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Abstract

Water stable isotopes in Greenland ice core data provide key paleoclimatic information, and have been compared with precipitation isotopic composition simulated by isotopically-enabled atmospheric models. However, post-deposition processes linked with snow metamorphism remain poorly documented. For this purpose, a monitoring of the isotopic composition ($\delta^{18}\text{O}$, δD) of surface water vapor, precipitation and samples of top (0.5 cm) snow surface has been conducted during two summers (2011–2012) at NEEM, NW Greenland. The measurements also include a subset of ^{17}O -excess measurements over 4 days, and the measurements span the 2012 Greenland heat wave. Our observations are consistent with calculations assuming isotopic equilibrium between surface snow and water vapor. We observe a strong correlation between surface vapor $\delta^{18}\text{O}$ and air temperature ($0.85 \pm 0.11 \text{ ‰ } ^\circ\text{C}^{-1}$ ($R = 0.76$) for 2012). The correlation with air temperature is not observed in precipitation data or surface snow data. Deuterium excess (d-excess) is strongly anti-correlated with $\delta^{18}\text{O}$ with a stronger slope for vapor than for precipitation and snow surface data. During nine 1–5 days periods between precipitation events, our data demonstrate parallel changes of $\delta^{18}\text{O}$ and d-excess in surface snow and surface vapor. The changes in $\delta^{18}\text{O}$ of the vapor are similar or larger than those of the snow $\delta^{18}\text{O}$. It is estimated that 6 to 20 % of the surface snow mass is exchanged with the atmosphere using the CROCUS snow model. In our data, the sign of surface snow isotopic changes is not related to the sign or magnitude of sublimation or condensation. Comparisons with atmospheric models show that day-to-day variations in surface vapor isotopic composition are driven by synoptic weather and changes in air mass trajectories and distillation histories. We suggest that, in-between precipitation events, changes in the surface snow isotopic composition are driven by these changes in surface vapor isotopic composition. This is consistent with an estimated 60 % mass turnover of surface snow per day driven by snow recrystallization processes under NEEM summer surface snow temperature gradients. Our findings

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mate signal only during snowfall events, and the covariance between precipitation and temperature (Persson et al., 2011); (ii) changes in relationships between surface and condensation temperature linked with changes in boundary layer dynamics; (iii) changes in moisture sources and distillation along air mass trajectories (Masson-Delmotte et al., 2005b). Simulations of water stable isotopes within regional or general circulation atmospheric models have been used to explore the drivers of changes in isotope–temperature relationships, and evaluate models against ice core data. Such studies have confirmed the importance of precipitation seasonality for glacial-interglacial changes and highlighted the role of changes in atmospheric circulation and moisture sources (Cuffey and Steig, 1998; Jouzel et al., 1997; Krinner et al., 1997).

Second-order parameters such as the d-excess ($d\text{-excess} = \delta D - 8 \times \delta^{18}O$) and more recently the ^{17}O -excess are expected, based on modeling, to preserve the signature of the moisture source. The reason for this is the isotopic composition of source moisture being controlled by kinetic effects at evaporation related to wind speed, sea surface temperature and relative humidity (for d-excess) or relative humidity (for ^{17}O -excess) (Merlivat and Jouzel, 1979; Landais et al., 2011; Johnsen et al., 1989). Measurements of d-excess in Greenland ice cores have therefore been used to infer present and past evaporation conditions and locate the main moisture sources (Johnsen et al., 1989; Masson-Delmotte et al., 2005b; Steen-Larsen et al., 2011). A few measurements of ^{17}O -excess conducted at the seasonal scale show a seasonal cycle in anti-correlation with respect to Greenland temperature and $\delta^{18}O$ (Landais et al., 2011). Existing ice core records have revealed a strong anti-correlation of d-excess with $\delta^{18}O$, reflecting the impact of changes in condensation temperature on (i) the ratio of equilibrium fractionation for δD and $\delta^{18}O$ (Merlivat and Nief, 1967; Ellehoj et al., 2013), and (ii) kinetic fractionation on ice crystals (Jouzel and Merlivat, 1984). However, ice cores have also depicted specific d-excess signals spanning both present and past decadal to millennial and orbital timescales, interpreted as reflecting changes in moisture source conditions (Steffensen et al., 2008; Masson-Delmotte et al., 2005a; Steen-Larsen et al., 2011). However simulating for instance glacial-interglacial changes in Greenland precipitation

d-excess consistent with ice core data remains difficult (Werner et al., 2001; Risi et al., 2010).

