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Isotopic Evidence for Lateral Flow and Diffusive Transport, but Not Sublimation, in a Sloped Seasonal Snowpack, Idaho, USA

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Key Points:
- Large event-specific isotopic steps are preserved throughout most of the winter and early spring
- Advection via sublimation and pervasive flow is minimal, while isotope diffusivity is ~10E−10 m2/s
- Melt channels and progressive loss of isotopically distinct upper layers indicate lateral flow

Supporting Information:
- Supporting Information S1
- Figures S1–S4
- Data Set S1

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Isotopic evidence for lateral flow and diffusive transport, but not sublimation, in a sloped seasonal snowpack, Idaho, USA

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Abstract Oxygen and hydrogen isotopes in snow were measured in weekly profiles during the growth and decline of a sloped subalpine snowpack, southern Idaho, 2011–2012. Isotopic steps (10‰, $\delta^{18}O$; 80‰, δD) were preserved relative to physical markers throughout the season, albeit with some diffusive smoothing. Melting stripped off upper layers without shifting isotopes within the snowpack. Meltwater is in isotopic equilibrium with snow at the top but not with snow at each respective collection height. Transport of meltwater occurred primarily along pipes and lateral flow paths allowing the snowpack to melt initially in reverse stratigraphic order. Isotope diffusivities are ~2 orders of magnitude faster than estimated from experiments but can be explained by higher temperature and porosity. A better understanding of how snowmelt isotopes change during meltout improves hydrograph separation methods, whereas constraints on isotope diffusivities under warm conditions improve models of ice core records in low-latitude settings.

1. Introduction

Over 1 billion people worldwide depend on snowpack-derived water every year for drinking and irrigation [*Bales et al.*, 2006], and snowmelt further contributes disproportionately to western U.S. groundwater resources [*Wilson et al.*, 1980]. Thus, understanding the mechanisms by which snow melts to become streamflow and groundwater, or is returned to the atmosphere by evaporation and sublimation, is immediately crucial to a large segment of the U.S. population and globally. Stable isotopes of oxygen and hydrogen have played key roles in understanding snowmelt-derived runoff generation [e.g., *Rodhe*, 1998; *Klaus and McDonnell*, 2013] and in quantifying atmospheric returns [*Gustafson et al.*, 2010; *Sokratov and Golubev*, 2009]. Both fields of inquiry, however, invoke many assumptions about how hydrologic and chemical processes control the isotopic evolution of a snowpack.

Early snowmelt runoff generation studies [e.g., *Hooper and Shoemaker*, 1986] demonstrated that meltwater from the snowpack does not represent local isotopic equilibrium. Rather, isotopes in interstitial water can be either enriched or depleted relative to equilibrium fractionation with adjacent snow [e.g., *Taylor et al.*, 2001; *Zhou et al.*, 2008; Figure S1], leading to increasingly complex models of snow versus melt isotopic behavior [*Feng et al.*, 2002; *Lee et al.*, 2010]. Flow of meltwater through snow further complicates isotopic distribution. Although early studies of flow through snow conceptualized a one-dimensional system, heterogeneous flow is predicted and apparent from ice layers, ice columns, and lysimeter data [e.g., *Colbeck*, 1979, 1991; *Marsh and Woo*, 1985]. For example, using dye tracer imaging, *Eiriksson et al.* [2013] demonstrated that lateral, or slope-parallel, flow can be significant. Even in combination with models, a major issue for understanding snowpack-derived water resources is that the isotopic contribution of a melting snowpack is not simply derivable from snowpack measurements alone.

The loss of snow water equivalent throughout the winter through sublimation can impact the isotopic composition of a snowpack. Estimates of snow sublimation range widely from less than a few percent to as much as 50% [*Berg*, 1986; *Pomeroy et al.*, 1998; *Earman et al.*, 2006; *Gustafson et al.*, 2010]. In principle, sublimation can be identified isotopically because water vapor lost from the snowpack surface should be replaced by water vapor from deeper in the snow column via interconnected pore space. That is, sublimation loss should induce an upward advective component that shifts the physical location of isotopic boundaries. This effect might have important implications for identifying the timing of climatic shifts in ice core records [*Satake and Kawada*, 1997; *Ekaykin et al.*, 2004; *Neumann and Waddington*, 2004].

