (ERA5-OAflux: $B = 2.155$, $C = 17.40$; GCM MME: $B = 6.23 \pm 1.57$, $C = 75.9 \pm 15.8$ — 2.4σ and 3.9σ from the MME mean, respectively). This is mathematically expected: B and C are not translation-invariant, so regressions over disjoint d -excess ranges (GCM: -15 to $+5\text{‰}$; ERA5: 8 to 16‰) shift predictably even when the underlying physics is identical. For an axis offset δ :

$$B_{\text{shifted}} = B - 2A\delta ; C_{\text{shifted}} = A\delta^2 - B\delta + C$$

Substituting $\delta \approx -13\text{‰}$ (the offset between the two ranges) yields predicted GCM values of $B \approx 9.07$ and $C \approx 90.4$, both within 1 – 2σ of the actual MME values. The apparent

769 disagreement in B and C therefore reflects the well-documented systematic offset in
770 GCM-simulated *d-excess*^{12,29,30} rather than a difference in underlying physics.

771 A residual disagreement remains. At the paleo *d-excess* operating range ($d \approx 12\text{‰}$), the
772 local sensitivity $dE/dd = 2Ad + B$ is $8.5 \text{ cm yr}^{-1} \text{‰}^{-1}$ from ERA5-Oaflux and 11.12 cm
773 $\text{yr}^{-1} \text{‰}^{-1}$ from the GCM MME — a $\sim 30\%$ difference. This could reflect either
774 non-quadratic structure in the true relationship (so that quadratic fits over different
775 *d-excess* ranges yield different effective sensitivities) or genuine model deficiencies in
776 simulated kinetic fractionation. We do not resolve this here; the residual is bounded,
777 affects only reconstructed amplitude (not timing or regime structure), and corresponds to
778 an amplitude uncertainty of approximately $\pm 30\%$ if GCM coefficients were used.

779 We adopt the ERA5+OAFlux calibration for paleo-reconstruction because its *d-excess*
780 range ($8\text{--}16\text{‰}$) matches the EDC paleo range ($11.7\text{--}14.2\text{‰}$) — applying GCM
781 coefficients would require extrapolation beyond their simulated domain — and because it
782 is anchored in three independent observational flux products (OAFlux, J-OFURO3,
783 HOAPS) and an in-situ ship dataset (Uemura et al., 2008), all yielding consistent
784 functional forms over the source region. The WISOMIP ensemble's role here is
785 structural: agreement on A confirms that the curvature is physically robust across
786 frameworks, while the calibration of magnitude rests on the observational chain.

787 **Supplementary Note 2. Use of d_{corr} without supersaturation correction**

788 We apply the evaporation transfer function directly to the EDC d_{corr} record without the
789 logarithmic supersaturation correction ($d\ln$) commonly used in paleoclimate applications
790^{20,22}. At EDC specifically, three findings from Landais et al.²⁰ demonstrate that $d\ln$ offers
791 no advantage over d_{corr} for source-region reconstruction.

792 First, d_{corr} and $d\ln$ correlate with source temperature equally well at EDC (both $R =$
793 0.93)²⁰, in sharp contrast to Dome F where $d_{\text{corr}}\text{--}T_{\text{source}}$ yields $R = 0.49$ while

794 $\text{dln-T}_{\text{source}}$ yields $R = 0.97$. The supersaturation distortion that dln corrects for is
795 therefore present at Dome F but largely absent at EDC.

796 Second, the EDC $\delta\text{D}-\delta^{18}\text{O}$ slope is 8.2 with no anticorrelation between d_{corr} and $\delta^{18}\text{O}$ —
797 the classical isotopic fingerprint of supersaturation enhancement is absent at this site.

798 Third, Landais et al.²⁰ themselves use classical *d-excess* rather than dln for their 800-kyr
799 EDC reconstruction because both definitions yield identical timing for all interglacial
800 maxima.

801 Beyond these prior results, the sensitivity of our transfer function provides an additional
802 methodological reason to avoid the correction. The local sensitivity $\partial E/\partial d \approx 8.5 \text{ cm yr}^{-1}$
803 ‰^{-1} at the paleo operating range ($d \approx 12\text{‰}$) means that any uncertainty introduced into d
804 by the supersaturation tuning, itself the dominant source of error in MCIM-based ΔT_{source}
805 reconstructions, increasing uncertainty by a factor of two, would propagate amplified into
806 reconstructed E . At EDC, where the correction does not improve source-signal recovery,
807 applying it would amplify rather than reduce reconstruction error.

808 **Supplementary Note 3. Tropical decoupling of *d-excess* from evaporation and its** 809 **non-applicability to EDC**

810 The *d-excess*–evaporation correlation weakens significantly in the tropics (Fig. 1D) for
811 three reasons: (i) high boundary-layer relative-humidity compresses the $(1 - h)$ dynamic
812 range that drives *d-excess* variability; (ii) the Uemura et al.¹⁴ calibration we adopt was
813 derived in the Southern Indian Ocean midlatitudes and is not expected to hold under
814 tropical thermodynamic conditions; and (iii) sub-cloud re-evaporation of falling rain and
815 convective downdrafts inject compositionally distinct moisture into the boundary layer,
816 overwriting the primary surface evaporation signal^{26, 55}. However, this tropical
817 decoupling does not affect our reconstruction; moisture transport to the EPICA Dome C
818 site is dominated by cold, high-latitude pathways where these confounding convective
819 processes are physically suppressed, thereby preserving the primary source evaporation
820 signal²⁷.

