https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

Review article: Stable water isotopic evolution at the glacier surface

Mingyu Dou[1, 2], Aibin Zhao [1], Deji [1*], Jing Gao [1, 3*]

[1] *State Key Laboratory of Tibetan Plateau Earth System, Resources and Environment, Institute of Tibetan Plateau Research, Chinese Academy of Sciences, Beijing 100101, China*

[2] *University of Chinese Academy of Sciences, Beijing, 100049, China*

[3] *Lanzhou University, Lanzhou 730000, China*

* Correspondence to: Deji, E-mail: deji@itpcas.ac.cn

Jing Gao, E-mail: gaojing@itpcas.ac.cn

Abstract: Stable water isotopes ($\delta^{18}O$ and δD) preserved in ice cores record past climate change. Reconstructing historical temperatures from ice cores relies fundamentally on the empirical linear relationship between stable water isotopes and local temperature. However, post-depositional processes at the glacier surface continuously alter initial isotopic signatures across diverse climatic regimes, yet their differential contributions across regions remain poorly understood. In this review, we synthesise findings from current field observations, laboratory experiments, and numerical modelling to elucidate the physical mechanisms governing isotopic evolution at the glacier surface. We systematically evaluate key post-depositional drivers, including sublimation–deposition, turbulent mixing, solid- and gas-phase diffusion, dry- and wet-snow metamorphism, wind-driven redistribution and pumping, and meltwater infiltration–refreezing. Previous observations from East Antarctica indicate post-depositional enrichments in near-surface $\delta^{18}O$ of 3.0–3.3‰ and depletions in *d*-excess of 3.4–3.5‰, while Greenland surface snow exhibits diurnal fluctuations of ~2.5‰ in $\delta^{18}O$ and ~13‰ in δD. These modifications reduce seasonal signals and weaken the local isotope-temperature relationship. We emphasise that future high-resolution field observations of isotopic compositions within snowpacks and atmospheric moisture must be prioritised. Such observational efforts are essential to elucidate key physical mechanisms, optimise the isotope-enabled atmosphere–snowpack models, and enhance our capacity to simulate interfacial isotopic exchange. Finally, we identify key research priorities to bridge existing knowledge gaps and propose an integrated observation-experiment-modelling framework to improve the calibration of water isotope proxies and enhance the precision of ice-core-based paleoclimatic reconstructions.

Keywords: stable water isotopes, post-depositional processes, sublimation–deposition, diffusion, ice core

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

1 Introduction

Paleoclimate can be reconstructed from ice-core archives (EPICA community members, 2004; Masson-Delmotte et al., 2006). They offer an essential long-term perspective required to distinguish anthropogenic forcing from natural variability and are indispensable for constraining climate sensitivity in future projections (Intergovernmental Panel on Climate Change (IPCC), 2021; Tierney et al., 2020). Historical temperature records spanning approximately 800,000 and 123,000 years have been obtained from the Dome C ice core and the North Greenland Ice Core Project (NGRIP) ice core, respectively (North Greenland Ice Core Project Members, 2004; Jouzel et al., 2007). High-resolution ice-core records also allow us to identify seasonal temperature variability in paleoclimate (Furukawa et al., 2017; Vinther et al., 2010), relying on a favourable linear correlation between stable water isotopes and temperature (Dansgaard, 1964; Merlivat and Jouzel, 1979; Steen-Larsen et al., 2014).

Newly fallen snow stores the isotopic signature of atmospheric moisture. However, before these signals are finally preserved in ice cores, they undergo continuous fractionation during multiple phase transitions at the atmosphere–snow–ice interface, thereby modifying the original isotopic signal (Kiemle et al., 2023). The post-depositional processes (Figure 1) include surface sublimation and deposition of vapour (Town et al., 2024), diffusion (Gkinis et al., 2021), snow metamorphism (Kopec et al., 2022), wind pumping (Fisher et al., 1983; Neumann and Waddington, 2004), wind deposition (King et al., 2004), and infiltration–refreeze (Ala-aho et al., 2017). Non-precipitation-driven fractionation mechanisms exist at the interface of surface snow and near-surface water vapour, with approximately 6–20% of surface snow being involved in isotopic interactions with the atmosphere (Steen-Larsen et al., 2014), inducing a synchronous diurnal cycle in surface snow isotopes at a depth of around 2–5 mm (Ritter et al., 2016). Snow pores also allow hydrogen and oxygen stable isotopes to diffuse along gradients in temperature and isotopic composition (Neumann and Waddington, 2004). Estimates suggest that 24–43% of the snow may be affected within one day (Ebner et al., 2017), progressively damping the seasonal signal archived in the snowpack (Van De Berg et al., 2013). Enhanced firn densification associated with snow metamorphism can further limit isotope diffusion by affecting both diffusion length and diffusion coefficient (Holme et al., 2018). In contrast, wind pumping can promote the transport of atmospheric moisture into the snowpack, thereby intensifying isotopic exchange processes (Colbeck, 1997; Neumann and Waddington, 2004). Wind deposition not only redistributes surface snow but also modulates the regional surface

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

mass and energy balance by amplifying snow sublimation (Amory and Kittel, 2019; Palm et al., 2017). Additionally, accurate paleoclimate reconstruction must also account for precipitation intermittency (Casado et al., 2018), as ice-core isotope records may contain discontinuous climatic signals due to intermittent precipitation. This substantially constrains the relationship between isotopic composition and temperature (Münch et al., 2021), introduces considerable noise, and decreases the temporal resolution available for extracting meaningful climate information (Casado et al., 2020; Persson et al., 2011).

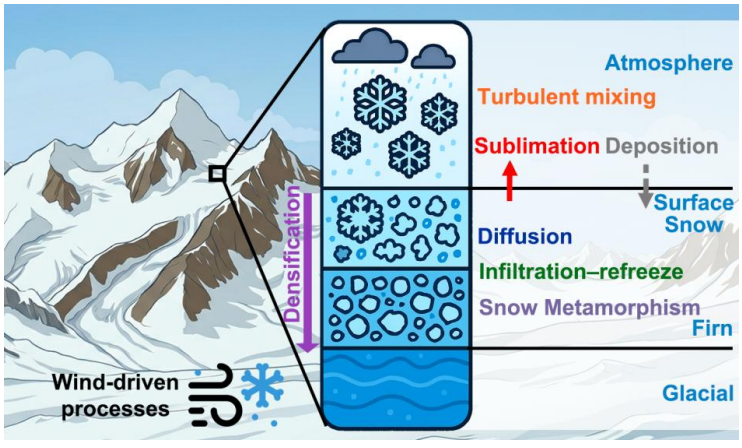


Figure 1. Primary post-depositional processes at the atmosphere–snow–ice interface.

Researchers typically select Antarctic, Greenland Ice Sheet, or high-altitude glaciers as sites for observation of post-depositional isotopic evolution at the atmosphere–snow–ice interface (Figure 2), where well-defined snow layers and long accumulation histories provide favourable conditions for monitoring sublimation, diffusion, and snow metamorphism. At East Antarctic sites such as Dome C, snowpack observations show that post-depositional processing systematically raises near-surface $\delta^{18}O$ and lowers deuterium excess (*d*-excess), with the imprint concentrated in the top centimetre of the snowpack (Ollivier et al., 2025).Continuous measurements of vapour isotopes revealed variations of up to 10‰ in both surface snow and vapour (Steen-Larsen et al., 2014), allowing isotopic signals to more strongly reflect local processes rather than weather-scale or precipitation inputs (Ritter et al., 2016). Additional evidence from the Greenland Ice Sheet shows that d-excess may decrease by up to 5‰ during summer following deposition (Town et al., 2024). Even at the daily scale, surface snow within the uppermost few millimetres exhibited diurnal fluctuations of ~2.5‰ in $\delta^{18}O$ and ~13‰ in δD, comparable to 10–20% of the seasonal isotopic variation or even of the magnitude of a glacial-interglacial transition (Madsen et al., 2019). Collectively, these findings highlight the

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.



importance of post-depositional processes and caution that they must be accounted for when interpreting ice-core stable isotope records (Jouzel and Masson-Delmotte, 2010).

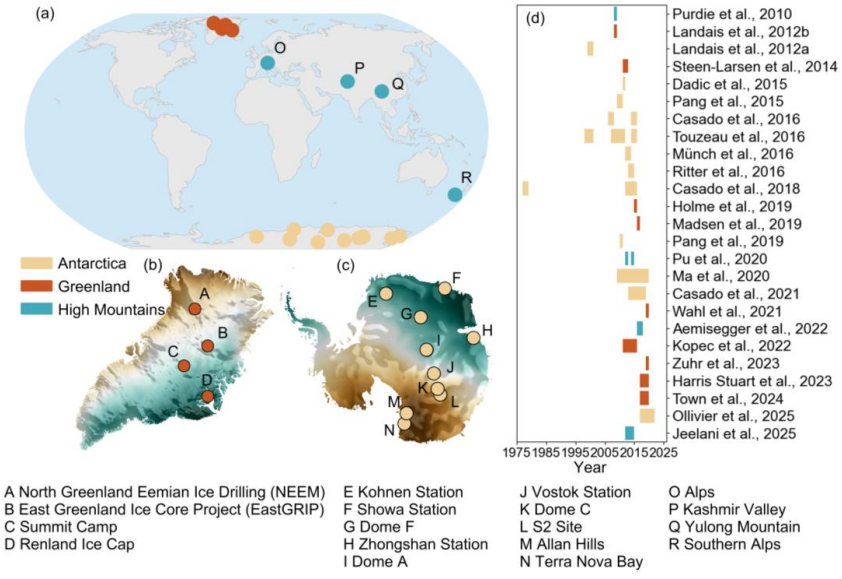


Figure 2. Spatial distribution of major observation regions for stable water isotopes at the atmosphere–snow–ice interface in Antarctica, Greenland, and high-mountain areas. (a-c) Distribution of observation and sampling sites. (d) Temporal distribution of sample collection.

The transmission of isotopic signals from precipitation into ice-core records is governed by cryospheric processes and their interactions with other Earth system components. Due to the multiphase, multi-component, and highly nonlinear nature of this system, the specific roles and relative contributions of these processes and their interactions to the final isotope records remain poorly understood, particularly across regions. To bridge this gap, we present an integrated overview of physical and isotopic processes at the atmosphere–snow–ice interface, synthesising current observations, laboratory investigations, and numerical modelling. We review fundamental stable water isotopic theory and current methodologies (Section 2), analyse post-depositional isotopic evolution (Section 3), and conclude with critical priorities and future directions.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

2 Theory and methods

2.1 Stable water isotopic theory

Molecules composed of different isotopes exhibit systematic differences in thermodynamic properties, such as zero-point energy and vibrational frequency, which lead to isotopic fractionation during natural phase transitions (Craig, 1961). The fractionation factor α is commonly used to characterise isotopic fractionation occurring during phase transitions, which is defined as

$$\alpha = \frac{R_A}{R_B} \quad (1)$$

where R_A and R_B represent the isotopic compositions assigned to the initial reactant and resulting product, respectively. Lighter isotopes (^{16}O and H) preferentially partition into the gas phase, whereas the residual liquid or solid phases become enriched in heavier isotopes (^{18}O and D) (Zhang et al., 2005). Isotopic fractionation is classified into mass-dependent and mass-independent processes (Thiemens, 1999). The former category can be further subdivided into equilibrium and non-equilibrium fractionation, reflecting differences in the underlying thermodynamic and transport mechanisms (Craig, 1961).

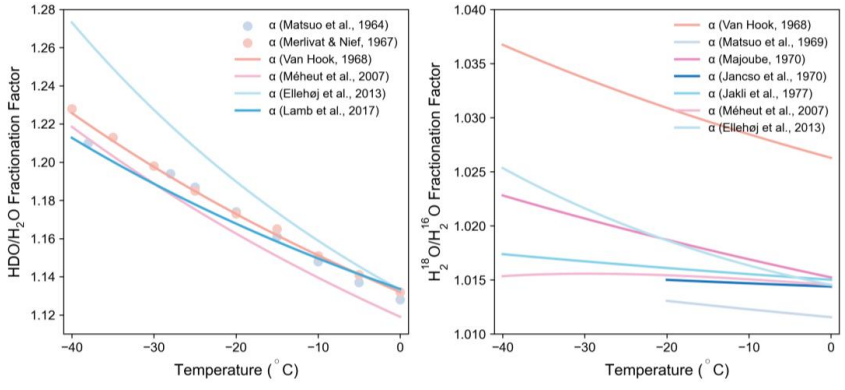


Figure 3. Temperature dependence of the gas–solid equilibrium fractionation factors of stable water isotopes. The left panel shows α for HDO/H₂O, and the right panel shows H₂¹⁸O/H₂¹⁶O, as reported in different studies.

Equilibrium fractionation refers to deviations in the distribution of different isotopes from an equiprobable distribution owing to differences in molecular structure and chemical bond properties (Knyazev et al., 1992). This occurs when the exchange rates of water molecules

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

between different phases are approximately equal, producing no net isotopic change (Urey, 1947). The fractionation factor is expressed as follows (Merlivat and Nief, 1967):

$$\alpha = \frac{e_{s,\text{light isotopes}}}{e_{s,\text{heavy isotopes}}} \tag{2}$$

where $e_{s,\text{ light isotopes}}$ and $e_{s,\text{ heavy isotopes}}$ represent the saturated vapour pressures of light and heavy isotopic water molecules, respectively. The gas–solid equilibrium fractionation factor (α_{eq}) reflects how stable water isotopes partition between the gas and solid phases at a given temperature (Figure 3) (Ellehoj et al., 2013; Jakli and Staschewski, 1977; Jancso et al., 1970; Majoube, 1970; Matsuo and Matsubaya, 1969; Merlivat and Nief, 1967).