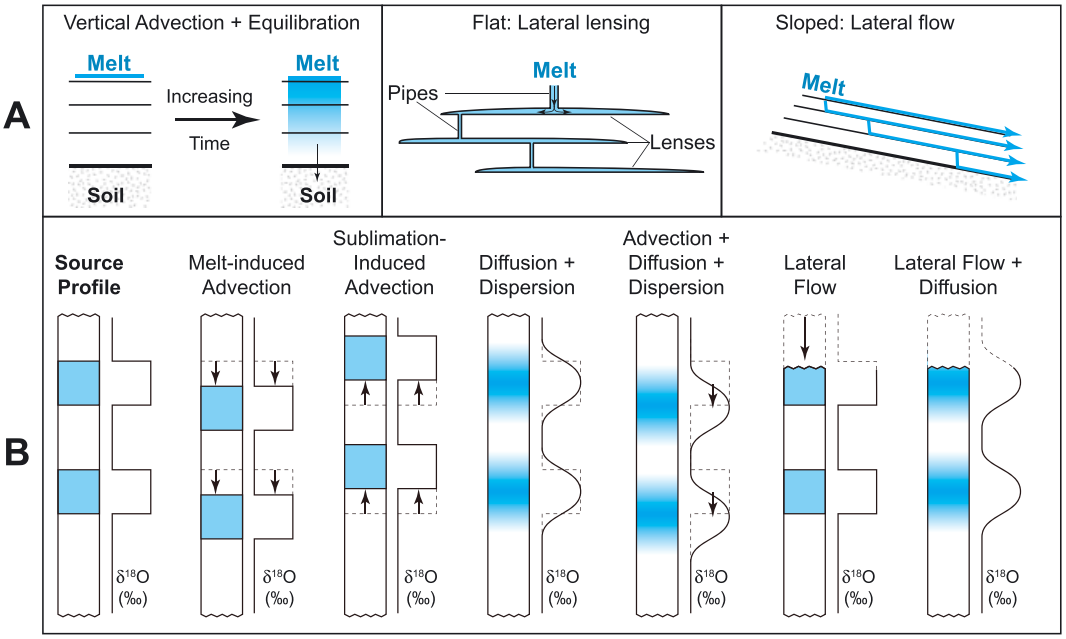

Figure 1. (a) Physical models of fluid movement through a snowpack. Pervasive flow: Meltwater in surface layers flows downward pervasively, advecting its isotope composition. Dispersion plus diffusion broadens any chemical/isotopic discontinuities. Pipes-and-lenses model [cf. *Colbeck*, 1991]: In flat-lying snowpacks, water bypasses sections of the snowpack along discrete channels (pipes) and lenses along relatively impermeable layers, with relatively little interaction with snow layers. Lateral flow: In sloped snowpacks, water bypasses sections of snowpack along channels (pipes) and exits system with relatively little isotopic interaction along either channels or sheets. (b) Models of isotopic behavior in snowpacks showing source profile with isotopically distinctive horizons, advective processes (downward pervasive flow versus upward sublimation), diffusion and dispersion, combined processes, and lateral flow with or without diffusion. Our data generally support the last model—local diffusion plus lateral flow.

Understanding the processes controlling the isotopic evolution of snow and meltwater can inform perceptions about snow hydrology. Yet relatively few direct studies of the temporal evolution of snowpack isotopes have been conducted, and these are limited spatially and temporally (typically 10–20 analyses per profile and 2–4 profiles per season). These studies emphasize how snowmelt differs isotopically from fresh snow [e.g., *Taylor et al.*, 2001], kinetic effects of liquid-solid exchange on isotope profiles [*Zhou et al.*, 2008], and the impact of rain, melting, and slope aspect on isotope profiles [*Taylor et al.*, 2001; *Unnikrishna et al.*, 2002; *Moran and Marshall*, 2009; *Lee et al.*, 2010; *Dahlke and Lyon*, 2013]. Overall, most isotope hydrologists view melt flow as "…the dominant process causing isotopic redistribution in temperate snowpacks" [*Taylor et al.*, 2001].

In this study, we measured the isotopic evolution of the southern Idaho snowpack as it developed through the winter season and melted during the spring. Our work is distinct from previous studies in that we investigated a sloped, relatively warm snowpack—typical of most regions of the western U.S.—and measured weekly profiles over a period of ~3 months. This period includes multiple accumulation and melting events. Our data show large isotopic variations and help address the following questions:

1. How does sublimation affect snow isotopes? Can we quantify the amount of sublimation?
2. How do melting events and downward percolation of liquid water affect isotopic profiles?
3. What general processes best describe the isotopic evolution of snowpacks?