Table S1.

Compilation of published empirical relationships between marine vapor *d-excess* and ocean surface conditions, derived from concurrent in-situ ship-based water vapor isotope measurements. For each study we list the regression equation, the humidity variable used (note that RH is defined differently across studies — normalized to saturation at SST in most cases, with minor variations indicated), the reported coefficient of determination (R^2) or correlation coefficient (R), the oceanic sector sampled, and the campaign duration. The Uemura et al. (2008) relationship is adopted in this study because its sampling domain (35°S–65°S, Southern Indian Ocean) most closely matches the established moisture-source region of the EPICA Dome C ice core.

Source	Empirical Relationship	Key Method/Variable	R^2 (or R)	Region	Duration
Pfahl & Wernli, ¹³	$d = 48.3\text{‰} - 0.53\text{‰}/\% \cdot \text{RH}_{2\text{mSST}}$	RH averaged at 2 m height, calculated w.r.t. saturation at SST ($\text{RH}_{2\text{mSST}}$)	$R = -0.82$	Mediterranean	2000–2006
Uemura et al. ¹⁴	$d = 0.45 \cdot \text{SST} - 0.42 \cdot h^* + 37.9$	Relative humidity normalized to saturation vapor pressure at the sea surface (h^*)	$R^2 = 0.70$ (3-variable fit)	35°S–65°S	04/01/2006–30/01/2006
Benetti et al. ¹¹	$d = -0.45 \cdot \text{Rhs} + 42$	Rhs (relative humidity w.r.t. SST)	$R^2 = 0.89$	26°N, 35°W	21/08/2012–10/09/2012
Pfahl & Sodemann ¹⁰	$d = 48.2\text{‰} - 0.54 \cdot \text{RH}$	RH with respect to saturation at the sea surface	Not specified	Global (implied)	Not specified
Bonne et al., ¹⁷	$d = -0.33 \cdot \text{RH} + 0.27 \cdot \text{SST} + 25.01$	RH_{sea} (relative humidity w.r.t. saturation at SST)	$R^2 = 0.76$	Whole Atlantic	Sep 2015–Jun 2017
Thurnherr et al., ¹⁶	$d = -0.4 \cdot h_{\text{SST}} + 36.8$	h_{SST} (specific humidity normalized to saturation specific humidity at SST)	$R = -0.73$	Antarctic to Europe	21/11/2016–11/04/2017

Table S2. Master sensitivity and stability analysis of the *d*-excess–evaporation relationship across various regions and independent datasets.

Statistics are provided for the regression of water vapor *d*-excess against three independent oceanic evaporation products: OAFlux (hybrid/synthesized), J-OFURO3 (satellite-derived latent heat flux), and HOAPS (satellite-derived evaporation). Results are presented for the global ocean (60N–60°S), the broad Southern Hemisphere Midlatitude ocean catchment (35°S–60°S), and three latitudinal migration scenarios within the Indian Ocean sector (30°E- 150°E): Northward (30°S–50°S), Central (35°S–55°S), and Southward (40°S–60°S) shifts. The table includes coefficients for both linear ($E=md+cE$) and quadratic ($E=ad^2+bd+cE$) fits. The "Stability (CV %)" row represents the Coefficient of Variation calculated across the three Indian Ocean latitudinal shifts, quantifying the spatial invariance of the proxy's sensitivity. R^2 : coefficient of determination; RMSE: root-mean-square error; All p-values $<10^{-23}$.

Region (Latitude Range)	R^2 (Lin)	Slope	R^2 (Quad)	Coeff. a	Coeff. b	Int. c	RMSE
OAFlux (cm/yr)							
60°N-60°S	0.879	8.95	0.897	0.305	0.737	20.49	15.62
35°S-60°S	0.923	7.97	0.959	0.342	0.716	19.24	6.99
30°S-50°S; 30°E- 150°E	0.914	8.26	0.926	0.208	3.096	14.02	9.8
35S-55S; 30°E- 150°E	0.949	7.81	0.965	0.266	2.155	17.4	6.52
40°S-60°S; 30°E- 150°E	0.941	7.83	0.968	0.349	1.365	17.15	5
Stability (CV %)	--	5.81%		25.7%	62.3%	12%	--
J-OFURO3 (W/m2)							
60N-60°S	0.82	6.86	0.838	0.235	0.537	26.27	15.61
35°S-60°S	0.889	6.1	0.922	0.256	0.678	25.75	7.52
30°S-50°S; 30°E- 150°E	0.875	6.3	0.884	0.148	2.63	20.06	9.53
35S-55S; 30°E- 150°E	0.914	5.84	0.929	0.193	1.748	25.21	7.1
40°S-60°S; 30°E- 150°E	0.902	6.02	0.926	0.258	1.243	24.36	5.95
Stability (CV %)	--	6.35%	--	20.5%	68.1%	11 %	--
HOAPS (mm/day)							
60N-60°S	0.702	0.215	0.737	0.0116	-0.0958	1.585	0.67
35°S-60°S	0.784	0.151	0.795	0.0039	0.0693	0.934	0.32
30°S-50°S; 30°E- 150°E	0.779	0.151	0.811	0.0069	-0.022	1.699	0.31
35S-55S; 30°E- 150°E	0.817	0.137	0.825	0.0033	0.0658	1.227	0.28
40°S-60°S; 30°E- 150°E	0.756	0.166	0.769	0.0057	0.0608	0.99	0.32

Stability (CV %)	--	9.81%	--	34.20%	123.50%	28.90%	--
------------------	----	-------	----	--------	---------	--------	----

Table S3: Sensitivity Analysis of Reconstructed Evaporation with moisture source temperature.