Temperature is a critical factor influencing isotopic equilibrium fractionation. As temperature decreases, the equilibrium fractionation factor rises, leading to greater separation of light and heavy isotopes between different phases or substances. Moreover, temperature also influences the time required to attain equilibrium. Lower temperatures slow isotopic exchange and consequently increase the time needed to reach equilibrium (Ellehoj et al., 2013; Merlivat and Nief, 1967; Urey, 1947). The relationship between light and heavy isotope ratios can be described using the equilibrium fractionation factor and time, as expressed below:

$$R_s = R_0\left[1 + (\alpha_{\text{eq}} - 1)\exp\left(-\alpha_{eq}^{\frac{t}{\tau}}\right)\right] \tag{3}$$

where R_s and R_0 represent the isotope ratios at a given time and the initial time, respectively, t represents time, and τ represents the characteristic time scale.

Non-equilibrium fractionation can be divided into kinetic fractionation and diffusion fractionation. Kinetic fractionation refers to the deviation in isotopic composition between products and reactants arising from the different reaction rates of light and heavy molecules during an irreversible or unidirectional process (Li et al., 2024). The parameterised expressions for the kinetic fractionation factor α_k of stable water isotopes, calculated from first-principles molecular dynamics theory over 190–500 K, are given as follows (Hellmann and Harvey, 2020):

$$\alpha_k = \frac{S}{\frac{\alpha_{eq}(S-1)}{D_{r,x}} + 1} \tag{4}$$

$$D_{r,\text{HDO}} = 0.98258 - \frac{0.02546}{T^*} + \frac{0.02421}{(T^*)^{5/2}} \tag{5}$$

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

$$D_{r,H_2{}^{17}O} = 0.98284 + \frac{0.003517}{(T^*)^{1/2}} - \frac{0.001996}{(T^*)^{5/2}} \quad (6)$$

$$D_{r,H_2{}^{18}O} = 0.96671 + \frac{0.007406}{(T^*)^{1/2}} - \frac{0.004861}{(T^*)^{3}} \quad (7)$$

where T represents temperature (K) and $T^* = T/100K$, S represents relative humidity, and $D_{r,x}$ represents the diffusivity ratio of x relative to H_2O.

Differences in molecular diffusivity constitute the fundamental mechanism of diffusion fractionation (Shang et al., 2024). The fractionation factor can be expressed as:

$$\alpha = \sqrt{\frac{m}{m^*}} \quad (8)$$

where m^* and m represent the masses of the light and heavy isotope molecules (kg), respectively. When diffusion occurs within an ambient air matrix, the fractionation factor must account for the molecular mass of air (m_{air}), which can be expressed as:

$$\alpha = \sqrt{\frac{m^*(m + m_{air})}{m(m^* + m_{air})}} \quad (9)$$

d-excess is a reliable indicator for assessing non-equilibrium effects and provides information on kinetic fractionation processes, including evaporation and local water vapour mixing (Dansgaard, 1964). Linear *d*-excess is used to quantify deviations of $\delta^{18}O$ and δD relationships from the Global Meteoric Water Line (GMWL), which can be expressed as:

$$d\text{-}excess = \delta D - 8\delta^{18}O \ (‰) \quad (10)$$

In this equation, 8 denotes the empirical slope parameter of the GMWL.

It has been shown that the scale effect associated with linear *d*-excess may result in a loss of explanatory power under conditions of extreme isotopic depletion (Dütsch et al., 2017). To avoid biases in linear *d*-excess in extremely dry and cold regions, exponential deuterium excess ($\tilde{d}$-excess) has been proposed (Uemura et al., 2012) and is expressed as follows:

$$\tilde{d}\text{-}excess = \frac{R^D}{(R^{^{18}O})^{8.8}} - 1 \ (‰) \quad (11)$$

where the coefficient of 8.8 represents the empirical parameter reflecting the meteoric δ^2H–$\delta^{18}O$ relationship in logarithmic or exponential space.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

2.2 Field observation methods

2.2.1 Observations at the atmosphere–snow interface

Accurate field-based measurements of isotope fluxes at the atmosphere–snow interface provide a crucial foundation for studying gas–solid isotopic fractionation and analysing its mechanisms (Figure 4, Table 1). The atmospheric isotope component can be determined directly from a dual-end-member mixing model based on the Keeling plot (KP) approach (Good et al., 2012). In this framework, laser spectroscopy is used to obtain the isotopic composition and concentration at multiple heights or from continuous single-point measurements and to estimate isotopic variations in locally sourced vapour through the intercept of a linear regression (Hu et al., 2021). However, the application of the KP requires several underlying assumptions, including two key conditions. The first is that vapour isotopes are determined only by mixing between background concentrations in the local source and the atmospheric boundary layer. The second is that the isotopic contributions from these components remain constant. Under complex or non-stationary atmospheric conditions, however, regression analyses based on statistical extrapolation may introduce substantial uncertainty (Yuan et al., 2020).

Atmospheric isotopic composition can also be estimated indirectly through water vapour flux, which characterises the exchange of moisture across the snow–atmosphere boundary. When combined with real-time isotope measurements, it enables quantitative determination of air–solid isotope fluxes. At present, methods commonly used to observe water vapour flux include evaporation pans, eddy covariance (EC), flux-gradient (FG) method, relaxed eddy accumulation (REA), and one-camera time lapse structure-from-motion (O-T-SfM).

The evaporation pan method is a traditional approach for measuring water vapour flux. In its standard form, the pan is packed with undisturbed natural snow and re-weighed at fixed intervals, so that the vapour flux can be derived from the measured changes in snow mass and density. Owing to its low cost and simple setup, the method remains useful for short campaigns, although the resulting records are neither spatially nor temporally continuous and respond strongly to local microclimate

Based on the turbulent energy cascade, EC allows direct quantification of interactions of vapour, energy, and momentum occurring at the snow–atmosphere boundary (Ma et al., 2024). EC determines water vapour flux through analysis of the covariance of high-frequency water

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

vapour concentrations with vertical wind velocity (Good et al., 2012), which places demanding requirements on measurement precision and temporal resolution. Recent developments in laser technology have enabled continuous measurements of isotopic vapour mixing ratios at high sampling frequencies (~10 Hz), allowing direct estimation of isotopic vapour fluxes (Griffis et al., 2010). Field observations require high-precision, high-temporal-resolution sensors and instruments with long-term stability. Consequently, such measurements are costly and require cospectral frequency-loss corrections to ensure accurate flux quantification (Wahl et al., 2021).

The FG method, grounded in Monin-Obukhov similarity theory (MOST), assumes that turbulent diffusivities can be parameterised by mean gradients and stable flux–gradient similarity relationships (Zhao et al., 2020). This gradient-based approach relies on concurrent observations collected at multiple heights. Therefore, paired analysers should be employed with regular calibration and maintenance to ensure high-precision paired measurements (Wahl et al., 2021).

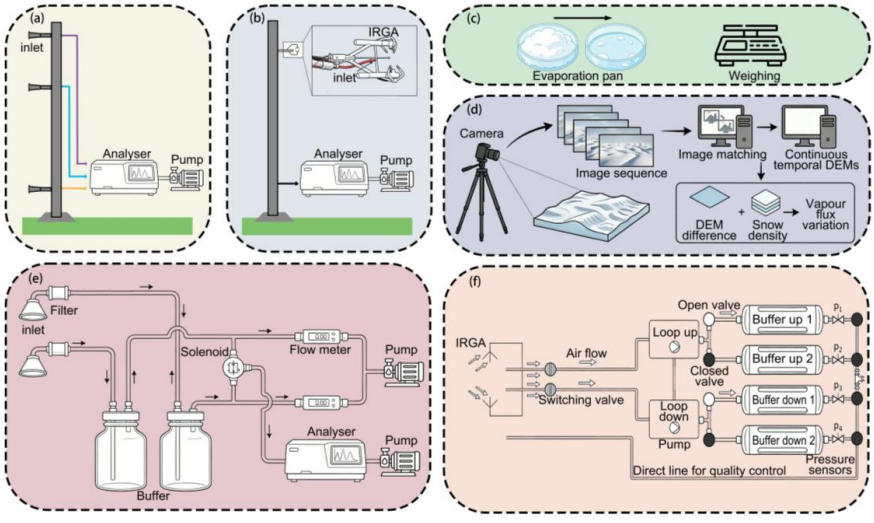


Figure 4. Schematic diagram of the primary observation methods for stable water isotopes at the atmosphere–snow–ice interface. (a) Keeling plot method (modified from Good et al., 2012). (b) Eddy covariance method (modified from Wahl et al., 2021). (c) Evaporation pan method. (d) One-camera time lapse structure-from-motion method (modified from Liu et al., 2024). (e) Flux-gradient method (modified from Zhao et al., 2020). (f) Relaxed eddy accumulation method (modified from Kunz et al., 2025).

REA employs a conditional sampling strategy to quantify turbulent trace-gas fluxes with

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

slow-response instrumentation. Vertical air motion is determined from high-frequency wind-speed signals. The sampling system consists of two independent inlet lines fitted with rapid-response valves and loop assemblies, enabling air collection into separate aluminium cylinder sets that represent updraught and downdraught air masses, respectively. This method indirectly infers vapour isotopic composition from concentration differences and statistical parameters. When atmospheric conditions deviate from steady state or lack sufficiently developed turbulence, the derived estimates can be subject to considerable bias (Kunz et al., 2025).

O-T-SfM acquires time-lapse images of the snow surface at fixed positions and intervals and generates continuous temporal digital elevation models (DEMs) through image matching. When combined with snow density, this approach allows estimation of spatiotemporal variations in vapour flux. The method can distinguish between blowing snow and surface sublimation based on snow-surface features captured in the imagery and can also provide information on the spatial heterogeneity of surface snow within the photographic domain (Kochanski et al., 2019). Because O-T-SfM relies on snow-surface bedforms for feature matching, it performs best under snowfall-free conditions and where bedforms are abundant but is limited in capturing melt processes and smooth fresh snow surfaces (Liu et al., 2024).

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

Table 1. Primary advantages and limitations of field observation methods at the atmosphere–snow interface

Measurement target	Methods	Main advantage	Main limitation	Typical field conditions
Atmospheric isotopic composition	KP	Conceptually simple; requires relatively simple instrumentation	Relies on simple models and assumptions; sensitive to atmospheric non-stationarity and regression uncertainty	Relatively homogeneous snow and ice surfaces under stable atmospheric conditions
Water vapour flux	Evaporation pans	Simple and inexpensive; easy to operate	Discontinuous measurements; sensitive to local microclimate	Short-term or low-cost field campaigns
	EC	High temporal resolution; enables direct measurement of turbulent vapour flux	Expensive and technically demanding; stringent requirements for sensor precision; requires corrections	Continuous high-resolution monitoring of snow–atmosphere exchange
	FG	Relates vertical concentration gradients to fluxes; compatible with relatively slow-response analysers	Requires measurements at multiple heights and regular calibration of paired instruments	Stable or moderately turbulent atmospheric boundary layers
	REA	Compatible with relatively slow-response analysers	Requires high-frequency vertical wind measurements and rapid switching valves; sensitive to non-stationary atmospheric conditions	Field observations using relatively slow-response analysers
	O-T-SfM	Non-contact and relatively low-cost; provides spatially distributed measurements; can distinguish blowing-snow redistribution from surface sublimation	Requires identifiable snow surface features; limited during snowfall or melt; requires photogrammetric processing and accurate georeferencing	Field observations requiring spatially resolved snow surface changes to distinguish sublimation from blowing-snow redistribution

2.2.2 Observations at the snow–ice interface

To investigate the transfer and preservation of isotopic signals from snowpack to glacial ice, integration of field sampling and laboratory analysis is required. Layered samples of snowfall and snowpack should be collected and immediately sealed in specialised airtight bags. After melting, the samples must be transferred to airtight bottles for subsequent laboratory

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

isotope analysis.

In addition, physical parameters of the snowpack should be measured simultaneously during sampling. Snow density is determined by collecting snow samples of known volume and weighing (Kopec et al., 2022). Infrared reflectance instruments, such as IceCube, are commonly employed to determine the snow specific surface area (SSA) (Harris Stuart et al., 2023). It is then possible to empirically relate the physical structure of snow and ice to their isotopic composition, thereby providing insights into cryosphere isotopic processes.

2.3 Laboratory experiments

Isotopic processes at the atmosphere–snow–glacier interface are influenced by multiple factors, making quantitative control of variables in field observations challenging. This complexity consequently limits understanding of certain processes. Laboratory experiments of key post-depositional processes have therefore become an effective way to clarify the underlying mechanisms. Previous studies have primarily focused on processes such as sublimation, snow metamorphism, and wind-driven snow.

2.3.1 Surface modification

Sublimation–deposition processes can substantially modify the isotopic signature preserved in snow. Controlled cold-box experiments demonstrated that sustained sublimation under dry-air conditions induces pronounced isotopic changes in snow (Figure 5a). Under controlled airflow and temperature, isotopically uniform samples became progressively enriched in $\delta^{18}O$ (by up to 8‰), while d-excess decreased by up to 25‰. The strongest modification occurred at the snow surface, and the imprint penetrated to approximately 3 cm depth within 4–6 days and sublimation-driven fractionation can thus rework the near-surface snowpack on timescales of days (Hughes et al., 2021). Such post-depositional modifications may subsequently be preserved in ice-core records.

Experiments using the Experimental Multiphasic Atmospheric Simulation Chamber (CESAM) have demonstrated the applicability of diffusion–deposition theory under controlled conditions, while also revealing limitations of classical models in estimating isotopic fractionation during deposition at very low temperatures. By inducing vapour deposition onto a cylindrical copper cold trap maintained below the chamber temperature (Figure 5b), the experiments showed a pronounced kinetic effect during ice deposition. Notably, ^{17}O-excess exhibited a stronger response to changes in kinetic fractionation intensity than d-excess (Casado

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

et al., 2016). These findings provide experimental constraints for evaluating diffusion–deposition models and for interpreting isotopic signatures formed under strongly kinetic deposition conditions.