The interpretation of the ice core data and the comparison with atmospheric model results implicitly rely on the assumption that the snowfall precipitation signal is perfectly preserved in the snow/ice matrix. However, post-deposition processes associated with wind scouring and firn isotopic diffusion are known to introduce a “post deposition noise” in the surface snow. Comparisons of isotopic records obtained from nearby shallow ice cores have allowed to estimate a “signal to noise” ratio with respect to a common “climate” signal (Fisher and Koerner, 1994, 1988; White et al., 1997; Steen-Larsen et al., 2011). Diffusion lengths of typically 7–10 cm have been diagnosed in Greenland ice cores based on the loss of magnitude of seasonal cycles in shallow ice cores (Johnsen et al., 2000), and statistical methods have been used to “back-diffuse” ice core signals for the purpose of identifying seasonal cycles for dating ice cores or for the correction of loss of amplitude for winter and summer water stable isotope signals (Johnsen, 1977). In parallel, numerical snow models have been developed in order to represent the surface snow metamorphism, a process associated with vapor-snow mass exchanges in the upper centimeters of the firn. Snow metamorphism affects changes in grain size, surface albedo, and more generally the surface snow energy budget and mass balance (Vionnet et al., 2012; Brun et al., 2011). Snow models are growingly incorporated in atmospheric – land surface models, or used for the coupling between atmospheric and ice sheet models (Rae et al., 2012), but none of them is yet equipped with the explicit modeling of water stable isotopes.

The motivation for our study is to investigate whether post-deposition has detectable impacts on (i) isotope–temperature relationships; (ii) d-excess vs. $\delta^{18}\text{O}$ relationships; and (iii) processes affecting surface snow isotopic composition in-between precipitation events.

For this purpose, a surface water isotope-monitoring program has been established at the NEEM site, NW Greenland, with the goal to improve the interpretation of the NEEM deep ice core through a better understanding of the processes controlling the

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water isotopic composition measured in the ice core record, at the event scale. In summer 2008, this program combined event and sub-event precipitation sampling for water stable isotope analysis, together with shallow ice core data, and water vapor monitoring using cryogenic trapping (Steen-Larsen et al., 2011). These first measurements showed parallel isotopic variations between vapor and snowfall. The surface water vapor isotopic composition was found to predominantly be close to isotopic equilibrium with surface snow (Steen-Larsen et al., 2011). The resolution of water vapor isotope observations was subsequently strongly improved thanks to continuous, in situ measurements using cavity-ringdown spectrometers (CRDS) during summer 2010 (Steen-Larsen et al., 2013). The day-to-day variability of the surface atmospheric water vapor $\delta^{18}\text{O}$ was in good agreement with the results from an atmospheric general circulation model, LMDZiso, nudged to atmospheric analyses. While the model did not capture the magnitude of vapor d-excess variations observed at NEEM, it showed that high d-excess events coincided with inflows of moisture originating from the Arctic. This finding demonstrated that large-scale atmospheric circulation changes drive day-to-day variations of NEEM surface water vapor. During clear-sky days, CRDS measurements conducted at heights from 1 to 13 m showed a strong diurnal variability in humidity and $d^{18}\text{O}$, interpreted to reflect the fact that the snow surface acts as a moisture source (sink) during the warming (cooling) phase.

Altogether these preliminary findings altogether have qualitatively evidenced interactions between the atmospheric water vapor and the snow surface, which has motivated further observations. Here, we report new data acquired at NEEM in summer 2011 and spring-summer 2012 which include a systematic monitoring of surface snow and water vapor, precipitation (only for 2011), and the first ^{17}O -excess measurements simultaneously conducted during 4 days on water vapor and surface snow. The CROCUS snow model (Vionnet et al., 2012) has been adapted to the NEEM site and used to calculate the snow-air net mass exchange.

This manuscript is organized in the following way. Section 2 describes the NEEM site, sampling strategy and analytical methods, as well as the set up for the CRO-

CUS model and the basis for vapor-surface snow equilibrium calculations. Section 3 describes the results of the new isotopic composition measurements and reports the $\delta^{18}\text{O}$ -temperature and d-excess- $\delta^{18}\text{O}$ relationships, as well as the ^{17}O -excess results. Section 4 is finally devoted to the discussion of our results and the comparison with CROCUS calculations, in order to qualitatively understand the processes controlling the isotopic composition of surface snow.