2. Conceptual Models of Snowpack Isotopic Evolution

Meltwater flow through a snowpack may be viewed in reference to pervasive continuum versus preferential flow end-members (Figure 1a). At a flat site, if snowpack responds to melting as a continuum (i.e., pervasive infiltration of liquid downward through the granular matrix), meltwater originating at the surface will permeate lower levels, causing isotopic reequilibration everywhere, before exiting the snowpack at its base. In

contrast, two forms of preferential flow can redistribute meltwater without affecting deeper layers. First, if discontinuous features dominate water movement in snowpacks (i.e., pipes, macropores, and stratigraphic boundaries) [see *Colbeck*, 1979, 1991], meltwater at the surface of a horizontal snowpack will flow downward along discrete pipes and accumulate in lenses with relatively little interaction with intervening layers. Second, in a sloped snowpack, such as is common in mountainous regions, meltwater will exit the snowpack by flowing laterally along specific layers, again with relatively little interaction with deep layers.

The isotopic impacts of these end-member models differ significantly (Figure 1b). Starting with an initial isotopic profile, continuum surface melting and sublimation shift isotopic anomalies uniformly downward and upward, respectively. Diffusion and dispersion smooth isotopic steps, blurring the initial isotopic stratigraphy, but the combination of these continuum processes should still shift isotopic anomalies. In contrast, surface melting in a physically discontinuous snowpack, either without diffusion or with diffusion, simply removes the upper portions of a profile without affecting the underlying snowpack. Thus, an isotopic profile will shorten but otherwise retain its internal isotopic character in terms of isotopic anomaly locations.

In the context of our three questions, these conceptual models predict the following: (1) Sublimation should shift isotopic boundaries upward by a distance proportional to the amount of sublimation. (2) Melting should advect isotopic boundaries downward if flow is pervasive but will not change profiles within the snowpack if flow is preferential. (3) Overall, a static isotopic profile throughout the winter season would support preferential flow; progressive smearing of profiles would indicate diffusion/dispersion (by amounts that can potentially be quantified), and upward or downward shifts of isotopic boundaries would support pervasive flow.

3. Materials and Methods

The Dry Creek Experimental Watershed (DCEW) is a small (28 km^2) watershed 16 km northeast of Boise, Idaho (Figure S2 in the supporting information) spanning elevations of 1000 to 2100 m in the foothills of the Boise Front. The higher elevations of the DCEW are classified as a moist, continental climate with dry summers [*Henderson-Sellers and Robinson*, 1986] and approximately half of the annual precipitation falls as snow. Within the upper reaches of the Dry Creek drainage basin are two instrumented weather/snowpack monitoring sites, the Bogus Basin SNOTEL at 1930 m elevation (U.S. Department of Agriculture-Natural Resources Conservation Service) and the Bogus Ridge study site at ~2100 m operated by the Cryosphere Geophysics and Remote Sensing group at Boise State University. The SNOTEL site has been reporting records of precipitation, snow depth, snow water equivalent, and temperature since 1999, while at the Boise State University installation, geophysical assessment of snow stratigraphy and weather station monitoring have been ongoing since 2010.

Initial snowfall in November 2011 at the Bogus Ridge study site was succeeded by a long dry interval until a major, ~1 m series of snowfall events 18–20 January. Snow from the earlier November storm had $\delta^{18}O$ and δD values of about −14‰ and −100‰ (V-SMOW), respectively, whereas basal snow for the mid-January event had $\delta^{18}O$ and δD values of −20 to −25‰ and −160 to −200‰, respectively. Recognition of this isotopic disparity in late January led to our more focused investigation through the remainder of the winter season. The early winter dry interval produced a persistent distinctive texture for November snow compared to mid-January snow, facilitating discrimination based on physical characteristics. Because of changes in physical conditions (temperature, etc.), even the mid-January event showed distinct internal physical and isotopic layering.