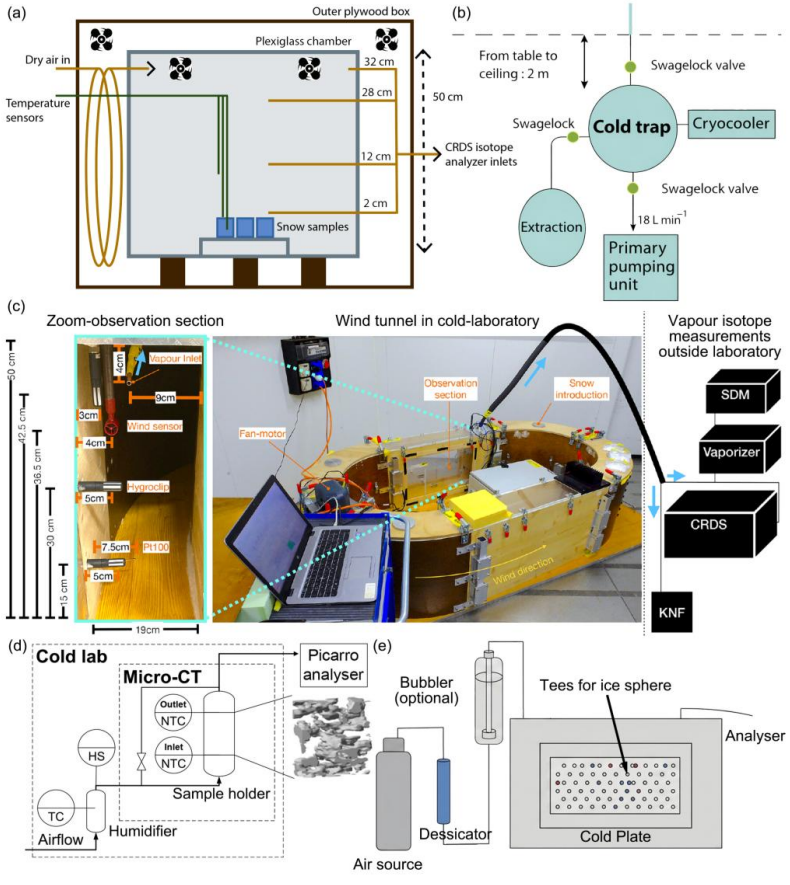


Figure 5. Schematic diagram of the primary laboratory experiments for stable water isotopes. (a–c) Surface modification processes, including the cold-box experiment (modified from Hughes et al., 2021), CESAM chamber experiment (modified from Casado et al., 2016), and ring wind tunnel experiment (modified from Wahl et al., 2024). (d) Cold-laboratory experiments on snowpack diffusion processes (modified from Ebner et al., 2017). (e) Solid ice sphere experiments on ice storage processes (modified from Bellagamba et al., 2024).

Moreover, wind-driven processes can directly alter the physical properties and isotopic characteristics of surface snow. Cold-laboratory ring wind tunnel experiments revealed coupled changes in snow microstructure and stable water isotopic composition during airborne snow transport. Under sustained wind conditions, snow was transported through both suspension and

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

saltation under different temperature conditions (Figure 5c), resulting in substantial changes in specific surface area (SSA) and isotopic composition.

Over 1.25–2.5 h of transport, SSA declined to 35–70% of its initial value, and within 3 h the isotopic signal shifted by up to +1.47‰ in $\delta^{18}O$, ±5.7‰ in δD, and −6.1‰ in d-excess (Wahl et al., 2024). These results demonstrate that wind-driven snow transport can induce concurrent physical and isotopic modifications of surface snow, providing important evidence for improving wind-blown snow models and interpreting isotopic signals preserved in polar ice cores.

2.3.2 Snowpack diffusion

Diffusion within the snowpack further alters the vertical distribution of stable water isotopes. Controlled experiments quantified isotope diffusion length and diffusivity using layered snow samples with contrasting isotopic compositions and different thicknesses. The experimental configuration was designed to minimise sampling-related uncertainties. The calculated and experimentally derived diffusion parameters differed by approximately 15% (Van Der Wel et al., 2011). These results provide an experimental basis for assessing the effects of differential isotope diffusion on ice-core isotope records.

Moreover, the snowpack is not a closed system but continuously exchanges stable water isotopes with the atmosphere, with post-depositional isotopic exchange strongly modulated by thermal conditions. Controlled cold-laboratory experiments demonstrated that isotope and temperature disequilibrium drives substantial $\delta^{18}O$ changes in snow (Figure 5d). Their experiments showed that positive temperature gradients were associated with enhanced sublimation, whereas negative gradients favoured net deposition, allowing the contrasting effects of these processes on isotopic evolution to be distinguished. Critically, the introduction of an additional thermal gradient within the snow sample induced complex interactions that enhanced recrystallisation and further modified the $\delta^{18}O$ signature (Ebner et al., 2017). These findings provide a physical basis for interpreting ice-core isotope records and refining climate models.

2.3.3 Ice storage

The pore structure of ice can still influence fractionation during sublimation. Experiments with suspended solid ice spheres showed that limited levels of ice porosity enhance the effective diffusivity of isotopes, thereby revealing new physical mechanisms relevant to interpretation of

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

ice-core isotope records (Figure 5e). In controlled experiments with known isotopic compositions, small but statistically significant fractionation was observed during ice sublimation. Importantly, the results demonstrate that even after pore-space closure, entrapped air bubbles can sustain effective isotope diffusion within the ice matrix (Bellagamba et al., 2024). This mechanism helps resolve why fractionation persists in low-porosity ice and provides a new physical basis for interpreting subtle isotopic variations in deep ice cores.

2.4 Model simulations

Field observations remain constrained by limited spatial extent and observation duration (Carroll et al., 2022), whereas laboratory experiments, although enabling quantitative investigation under controlled conditions, cannot fully reproduce the complex interactions among environmental factors in natural snow and ice (Hughes et al., 2021). To comprehensively explore the influence and relative contribution of different post-depositional processes to ice-core isotope records, isotope-enabled model simulations have therefore become essential.

Isotope-enabled atmospheric general circulation models, such as LMDZ5Aiso, LMDZ6Aiso, iCESM1, ECHAM5-JSBACH-wiso, ECHAM5-wiso, and ECHAM6-wiso, provide the large-scale climatic background and the primary isotopic forcing, including isotopic compositions of moisture, cloud water, and precipitation (Werner et al., 2011), which serve as upper-boundary conditions for snow isotope models that simulate post-depositional changes (Dutrievoz et al., 2025; Goursaud et al., 2018; Ollivier et al., 2025).

Statistical and empirical models are effective for describing simple patterns of snow isotopic composition. The Snow Isotopic Signal Generator (SISG) simulates snowpack stratification by sequentially incorporating individual snowfall events until a predefined surface-depth threshold is reached. The model indirectly assesses the combined effects of processes such as diffusion, sublimation, and snow metamorphism by comparing simulated snow isotopic compositions with observed values (Figure 6a). However, models based solely on precipitation inputs fail to fully reproduce the observed isotopic variability and key features (Ollivier et al., 2025). Based on a continuous observational dataset collected at the East Greenland Ice Core Project (EastGRIP) during the 2017–2019 summer seasons, an empirical decay model was developed to characterise rapid SSA reduction associated with snow metamorphism, providing a quantitative basis for linking SSA evolution with isotopic fractionation. Because it was derived from fits to observational data, the model has spatiotemporal limitations and does not incorporate more complex mechanisms of snow

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

metamorphism (Harris Stuart et al., 2023).

To elucidate the fundamental processes occurring at the snow–atmosphere interface and the mechanisms governing isotopic exchange, some studies have proposed idealised models. Within a one-dimensional box-model framework, the gas–solid interaction processes were decomposed into three key physical reservoirs: the free-atmospheric layer, the boundary layer, and the surface snow layer (Ma et al., 2024; Madsen et al., 2019). The model quantified vapour and isotope fluxes between these adjacent units to simulate interfacial processes. This idealised model reproduces the core characteristics of isotopic variability during atmosphere–snow vapour exchange (Figure 6b). However, its applicability to complex environmental conditions and long-term processes is limited by its simplifying assumptions and by the lack of explicit representation of external air-mass mixing (Ritter et al., 2016).

Furthermore, a numerical mass-balance snow-surface model has been developed to reproduce the isotopic changes of the uppermost snow layer (0–0.5 cm) and to compare the results with observations. This model incorporates isotopic fractionation associated with sublimation and deposition, as well as diffusion within the snowpack, while omitting additional processes including snow metamorphism and wind-driven snow (Wahl et al., 2022). Building on this approach, the SNOWISO model was developed with two components: a surface cell of variable height and an underlying snowpack represented by grid cells of constant spacing (Figure 6c). The model incorporates precipitation and isotopic input, isotopic fractionation during vapour exchange with the atmosphere, isotope diffusion between all grid cells, and release of a new snowpack cell from the surface cell, thereby allowing dynamic simulation of isotopic processes at the atmosphere–snow interface (Dietrich et al., 2023).

The isotope-enabled Community Firn Model (Iso-CFM) combined with an ice-physics model was designed to simulate isotope diffusion and fractionation processes within the snowpack. Iso-CFM calculates attenuation and smoothing of isotopic signals using Fick's second law and a Gaussian filter to estimate diffusion length at different depths and to quantify the intensity of isotope diffusion processes (Stevens et al., 2020). Iso-CFM requires only temperature and accumulation as inputs, without prior information on the isotopic composition of precipitation (Figure 6d). Consequently, its results are not directly affected by factors that may bias the $\delta^{18}O$ signal, such as kinetic effects, atmospheric inversions, or precipitation intermittency. Nevertheless, sufficiently high-resolution isotope measurements remain crucial for accurate model application, while adjustments for ice-flow thinning and ice diffusion are

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

still required (Gkinis et al., 2021). The ice-physics model, based on a Rayleigh distillation framework and incorporating solid-phase diffusion within ice grains, describes isotopic fractionation in freshly deposited snow. However, uncertainty in the results remains substantial because parameters such as the solid-phase diffusion coefficient of ice particles and the sublimation rate are difficult to measure accurately (Hu et al., 2022).

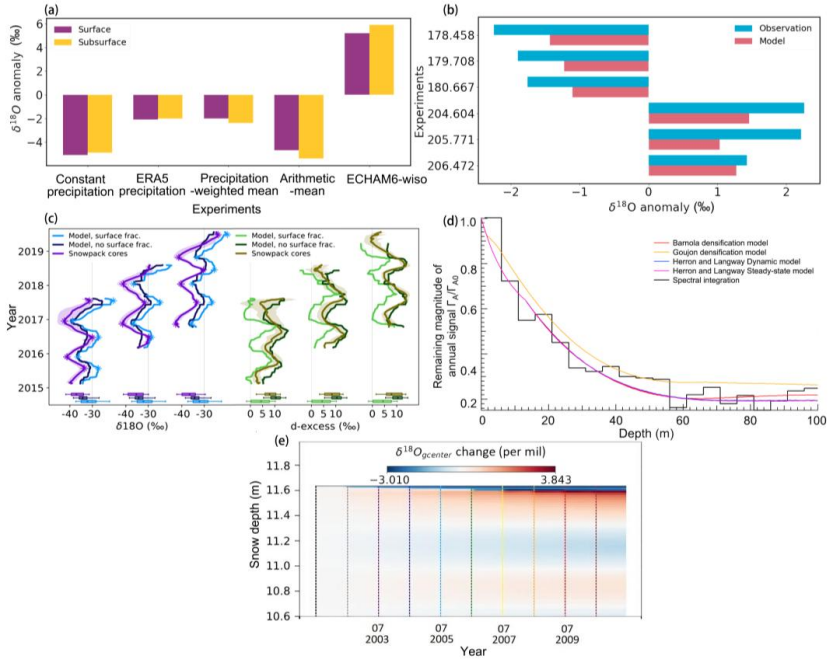


Figure 6. Evaluation of major model simulations against observations for stable water isotopes. (a) $\delta^{18}O$ anomalies of surface and subsurface snow simulated by the SISG model (modified from Ollivier et al., 2025), using respective daily precipitation inputs. (b) $\delta^{18}O$ anomalies at different depths (cm) simulated by the one-dimensional box-model framework model (modified from Madsen et al., 2019). (c) $\delta^{18}O$ and *d*-excess simulated by the SNOWISO model (modified from Dietrich et al., 2023), compared with three-year averaged snowpack core observations (2017–2019) at the EastGRIP site. (d) Attenuation of the annual $\delta^{18}O$ signal simulated by the Iso-CFM (modified from Gkinis et al., 2021), using different firn densification models and the annual $\delta^{18}O$ signal attenuation derived from spectral peak analytical integration. (e) $\delta^{18}O$ anomalies of snowpack simulated by SURFEX–Crocus v8.0 (modified from Touzeau et al., 2018), driven by vapour phase diffusion over 10 years.

Systematic simulation of isotopic processes at the atmosphere–snow–ice interface using integrated, coupled energy–mass balance models provides an effective tool for understanding isotope-exchange mechanisms. SNOWPACK is a one-dimensional, physically based snowpack

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

model that integrates sublimation, deposition, water transport, and snow metamorphism (Wever et al., 2014). These coupled processes allow the model to capture complex isotopic variations. However, model complexity and the high requirements for input data limit its applicability in low-accumulation environments and during snowmelt periods (Aemisegger et al., 2022). SURFEX–Crocus v8.0 is a detailed one-dimensional snowpack model that represents the snow cover as multiple layers with dynamically evolving thicknesses (Figure 6e). The model includes modules for snowfall, updating snow layering, metamorphism, snow compaction, wind drift, snow albedo, latent and sensible energy and mass fluxes, the vertical snow temperature, and surface sublimation and deposition. This model implements two types of vapour-phase diffusion controlled by isotopic and thermal gradients. The isotopic composition at the grain surface evolves through exchange with two compartments: sublimation–deposition with vapour in the pore space and solid diffusion within the grain centre. However, processes such as wind pumping, reworking of the uppermost snow layer, and hoar deposition are not currently included in SURFEX–Crocus v8.0 (Touzeau et al., 2018).