2 Methods

2.1 NEEM site description

The sampling and measurements were carried out as part of the international deep drilling program conducted at NEEM, NW Greenland (77.45° N 51.05° W, 2484 m.a.s.l.), from 2007 to 2012, providing climatic and glaciological information back to the last interglacial period (NEEM Community members, 2013; Steen-Larsen et al., 2011). An automatic weather station was installed at the NEEM site in 2006 to supply meteorological observations. Air temperature and relative humidity (post corrected to respect to ice) were measured using Campbell Sci. HMP45C ($\pm 0.1^\circ\text{C}$ and $\pm 5\% < 90\% \text{ RH}$ and $\pm 10\% > 90\% \text{ RH}$), wind direction and speed using RM Young propeller-type vane ($\pm 5^\circ$ and $\pm 0.1 \text{ ms}^{-1}$), and station pressure using Vaisala PTB101B ($\pm 0.1 \text{ mb}$) (Steffen and Box, 2001; Steffen et al., 1996). The estimated mean summer (JJA) temperature at NEEM is $\sim -11 \pm 5^\circ\text{C}$ (1σ based on 2006–2011), but the summers of 2011 and 2012 were found to be significantly warmer ($\sim -7^\circ\text{C}$) than average. The annual mean accumulation rate from 1964 to 2005 is estimated at 20 cm a^{-1} (water equiv.), with a large fraction (between 2.5 and 4.5) of precipitation occurring in JJA compared to DJF (Steen-Larsen et al., 2011). During the field campaigns 2009–2011 a thermistor string was installed in the top snowpack between the surface and two meters depth. Extra sensors were installed in the top snow layer throughout the season. The resistances of the thermistors were recorded using a Pico-technology 24-bit data logger.

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measurements of the drift-standard with known isotopic composition used to correct the drift of the instrument are shown in the Supplement Figs. S1 and S2. We do not expect that the data gaps due to drift-calibrations produce any significant effect on the analysis. While the vertical gradients of water stable isotopes were investigated in Steen-Larsen et al. (2013), it was shown that they only depict diurnal gradients, and that water vapor isotopes co-vary at the day-to-day scale at different heights. This was also verified for our new measurements (not shown) and we will here only describe and discuss the measurements obtained at 3 m height (Table 1). Note that the 2011 sampling was initiated on Day 185 (5 July) (Noon UTC on 1 January is Day 0.5), while the 2012 sampling covers a longer time period, starting on Day 141 (21 May). In both years, measurements stopped at Day 215 (4/5 August) due to the closure of the summer camp. Hereafter, vapor measurements are reported as $\delta^{18}\text{O}_v$ and d-excess_v.

2.3 Precipitation and snow surface samples

A white table with ~20 cm high sides covering an area of ~0.7 m² made out of opaque Plexiglas was installed on the edge of the clean air sector next to the atmospheric vapor station in order to collect precipitation. The table was installed at a height of ~1.5 m to limit the collection of blowing snow. Precipitation was collected in 2011 on event and sub-event basis as reported in Steen-Larsen et al. (2011), leading to 41 samples (Table 1). We have discarded two outliers with very low d-excess_p of ~ -10‰.

A designated area (5 × 5 m) was marked off from which the snow surface samples were collected. The snow surface samples were collected from the top ~0.5 cm of the snow surface. All samples were collected within the designated area but never at any previous sampled place. A sample was collected by taking surface area from a 15 × 15 cm area every 12 h (a few (25) samples were only collected every 24 h in beginning of 2012). Altogether, 51 samples were collected in 2011, and 122 samples in 2012.

Snowfall and surface snow samples were melted in sealed plastic bags before being transferred to a vial, which was kept frozen until measurements. The precipitation and

snow surface samples were measured using a Picarro inc. liquid analyzer at Laboratoire des Sciences du Climat et de l'Environnement (LSCE), Gif-Sur-Yvette and Centre for Ice and Climate (CIC), Copenhagen (see Table 1). Hereafter, precipitation measurements are reported as $\delta^{18}\text{O}_p$ and $d\text{-excess}_p$, and surface snow measurements as $\delta^{18}\text{O}_s$ and $d\text{-excess}_s$.

2.4 ^{17}O -excess of snow surface and atmospheric water vapor

During the period from 11 to 14 July 2011, a specific sampling was conducted to explore day-to-day variations in ^{17}O -excess of snow surface and atmospheric water vapor. Using a cryogenic trapping system similar to Steen-Larsen et al. (2011), water vapor from respectively 1 and 10 m above the snow surface was collected with the aim of measuring ^{17}O -excess. Vapor trapping was conducted during 6 h, leading to 2 samples per level per day. Because results are very similar at the two heights, we only report here the 10 m data. ^{17}O -excess measurements were also conducted on the corresponding subset of surface snow samples. The ^{17}O -excess of the snow surface and atmospheric water vapor samples were measured at Laboratoire des Sciences du Climat et de l'Environnement (LSCE) using the fluorination technique (Barkan and Luz, 2005; Landais et al., 2012) (Table 1). The same notation is used for ^{17}O -excess as for $\delta^{18}\text{O}$ and $d\text{-excess}$ to report vapor data ($^{17}\text{O}\text{-excess}_v$) and surface snow data ($^{17}\text{O}\text{-excess}_s$).