We excavated conventional snow pits [*Fierz et al.*, 2009] on an approximately weekly basis. To avoid biasing flow behavior, pits were sequentially excavated in an upslope, roughly east-to-west direction and were separated by at least 2 m, with walls oriented so no direct sunlight could disturb observations. Each pit was filled in after measurements. We recorded snow height, hardness (Figure S3 and Data Set S1), temperature, density, relative dielectric permittivity/water content, grain sizes, and grain shapes. Isotope analyses were collected in 2 cm intervals, excepting the first profile which sampled at 5 cm intervals above 30 cm height. Fifteen snowpack isotopic profiles were analyzed between January and April 2012, yielding over 2000 individual samples (Data Set S1). We also monitored the dielectric relative permittivity of the snowpack at 1 to 6 h time steps with an upward looking Ground Penetrating Radar (upGPR) ([*Heilig et al.*, 2009, 2010] and see also *Heilig et al.* [2012, 2015] for radar details at this site). Isotope analyses were collected using a Los Gatos Research Liquid Water Isotope Analyzer and are reported relative to Vienna-Standard Mean Ocean Water (V-SMOW).

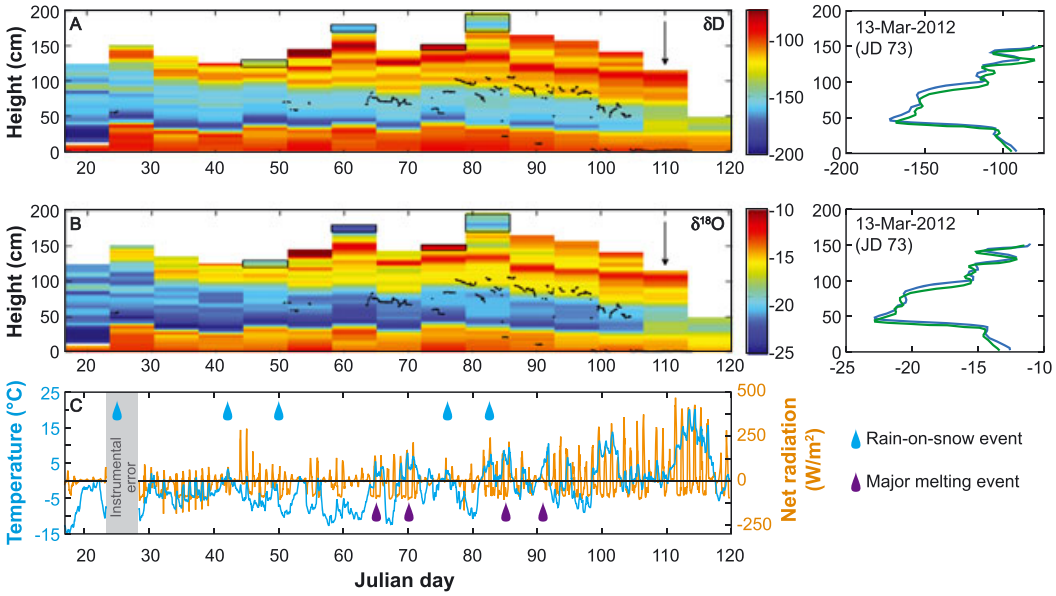

Figure 2. Isotope data for 2011–2012 snowpack. (a and b) δD and $\delta^{18}O$, showing preservation of low-δD, low-$\delta^{18}O$ interval through nearly the entire season. Profiles demonstrate reproducibility of measurements in duplicate transects. Boxes show portions of the snowpack that were removed during subsequent melting events without isotopic impact. Arrows show time when isotopic exchange and homogenization is first obvious. Black dots show upGPR-derived location of the meltwater front. (c) Temperature and net radiation records; prominent rain-on-snow and melting events during snowpack accumulation are identified.

Analytical reproducibility was approximately ±1‰ for δD and 0.15‰ for $\delta^{18}O$ (2σ).The depth of the meltwater front from upGPR data was located as described by *Wever et al.* [2015].