3 Post-depositional processes and water isotopic evolution

Post-depositional processes encompass the period from snow deposition to its densification into glacial ice, during which a range of physical processes modify the original isotopic signals. Based on field observations and laboratory experiments, several major isotopic exchange processes have been identified. At the snow–atmosphere interface, isotopic exchange processes involve interactions between the surface snow and the overlying atmosphere, encompassing two processes: turbulent mixing and sublimation–deposition. The isotopic characteristics of the snowpack continue to evolve during firnification. In the solid phase, this primarily involves solid-phase isotope diffusion. In the gas phase, isotopic gradient diffusion occurs, whereas snow metamorphism takes place at the snowpack solid–gas interface. Furthermore, post-depositional processes affecting snow isotopic compositions may also be induced by external forces, including wind-driven and infiltration–refreeze processes (Table 2).

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

444 Table 2. Favourable conditions and primary controls of main post-depositional processes

Main post-depositional processes	Interaction interface	Favourable conditions	Primary controls
Sublimation–deposition	Snow–atmosphere interface	Cold and dry snow surfaces exposed to the atmosphere	Temperature, relative humidity, wind speed, and surface energy balance
Turbulent mixing		Vertical wind-exposed snow surfaces with strong near-surface turbulence	Wind speed, topography, surface friction, wind shear, and atmospheric stability
Solid-phase diffusion	Snowpack solid phase	Snow crystal interiors and dense snow or ice layers	Snow temperature, grain size, and molecular diffusivity
Snow metamorphism	Snowpack solid–gas interface	Snowpacks under strong temperature gradients or near-melting conditions	Snow temperature gradient, SSA, snowpack pore structure, and water content
Isotopic gradient diffusion	Snowpack gas phase	Snowpacks with pronounced isotopic gradients	Snowpack isotopic gradient, porosity, tortuosity, density, and effective diffusion coefficient
Wind-driven processes	Snow surface and snowpack	Strongly wind-exposed and porous snowpacks	Wind speed and direction, surface roughness, snow-grain mobility, and snowpack pore structure
Infiltration–refreeze	Snowpack liquid–solid interface	Snowpacks containing liquid water, particularly in the ablation zones of ice sheets or glaciers and during seasonal snowmelt	Snow temperature, snowpack permeability, water content, and subsurface ice crusts

445

446 **3.1 Sublimation–deposition**

447 Sublimation initially produces a layer of saturated vapour at the surface of snow crystals
448 (Figure 7) (Jambon-Puillet et al., 2018; Touzeau et al., 2018), which is controlled mainly by
449 temperature and vapour pressure and typically accelerates as temperature rises and vapour
450 pressure decreases. When atmospheric vapour pressure is lower than that at the snow-crystal
451 surface, a vapour-pressure gradient drives the continuous transport of water molecules from the
452 snow-crystal surface into the atmosphere. This causes the vapour layer at the snow-crystal
453 surface to become unsaturated, disrupting the original gas–solid equilibrium and sustaining
454 sublimation. Conversely, when atmospheric vapour pressure exceeds that at the snow-crystal

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

surface, deposition occurs through the transfer of vapour and, under some conditions, liquid droplets to the surface. Furthermore, sublimation–deposition affects the near-surface energy balance and indirectly influences temperature and humidity, which in turn feed back on this process (Wahl et al., 2024). The rate of sublimation–deposition can be quantified by the latent heat flux, according to the following relationship (Lundquist et al., 2024):

$$Q = L_s \cdot E \tag{12}$$

where E, Q, and L_s represent the sublimation rate (kg m^{-2} s^{-1}), latent heat flux (W m^{-2}) and latent heat of sublimation (J kg^{-1}), respectively. A positive latent heat flux indicates sublimation, during which snow crystals lose mass. In contrast, a negative value signifies deposition, whereby vapour is added to the snowpack.

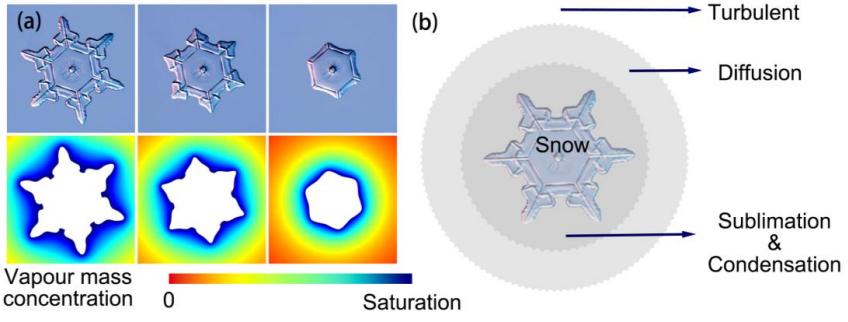


Figure 7. Schematic depiction of snowflake sublimation and gas–solid interactions at the crystal surface during sublimation and deposition (modified from Jambon-Puillet et al., 2018). (a) Time sequence illustrating the evaporation of a snowflake. Shown are snapshots from a three-dimensional reconstruction of the same snowflake based on a constant thickness assumption and representative model parameters. Colours indicate the water vapour concentration field, with values spanning from saturation conditions (blue) to zero (red). (b) Schematic depiction of snowflake and surrounding structure.

Methods used to calculate latent heat flux include eddy covariance (EC), the bulk aerodynamic flux method (BF), the Bowen ratio–energy balance method (BR), and the aerodynamic profile method (AP) (Garratt, 1994; Hock, 2005; Sexstone et al., 2016). Their underlying formulations can be expressed as follows:

$$Q_{E-\mathrm{EC}} = \rho L_s \overline{w'\,q'} \tag{13}$$

where ρ represents air density (kg m^{-3}) and $\overline{w'\,q'}$ represents the covariance between vertical wind speed (w') and specific humidity (q').

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

$$Q_{E\text{-}BF} = \rho L_s C_E u_r (q_s - q_r) \quad (14)$$

where q_s and q_r represent surface and reference height specific humidity. C_E represents the latent heat exchange coefficient, and u_r represents wind speed at the reference level.

$$Q_{E\text{-}AP} = -\rho L_s C_E \Delta u \, \Delta q \quad (15)$$

where Δu and Δq represent the wind-speed gradient and the specific-humidity gradient between measurement heights, respectively.

$$Q_{E\text{-}BR} = \frac{Q^* - \frac{dU}{dt}}{1 + \beta} \quad (16)$$

where Q^* represents net radiative flux (W m^{-2}), dU/dt represents the snowpack thermal energy storage term (W m^{-2}), and β represents the Bowen ratio.

Isotopic exchange occurring between snow crystals and the saturated vapour layer during sublimation–deposition is generally considered an equilibrium fractionation process. Molecular diffusion between the snow-crystal surface and the near-surface atmosphere causes non-equilibrium kinetic fractionation. Light isotopes preferentially diffuse from the saturated vapour layer into the atmosphere, whereas heavy isotopes gradually accumulate, thereby increasing the heavy isotope content of snow crystals. This process is governed primarily by relative humidity and wind speed in the near-surface atmosphere, which together regulate the intensity of vapour diffusion and determine the magnitude of kinetic fractionation (Hu et al., 2022). At Dome C, the net snow sublimation rate was approximately 3.1–3.7 mm yr^{-1}, accounting for 12–15% of the annual surface mass balance (Ollivier et al., 2025). Such high-frequency, shallow variations in isotopic composition are likely to weaken the stability of the isotope-temperature relationship on seasonal and annual timescales.

3.2 Turbulent mixing

At the interface between the near-surface layer and the atmospheric boundary layer, water molecules are transported primarily by turbulent mixing, which is much faster than molecular diffusion (Garratt, 1994). Through turbulent mixing, water molecules can be dispersed non-selectively into the atmosphere over a broad spatial range, thereby altering near-surface humidity and continuously driving sublimation or deposition processes (Wang et al., 2019). Turbulence is influenced by multiple factors, including wind speed, temperature, humidity, topography, surface friction, atmospheric stability, wind shear, and larger-scale weather systems

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

(Wanner et al., 2024). During the day, within the boundary layer, rising temperatures enhance convection and intensify turbulent mixing (Casado et al., 2018). At night, as temperatures fall, turbulent activity weakens, the boundary layer becomes more stable, and turbulent diffusion is reduced. The turbulent diffusion coefficient $K_H(z,t)$ describes the diffusion rate of substances in turbulent flow. Within the framework of Monin-Obukhov similarity theory, it can be expressed as follows (Madsen et al., 2019):

$$K_H(z,t) = \frac{ku^* z' \left(1 - \frac{z'}{h}\right)^{1.5}}{\phi_m\left(\frac{z}{L}\right)\phi_H\left(\frac{z}{L}\right)} \tag{17}$$

where k denotes the von Kármán constant, u^* is the friction velocity, z corresponds to the vertical distance above the snow surface, ϕ_m and ϕ_H represent nondimensional stability functions, L represents the Monin-Obukhov length, and h corresponds to the total depth of the atmospheric boundary layer.

Turbulent diffusion is a macroscale eddy-transport process in the atmosphere and therefore does not directly cause stable water isotopic fractionation (Kim and Lee, 2011; Lee et al., 2009). By ventilating the interfacial layer, turbulence steepens the vapour-pressure gradient that drives molecular diffusion away from the snow surface (Bintanja, 2001), thereby strengthening fractionation during the molecular-diffusion stage. Yet when mixing becomes vigorous enough to homogenise the near-surface vapour, the same process erodes the gradient and weakens fractionation. Turbulence additionally modulates snow-surface temperature through the surface energy balance, changing the equilibrium fractionation factor that governs interfacial exchange. Continuous vapour isotope measurements reveal coupled changes in surface-snow and atmospheric vapour isotopic signatures with amplitudes of up to 10‰ over 5 days at the Kohnen Station in East Antarctica, such that isotopic signals may reflect local processes more strongly than weather-scale variability or precipitation inputs (Ritter et al., 2016).

3.3 Solid-phase diffusion

Solid-phase diffusion is based on the migration of water molecules through the crystal lattice via a vacancy mechanism in a self-diffusion process that has no preferred direction and is very slow (Touzeau et al., 2018). The Arrhenius law describes the temperature (T) sensitivity of the coefficient (D).

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

$$D = D_0 \exp\left(-\frac{Q}{RT}\right) \quad (18)$$

where R denotes the universal gas constant (8.314 J mol^{-1} K^{-1}), D_0 represents the maximum diffusion coefficient (m^2 s^{-1}), and Q represents the activation energy for diffusion (J mol^{-1}). Based on this relationship, the empirical expression of molecular self-diffusion in ice can be expressed as (Johnsen et al., 2000; Ramseier, 1967):

$$D_{\text{ice}} = 3.96 \cdot 10^4 \exp\left(-\frac{7273}{T}\right) \quad (19)$$

The diffusion time of molecules in the solid phase can be estimated using the characteristic-time expression derived from Fick's second law:

$$\Delta t_{\text{sol}} = \frac{R_{\text{m}}^2}{D_{\text{ice}}} \quad (20)$$

where R_{m} is the mean grain radius (m). Theoretically, when the temperature is -20°C, the snow-crystal diffusion coefficient is 1.32×10^{-8} m^2 yr^{-1}, implying that water molecules require 0.76 yr to traverse a grain with a radius of 0.1 mm, which corresponds to the timescale of solid-phase diffusion. In addition, snow crystals may contain tiny, closed air bubbles formed during crystal growth, which create enhanced diffusion pathways. Water molecules lost from the snow-crystal surface through sublimation can therefore be replenished more rapidly via pore pathways than through diffusion within the crystal lattice (Bellagamba et al., 2024). The solid-phase diffusion experiment demonstrated that the residual ice became progressively enriched in heavy isotopes. When roughly half of the sphere's mass had been lost, $\delta^{18}O$ and δD values had increased by approximately 0.5‰ and 3.5–5.5‰, respectively (Bellagamba et al., 2024).

The isotopic composition of snow crystals is heterogeneous, and their microstructure can be approximated as a concentric two-layer structure comprising an inner core and an outer shell (Touzeau et al., 2018). A degree of isolation exists between the inner core and the outer shell, and this separation intensifies as the grain grows. These layers may also form at different times or under different environmental conditions (Lu and DePaolo, 2016). Non-directional solid-phase diffusion homogenises the isotopic composition within individual snow crystals.

3.4 Snow metamorphism

Snow exhibits a bicontinuous porous structure composed of a fully connected network of snow crystals and air-filled pore space (Löwe et al., 2011), and its microstructure determines

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

properties such as strength and albedo. Snow metamorphism refers to the continuous recrystallisation of the snow microstructure and is classified into dry-snow and wet-snow metamorphism depending on whether liquid water is present (Moure and Fu, 2024).

3.4.1 Dry-snow metamorphism

Dry-snow metamorphism involves substantial structural changes in snow crystals through rapid sintering and recrystallisation, resulting in altered physical properties. This phenomenon can be divided into isothermal metamorphism, temperature-gradient metamorphism, and variable temperature-gradient metamorphism.

Isothermal snow metamorphism refers to the evolution of snow microstructure towards thermodynamic equilibrium under constant-temperature conditions. Differences in the curvature of snow crystals lead to slight variations in temperature and equilibrium vapour pressure (Colbeck, 1983; Legagneux and Domine, 2005). Convex points have higher equilibrium vapour pressure and sublimate more readily, whereas concave points preferentially deposit vapour (Aemisegger et al., 2022; Löwe et al., 2011). Under these conditions, the curvature effect is the dominant factor, driven by the reduction in surface free energy at the complex ice–air interface (Ebner et al., 2017). Through microscale sublimation–diffusion–deposition processes, isothermal snow metamorphism causes snow crystals to become larger and smoother, together with a gradual reduction in SSA, while the density of the snowpack remains essentially constant (Ebner et al., 2015). Isotopic fractionation in snow layers under isothermal conditions is limited in magnitude and can be approximated as an equilibrium process. Lighter isotopes at convex points are more likely to enter the vapour phase. Through short-range diffusion, these molecules are redeposited at concave points, accompanied by further isotopic fractionation, during which heavier isotopes are preferentially incorporated into the solid phase.