The table summarizes the mean evaporation anomaly ($\Delta E\%$) and its coupling strength with moisture source temperature (R^2) across 15 sensitivity scenarios, utilizing three independent observational flux products (OAFlux, J-OFURO3, HOAPS) across five distinct calibration regions (from Table-1). R^2 values represent the correlation between reconstructed evaporation (using each calibration) and moisture source temperature, not between evaporation products. The near-identical R^2 values across calibrations demonstrate robustness of the regime-dependent coupling pattern. Results are categorized into three thermal regimes based on moisture source temperature anomalies (ΔT_{source}): Interglacial ($\Delta T_{\text{source}} \geq -1.0^\circ\text{C}$), Intermediate ($-3.5 < \Delta T_{\text{source}} < -1.0^\circ\text{C}$), and Glacial ($\Delta T_{\text{source}} \leq -3.5^\circ\text{C}$). Please note that while the primary linear sensitivity remains remarkably stable across all regions (CV<6% for OAFlux and J-OFURO3), the individual quadratic coefficients show high Coefficients of Variation. This variability is physically and mathematically anticipated due to the inherent multicollinearity and parameter coupling between first- and second-order terms required to fit the non-linear scaling of saturation vapor pressure across diverse temperature regimes.

Evaporation Dataset	Calibration Region	Interglacial (IG)		Intermediate (INT)		Glacial (G)	
		$\Delta E(\%)$	R^2	$\Delta E(\%)$	R^2	$\Delta E(\%)$	R^2
OAFlux	60°N-60°S	11.3	0.231	-3.8	0.100	-15.2	0.568
	35°S-60°S	11.9	0.231	-4.0	0.100	-15.9	0.568
	30°S-50°S;30°E- 150°E	9.8	0.230	-3.5	0.099	-13.8	0.571
	35S-55S; 30°E- 150°E	10.5	0.231	-3.6	0.099	-14.5	0.570
	40°S-60°S; 30°E- 150°E	11.8	0.231	-4.0	0.099	-15.9	0.569
J-OFURO3	60°N-60°S	9.8	0.231	-3.3	0.100	-13.1	0.568
	35°S-60°S	10.1	0.231	-3.4	0.100	-13.5	0.569
	30°S-50°S;30°E- 150°E	8.5	0.230	-3.0	0.098	-12.0	0.571
	35S-55S; 30°E- 150°E	8.8	0.231	-3.1	0.099	-12.2	0.570
	40°S-60°S; 30°E- 150°E	10.0	0.231	-3.4	0.099	-13.6	0.569
HOAPS	60°N-60°S	10.2	0.233	-3.1	0.102	-12.5	0.563
	35°S-60°S	7.1	0.230	-2.5	0.098	-10.1	0.571
	30°S-50°S;30°E- 150°E	6.8	0.232	-2.2	0.100	-8.8	0.567
	35S-55S; 30°E- 150°E	6.0	0.230	-2.1	0.098	-8.5	0.571
	40°S-60°S; 30°E- 150°E	8.0	0.231	-2.8	0.099	-11.1	0.570

ENSEMBLE	Mean ± Std	9.4±1.8		-3.2±0.6		-12.7±2.3	
----------	------------	---------	--	----------	--	-----------	--

Table S4.

Quadratic transfer function coefficients ($E = Ad^2 + Bd + C$) derived independently for each of the seven WISOMIP isotope-enabled GCMs over the EDC moisture source region (35°S–60°S, 30°E–150°E), using GCM-simulated evaporation (ev , cm yr^{-1}) regressed against the GCM-simulated d -excess of the evaporative flux. R^2 values refer to the goodness of fit of the quadratic regression for each model. The Multi-Model Ensemble mean (MME mean) and inter-model standard deviation ($\text{MME} \pm 1\sigma$) are reported across all seven models. The ERA5+OAF flux transfer function coefficients (this study) are shown for comparison; the quadratic coefficient A falls within the GCM $\text{MME} \pm 1\sigma$ spread, confirming that the nonlinear functional form is physically robust and independent of the reanalysis calibration dataset.