4. Results

4.1. Profile Evolution

All profiles show a dramatic shift from high δD and $\delta^{18}O$ at the snowpack base (November snow) to low δD and $\delta^{18}O$ through the middle (below 100 cm) and upper snowpack (Figure 2). Starting J.D. 27, high δD and $\delta^{18}O$ snow was added, producing a distinct tongue of low δD and $\delta^{18}O$ snow. Such a profile may be typical inasmuch as early- and late-season precipitation occurs at higher temperatures than during midseason, and isotopes in precipitation commonly correlate with seasonal temperature [*Rozanski et al.*, 1992]. A similar profile (isotopically high-low-high) was reported for a snowpack in Japan [*Zhou et al.*, 2008]. Periodically, new events added snow with distinctive δD and $\delta^{18}O$ values, but it is especially important that several of these (boxes, Figure 2) disappear from later profiles. This loss reflects rain-on-snow and/or melting events that apparently removed the top portions of the snowpack without affecting compositions below. For example, despite the occurrence of rain-on-snow initiating approximately J.D. 25, and melting initiating approximately J.D. 66, the low δD and $\delta^{18}O$ tongue persisted nearly intact until J.D. 110 (arrows, Figure 2), when its isotopic values began shifting upward rapidly. The latter isotopic homogenization reflects the initiation of wholesale melting and percolation of the melting front down to the soil surface (Figures 2a and 2b), accompanied by a dramatic reduction in height between J.D. 108 and J.D. 117.

Stacked profiles (Figure 3) allow close comparison of the latter part of the season as sporadic then progressive melting proceeded. Correlation of physically and isotopically distinctive layers (purple dashed lines and black thin lines, Figure 3) illustrates compaction of the snowpack by ~20% between J.D. 80 and J.D. 108. Importantly, physical correlation of snow stratigraphy (snow hardness, purple dashed lines, Figure 3) approximately parallels isotopic correlations.

Spatial heterogeneity in snow height and melt conditions at this site [*Heilig et al.*, 2012, 2015] corresponds with isotopic observations. For example, a thick low-δD, low-$\delta^{18}O$ section between 10 and 30 cm for the J.D. 20 profile is considerably thinner in later profiles because high-δD, high-$\delta^{18}O$ November snow was

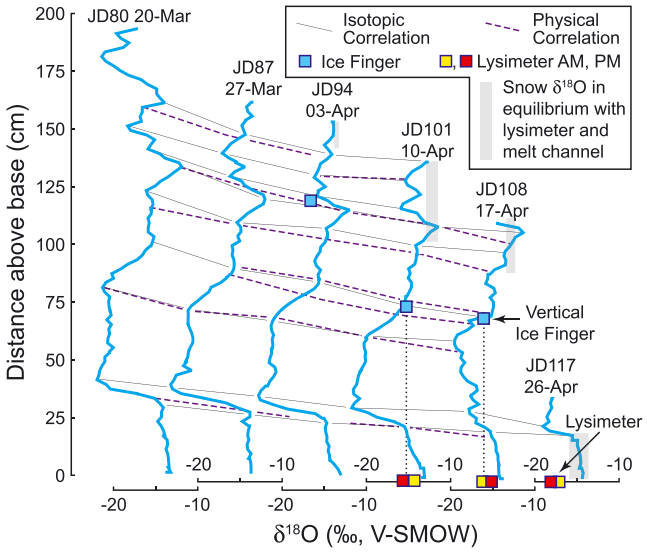

Figure 3. Isotope data for winter/spring 2012, showing physical (thin lines) and isotopic correlations (dashed lines) between profiles, and evolution of snowpack isotopic profile and height. Vertical ice finger locations and compositions (blue boxes), lysimeter compositions (red and yellow boxes), and compositions of snow in equilibrium with ice finger and lysimeter compositions also shown.

extremely thin at the J.D. 20 location. Evidently, the November snow surface had local topography, so more low-δD, low-$\delta^{18}O$ snow accumulated in January at the J.D. 20 profile, leading to a thicker layer than elsewhere. Similarly, snow hardness was measured laterally to collection of isotopes, which led to slight discrepancies in isotopic versus snow hardness correlations. The consistency of profiles, as correlated using physical data, suggests these variations did not adversely affect our study.

4.2. Melting

Late in the season (J.D. 94, J.D. 101, and J.D. 108), vertical ice fingers, representing in situ frozen meltwater, were discovered and sampled at ~120 and 70 cm above snowpack base (Figure 3). The fractionation between this former liquid and local snow, $\Delta(\text{liquid} - \text{snow})$, is approximately −1.5‰ for deuterium and 0.0‰ for oxygen, which is strongly out of isotopic partitioning equilibrium (at 0°C, $\Delta D \sim -21.2‰$ and $\Delta^{18}O \sim -2.9‰$) [*Lehmann and Siegenthaler*, 1991]. This water composition, however, corresponds closely with snowmelt exiting the base of the snowpack (lysimeter values, Figure 3). The snowmelt, in turn, is out of partitioning equilibrium with basal snow in profiles J.D. 101 and J.D. 108 (-1.7 ± 0.8 to $-1.9 \pm 0.4‰$ $\delta^{18}O$ and -10 ± 5 to $-12 \pm 2‰$ δD, 2 SE) but approximates equilibrium for the final profile, J.D. 117 ($\Delta D = -25 \pm 5‰$, $\Delta^{18}O = -3.4 \pm 0.7‰$, 2 SE).