In practice, temperature gradients are common within snow layers, driven by solar heating and radiation cooling (Bartels-Rausch et al., 2014). Temperature-gradient metamorphism may completely renew snow crystals, meaning that all water molecules within a given crystal are cycled through sublimation–deposition processes (Pinzer et al., 2012). These processes are driven by vapour-pressure gradients that develop in response to temperature differences across the pore space. Vapour produced by sublimation in warmer snow layers, where vapour pressure is higher, diffuses through the pore space down and deposits in colder snow layers. As a result, grains at warmer surfaces tend to develop rounded morphologies, whereas colder surfaces

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

favour the formation of faceted grain structures (Calonne et al., 2014). Under such temperature gradients, isotopic fractionation in snow layers constitutes a kinetic, non-equilibrium process. Lighter isotopes preferentially sublimate from warmer surfaces and diffuse towards colder regions, where heavier isotopes are preferentially deposited. This drives systematic isotopic redistribution along the gradient, with fractionation that is more pronounced and spatially more extensive than under isothermal conditions (Pinzer and Schneebeli, 2009).

Diurnal variations in the temperature gradient in shallow snow layers, driven by radiative heating and cooling, lead to variable temperature-gradient metamorphism under natural conditions. Owing to the fluctuating direction and intensity of the gradient, snow crystals cannot develop pronounced faceted structures, ultimately producing a coarser snow structure with reduced structural continuity (Pinzer and Schneebeli, 2009). Such metamorphism influences the diurnal evolution of hydrogen and oxygen stable isotopes but is mainly confined to the upper ~1 cm layer of the snowpack (Ritter et al., 2016).

Observations from Dome C, Antarctica, indicate that dry-snow metamorphism introduces an additional isotopic imprint. The $\delta^{18}O$–temperature slope decreases from 0.46‰ $°C^{-1}$ under precipitation-dominated conditions to 0.24‰ $°C^{-1}$ when metamorphism prevails. In the Vostok snowpit, principal component analysis also indicates that 75% of the total variance corresponds to an isotopic signature characterised by a d-excess/$\delta^{18}O$ ratio of −3.2, suggesting that a substantial portion of the signal is attributable to post-depositional metamorphic processes (Casado et al., 2021).

3.4.2 Wet-snow metamorphism

Wet-snow metamorphism occurs when temperatures approach the melting point. The presence of liquid water modifies the thermodynamic state of the snowpack and strengthens the mechanical and thermal linkages among ice grains through capillary bridging, thereby accelerating the coarsening and densification of snow crystals. Consequently, wet-snow metamorphism is several orders of magnitude faster than dry-snow metamorphism (Brun, 1989; Moure and Fu, 2024). Wet-snow metamorphism is characterised by homogeneous coarsening together with localised crystal growth in regions associated with preferential water infiltration and a slowdown in wet-snow metamorphism during melt–refreeze conditions (Avanzi et al., 2017).

Under wet-snow metamorphic conditions, snow exhibits $\delta^{18}O$ enrichment relative to the original precipitation signal, with surface snow showing an increase of up to 2.5‰ over a four-

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

day period (Sokratov et al., 2024). Compared with dry-snow metamorphism, wet-snow metamorphism generates a more uniform isotopic distribution throughout the snowpack owing to extensive mixing of isotopes by percolating meltwater.

3.5 Isotopic gradient diffusion

Pore-space diffusion along isotopic gradients is the principal mechanism that smooths the isotope signals of snow layers, and its strength depends on snow porosity, tortuosity, and density as well as on the effective diffusion coefficient (D_s) (Touzeau et al., 2018). The isotopic signature preserved in the snowpack reflects the combined influences of the original snowfall signal, post-depositional snow–atmosphere exchange, and factors such as temperature within the snowpack. Together, these processes produce pronounced spatiotemporal isotopic heterogeneity, generating isotopic gradients within the snowpack that drive vapour diffusion through pore spaces and exchange with the solid phase.

In modelling terms, this diffusion is commonly described by Fick's law, whereby changes in isotope concentration at a given point are determined by the surrounding isotopic gradient and the diffusion properties of the medium (Neumann and Waddington, 2004; Touzeau et al., 2018).

$$F^{18}(n+1 \to n) = -2D_{eff}(t, n \to n+1)\,\frac{D_{18}}{D}\,\frac{C_{v(t,n)}\,c^{18}{}_{vap,ini}(t,n) - C_{v(t,n+1)}\,c^{18}{}_{vap,ini}(t,n+1)}{dz(t,n) + dz(t,n+1)} \quad (21)$$

where $F^{18}(n+1 \to n)$ represents the flux of water molecules containing $\delta^{18}O$ from layer $n+1$ to layer n (kg m^{-2} s^{-1}), $D_{eff}(t, n \to n+1)$ represents the effective interfacial diffusivity between layers n and $n+1$ (m^2 s^{-1}), D_{18}/D represents the ratio of the diffusivity of the heavy isotopologue to that of the light isotopologue, $C_v(t,n)$ represents the saturated vapour mass concentration in the snow pore space (kg m^{-3}), $c^{18}_{vap,ini}(t,n)$ represents the mass fraction of a given isotopologue in the initial vapour, and $dz(t,n)$ represents the snow layer thickness (m).

At Dome C, isotope diffusion smooths the original seasonal signal driven by temperature changes, producing $\delta^{18}O$ fluctuations with a typical length scale of approximately 20 cm within the snowpack (Ma et al., 2020; Casado et al., 2018). Comparisons between snow-core and ice-core $\delta^{18}O$ signals from East Antarctica further revealed 55–70% signal attenuation over five years, which is attributed to isotope diffusion. Based on a firn diffusion model, the diffusion length is estimated to be approximately 6 cm. Collectively, these observations confirm that

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

diffusion smooths the $\delta^{18}O$ variability in the snowpack (Mahalinganathan et al., 2022). As snow compacts, pore spaces diminish and vapour-diffusion pathways shorten, gradually suppressing gas–solid exchange and isotopic fractionation.

3.6 Wind-driven processes

Wind-driven snow includes drifting snow, which mainly involves creep and saltation in the near-surface layer below approximately 2 m, and blowing snow transported at greater heights (Déry and Taylor, 1996). These processes are controlled chiefly by topography and wind (Melo et al., 2022). At the macroscale, wind-driven snow most obviously redistributes snow mass, which directly affects isotopic composition. At the microscale, snow crystals collide and fracture, undergoing fragmentation and compaction, while also experiencing multiple sublimation–deposition cycles at the surfaces of blowing snow (Wahl et al., 2024; Walter et al., 2024). These processes alter physical characteristics, including size, shape, density, and optical characteristics. Snow-crystal breakage increases the SSA, thereby promoting isotopic interaction between the atmospheric vapour and surface snow (Wahl et al., 2024). Under warmer air-temperature conditions, wind tunnel experiments exhibited the strongest isotopic fractionation in snow, with $\delta^{18}O$ and δD increasing at 0.006 and 0.011‰ min^{-1}, respectively, and d-excess decreasing at 0.049‰ min^{-1} (Wahl et al., 2024).

Interaction between wind and a rough snow surface also triggers wind pumping (Touzeau et al., 2018), whereby small-scale high- and low-pressure zones are generated as airflow passes over an irregular snow surface. This process enhances vapour exchange and intensifies sublimation (Ebner et al., 2017). Wind pumping enhances airflow through pore spaces, thereby promoting snow metamorphism, which primarily influences low-density snow layers (Drake et al., 2017). The potential magnitude of this effect is substantial: model experiments that resolve forced ventilation and pore-space diffusion indicate that sustained winds of 5–10 m/s can drive an annual-mean $\delta^{18}O$ enrichment of 3–7‰ across East Antarctica (Town et al., 2008).

Moreover, advection affects vapour saturation and air temperature, thereby indirectly influencing isotopic fractionation in surface snow. Advection of dry air reduces local vapour saturation and enhances sublimation (Bintanja, 2001; Hu et al., 2022). By contrast, cold-air advection lowers snow temperature by removing sensible heat, thereby inhibiting sublimation and reducing the intensity of isotopic fractionation (Harder et al., 2019).

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

3.7 Infiltration–refreeze

At higher temperatures, precipitation may occur as rainfall or existing surface snow may melt. Infiltration of rainwater or meltwater can homogenise isotope records within the snowpack (He et al., 2006). This process reduces seasonal isotopic signals, can lead to relative enrichment in heavy isotopes, and introduces temporal discontinuities into isotope records (Moran and Marshall, 2009). Snow begins to melt when the temperature rises to 0°C, forming a thin liquid-water film on the crystal surface. This process corresponds to equilibrium fractionation under ice–liquid equilibrium conditions. As temperature rises further, snow melts rapidly and isotopic fractionation becomes very limited (Taylor et al., 2001). As meltwater percolates through the snowpack, isotopic exchange occurs between the meltwater and snow crystals, shifting the meltwater composition towards lower abundances of heavy isotopes (Ala-aho et al., 2017). Observations in the Swiss Alps show that during meltwater formation, the preferential loss of light isotopes enriches the remaining snow by 3‰ in $\delta^{18}O$ and 25‰ in δD (Aemisegger et al., 2022). When temperature falls below 0°C, the meltwater begins to refreeze, with heavy isotopes preferentially partitioning into the solid phase and further depleting the residual liquid water (Ala-aho et al., 2017). Observations from Haig Glacier in the Canadian Rockies show that mean $\delta^{18}O$ increases from −19.2‰ in wet snow to −18.6‰ in glacier ice, revealing the role of infiltration–refreeze processes in snow isotopic enrichment (Penn et al., 2023).

The infiltration–refreeze process modifies the vertical isotopic profile of the snowpack while simultaneously complicating the interpretation of ice-core isotope records (Moran and Marshall, 2009). Moreover, infiltration–refreezing of meltwater can form ice crusts within the snowpack (Quéno et al., 2020). Densification of these ice crusts restricts liquid infiltration and gas diffusion through pore spaces, thereby limiting isotopic exchange between adjacent permeable snow layers separated by relatively impermeable crusts (Langway et al., 1993).

4 Challenges and future perspectives

Post-depositional processes have been identified as important controls on the isotope records preserved in ice cores. However, the contributions and effects of these processes across different timescales remain poorly understood, thereby reducing the interpretability of isotope records for paleoclimate studies. Existing studies are limited in their ability to support high-precision physical modelling because of coarse spatial resolution and temporal discontinuities

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

in observations of surface and subsurface snow. Observations with high vertical resolution and broader spatial coverage are therefore required to disentangle the key physical processes and mechanisms. The responsiveness of *d*-excess and ^{17}O-excess to differences between equilibrium and kinetic fractionation makes them indispensable natural tracers to quantify and separate the contributions of individual post-depositional processes to isotope records. Future research should therefore focus more strongly on second-order stable water isotope variables.

Isotope-enabled atmospheric general circulation models have reached a relatively mature stage, and these models effectively capture isotopic processes during vapour transport. By contrast, fully physical isotope-enabled atmosphere–snow–ice models remain in an exploratory phase. Further development of these models is required to improve simulations of isotopic evolution at the atmosphere–snow–ice interface. Moreover, machine learning has demonstrated strong capabilities in practical applications, including remote-sensing interpretation and spatial data mining. Machine learning is well suited to modelling the complex, non-linear relationships and coupling mechanisms among multiple variables and is therefore potentially valuable for describing intricate geoscientific processes that are difficult to represent using traditional parametric methods. Future research should integrate machine learning with isotope modelling to enable more rigorous and robust investigation of stable water isotope processes at the atmosphere–snow–ice interface.

Ice cores over the Tibetan Plateau have been used to reconstruct regional climatic and environmental evolution since the Last Interglacial. Although the underlying physical processes and isotopic fractionation mechanisms are similar to those in polar regions, their magnitudes and manifestations differ because of contrasting climatic conditions. Antarctica and Greenland are characterised by arid, frigid climates with low accumulation rates, whereas the Tibetan Plateau exhibits more pronounced seasonal variability and a more complex moisture supply. Compared with polar regions, snow on the Tibetan Plateau is more frequently present as wet snow, rendering isotopic fractionation after snow deposition more complex. Relative to the post-depositional processes dominated by sublimation–deposition and vapour diffusion within the snowpack under polar dry-snow conditions, wet-snow environments are additionally regulated by processes such as wet-snow metamorphism, infiltration–refreeze, and liquid-water transport. The integrated effects of these mechanisms increase the complexity of snowpack isotopic composition, posing substantial challenges for robust interpretation. Consequently, when reconstructing paleoclimate from the isotope records of Tibetan Plateau ice cores, the effects of wet-snow metamorphism and liquid-water processes must be considered particularly

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

carefully.

The role of vapour exchange and phase transitions at the atmosphere–snow–ice interface, together with the associated isotopic fractionation, cannot be ignored when interpreting paleoclimate signals preserved in ice-core records. Future research should place greater emphasis on the potential of *d*-excess and ^{17}O-excess to quantify individual post-depositional processes and should aim to establish an integrated observation-experiment-modelling framework. By combining multisource datasets and comparing processes across different regions, such an approach could enable quantitative characterisation of the contributions of individual processes and their coupled effects on ice-core isotope records across multiple timescales.

CRediT authorship contribution statement

Mingyu Dou: Writing–review & editing, Writing–original draft, Methodology, Investigation, Formal analysis, Conceptualisation. **Jing Gao:** Writing–review & editing, Supervision, Resources, Funding acquisition, Conceptualisation. **Aibin Zhao:** Writing–review & editing. **Deji:** Writing–review & editing, Supervision, Conceptualisation.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Acknowledgements

This work was funded by the Science and Technology Projects of Xizang Autonomous Region, China (Grant No. XZ202501ZY0081) and The Second Tibetan Plateau Scientific Expedition and Research (Grant Nos. 2024QZKK0400 and 2024QZKK0100).