Model	A	B	C	R^2
CAM5	0.177	6.4078	66.97	0.792
ECHAM	0.07	4.2783	75.67	0.738
GISS	0.1227	5.8459	75.09	0.697
GSM	0.3749	8.5385	79.26	0.793
LMZ	0.3067	5.6071	48.49	0.8
MIROC	0.3189	8.0421	87.77	0.864
NICAM	0.0838	4.8603	98.39	0.942
MME mean	0.207	6.2257	75.95	
$\text{MME} \pm 1\sigma$	± 0.12	± 1.57	± 15.76	
ERA5	0.266	2.155	17.4	0.952

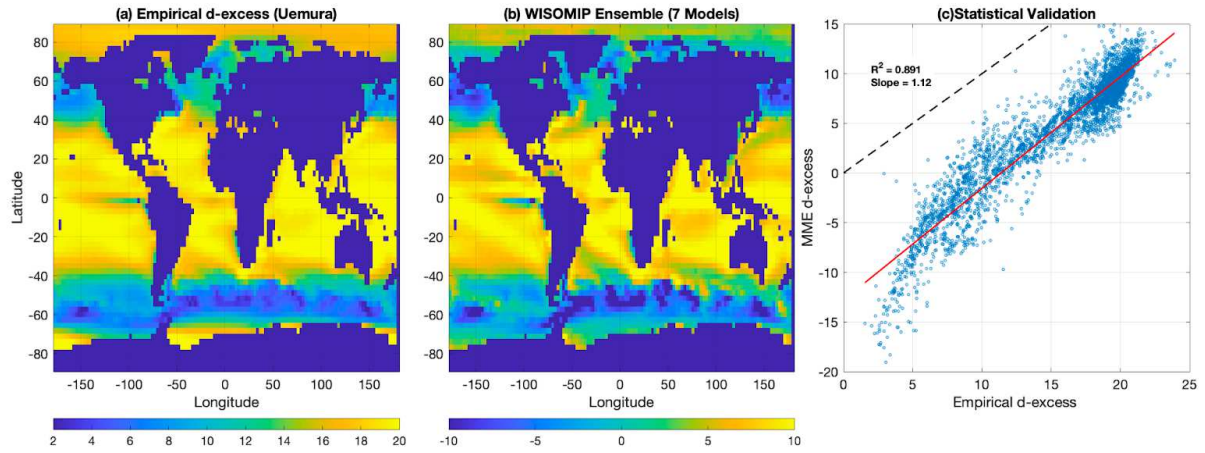


Fig. S1: Global Validation of the d-excess Proxy against the WISOMIP iGCM Ensemble.

Spatial comparison and statistical validation between the (a) empirically derived oceanic d -excess (Uemura et al., 2008) and the (b) Multi-Model Ensemble (MME) mean from seven nudged, isotope-enabled General Circulation Models (iGCMs) from the Water Isotope Model Intercomparison Project (WISOMIP; Bong et al., 2025). One model (CAM6) was excluded from the MME as a numerical outlier ($R^2 \approx 0.01$). (c) Linear regression and scatter validation for the global ocean (60°S to 60°N). All data were down-sampled to a common 128×64 Gaussian grid, and land areas were masked.