4.3. Small-Scale Variations

Some portions of the profile retained virtually constant compositions through most of the season, including a section between ~55 and 80 cm height above the snowpack base (Figure 4a) and the basal layer (Figure 4b). In contrast, regions with pronounced isotopic anomalies showed small but systematic shifts through time, including the region of minimum δD and $\delta^{18}O$ at ~40 cm height ("Min-δ," Figures 4a and 4c) and a positive anomaly ("High-δ," Figure 4d) that first formed at ~150 cm height and ultimately dropped to ~100 cm height during snowpack compaction. Because isotopic markers did not shift up or down relative to physical markers, these anomalies permit quantification of diffusive processes including placing limits on effective diffusion rates.

5. Interpretations

5.1. Preferential Flow

The remarkably constant isotopic stratigraphy and close correspondence with physical stratigraphy (Figures 2 and 3) strongly suggest that continuum processes including pervasive meltwater advection and sublimation were minimal in this snowpack. For example, if melting affected snowpacks in a continuum fashion, loss of ~30 cm of snow after J.D. 80 and 25 cm after J.D. 101 should have shifted isotopic boundaries downward relative to physical boundaries by 25–30 cm. Yet isotopic boundaries moved negligibly between J.D. 80 and J.D. 101 and perhaps moved upward slightly between J.D. 101 and J.D. 108. This loss of upper portions of the snowpack through melting without major isotopic consequences deeper in our profiles supports the view that melt propagates through discontinuous, preferential pathways, such as pipes and lenses, or as lateral flow along stratigraphic boundaries [e.g., *Colbeck*, 1979, 1991; *Marsh and Woo*, 1985] until shortly before melt out. Such a process has been imaged directly in warm snowpacks [*Humphrey et al.*, 2012; *Winski et al.*, 2012; *Eiriksson et al.*, 2013] (Figure S4) and apparently has little isotopic impact on the residual snowpack until wholesale melting late in the season.