Data availability

No new data were created during this study. All figures within this manuscript were generated using data from openly available datasets referenced in the figure captions.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

References

Aemisegger, F., Trachsel, J., Sadowski, Y., Eichler, A., Lehning, M., Avak, S., and Schneebeli, M.: Fingerprints of Frontal Passages and Post-Depositional Effects in the Stable Water Isotope Signal of Seasonal Alpine Snow, J. Geophys. Res.-Atmos., 127, e2022JD037469, https://doi.org/10.1029/2022JD037469, 2022.

Ala-aho, P., Tetzlaff, D., McNamara, J. P., Laudon, H., Kormos, P., and Soulsby, C.: Modeling the isotopic evolution of snowpack and snowmelt: Testing a spatially distributed parsimonious approach, Water Resour. Res., 53, 5813–5830, https://doi.org/10.1002/2017WR020650, 2017.

Amory, C. and Kittel, C.: Brief communication: Rare ambient saturation during drifting snow occurrences at a coastal location of East Antarctica, The Cryosphere, 13, 3405–3412, https://doi.org/10.5194/tc-13-3405-2019, 2019.

Avanzi, F., Petrucci, G., Matzl, M., Schneebeli, M., and De Michele, C.: Early formation of preferential flow in a homogeneous snowpack observed by micro-CT, Water Resour. Res., 53, 3713–3729, https://doi.org/10.1002/2016WR019502, 2017.

Bartels-Rausch, T., Jacobi, H.-W., Kahan, T. F., Thomas, J. L., Thomson, E. S., Abbatt, J. P. D., Ammann, M., Blackford, J. R., Bluhm, H., Boxe, C., Domine, F., Frey, M. M., Gladich, I., Guzmán, M. I., Heger, D., Huthwelker, Th., Klán, P., Kuhs, W. F., Kuo, M. H., Maus, S., Moussa, S. G., McNeill, V. F., Newberg, J. T., Pettersson, J. B. C., Roeselová, M., and Sodeau, J. R.: A review of air–ice chemical and physical interactions (AICI): liquids, quasi-liquids, and solids in snow, Atmos. Chem. Phys., 14, 1587–1633, https://doi.org/10.5194/acp-14-1587-2014, 2014.

Bellagamba, A. W., Berkelhammer, M., Hamed, Y., Pearce, K., and Steen-Larsen, H. C.: Isotopic Fractionation during Sublimation of Low Porosity Ice, Chem. Geol., 670, 122445, https://doi.org/10.1016/j.chemgeo.2024.122445, 2024.

Bintanja, R.: Modelling snowdrift sublimation and its effect on the moisture budget of the atmospheric boundary layer, Tellus A, 53, 215, https://doi.org/10.3402/tellusa.v53i2.12189, 2001.

Brun, E.: Investigation on Wet-Snow Metamorphism in Respect of Liquid-Water Content, Ann. Glaciol., 13, 22–26, https://doi.org/10.3189/S0260305500007576, 1989.

Calonne, N., Flin, F., Geindreau, C., Lesaffre, B., and Rolland Du Roscoat, S.: Study of a temperature gradient metamorphism of snow from 3-D images: time evolution of microstructures, physical properties and their associated anisotropy, The Cryosphere, 8, 2255–2274, https://doi.org/10.5194/tc-8-2255-2014, 2014.

Carroll, R. W. H., Deems, J., Sprenger, M., Maxwell, R., Brown, W., Newman, A., Beutler, C., and Williams, K. H.: Modeling Snow Dynamics and Stable Water Isotopes Across Mountain Landscapes, Geophys. Res. Lett., 49, e2022GL098780, https://doi.org/10.1029/2022GL098780, 2022.

Casado, M., Cauquoin, A., Landais, A., Israel, D., Orsi, A., Pangui, E., Landsberg, J., Kerstel, E., Prie, F., and Doussin, J.-F.: Experimental determination and theoretical framework of kinetic fractionation at the water vapour–ice interface at low temperature, Geochim. Cosmochim. Acta, 174, 54–69, https://doi.org/10.1016/j.gca.2015.11.009, 2016.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.



Casado, M., Landais, A., Picard, G., Münch, T., Laepple, T., Stenni, B., Dreossi, G., Ekaykin, A., Arnaud, L., Genthon, C., Touzeau, A., Masson-Delmotte, V., and Jouzel, J.: Archival processes of the water stable isotope signal in East Antarctic ice cores, The Cryosphere, 12, 1745–1766, https://doi.org/10.5194/tc-12-1745-2018, 2018.

Casado, M., Münch, T., and Laepple, T.: Climatic information archived in ice cores: impact of intermittency and diffusion on the recorded isotopic signal in Antarctica, Clim. Past, 16, 1581–1598, https://doi.org/10.5194/cp-16-1581-2020, 2020.

Casado, M., Landais, A., Picard, G., Arnaud, L., Dreossi, G., Stenni, B., and Prié, F.: Water Isotopic Signature of Surface Snow Metamorphism in Antarctica, Geophys. Res. Lett., 48, e2021GL093382, https://doi.org/10.1029/2021GL093382, 2021.

Colbeck, S. C.: Theory of metamorphism of dry snow, J. Geophys. Res., 88, 5475–5482, https://doi.org/10.1029/JC088iC09p05475, 1983.

Colbeck, S. C.: Model of wind pumping for layered snow, J. Glaciol., 43, 60–65, https://doi.org/10.3189/S002214300000280X, 1997.

Craig, H.: Isotopic Variations in Meteoric Waters, Science, 133, 1702–1703, https://doi.org/10.1126/science.133.3465.1702, 1961.

Dadic, R., Schneebeli, M., Bertler, N. A. N., Schwikowski, M., and Matzl, M.: Extreme snow metamorphism in the Allan Hills, Antarctica, as an analogue for glacial conditions with implications for stable isotope composition, J. Glaciol., 61, 1171–1182, https://doi.org/10.3189/2015JoG15J027, 2015.

Dansgaard, W.: Stable isotopes in precipitation, Tellus, 16, 436–468, https://doi.org/10.3402/tellusa.v16i4.8993, 1964.

Déry, S. J. and Taylor, P. A.: Some aspects of the interaction of blowing snow with the atmospheric boundary layer, Hydrol. Process., 10, 1345–1358, https://doi.org/10.1002/(SICI)1099-1085(199610)10:10<1345::AID-HYP465>3.0.CO;2-2, 1996.

Dietrich, L. J., Steen-Larsen, H. C., Wahl, S., Jones, T. R., Town, M. S., and Werner, M.: Snow-Atmosphere Humidity Exchange at the Ice Sheet Surface Alters Annual Mean Climate Signals in Ice Core Records, Geophys. Res. Lett., 50, e2023GL104249, https://doi.org/10.1029/2023GL104249, 2023.

Drake, S. A., Selker, J. S., and Higgins, C. W.: Wind enhances differential air advection in surface snow at sub-meter scales, The Cryosphere, 11, 2075–2087, https://doi.org/10.5194/tc-11-2075-2017, 2017.

Dutrievoz, N., Agosta, C., Risi, C., Vignon, É., Nguyen, S., Landais, A., Fourré, E., Leroy-Dos Santos, C., Casado, M., Masson-Delmotte, V., Jouzel, J., Dubos, T., Ollivier, I., Stenni, B., Dreossi, G., Masiol, M., Minster, B., and Prié, F.: Antarctic Water Stable Isotopes in the Global Atmospheric Model LMDZ6: From Climatology to Boundary Layer Processes, J. Geophys. Res.-Atmos., 130, e2024JD042073, https://doi.org/10.1029/2024JD042073, 2025.

Dütsch, M., Pfahl, S., and Sodemann, H.: The impact of nonequilibrium and equilibrium fractionation on two different deuterium excess definitions, J. Geophys. Res.-Atmos., 122, 12732–12746, https://doi.org/10.1002/2017JD027085, 2017.

Ebner, P. P., Schneebeli, M., and Steinfeld, A.: Tomography-based monitoring of isothermal

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.



snow metamorphism under advective conditions, The Cryosphere, 9, 1363–1371, https://doi.org/10.5194/tc-9-1363-2015, 2015.

Ebner, P. P., Steen-Larsen, H. C., Stenni, B., Schneebeli, M., and Steinfeld, A.: Experimental observation of transient δ18O interaction between snow and advective airflow under various temperature gradient conditions, The Cryosphere, 11, 1733–1743, https://doi.org/10.5194/tc-11-1733-2017, 2017.

Ellehoj, M. D., Steen-Larsen, H. C., Johnsen, S. J., and Madsen, M. B.: Ice-vapor equilibrium fractionation factor of hydrogen and oxygen isotopes: Experimental investigations and implications for stable water isotope studies, Rapid Commun. Mass Spectrom., 27, 2149–2158, https://doi.org/10.1002/rcm.6668, 2013.

EPICA community members: Eight glacial cycles from an Antarctic ice core, Nature, 429, 623–628, https://doi.org/10.1038/nature02599, 2004.

Fisher, D. A., Koerner, R. M., Paterson, W. S. B., Dansgaard, W., Gundestrup, N., and Reeh, N.: Effect of wind scouring on climatic records from ice-core oxygen-isotope profiles, Nature, 301, 205–209, https://doi.org/10.1038/301205a0, 1983.

Furukawa, R., Uemura, R., Fujita, K., Sjolte, J., Yoshimura, K., Matoba, S., and Iizuka, Y.: Seasonal-scale dating of a shallow ice core from Greenland using oxygen isotope matching between data and simulation, J. Geophys. Res.-Atmos., 122, 10873–10887, https://doi.org/10.1002/2017JD026716, 2017.

Garratt, J. R.: Review: the atmospheric boundary layer, Earth-Sci. Rev., 37, 89–134, https://doi.org/10.1016/0012-8252(94)90026-4, 1994.

Gkinis, V., Holme, C., Kahle, E. C., Stevens, M. C., Steig, E. J., and Vinther, B. M.: Numerical experiments on firn isotope diffusion with the Community Firn Model, J. Glaciol., 67, 450–472, https://doi.org/10.1017/jog.2021.1, 2021.

Good, S. P., Soderberg, K., Wang, L., and Caylor, K. K.: Uncertainties in the assessment of the isotopic composition of surface fluxes: A direct comparison of techniques using laser-based water vapor isotope analyzers, J. Geophys. Res.-Atmos., 117, 2011JD017168, https://doi.org/10.1029/2011JD017168, 2012.

Goursaud, S., Masson-Delmotte, V., Favier, V., Orsi, A., and Werner, M.: Water stable isotope spatio-temporal variability in Antarctica in 1960–2013: observations and simulations from the ECHAM5-wiso atmospheric general circulation model, Clim. Past, 14, 923–946, https://doi.org/10.5194/cp-14-923-2018, 2018.

Griffis, T. J., Sargent, S. D., Lee, X., Baker, J. M., Greene, J., Erickson, M., Zhang, X., Billmark, K., Schultz, N., Xiao, W., and Hu, N.: Determining the Oxygen Isotope Composition of Evapotranspiration Using Eddy Covariance, Bound.-Layer Meteor., 137, 307–326, https://doi.org/10.1007/s10546-010-9529-5, 2010.

Harder, P., Pomeroy, J. W., and Helgason, W. D.: A simple model for local-scale sensible and latent heat advection contributions to snowmelt, Hydrol. Earth Syst. Sci., 23, 1–17, https://doi.org/10.5194/hess-23-1-2019, 2019.

Harris Stuart, R., Faber, A.-K., Wahl, S., Hörhold, M., Kipfstuhl, S., Vasskog, K., Behrens, M., Zuhr, A. M., and Steen-Larsen, H. C.: Exploring the role of snow metamorphism on the isotopic composition of the surface snow at EastGRIP, The Cryosphere, 17, 1185–1204, https://doi.org/10.5194/tc-17-1185-2023, 2023.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

He, Y., Pang, H., Theakstone, W. H., Zhang, D., Lu, A., Song, B., Yuan, L., and Ning, B.: Spatial and temporal variations of oxygen isotopes in snowpacks and glacial runoff in different types of glacial area in western China, Ann. Glaciol., 43, 269–274, https://doi.org/10.3189/172756406781811943, 2006.

Hellmann, R. and Harvey, A. H.: First-Principles Diffusivity Ratios for Kinetic Isotope Fractionation of Water in Air, Geophys. Res. Lett., 47, e2020GL089999, https://doi.org/10.1029/2020GL089999, 2020.

Hock, R.: Glacier melt: a review of processes and their modelling, Prog. Phys. Geogr., 29, 362–391, https://doi.org/10.1191/0309133305pp453ra, 2005.

Holme, C., Gkinis, V., and Vinther, B. M.: Molecular diffusion of stable water isotopes in polar firn as a proxy for past temperatures, Geochim. Cosmochim. Acta, 225, 128–145, https://doi.org/10.1016/j.gca.2018.01.015, 2018.

Hu, J., Yan, Y., Yeung, L. Y., and Dee, S. G.: Sublimation Origin of Negative Deuterium Excess Observed in Snow and Ice Samples From McMurdo Dry Valleys and Allan Hills Blue Ice Areas, East Antarctica, J. Geophys. Res.-Atmos., 127, e2021JD035950, https://doi.org/10.1029/2021JD035950, 2022.

Hu, Y., Xiao, W., Wei, Z., Welp, L. R., Wen, X., and Lee, X.: Determining the Isotopic Composition of Surface Water Vapor Flux From High-Frequency Observations Using Flux-Gradient and Keeling Plot Methods, Earth Space Sci., 8, e2020EA001304, https://doi.org/10.1029/2020EA001304, 2021.

Hughes, A. G., Wahl, S., Jones, T. R., Zuhr, A., Hörhold, M., White, J. W. C., and Steen-Larsen, H. C.: The role of sublimation as a driver of climate signals in the water isotope content of surface snow: laboratory and field experimental results, The Cryosphere, 15, 4949–4974, https://doi.org/10.5194/tc-15-4949-2021, 2021.