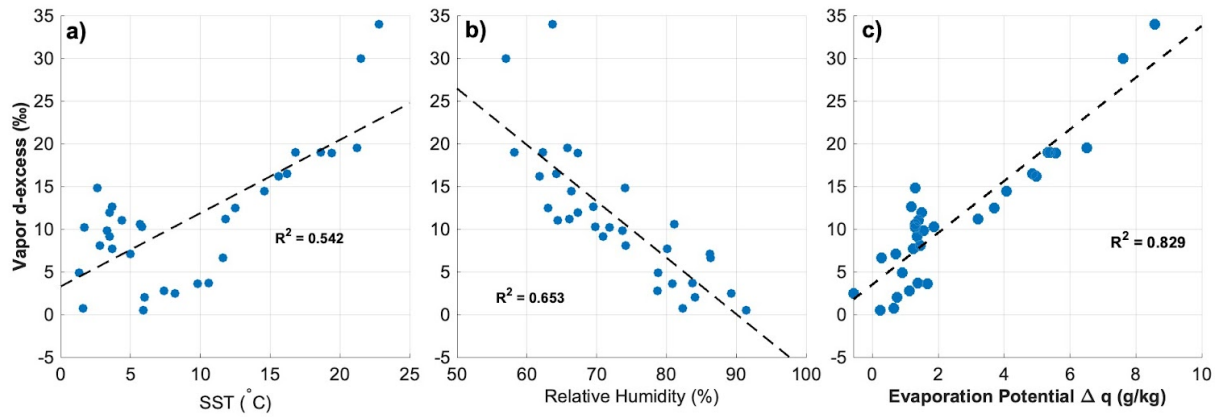
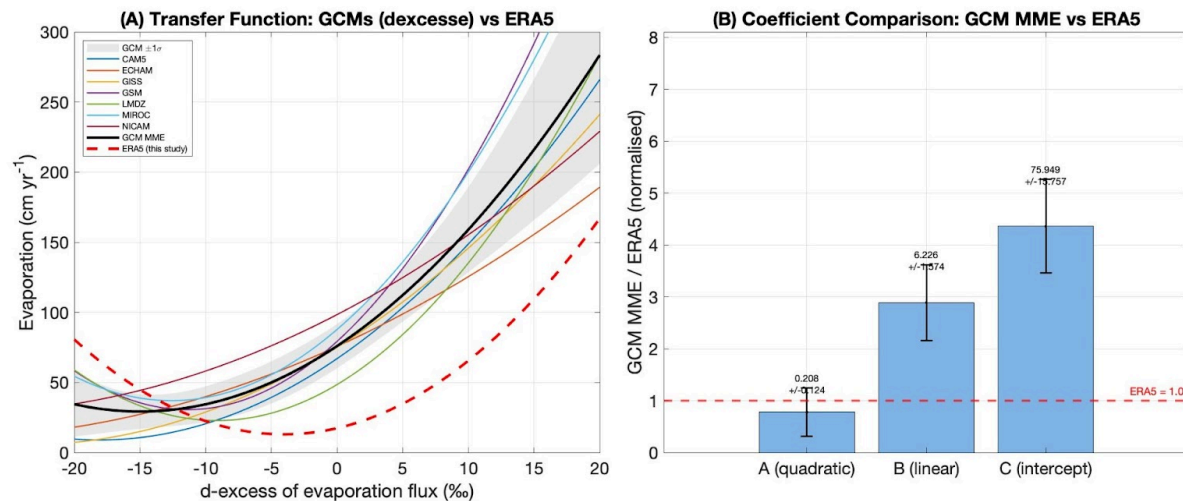
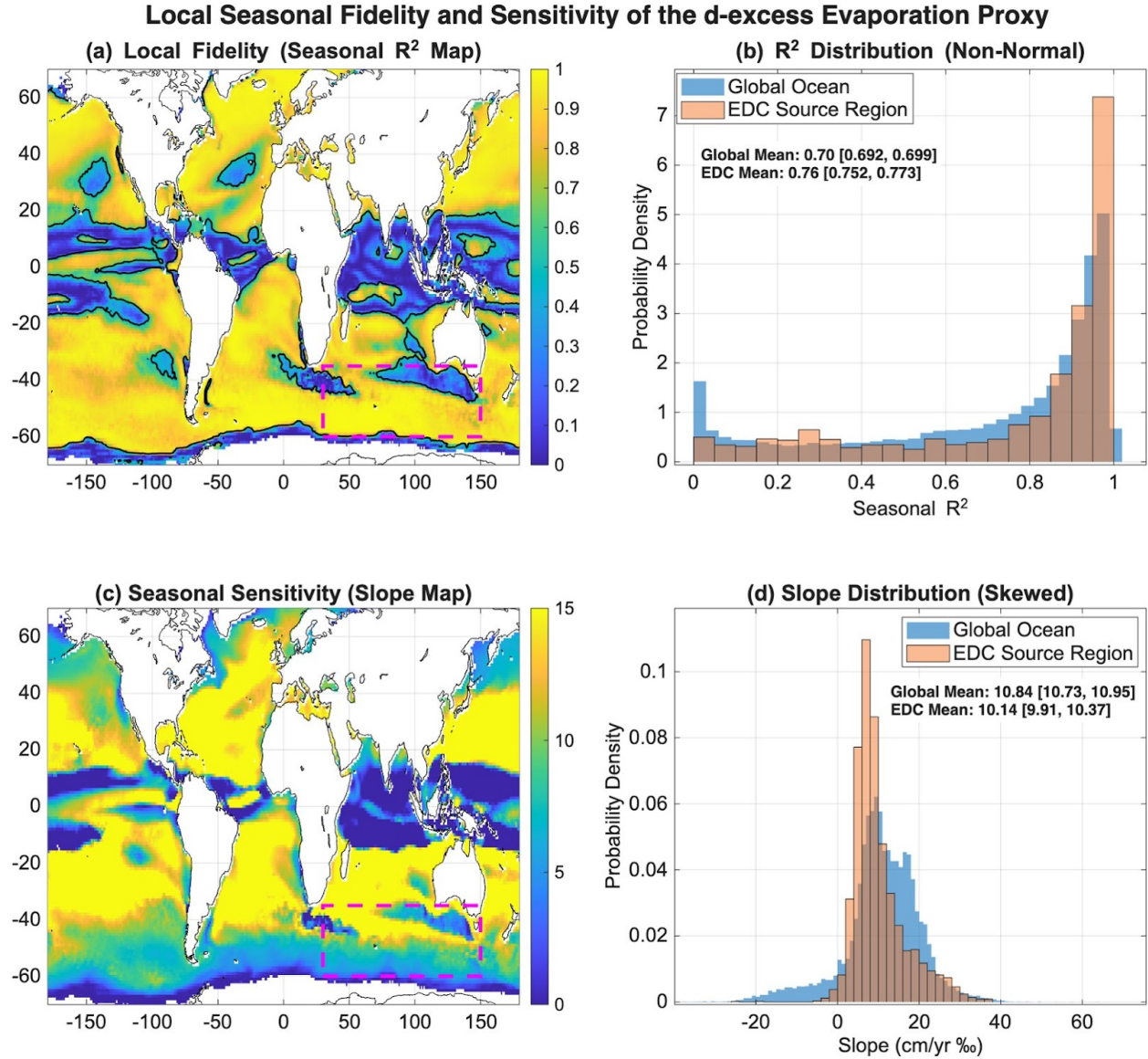


Fig. S2. Observational evidence of d-excess as proxy of evaporation rate. Observational validation of water vapor d-excess as a proxy for oceanic evaporation potential over 30°S–60°S ocean; data from Uemura et al., 2008). Scatter plots show d-excess against (a) sea surface temperature (SST), (b) relative humidity (RH), and (c) the thermodynamic evaporative potential (Δq).