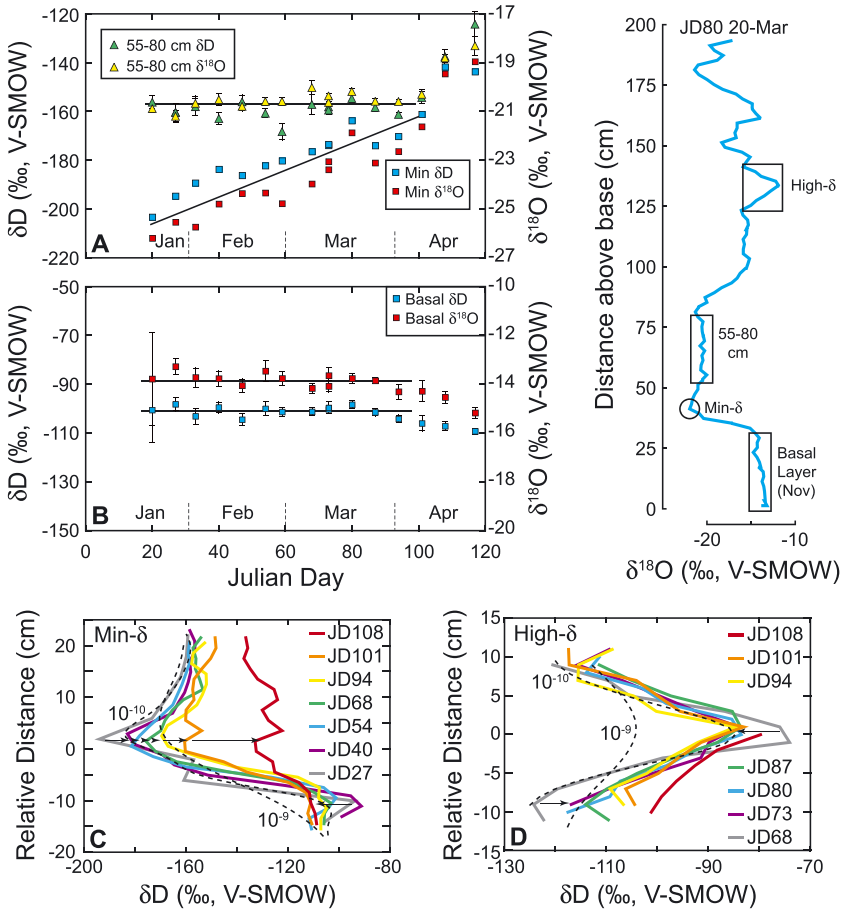

Figure 4. Details of isotopic trends in profiles. Profile at right shows regions of interest. (a) $\delta^{18}O$ and δD for isotopic minimum located at ~40 cm height shows systematic increase through time, whereas average composition for compositionally uniform horizon at 55–80 cm above snowpack base remains invariant until late spring. Error bars are ±2 SE. (b) Mean $\delta^{18}O$ and δD for basal layer of snow deposited in November 2011 shows minimal changes until late spring. Error bars are ±2 SE. (c and d) Evolution of δD profile in Min-δ and High-δ regions showing gradual smoothing. Dashed lines show predicted profiles after 40 days of diffusive smoothing from initial profile (e.g., from J.D. 27 to J.D. 68 in Figure 4c and from J.D. 68 to J.D. 108 in Figure 4d); values are in m^2/s. The J.D. 108 profile in Figure 4d is incomplete and shifted because of melting.

One other study that examined changes to isotope profiles during melting [*Zhou et al.*, 2008] similarly showed that the upper layers simply disappeared from profiles without major change to isotopes lower down, presumably reflecting the same preferential flow process we propose. Our view is supported by ice fingers (former melt) at ~70 and 100 cm height, the consistency of the melt channel composition with melt exiting the base of the snowpack, the location of the meltwater front (from upGPR data; Figures 2a and 2b), and equilibrium partitioning of melt isotopes with snow near or at the top of the snowpack (Figure 3).

5.2. Diffusion

Progressive convergence of compositions and flattening of isotopic profiles (Figure 4) implies isotopic exchange between adjacent layers. Although these changes presumably reflect diffusional equilibration between snow and interstitial water-saturated air, effective diffusivities (D_{eff}) could not have been constant because this would produce steadily decreasing rates of compositional change. For the Min-δ anomaly, the nearly constant rate of change relative to the 55–80 cm layer above (Figure 4a) implies that D_{eff} must have increased steadily through time. Conversely, after initial flattening between J.D. 68 and J.D. 73, compositional stasis of the High-δ anomaly until J.D. 101 implies that D_{eff} must have decreased.

Although not uniquely modelable, the data do place broad constraints on D_{eff} of $\sim 1 \times 10^{-9}$ to 1×10^{-10} m^2/s. Hydrogen and oxygen isotope profiles do not resolve significant differences in D_{eff} (i.e., $D_{eff,O} \sim D_{eff,H}$), as also shown in experiments [*Van der Wel et al.*, 2011]. Our empirical estimates are about 2 orders of magnitude

faster than experimental and theoretical estimates, which range from ~2×10^{-11} (experimental) [*Pohjola et al.*, 2007; *Van der Wel et al.*, 2011] to ~2×10^{-12} m^2/s (theoretical) [*Johnsen et al.*, 2000]. The discrepancies between our data and previous estimates probably reflect a combination of parameters. Simplifying expressions from *Johnsen et al.* [2000], D_{eff} (m^2/s) can be expressed as

$$D_{\mathrm{eff}} \sim 3.5 \cdot T^{0.94} e^{-6133/T} \left[1 - 1.3 \cdot \left(\frac{\rho}{\rho_{\mathrm{ice}}}\right)^2\right] \left(\frac{1}{\rho} - \frac{1}{\rho_{\mathrm{ice}}}\right) \quad (1)$$

where T is in kelvins and ρ is in kg/m^3 ($\rho_{ice} = 917\ kg/m^3$). This equation ignores variations in atmospheric pressure and differences in diffusivities for $HD^{16}O$ versus $H_2{}^{18}O$ [*Johnsen et al.*, 2000; *Van der Wel et al.*, 2011]. Much higher temperatures (0°C versus −19 to −32°C) and lower density (300–350 versus 400–600 kg/m^3) in our natural system increase calculated D_{eff} by 1–2 orders of magnitude. Thus, our results broadly support extrapolation of theoretical calculations to elevated temperatures. The differences in diffusive reequilibration patterns between the Min-δ and High-δ regions (increasing versus decreasing D_{eff}), however, point to the need to further explore diffusive reequilibration of snow isotopes in natural settings, focusing on physical parameters such as grain size, density, and temperature.