2.5 Calculations of equilibrium between surface snow and water vapor isotopic composition

In order to investigate the relationship between surface water vapor and snow, we use the observed air temperature (see Sect. 2.1) and vapor isotopic composition (see Sect. 2.2) (integrated 12 h back) to estimate the expected isotopic composition of the snow surface at equilibrium with the water vapor. The calculation is performed (i) using the fractionation coefficients for liquid water extrapolated below 0°C (Majoube, 1971),

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then (ii) using the fractionation coefficients for ice (Majoube, 1970) (Merlivat and Nief, 1967) and (iii) new fractionation coefficients for ice determined by Ellehoj et al. (2013). The reason for making calculations for liquid water is due to summer temperatures at NEEM surface close to 0 °C (Fig. 1). Results are reported in Sect. 3.8.

2.6 LMDZiso

The measurements are compared with simulations using the atmospheric general circulation model LMDZ (Hourdin et al., 2006). It is enabled with water isotopes (Risi et al., 2010) and the simulated water isotopic distribution has been validated at various time scales globally (Risi et al., 2010, 2013) and over Greenland (Steen-Larsen et al., 2011; Ortega et al., 2013). The details of isotopic implementation are given in Risi et al. (2010).

LMDZ is used here with resolution of 2.5° in latitude, 3.75° in longitude and 39 vertical levels. The first atmospheric layer is 60 m thick. The simulated winds are nudged towards those of the ECMWF operational analyses, allowing the model to reproduce the day-to-day large-scale atmospheric conditions (Risi et al., 2010; Yoshimura et al., 2008).

Snow is represented as a single, vertically homogeneous layer. Its inputs are snowfall and frost and its outputs are melt, sublimation and calving. Melt and sublimation are assumed not to fractionate. Frost is formed in equilibrium with the water vapor of the first atmospheric layer. Calving occurs whenever the snow height exceeds a maximum capacity of 3 m. In practice, at NEEM, the snow layer is almost always at its maximum capacity. Therefore, the isotopic composition of the snow varies very little at the daily scale in the model, since the snow height variations through snowfall or through frost are several orders of magnitudes smaller than the total snow layer height.

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2.7 CROCUS

We run the detailed snowpack model CROCUS (Vionnet et al., 2012) to drive the energy and mass fluxes exchanged between the atmosphere and the snow surface as well as the snow grain metamorphism. The meteorological forcing required by the model were extracted from ERA-interim (Dee et al., 2011) and projected onto NEEM as in Brun et al. (2013). The precipitation rate was multiplied by a factor 2 in order to match the annual accumulation recorder at NEEM. This adjustment does not significantly change the simulated heat and vapor surface fluxes. ERA-interim air temperature and humidity were compared with observations from the GC-NET station at NEEM (Steffen and Box, 2001; Steen-Larsen et al., 2011). A strong linear correlation of ERA-interim temperature vs. observations is found with $< 1^{\circ}\text{C}$ deviation to the 1 : 1 line for the range -20°C to 0°C . Similar strong linear correlation between ERA-interim air humidity and observation is found with < 100 ppmv deviation to the 1 : 1 lines for the range 2000 ppmv to 8000 ppmv. An evaluation of the snow surface temperature simulated with CROCUS was also carried out. As part of the snow surface monitoring program we installed thermistor probes from the surface and to a depth of ~ 2 m. However due to solar heating of the top thermistors and uncertainty in vertical position we use surface temperature estimates from MODIS MOD11_L2 from TERRA and MYD11_L2 from AQUA (Wan, 2009) for clear sky days and find a very good agreement to surface temperature estimates from CROCUS (See figure S3 in Supplement). This performance makes it possible to reliably calculate the vapor fluxes exchanged between the atmosphere and the surface snow layers.

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3 Results

3.1 NEEM climate during summer 2011 and spring-summer 2012

The mean air temperature during summer 2011 (day 185–216) and spring-summer 2012 (day 140–216) is respectively -6.9°C and -6.8°C (See Fig. 1). Diurnal cycles have a magnitude of $\sim 5\text{--}10^{\circ}\text{C}$ during clear sky days, but with a reduced magnitude during precipitation events, due to a reduced nocturnal minimum probably due to the greenhouse effect of clouds. Daily averaged temperatures (Fig. 1) show day-to-day variations associated with changes in large-scale atmospheric circulation, well captured by the LMDZiso nudged simulation (Fig. 1). The summer surface air temperatures reached for several episodes levels close to or slightly above 0°C during both 2011 and 2012. Particularly remarkable are two events in 2012 around day 191–196 (10–15 July) and around day 209 (28 July) during which melting of the surface layer was observed to occur (indicated on Fig. 1 with grey band). As reported in Steen-Larsen et al. (2011) based on satellite microwave data, starting in 1987, melt has only occurred in summer 2005 (24 July 2005, 25 July 2005). This 2005 event was likely been caused by the same cyclonic event and not like 2012 from two separate events. The 2012 measurements span the spring-summer transition around day 147–150. This transition was not recorded in the 2010 or 2011 datasets, as measurements were not initiated before the summer period had started. During the transition over these days the air temperature increased by $\sim 20\text{--}25^{\circ}\text{C}$ (minimum to maximum)

3.2 Variability of $\delta^{18}\text{O}_v$ and d-excess_v during summer 2011 and spring-summer 2012

We describe the 2011 and 2012 vapor $\delta^{18}\text{O}_v$ and d-excess_v data, report the mean value and range of variations (Table 2), and highlight two remarkable events recorded in spring and summer 2012.