924

Fig. S3. Independent validation of the ERA5+OAFflux transfer function using the WISOMIP GCM ensemble. **(A)** Quadratic transfer functions $E = Ad^2 + Bd + C$ fitted independently to each of the seven WISOMIP models (coloured curves) over the EDC moisture source region (35°S – 60°S , 30°E – 150°E), using the d-excess of the evaporative flux and surface evaporation extracted from ocean grid cells. The GCM multi-model ensemble mean (MME; black solid) and $\pm 1\sigma$ spread (grey shading) are shown alongside the ERA5+OAFflux calibration curve (red dashed). **(B)** Transfer function coefficients A (quadratic), B (linear), and C (intercept) for the GCM MME normalised to the corresponding ERA5 value (ERA5 = 1.0, red dashed line). Error bars show the GCM $\pm 1\sigma$ inter-model spread normalised by the ERA5 coefficient. The quadratic coefficient A — which governs the nonlinear shape of the transfer function — is consistent between the GCM MME and ERA5 (within 1σ), while B and C differ significantly due to the different d-excess ranges over which the two frameworks are calibrated (see Methods). Coefficient values for individual models are given in Table S4.



939

940 **Fig. S4. Local seasonal fidelity and sensitivity of the *d*-excess–evaporation**
 941 **relationship.** (a) Global map of the local coefficient of determination (R^2) calculated
 942 from the 12-month climatological cycles of vapor *d*-excess and oceanic evaporation at
 943 each grid point. Black contours delineate regions of high seasonal fidelity ($R^2 > 0.5$). (b)
 944 Probability density functions (PDF) of seasonal R^2 for the global ocean (60°N–60°) and
 945 the integrated EDC moisture source region E. (c) Global map of regression slopes (cm/yr
 946 per ‰), representing the seasonal sensitivity of the proxy. (d) PDF comparison of
 947 regression slopes for the global ocean (gray) and the EDC source region (red). In panels
 948 (b) and (d), mean values and 95% non-parametric confidence intervals (in brackets) were

estimated via 1,000 bootstrap iterations to account for the skewed, non-Gaussian nature of the spatial distributions. The magenta dashed boxes indicate the integrated EDC moisture source catchment.

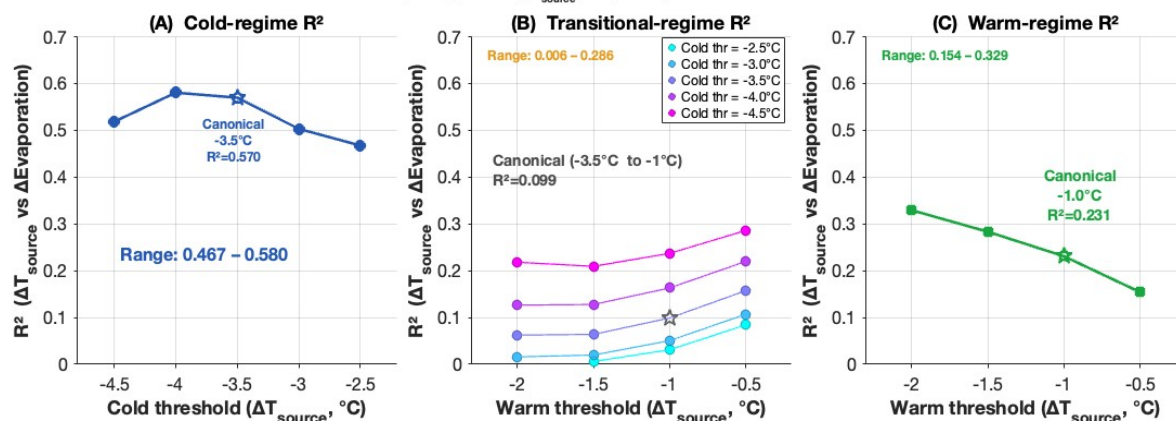


Fig. S5. Threshold sensitivity analysis of regime-dependent R^2 (ΔT_{source} , $\Delta \text{Evaporation}$). (A) Cold-regime R^2 as a function of the cold threshold. The first point represent R^2 when $\Delta T_{\text{source}} < -4.5^\circ\text{C}$, second point represents $\Delta T_{\text{source}} < -4^\circ\text{C}$ and so on. (B) Transitional-regime R^2 as a function of the warm threshold, with separate curves for each cold-threshold value, colors indicate cold threshold from -2.5°C (cyan) to -4.5°C (magenta). (C) Warm-regime R^2 as a function of the warm threshold. Stars mark canonical thresholds (cold = -3.5°C , warm = -1.0°C). Cold-regime R^2 (0.467–0.580) exceeds warm-regime R^2 (0.154–0.329) across all 19 valid threshold combinations, and transitional R^2 remains low throughout (0.006–0.286), confirming that the regime-dependent coupling hierarchy is robust to threshold choice.