Intergovernmental Panel on Climate Change (IPCC): Climate Change 2021 – The Physical Science Basis: Contribution of Working Group I to the Sixth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA, https://doi.org/10.1017/9781009157896, 2021.

Jakli, G. and Staschewski, D.: Vapour pressure of H218O ice (–50 to 0 °C) and H218O water (0 to 170 °C), J. Chem. Soc., Faraday Trans. 1, 73, 1505, https://doi.org/10.1039/f19777301505, 1977.

Jambon-Puillet, E., Shahidzadeh, N., and Bonn, D.: Singular sublimation of ice and snow crystals, Nat. Commun., 9, 4191, https://doi.org/10.1038/s41467-018-06689-x, 2018.

Jancso, G., Pupezin, J., and Van Hook, W. A.: Vapour Pressure of H218O Ice (I) (−17 °C to 0 °C) and H218O Water (0 °C to 16 °C), Nature, 225, 723, https://doi.org/10.1038/225723a0, 1970.

Jeelani, G., Lone, S. A., Padhya, V., Deshpande, R. D., Lone, A., and Rozanski, K.: Stable water isotopes in snow across the Kashmir Himalaya: constraints on hydrometeorological processes, Clim. Dynam., 63, 235, https://doi.org/10.1007/s00382-025-07726-4, 2025.

Johnsen, S. J., Clausen, H. B., Duffey, K. M., Hoffmann, G., Schwander, J., and Cryts, T. T.: Diffusion of stable isotopes in polar firn and ice: the isotope effect in firn diffusion, in: Physics of Ice Core Records, edited by: Hondoh, T., Hokkaido University Press, Sapporo, Japan, 121–140, https://doi.org/10.7916/D8KW5D4X, 2000.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

Jouzel, J. and Masson-Delmotte, V.: Paleoclimates: what do we learn from deep ice cores?, WIREs Clim. Change, 1, 654–669, https://doi.org/10.1002/wcc.72, 2010.

Jouzel, J., Masson-Delmotte, V., Cattani, O., Dreyfus, G., Falourd, S., Hoffmann, G., Minster, B., Nouet, J., Barnola, J. M., Chappellaz, J., Fischer, H., Gallet, J. C., Johnsen, S., Leuenberger, M., Loulergue, L., Luethi, D., Oerter, H., Parrenin, F., Raisbeck, G., Raynaud, D., Schilt, A., Schwander, J., Selmo, E., Souchez, R., Spahni, R., Stauffer, B., Steffensen, J. P., Stenni, B., Stocker, T. F., Tison, J. L., Werner, M., and Wolff, E. W.: Orbital and Millennial Antarctic Climate Variability over the Past 800,000 Years, Science, 317, 793–796, https://doi.org/10.1126/science.1141038, 2007.

Kiemle, S., Heck, K., Coltman, E., and Helmig, R.: Stable Water Isotopologue Fractionation During Soil-Water Evaporation: Analysis Using a Coupled Soil-Atmosphere Model, Water Resour. Res., 59, e2022WR032385, https://doi.org/10.1029/2022WR032385, 2023.

Kim, K. and Lee, X.: Isotopic enrichment of liquid water during evaporation from water surfaces, J. Hydrol., 399, 364–375, https://doi.org/10.1016/j.jhydrol.2011.01.008, 2011.

King, J. C., Anderson, P. S., Vaughan, D. G., Mann, G. W., Mobbs, S. D., and Vosper, S. B.: Wind-borne redistribution of snow across an Antarctic ice rise, J. Geophys. Res.-Atmos., 109, 2003JD004361, https://doi.org/10.1029/2003JD004361, 2004.

Knyazev, D. A., Myasoedov, N. F., and Bochkarev, A. V.: The theory of the equilibrium isotope effects of hydrogen, Russ. Chem. Rev., 61, 204–220, https://doi.org/10.1070/RC1992v061n02ABEH000941, 1992.

Kochanski, K., Anderson, R. S., and Tucker, G. E.: The evolution of snow bedforms in the Colorado Front Range and the processes that shape them, The Cryosphere, 13, 1267–1281, https://doi.org/10.5194/tc-13-1267-2019, 2019.

Kopec, B. G., Feng, X., Osterberg, E. C., and Posmentier, E. S.: Climatological Significance of δD-δ18O Line Slopes From Precipitation, Snow Pits, and Ice Cores at Summit, Greenland, J. Geophys. Res.-Atmos., 127, e2022JD037037, https://doi.org/10.1029/2022JD037037, 2022.

Kunz, A.-K., Borchardt, L., Christen, A., Della Coletta, J., Eritt, M., Gutiérrez, X., Hashemi, J., Hilland, R., Jordan, A., Kneißl, R., Legendre, V., Levin, I., Preunkert, S., Rubli, P., Stagakis, S., and Hammer, S.: A relaxed eddy accumulation flask sampling system for 14 C-based partitioning of fossil and non-fossil CO2 fluxes, Atmos. Meas. Tech., 18, 5349–5373, https://doi.org/10.5194/amt-18-5349-2025, 2025.

Lamb, K. D., Clouser, B. W., Bolot, M., Sarkozy, L., Ebert, V., Saathoff, H., Möhler, O., and Moyer, E. J.: Laboratory measurements of HDO/H_2O isotopic fractionation during ice deposition in simulated cirrus clouds, P. Natl. Acad. Sci. USA, 114, 5612–5617, https://doi.org/10.1073/pnas.1618374114, 2017.

Landais, A., Ekaykin, A., Barkan, E., Winkler, R., and Luz, B.: Seasonal variations of ^{17}O-excess and d-excess in snow precipitation at Vostok station, East Antarctica, J. Glaciol., 58, 725–733, https://doi.org/10.3189/2012JoG11J237, 2012a.

Landais, A., Steen-Larsen, H. C., Guillevic, M., Masson-Delmotte, V., Vinther, B., and Winkler, R.: Triple isotopic composition of oxygen in surface snow and water vapor at NEEM (Greenland), Geochim. Cosmochim. Acta, 77, 304–316, https://doi.org/10.1016/j.gca.2011.11.022, 2012b.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

Langway, C. C., Shoji, H., Mitani, A., and Clausen, H. B.: Transformation process observations of polar firn to ice, Ann. Glaciol., 18, 199–202, https://doi.org/10.3189/S0260305500011502, 1993.

Lee, X., Griffis, T. J., Baker, J. M., Billmark, K. A., Kim, K., and Welp, L. R.: Canopy-scale kinetic fractionation of atmospheric carbon dioxide and water vapor isotopes, Global Biogeochem. Cy., 23, 2008GB003331, https://doi.org/10.1029/2008GB003331, 2009.

Legagneux, L. and Domine, F.: A mean field model of the decrease of the specific surface area of dry snow during isothermal metamorphism, J. Geophys. Res.-Earth, 110, 2004JF000181, https://doi.org/10.1029/2004JF000181, 2005.

Li, W., Luo, X., Ji, Z., and Liu, C.: Kinetic fractionation of Mg isotopes during chemical diffusion in aqueous solutions: A reappraisal, Fundam. Res., S2667325824004643, https://doi.org/10.1016/j.fmre.2024.11.010, 2024.

Liu, J., Chen, R., Ma, S., Han, C., Ding, Y., Guo, S., and Wang, X.: The challenge of monitoring snow surface sublimation in winter could be resolved with structure-from-motion photogrammetry, J. Hydrol., 630, 130733, https://doi.org/10.1016/j.jhydrol.2024.130733, 2024.

Löwe, H., Spiegel, J. K., and Schneebeli, M.: Interfacial and structural relaxations of snow under isothermal conditions, J. Glaciol., 57, 499–510, https://doi.org/10.3189/002214311796905569, 2011.

Lu, G. and DePaolo, D. J.: Lattice Boltzmann simulation of water isotope fractionation during ice crystal growth in clouds, Geochim. Cosmochim. Acta, 180, 271–283, https://doi.org/10.1016/j.gca.2015.11.048, 2016.

Lundquist, J. D., Vano, J., Gutmann, E., Hogan, D., Schwat, E., Haugeneder, M., Mateo, E., Oncley, S., Roden, C., Osenga, E., and Carver, L.: Sublimation of Snow, B. Am. Meteorol. Soc., 105, E975–E990, https://doi.org/10.1175/BAMS-D-23-0191.1, 2024.

Ma, T., Li, L., Shi, G., and Li, Y.: Acquisition of Post-Depositional Effects on Stable Isotopes (δ18O and δD) of Snow and Firn at Dome A, East Antarctica, Water, 12, 1707, https://doi.org/10.3390/w12061707, 2020.

Ma, T., Jiang, Z., Ding, M., He, P., Li, Y., Zhang, W., and Geng, L.: A model framework for atmosphere–snow water vapor exchange and the associated isotope effects at Dome Argus, Antarctica – Part 1: The diurnal changes, The Cryosphere, 18, 4547–4565, https://doi.org/10.5194/tc-18-4547-2024, 2024.

Madsen, M. V., Steen-Larsen, H. C., Hörhold, M., Box, J., Berben, S. M. P., Capron, E., Faber, A.-K., Hubbard, A., Jensen, M. F., Jones, T. R., Kipfstuhl, S., Koldtoft, I., Pillar, H. R., Vaughn, B. H., Vladimirova, D., and Dahl-Jensen, D.: Evidence of Isotopic Fractionation During Vapor Exchange Between the Atmosphere and the Snow Surface in Greenland, J. Geophys. Res.-Atmos., 124, 2932–2945, https://doi.org/10.1029/2018JD029619, 2019.

Mahalinganathan, K., Thamban, M., Ejaz, T., Srivastava, R., Redkar, B. L., and Laluraj, C. M.: Spatial variability and post-depositional diffusion of stable isotopes in high accumulation regions of East Antarctica, Front. Earth Sci., 10, 925447, https://doi.org/10.3389/feart.2022.925447, 2022.

Majoube, M.: Fractionation Factor of 18O between Water Vapour and Ice, Nature, 226, 1242, https://doi.org/10.1038/2261242a0, 1970.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

Masson-Delmotte, V., Dreyfus, G., Braconnot, P., Johnsen, S., Jouzel, J., Kageyama, M., Landais, A., Loutre, M.-F., Nouet, J., Parrenin, F., Raynaud, D., Stenni, B., and Tuenter, E.: Past temperature reconstructions from deep ice cores: relevance for future climate change, Clim. Past, 2, 145–165, https://doi.org/10.5194/cp-2-145-2006, 2006.

Matsuo, S., Kuniyoshi, H., and Miyake, Y.: Vapor pressure of ice containing D_2O, Science, 145, 1454–1455, https://doi.org/10.1126/science.145.3639.1454, 1964.

Matsuo, S. and Matsubaya, O.: Vapour Pressure of H218O Ice, Nature, 221, 463–464, https://doi.org/10.1038/221463a0, 1969.

Melo, D. B., Sharma, V., Comola, F., Sigmund, A., and Lehning, M.: Modeling Snow Saltation: The Effect of Grain Size and Interparticle Cohesion, J. Geophys. Res.-Atmos., 127, e2021JD035260, https://doi.org/10.1029/2021JD035260, 2022.

Merlivat, L. and Jouzel, J.: Global climatic interpretation of the deuterium-oxygen 18 relationship for precipitation, J. Geophys. Res., 84, 5029–5033, https://doi.org/10.1029/JC084iC08p05029, 1979.

Merlivat, L. and Nief, G.: Fractionnement isotopique lors des changements d'état solide-vapeur et liquide-vapeur de l'eau à des températures inférieures à 0 °C, Tellus, 19, 122–127, https://doi.org/10.1111/j.2153-3490.1967.tb01465.x, 1967.

Méheut, M., Lazzeri, M., Balan, E., and Mauri, F.: Equilibrium isotopic fractionation in the kaolinite, quartz, water system: Prediction from first-principles density-functional theory, Geochim. Cosmochim. Acta, 71, 3170–3181, https://doi.org/10.1016/j.gca.2007.04.012, 2007.

Moran, T. and Marshall, S.: The effects of meltwater percolation on the seasonal isotopic signals in an Arctic snowpack, J. Glaciol., 55, 1012–1024, https://doi.org/10.3189/002214309790794896, 2009.

Moure, A. and Fu, X.: A Phase-Field Model for Wet Snow Metamorphism, Cryst. Growth Des., 24, 7808–7821, https://doi.org/10.1021/acs.cgd.4c00539, 2024.

Münch, T., Kipfstuhl, S., Freitag, J., Meyer, H., and Laepple, T.: Regional climate signal vs. local noise: a two-dimensional view of water isotopes in Antarctic firn at Kohnen Station, Dronning Maud Land, Clim. Past, 12, 1565–1581, https://doi.org/10.5194/cp-12-1565-2016, 2016.

Münch, T., Werner, M., and Laepple, T.: How precipitation intermittency sets an optimal sampling distance for temperature reconstructions from Antarctic ice cores, Clim. Past, 17, 1587–1605, https://doi.org/10.5194/cp-17-1587-2021, 2021.

Neumann, T. A. and Waddington, E. D.: Effects of firn ventilation on isotopic exchange, J. Glaciol., 50, 183–194, https://doi.org/10.3189/172756504781830150, 2004.

North Greenland Ice Core Project members: High-resolution record of Northern Hemisphere climate extending into the last interglacial period, Nature, 431, 147–151, https://doi.org/10.1038/nature02805, 2004.

Ollivier, I., Steen-Larsen, H. C., Stenni, B., Arnaud, L., Casado, M., Cauquoin, A., Dreossi, G., Genthon, C., Minster, B., Picard, G., Werner, M., and Landais, A.: Surface processes and drivers of the snow water stable isotopic composition at Dome C, East Antarctica – a multi-dataset and modelling analysis, The Cryosphere, 19, 173–200, https://doi.org/10.5194/tc-19-173-2025, 2025.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

Palm, S. P., Kayetha, V., Yang, Y., and Pauly, R.: Blowing snow sublimation and transport over Antarctica from 11 years of CALIPSO observations, The Cryosphere, 11, 2555–2569, https://doi.org/10.5194/tc-11-2555-2017, 2017.