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from $\sim 10\text{‰}$ to $\sim 46\text{‰}$ with a similar average of $\sim 23\text{‰}$. The range and mean value is comparable to observations from 2010 reported in Steen-Larsen et al. (2013) where a mean value of $\sim 26\text{‰}$ was observed together with observations of large d-excess values ($> 40\text{‰}$). Data from 2010, 2011, and 2012 show all a general anti-correlation between $\delta^{18}\text{O}_v$ and d-excess_v for synoptic events.

Based on the anomaly of the spring signal in 2012, we isolate this period when investigating $\delta^{18}\text{O}$ -temperature (Sect. 3.6) or d-excess- $\delta^{18}\text{O}$ (Sect. 3.7) relationships.

3.3 Variability of $\delta^{18}\text{O}_p$ and d-excess_p during summer 2011

The precipitation samples collected during summer 2011 cover four precipitation events (Fig. 1). The $\delta^{18}\text{O}_p$ and d-excess_p of the samples vary respectively between $\sim -35\text{‰}$ and $\sim -18\text{‰}$ with a mean value (arithmetic mean) of $\sim -26\text{‰}$ and between $\sim 2\text{‰}$ and $\sim 20\text{‰}$ with a mean value of $\sim 11\text{‰}$ (Table 2). No systematic pattern of change can be observed during precipitation events regarding the trends in the isotopic composition of precipitation or vapor.

3.4 Variability of $\delta^{18}\text{O}_s$ and d-excess_s during summer 2011 and spring-summer 2012

During summer 2011, $\delta^{18}\text{O}_s$ varies between $\sim -33\text{‰}$ and $\sim -21\text{‰}$ with an average value of $\sim -25\text{‰}$. The range of values is more depleted compared to $\delta^{18}\text{O}_p$, and the mean value is 1‰ lower than the average $\delta^{18}\text{O}_p$. A larger range of variations is observed in spring-summer 2012, with $\delta^{18}\text{O}_s$ varying between $\sim -36\text{‰}$ and $\sim -15\text{‰}$ with an average value of -24‰ (Similar value is obtained when averaging over same data collection period in 2011).

For d-excess_s, summer 2011 is characterized by values between $\sim 7\text{‰}$ and $\sim 19\text{‰}$ with a mean level of $\sim 12\text{‰}$. The range is smaller than depicted in the precipitation data (consistent with different sampling durations), and the mean value is identical. A larger range of variations is observed in spring-summer 2012, from $\sim 3\text{‰}$ and $\sim 24\text{‰}$ with

composition of snow surface and temperature, we use the mean air temperature averaged over 12 h prior to collection of the snow surface sample. The result does not depend on the integration time (tested over durations of 1 to 48 h). No significant correlation is detected between $\delta^{18}\text{O}_s$ and air temperature during 2011. In summer 2012, a significant correlation is associated with a slope of $\sim 0.31 \pm 0.09 \text{‰} \cdot \text{°C}^{-1}$ ($R = 0.32$) (Fig. 2, Table 2) due to the presence of snow surface samples collected before the spring-summer transition.

3.7 d-excess- $\delta^{18}\text{O}$ relationships in vapor, precipitation and surface snow in summer 2011 and 2012

For summer vapor data (excluding the 2012 spring-summer transition), daily d-excess_v is anti-correlated with $\delta^{18}\text{O}_v$ with slopes of $\sim -1.11 \pm 0.04 \text{‰}$ per ‰ in 2011 ($R = -0.76$) and $\sim -1.03 \pm 0.01 \text{‰}$ per ‰ in 2012 ($R = -0.86$) (Fig. 3). Similar slopes are obtained from hourly data. By contrast, a weaker relationship associated with a smaller slope emerges from 2011 surface snow data ($\sim -0.31 \pm 0.15 \text{‰}$ per ‰, $R = -0.27$) as well as 2012 surface snow data ($\sim -0.44 \pm 0.07 \text{‰}$ per ‰, $R = -0.53$). The latter is closer to the relationship obtained from precipitation data both from our 2011 dataset ($\sim -0.47 \pm 0.11 \text{‰}$ per ‰, $R = -0.57$) and reported from 2008 precipitation data by Steen-Larsen et al., 2011 (-0.43 ± 0.14 , $R = -0.58$).