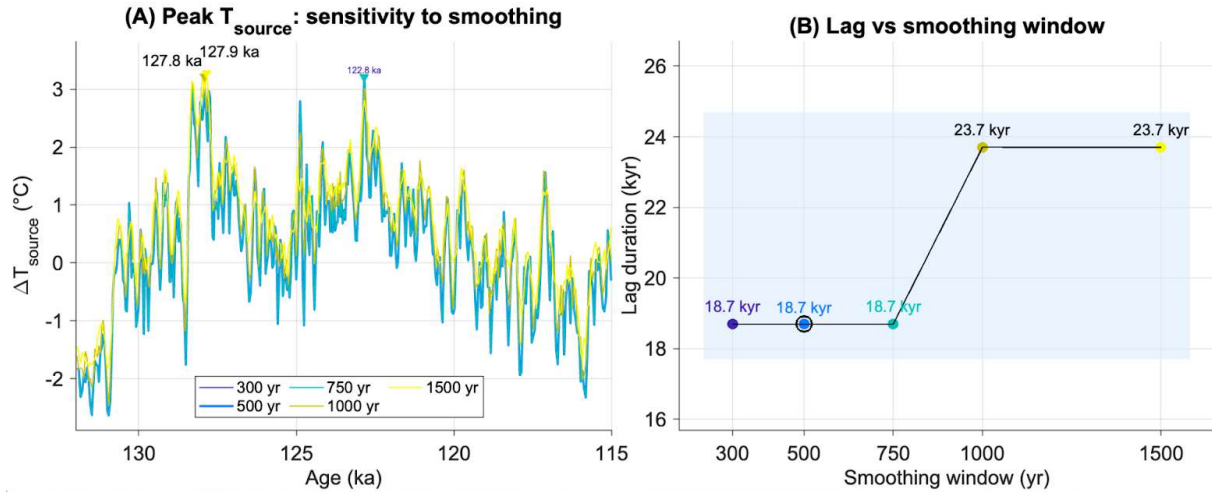
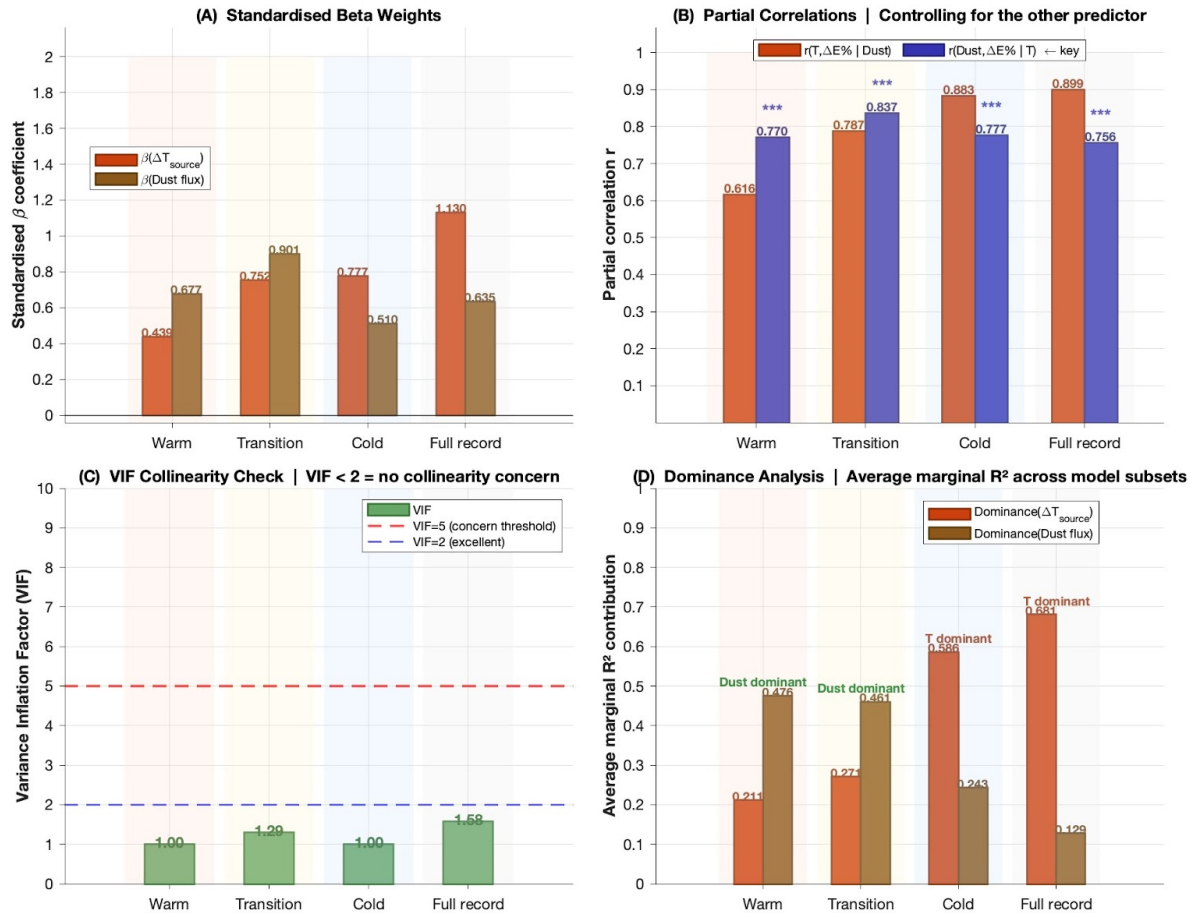


Fig. S6. Sensitivity of the reconstructed hydrological lag to the choice of Savitzky-Golay smoothing window applied to the raw EDC isotope records prior to reconstruction. For each of five smoothing windows (300–1500 yr), the full reconstruction pipeline was repeated independently: raw $\delta^{18}\text{O}$ and δD records were smoothed, seawater-corrected, converted to corrected d-excess (d_{corr}), and passed through the quadratic transfer function ($E = 0.266d^2 + 2.155d + 17.4$) to yield $\Delta E(\%)$ and ΔT_{source} . Peak ages were identified within the 100–150 ka window, identical to the main reconstruction. (A) Reconstructed ΔT_{source} over 115–132 ka for each smoothing window; filled triangles mark the peak source temperature age for each window, with the age labelled. The shift from 122.8 ka (300–500 yr) to 127.9 ka (1000–1500 yr) reflects the blending of two closely spaced MIS 5e warm episodes under heavier smoothing. (B) Lag duration as a function of smoothing window; the shaded band indicates the full range (18.7–23.7 kyr). The circled point denotes the 500 yr window used in the main reconstruction, consistent with Landais et al.²⁰. The peak evaporation age (104.2 ka) is invariant across all windows (not shown).