Pang, H., Hou, S., Landais, A., Masson-Delmotte, V., Prie, F., Steen-Larsen, H. C., Risi, C., Li, Y., Jouzel, J., Wang, Y., He, J., Minster, B., and Falourd, S.: Spatial distribution of ^{17}O-excess in surface snow along a traverse from Zhongshan station to Dome A, East Antarctica, Earth Planet. Sci. Lett., 414, 126–133, https://doi.org/10.1016/j.epsl.2015.01.014, 2015.

Pang, H., Hou, S., Landais, A., Masson-Delmotte, V., Jouzel, J., Steen-Larsen, H. C., Risi, C., Zhang, W., Wu, S., Li, Y., An, C., Wang, Y., Prie, F., Minster, B., Falourd, S., Stenni, B., Scarchilli, C., Fujita, K., and Grigioni, P.: Influence of summer sublimation on δD, $\delta^{18}O$, and $\delta^{17}O$ in precipitation, East Antarctica, and implications for climate reconstruction from ice cores, J. Geophys. Res.-Atmos., 124, 7339–7358, https://doi.org/10.1029/2018JD030218, 2019.

Penn, K., Marshall, S. J., and Sinclair, K. E.: Seasonal enrichment of heavy isotopes in meltwater runoff from Haig Glacier, Canadian Rocky Mountains, Front. Earth Sci., 11, 1125877, https://doi.org/10.3389/feart.2023.1125877, 2023.

Persson, A., Langen, P. L., Ditlevsen, P., and Vinther, B. M.: The influence of precipitation weighting on interannual variability of stable water isotopes in Greenland, J. Geophys. Res., 116, D20120, https://doi.org/10.1029/2010JD015517, 2011.

Pinzer, B. R. and Schneebeli, M.: Snow metamorphism under alternating temperature gradients: Morphology and recrystallization in surface snow, Geophys. Res. Lett., 36, 2009GL039618, https://doi.org/10.1029/2009GL039618, 2009.

Pinzer, B. R., Schneebeli, M., and Kaempfer, T. U.: Vapor flux and recrystallization during dry snow metamorphism under a steady temperature gradient as observed by time-lapse micro-tomography, The Cryosphere, 6, 1141–1155, https://doi.org/10.5194/tc-6-1141-2012, 2012.

Pu, T., Kong, Y., Wang, S., Shi, X., Wang, K., Niu, H., and Chen, P.: Modification of stable isotopes in snow and related post-depositional processes on a temperate glacier of Mt. Yulong, southeast Tibetan Plateau, J. Hydrol., 584, 124675, https://doi.org/10.1016/j.jhydrol.2020.124675, 2020.

Purdie, H., Bertler, N., Mackintosh, A., Baker, J., and Rhodes, R.: Isotopic and elemental changes in winter snow accumulation on glaciers in the Southern Alps of New Zealand, J. Climate, 23, 4737–4749, https://doi.org/10.1175/2010JCLI3701.1, 2010.

Quéno, L., Fierz, C., Van Herwijnen, A., Longridge, D., and Wever, N.: Deep ice layer formation in an alpine snowpack: monitoring and modeling, The Cryosphere, 14, 3449–3464, https://doi.org/10.5194/tc-14-3449-2020, 2020.

Ramseier, R. O.: Self-Diffusion of Tritium in Natural and Synthetic Ice Monocrystals, J. Appl. Phys., 38, 2553–2556, https://doi.org/10.1063/1.1709948, 1967.

Ritter, F., Steen-Larsen, H. C., Werner, M., Masson-Delmotte, V., Orsi, A., Behrens, M., Birnbaum, G., Freitag, J., Risi, C., and Kipfstuhl, S.: Isotopic exchange on the diurnal scale between near-surface snow and lower atmospheric water vapor at Kohnen station, East Antarctica, The Cryosphere, 10, 1647–1663, https://doi.org/10.5194/tc-10-1647-

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.



2016, 2016.

Sexstone, G. A., Clow, D. W., Stannard, D. I., and Fassnacht, S. R.: Comparison of methods for quantifying surface sublimation over seasonally snow-covered terrain, Hydrol. Process., 30, 3373–3389, https://doi.org/10.1002/hyp.10864, 2016.

Shang, B., Gao, J., Chen, G., and Wu, Y.: Stable isotopes in atmospheric water vapour: Patterns, mechanisms and perspectives, Sci. China Earth Sci., 67, 3789–3813, https://doi.org/10.1007/s11430-023-1410-6, 2024.

Sokratov, S. A., Komarov, A. Yu., Vasil'chuk, Yu. K., Budantseva, N. A., Vasil'chuk, D. Yu., Seliverstov, Yu. G., Grebennikov, P. B., and Frolov, D. M.: Spatial-Temporal Variability of the δ18O Values and the Snow Cover Structure on the Territory of the Meteorological Observatory of the Moscow State University, Water Resour., 51, S89–S99, https://doi.org/10.1134/S0097807824701367, 2024.

Steen-Larsen, H. C., Masson-Delmotte, V., Hirabayashi, M., Winkler, R., Satow, K., Prié, F., Bayou, N., Brun, E., Cuffey, K. M., Dahl-Jensen, D., Dumont, M., Guillevic, M., Kipfstuhl, S., Landais, A., Popp, T., Risi, C., Steffen, K., Stenni, B., and Sveinbjörnsdottír, A. E.: What controls the isotopic composition of Greenland surface snow?, Clim. Past, 10, 377–392, https://doi.org/10.5194/cp-10-377-2014, 2014.

Stevens, C. M., Verjans, V., Lundin, J. M. D., Kahle, E. C., Horlings, A. N., Horlings, B. I., and Waddington, E. D.: The Community Firn Model (CFM) v1.0, Geosci. Model Dev., 13, 4355–4377, https://doi.org/10.5194/gmd-13-4355-2020, 2020.

Taylor, S., Feng, X., Kirchner, J. W., Osterhuber, R., Klaue, B., and Renshaw, C. E.: Isotopic evolution of a seasonal snowpack and its melt, Water Resour. Res., 37, 759–769, https://doi.org/10.1029/2000WR900341, 2001.

Thiemens, M. H.: Mass-Independent Isotope Effects in Planetary Atmospheres and the Early Solar System, Science, 283, 341–345, https://doi.org/10.1126/science.283.5400.341, 1999.

Tierney, J. E., Poulsen, C. J., Montañez, I. P., Bhattacharya, T., Feng, R., Ford, H. L., Hönisch, B., Inglis, G. N., Petersen, S. V., Sagoo, N., Tabor, C. R., Thirumalai, K., Zhu, J., Burls, N. J., Foster, G. L., Goddéris, Y., Huber, B. T., Ivany, L. C., Kirtland Turner, S., Lunt, D. J., McElwain, J. C., Mills, B. J. W., Otto-Bliesner, B. L., Ridgwell, A., and Zhang, Y. G.: Past climates inform our future, Science, 370, eaay3701, https://doi.org/10.1126/science.aay3701, 2020.

Touzeau, A., Landais, A., Morin, S., Arnaud, L., and Picard, G.: Numerical experiments on vapor diffusion in polar snow and firn and its impact on isotopes using the multi-layer energy balance model Crocus in SURFEX v8.0, Geosci. Model Dev., 11, 2393–2418, https://doi.org/10.5194/gmd-11-2393-2018, 2018.

Touzeau, A., Landais, A., Stenni, B., Uemura, R., Fukui, K., Fujita, S., Guilbaud, S., Ekaykin, A., Casado, M., Barkan, E., Luz, B., Magand, O., Teste, G., Le Meur, E., Baroni, M., Savarino, J., Bourgeois, I., and Risi, C.: Acquisition of isotopic composition for surface snow in East Antarctica and the links to climatic parameters, The Cryosphere, 10, 837–852, https://doi.org/10.5194/tc-10-837-2016, 2016.

Town, M. S., Warren, S. G., Walden, V. P., and Waddington, E. D.: Effect of atmospheric water vapor on modification of stable isotopes in near-surface snow on ice sheets, J. Geophys. Res.-Atmos., 113, 2008JD009852, https://doi.org/10.1029/2008JD009852, 2008.

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.

Town, M. S., Steen-Larsen, H. C., Wahl, S., Faber, A.-K., Behrens, M., Jones, T. R., and Sveinbjornsdottir, A.: Post-depositional modification on seasonal-to-interannual timescales alters the deuterium-excess signals in summer snow layers in Greenland, The Cryosphere, 18, 3653–3683, https://doi.org/10.5194/tc-18-3653-2024, 2024.

Uemura, R., Masson-Delmotte, V., Jouzel, J., Landais, A., Motoyama, H., and Stenni, B.: Ranges of moisture-source temperature estimated from Antarctic ice cores stable isotope records over glacial–interglacial cycles, Clim. Past, 8, 1109–1125, https://doi.org/10.5194/cp-8-1109-2012, 2012.

Urey, H. C.: The thermodynamic properties of isotopic substances, J. Chem. Soc., 562–581, https://doi.org/10.1039/jr9470000562, 1947.

Van De Berg, W. J., Van Den Broeke, M. R., Van Meijgaard, E., and Kaspar, F.: Importance of precipitation seasonality for the interpretation of Eemian ice core isotope records from Greenland, Clim. Past, 9, 1589–1600, https://doi.org/10.5194/cp-9-1589-2013, 2013.

Van Der Wel, L. G., Gkinis, V., Pohjola, V. A., and Meijer, H. A. J.: Snow isotope diffusion rates measured in a laboratory experiment, J. Glaciol., 57, 30–38, https://doi.org/10.3189/002214311795306727, 2011.

Van Hook, W. A.: Vapor pressures of the isotopic waters and ices, J. Phys. Chem., 72, 1234–1244, https://doi.org/10.1021/j100850a028, 1968.

Vinther, B. M., Jones, P. D., Briffa, K. R., Clausen, H. B., Andersen, K. K., Dahl-Jensen, D., and Johnsen, S. J.: Climatic signals in multiple highly resolved stable isotope records from Greenland, Quat. Sci. Rev., 29, 522–538, https://doi.org/10.1016/j.quascirev.2009.11.002, 2010.

Wahl, S., Steen-Larsen, H. C., Reuder, J., and Hörhold, M.: Quantifying the Stable Water Isotopologue Exchange Between the Snow Surface and Lower Atmosphere by Direct Flux Measurements, J. Geophys. Res.-Atmos., 126, e2020JD034400, https://doi.org/10.1029/2020JD034400, 2021.

Wahl, S., Steen-Larsen, H. C., Hughes, A. G., Dietrich, L. J., Zuhr, A., Behrens, M., Faber, A.-K., and Hörhold, M.: Atmosphere-Snow Exchange Explains Surface Snow Isotope Variability, Geophys. Res. Lett., 49, e2022GL099529, https://doi.org/10.1029/2022GL099529, 2022.

Wahl, S., Walter, B., Aemisegger, F., Bianchi, L., and Lehning, M.: Identifying airborne snow metamorphism with stable water isotopes, The Cryosphere, 18, 4493–4515, https://doi.org/10.5194/tc-18-4493-2024, 2024.

Walter, B., Weigel, H., Wahl, S., and Löwe, H.: Wind tunnel experiments to quantify the effect of aeolian snow transport on the surface snow microstructure, The Cryosphere, 18, 3633–3652, https://doi.org/10.5194/tc-18-3633-2024, 2024.

Wang, Z., Huang, N., and Pähtz, T.: The Effect of Turbulence on Drifting Snow Sublimation, Geophys. Res. Lett., 46, 11568–11575, https://doi.org/10.1029/2019GL083636, 2019.

Wanner, L., Jung, M., Paleri, S., Butterworth, B. J., Desai, A. R., Sühring, M., and Mauder, M.: Towards Energy-Balance Closure with a Model of Dispersive Heat Fluxes, Bound.-Layer Meteor., 190, 25, https://doi.org/10.1007/s10546-024-00868-8, 2024.

Werner, M., Langebroek, P. M., Carlsen, T., Herold, M., and Lohmann, G.: Stable water isotopes in the ECHAM5 general circulation model: Toward high-resolution isotope

https://doi.org/10.5194/egusphere-2026-6263
Preprint. Discussion started: 6 November 2026
© Author(s) 2026. CC BY 4.0 License.



modeling on a global scale, J. Geophys. Res., 116, D15109, https://doi.org/10.1029/2011JD015681, 2011.

Wever, N., Fierz, C., Mitterer, C., Hirashima, H., and Lehning, M.: Solving Richards Equation for snow improves snowpack meltwater runoff estimations in detailed multi-layer snowpack model, The Cryosphere, 8, 257–274, https://doi.org/10.5194/tc-8-257-2014, 2014.

Yuan, Y., Du, T., Wang, H., and Wang, L.: Novel Keeling-plot-based methods to estimate the isotopic composition of ambient water vapor, Hydrol. Earth Syst. Sci., 24, 4491–4501, https://doi.org/10.5194/hess-24-4491-2020, 2020.

Zhang, X., Tian, L., and Liu, J.: Fractionation mechanism of stable isotope in evaporating water body, J. Geogr. Sci., 15, 375–384, https://doi.org/10.1007/BF02837526, 2005.

Zhao, J.-Y., Xiao, W., Zhang, M., Wang, J.-Y., Wen, X.-F., and Lee, X.-H.: Applications and prospect of the flux-gradient method in measuring the greenhouse gases and isotope fluxes, Chin. J. Plant Ecol., 44, 305–317, https://doi.org/10.17521/cjpe.2019.0227, 2020.

Zuhr, A. M., Wahl, S., Steen-Larsen, H. C., Hörhold, M., Meyer, H., and Laepple, T.: A snapshot on the buildup of the stable water isotopic signal in the upper snowpack at EastGRIP on the Greenland Ice Sheet, J. Geophys. Res.-Earth, 128, e2022JF006767, https://doi.org/10.1029/2022JF006767, 2023.