3.8 Comparison between measured surface snow isotopic composition and values calculated from isotopic equilibrium with surface water vapor

To study the relationship between the surface snow and water vapor isotopic composition we compare the calculated surface snow isotopic composition, using the method described in 2.7, to the observed data in Fig. 4, with the mean summer isotopic values summarized on Table 3.

The observed $\delta^{18}\text{O}_s$ variations appear to lie in between the theoretical calculations of a surface snow at equilibrium with vapor using liquid water and ice. During precipita-

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tion events, a better agreement is obtained using fractionation with water, while a better agreement is found in general for calculations of fractionation with ice in-between precipitation events.

The observed d-excess_s data also lie between the results obtained when using respectively fractionation with liquid water and with ice, but a much better fit is obtained when using the new ice fractionation coefficients (Ellehoj et al., 2013). Calculations with the old ice fractionation coefficients led to values much lower than observed.

From this comparison, the hypothesis of equilibrium between vapor and surface snow cannot be rejected. We note that equilibrium between surface vapor and precipitation was also suggested from 2008 $\delta^{18}\text{O}$, d-excess and ^{17}O -excess data (Steen-Larsen et al., 2011; Landais et al., 2012).

4 Discussion

4.1 Robustness of findings when comparing different years:

1. We find similar mean temperature, $\delta^{18}\text{O}$, and d-excess values in vapor and surface snow for 2011 and 2012;
2. Comparing results from different years a comparable relationship between d-excess and $\delta^{18}\text{O}$, $\delta^{18}\text{O}$ -temperature in vapor (2008, 2010, 2011, 2012), precipitation (2008, 2011), and surface snow (2011 and 2012).
3. We conclude that different $\delta^{18}\text{O}$ -temperature and d-excess- $\delta^{18}\text{O}$ slopes exist between vapor and surface snow
4. Our observations indicate that surface snow is at equilibrium with surface water vapor (or vice versa)

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5. The good agreement between LMDZiso simulations of day-to-day variations in $\delta^{18}\text{O}_v$ indicates that water vapor isotope variations are driven by large-scale circulation features.

We will therefore subsequently assume that the same processes are a play for several summers despite different meteorological conditions. We seek to investigate what drives the observed variation of surface snow isotopic composition in-between snowfall events by: comparing magnitudes of change in surface snow isotopes with simultaneous change in water vapor isotope (Sect. 4.2); investigate possible processes that can explain the observed variation (Sect. 4.3).

4.2 Magnitude of isotopic changes in surface snow and surface water vapor

We focus on changes in snow surface isotope values occurring in-between precipitation events. We have for one period (between day 191 and 194 of 2011) concomitant $\delta^{18}\text{O}$, d-excess and ^{17}O -excess data. During this period, $\delta^{18}\text{O}_s$ increases by $\sim 7\%$ in parallel with a $\sim 8\%$ increase in $\delta^{18}\text{O}_v$. Similarly, d-excess_s decreases by $\sim 6\%$ while d-excess_v decreases with $\sim 10\%$. Finally, ^{17}O -excess_s increases by ~ 15 ppm while ^{17}O -excess_v increases by ~ 40 ppm. During this event, the isotopic composition of the surface snow varies in parallel with that of water vapor, albeit with a damped magnitude.

In order to test if this finding is valid for all periods in-between precipitation, Table 3 shows a systematic investigation of the sign and magnitude of changes in $\delta^{18}\text{O}_s$, $\delta^{18}\text{O}_v$, d-excess_s and d-excess_v. Note that no lead or lag time could be identified for the simultaneous changes in the surface snow and water vapor isotopic composition seen in Fig. 1 and documented in Table 4, due to the time resolution of the surface snow measurements.

Table 4 demonstrates that changes in snow surface and water vapor isotopes systematically occur in the same direction, with the exception of two periods (2011, days 204–207 and 211–214) where significant changes are detected in d-excess_s but no

change is detected in d-excess_v. Magnitudes of observed changes in $\delta^{18}\text{O}_s$ are either similar (2 situations out of 9) or predominantly smaller (7 cases out of 9) than changes in the $\delta^{18}\text{O}_v$. Results are more ambiguous for d-excess_s, which is possibly due to the magnitude of the signal with respect to the larger analytical uncertainty associated with water vapor measurements.

From this co-evolution emerges the question that we now aim to address: are the changes in the snow surface isotopes caused by changes in the water vapor isotopic composition, or are the water vapor isotope changes caused by changes in the snow surface isotopic composition?