5.3. Implications

Preferential flow is important for understanding the isotopic evolution of snowmelt. Commonly, it is assumed that melt interacts continuously with snow, so that the isotope composition of meltwater reflects the lower part of a snowpack, which evolves isotopically as melting progresses. Even independent of our data, this view cannot be wholly correct because studies of coexisting snow and water within a profile show that isotope compositions are not commonly in partitioning equilibrium (Figure S1) [*Taylor et al.*, 2001; *Zhou et al.*, 2008]. Significantly, our data suggest that the earliest meltwater isotopes preferentially represent the upper (melting) portion of a snowpack. In this respect, we predict that during melting, high $\delta^{18}O$-high δD liquid is produced first, followed by low $\delta^{18}O$-low δD; i.e., the evolution of meltwater is the reverse of the original stratigraphy. Indeed, our lysimeter data evolved toward lower $\delta^{18}O$ and δD between J.D. 101 and J.D. 117 as the top of the snowpack melted (Figure 3), ultimately reaching values below the overlying snowpack.

Several studies emphasize increasing $\delta^{18}O$ values during meltout [e.g., *Taylor et al.*, 2001, 2002; *Laudon et al.*, 2002, 2004; *Carey and Quinton*, 2004] and explain this trend in terms of isotopic distillation: at equilibrium $\Delta^{18}O(\text{water-snow}) < 0$, so preferential loss of low $\delta^{18}O$ water drives residual snowpack and later-formed water to higher values. In fact, earliest melt waters can show high $\delta^{18}O$ variability or even decreasing trends [*Taylor et al.*, 2001, 2002], which we propose reflects the loss of upper layers of the snowpack and preferential flowthrough with little isotopic exchange. This view is consistent with studies showing that rain on snow can contribute directly to snow lysimeter outflow [*Maclean et al.,* 1995] and streamflow [*Eiriksson et al.*, 2013], implying that a substantial portion of liquid water can traverse the snowpack without complete equilibration. Continuum equilibrium fractionation processes between snow and water [*Taylor et al.*, 2001] probably occur only after the snowpack, and its isotopes are substantially homogenized (e.g., ≥J.D. 110, Figure 2), yielding the common late-stage trend toward increasing $\delta^{18}O$.

With respect to predicting water recharge, consider an isotopically zoned snowpack with high $\delta^{18}O$ values at top and bottom (higher temperature, early- and late-season precipitation) and low $\delta^{18}O$ in the middle (lower temperature, midseason precipitation). Initial melt should evolve in reverse stratigraphic order, and if the ground is still frozen, high $\delta^{18}O$ water from the top of the snowpack runs off directly into streams. As melting continues and the ground warms, low $\delta^{18}O$ water, derived either directly from the middle of the snowpack or through equilibrium partitioning between snow and water, may contribute preferentially toward saturating upper soil levels and displacing older water. Late-stage high $\delta^{18}O$ water may again contribute preferentially to streamflow and/or force low $\delta^{18}O$ soil water deeper. Thus, streamflow could see the highest $\delta^{18}O$ waters, soils the lowest $\delta^{18}O$. Future hydrograph separation studies may need to account for these differences in the timing of melting and how the melt interacts (or not) with the snowpack.

Studies of diffusive smoothing of isotopes mainly focus on polar records where temperatures are extremely low. This work generally shows that isotopic smoothing does occur with a decrease in isotopic amplitude up to an order of magnitude [e.g., *Johnsen et al.*, 2000]. In the context of more temperate ice records

[e.g., *Thompson et al.*, 1985], which are subject to melting in today's warming climate [e.g., *Thompson et al.*, 2006], diffusive smoothing should be much greater. Better understanding of diffusion rates in snow at relatively high temperatures would help extract additional seasonal information from these records, much as has been possible with polar ice cores.

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