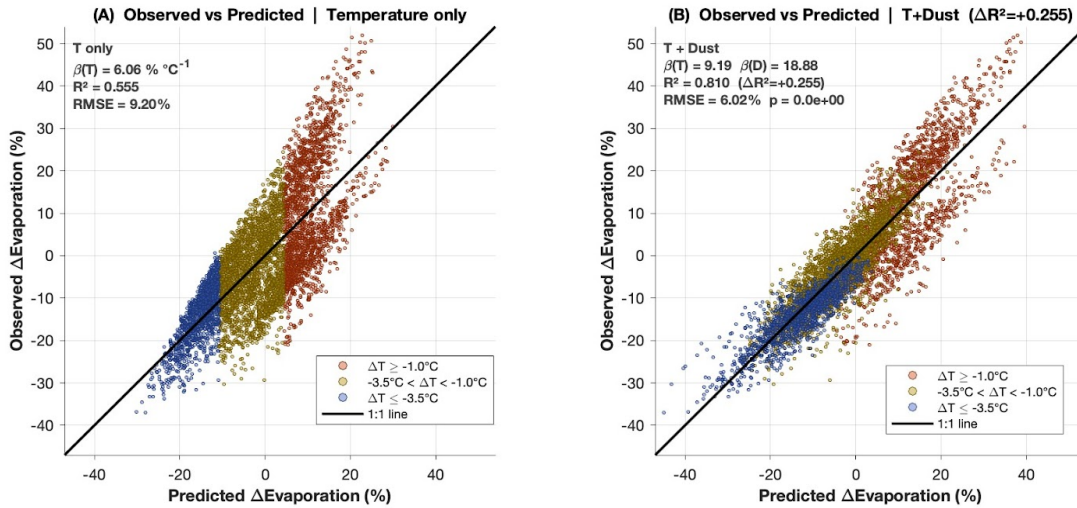
986

987 **Fig. S7 Statistical robustness of the wind-speed contribution to evaporation.**

988 **(A)** Standardised regression coefficients $\beta(\Delta T_{\text{source}})$ (orange) and $\beta(\text{Dust flux})$ (brown)
 989 from the combined model fitted within each thermal regime and the full 800-kyr record.
 990 Standardisation enables direct comparison of effect magnitude on a common scale.
 991 Background shading identifies thermal regimes. **(B)** Partial correlations between source
 992 temperature and $\Delta\text{Evaporation}$ controlling for dust flux (orange; $r(T, \Delta E\% | \text{Dust})$) and
 993 between dust flux and $\Delta\text{Evaporation}$ controlling for source temperature (blue; $r(\text{Dust},$
 994 $\Delta E\% | T)$) by regime. Asterisks denote significance of $r(\text{Dust}, \Delta E\% | T)$ ($p < 0.001$). **(C)**
 995 Variance Inflation Factors (VIF) by regime. Dashed lines mark VIF=5 (concern
 996 threshold, red) and VIF=2 (excellent, blue). All values below 2 confirm negligible
 997 collinearity between temperature and dust flux predictors. **(D)** Average marginal R^2

998 contribution of each predictor across all possible model subsets (dominance analysis) by
 999 regime. Labels indicate the dominant predictor in each regime.

1000



1001

1002 **Fig. S8 Multiple regression model performance diagnostics.**

1003 **(A)** Observed versus model-predicted $\Delta\text{Evaporation (\%)}$ using source temperature alone
 1004 (Model 1: $\Delta E\% = \beta_0 + \beta_1\Delta T_{\text{source}} + \epsilon$). Points are coloured by thermal regime: warm
 1005 ($\Delta T_{\text{source}} \geq -1.0^\circ\text{C}$, orange), transition ($-3.5^\circ\text{C} < \Delta T_{\text{source}} < -1.0^\circ\text{C}$, gold), and cold
 1006 ($\Delta T_{\text{source}} \leq -3.5^\circ\text{C}$, blue). The black line denotes the 1:1 reference. Inset shows regression
 1007 slope $\beta(T)$, R^2 , and RMSE. **(B)** As in (A) but using the combined temperature and dust
 1008 flux model (Model 2: $\Delta E\% = \beta_0 + \beta_1\Delta T_{\text{source}} + \beta_2\log(\text{Dust}) + \epsilon$; Lambert et al.⁸). Inset
 1009 shows $\beta(T)$, $\beta(D)$, combined R^2 , incremental ΔR^2 over Model 1, RMSE, and F-test
 1010 p-value for the improvement.

1011

1012

Supplementary Files

This is a list of supplementary files associated with this preprint. Click to download.

- [SupplementaryInformation.pdf](#)