4.3 Possible causes for change in snow surface and water vapor isotopes

We will resort to only discuss changes in-between precipitation events. If changes in the snow surface isotopes lead changes in the water vapor isotopic composition, we need to explain why the isotopic composition of the snow surface isotopes is modified. Our first hypothesis is that the surface snow isotopic composition is affected by the isotopic composition of the firn below the top layer, itself reflecting the isotopic composition, at the time of campaign, of spring or winter snowfall. If this would be the case, one would expect the surface snow isotopic composition to show a systematic decrease towards more depleted spring values. However, our observations depict 3 out of 9 cases where $\delta^{18}\text{O}_v$ is increasing, and 6 cases out of 9 when it is decreasing. A second hypothesis could be that, for a given $\delta^{18}\text{O}_s$, changes in surface air temperature affect the isotopic composition of the water vapor formed at equilibrium with this snow. However, a 10 °C warming only leads to a change of ~2 ‰ in $\delta^{18}\text{O}_v$ under the assumption of isotopic equilibrium. This mechanism is therefore unable to account for the magnitude of $\delta^{18}\text{O}_s$ changes (Table 4). Finally, we cannot identify any mechanism, which could explain why changes in $\delta^{18}\text{O}_s$ (alone) can lead to a larger change $\delta^{18}\text{O}_v$. We now investigate how mass fluxes between surface snow and atmosphere may explain this finding.

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would be expected based on the preceding precipitation event. We do therefore not find support for the hypothesis that sublimation causes sampling of the snow surface to be similar to playing a tape recorder in reverse.

We also hypothesize that surface snow isotopic composition is reflecting surface hoar formation. During NEEM field campaigns, such surface hoar formation was observed during clear-sky nights in response to condensation of water vapor in the air on to the cold snow surface. Similarly, condensation (negative mass flux from surface to atmosphere) is also simulated by CROCUS on days with a clear diurnal cycle. In these cases the simulated diurnal amount of sublimation is much larger than the simulated diurnal amount of condensation. It is therefore likely that the night frost would vanish throughout the day. The ice self-diffusion has a value of $\sim 5 \times 10^{-8} \text{ m}^2 \text{ yr}^{-1}$ at -20°C (Whillans and Grootes, 1985) and the characteristic time for diffusion in a grain in the surface layer is about 10^{-2} yr (Waddington et al., 2002). We therefore do not expect the snow surface to take up a significant isotopic signal from the surface hoar formed during the night. We do notice that, during clear-sky days, the diurnal cycle of sublimation/condensation is reflected in the water vapor diurnal cycles of both $\delta^{18}\text{O}_v$ and $d\text{-excess}_v$ as also observed by Steen-Larsen et al. (2013).

At the day-to-day scale, our final hypothesis is that changes in snow surface isotopic composition are driven by synoptic changes in the atmospheric water vapor isotopic composition. First, the day-to-day variability of $\delta^{18}\text{O}_v$ is well captured by the LMDZiso atmospheric general circulation model, nudged to ECMWF operational analysis wind fields (Fig. 1). We therefore conclude that changes in the snow surface isotopes do not drive the variability in the water vapor isotopes, which instead is driven by changes in large-scale winds and moisture advection. The fact that changes in $\delta^{18}\text{O}_v$ in-between precipitation events are always greater than or equal to the changes in $\delta^{18}\text{O}_s$ (Table 3) supports the hypothesis that the snow surface isotopic composition takes up part of the atmospheric water vapor signal.

So far, we do not have data available to constrain the nature of the process at play. For surface snow subjected to a temperature gradient of $0.5^\circ \text{C cm}^{-1}$, Pinzer

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et al. (2012) showed that the characteristic residence time for a snow crystal to stay in place before being sublimated is 2–3 days. This corresponds to a mass turnover time of ~60 % of total ice mass per day (Pinzer et al., 2012). Based on model outputs from CROCUS and our in-situ temperature observations, such a temperature gradient occurs in the top 5 cm of the NEEM snow pack on clear-sky days. Larger temperature gradients in the top 1–2 cm of the snow pack cannot be excluded. We speculate that wind pumping (Clarke and Waddington, 1991; Neumann and Waddington, 2004) can cause a continuous replacement of the interstitial water vapor in the top snow layer. Due to the continuous recrystallization described by Pinzer et al. (2012), we hypothesize that this process leads to an imprint of changes in surface water vapor isotopic composition into surface snow. Further investigations including controlled laboratory experiments and isotopic modeling are needed to understand how metamorphism processes can impact the $\delta^{18}\text{O}$ -temperature and d-excess- $\delta^{18}\text{O}$ relationships.

5 Conclusions

During the two warm summers 2011 and 2012 at NEEM, continuous measurements of surface water vapor combined with isotopic measurements of snow samples collected every 12–24 h from the top 0.5 cm of the snow surface reveal parallel variations in-between precipitation events, with larger variations in the vapor $\delta^{18}\text{O}_v$ than in the snow surface $\delta^{18}\text{O}_s$. We also report positive correlations between $\delta^{18}\text{O}$ and temperature and negative correlations between d-excess and $\delta^{18}\text{O}$ in the vapor data, but weaker correlations as well as different slopes in precipitation and surface snow data, a finding confirmed by results obtained from datasets acquired during earlier years at NEEM. We note a decoupling between vapor $\delta^{18}\text{O}$ and temperature during the warmest days such as the summer 2012 heat wave. Changes in surface air temperature during the summer only account for a tiny fraction of the variability in precipitation and surface snow isotopic composition. This finding is consistent with investigations of shallow ice core data, which show much weaker relationships between Greenland summer $\delta^{18}\text{O}$

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face snow due to snow metamorphism associated with a strong temperature gradient in the upper centimeters of the snow surface. Laboratory experiments conducted by injecting an isotopically known vapor into a snow disk of known isotopic composition and different temperature gradients are needed in order to quantify the magnitude and rates of changes of the isotopic processes occurring during controlled snow metamorphism. Such laboratory experiments would also allow validating the implementation of water stable isotopes in snow models such as CROCUS. Similar monitoring frameworks in different places could investigate the validity of our findings for other sites. If our interpretation of the observed signals is correct, changes in surface snow isotopic composition are expected to be significant (i) if large day-to-day surface changes in water vapor occur in-between precipitation events, (ii) wind pumping is efficient, (ii) snow metamorphism is enhanced by large temperature gradients in the upper first centimeters of the snow. Due to dry and cold conditions, Central East Antarctica would be an excellent case study, albeit with the challenge to measure accurately the isotopic composition of water vapor at very low concentrations, even in summer where metamorphism is expected to be larger.

Our hypothesis that the surface snow isotopic composition is affected by isotopic exchanges with the atmospheric water vapor in-between precipitation events has also implications for the interpretation of ice core records. Indeed, classically, ice core stable isotope records are interpreted as reflecting precipitation-weighted signals, and compared to observations (e.g. station data, atmospheric reanalysis) and atmospheric model results for precipitation (using $\delta^{18}\text{O}$ precipitation data, or precipitation-weighted temperature), ignoring such snow-vapor exchanges. Recording a surface climate signal even when no precipitation is deposited suggests a more continuous archiving process than previously thought, but makes the comparison with atmospheric simulations more challenging. It has long been known that processes in the surface snow damp the signal associated with each precipitation event. Mathematical calculations of so-called isotopic diffusion require being reconciled with the physical understanding of the processes at play. Our findings also challenge the use of purely statistical

back-diffusion calculations in order to restore the full magnitude of seasonal variations, a method classically applied for identifying seasonal cycles in damped isotopic signals.

Supplementary material related to this article is available online at
<http://www.clim-past-discuss.net/9/6035/2013/cpd-9-6035-2013-supplement.pdf>.

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Table 1. Overview of data collected during the 2011 and 2012 field campaign.

Sample type	Time period	Isotopic measurement type	Number of samples	Precision and accuracy	Measurement type	Place of measurement
Picarro water vapor isotope	Day 185 to Day 215 of 2011	δD_v , $\delta^{18}O_v$ continuously from 3 m above snow surface interrupted by calibration 30 min every 6 h	~ hourly resolution	$\delta^{18}O_v = 0.23\text{‰}$ $\delta D_v = 1.4\text{‰}$ (Steen-Larsen et al., 2013)	Picarro CRDS analyzer	NEEM
Picarro water vapor isotope	Day 141 to Day 215 of 2012	δD_v , $\delta^{18}O_v$ continuously from 3 m above snow surface interrupted by calibration 30 min every 6 h	~ 15 min resolution	$\delta^{18}O_v = 0.23\text{‰}$ $\delta D_v = 1.4\text{‰}$ (Steen-Larsen et al., 2013)	Picarro CRDS analyzer	NEEM
Cryogenic collected vapor	Day 191 to 194 of 2012	^{17}O -excess from 6 h trapping of 10 m height vapor (2 samples per day)	7 samples	6 ppm (Landais et al., 2012)	IRMS using a flourination line	LSCE
Precipitation samples	Day 189 to Day 210 of 2011	δD_p , $\delta^{18}O_p$ Sub-event resolution	41 samples	$\delta^{18}O_p = 0.1\text{‰}$ $\delta D_p = 1.0\text{‰}$	Picarro CRDS analyzer	LSCE
Snow surface samples	Day 188 to Day 215 of 2011	δD_s , $\delta^{18}O_s$ every 12 h	51 samples	$\delta^{18}O_s = 0.1\text{‰}$ $\delta D_s = 1.0\text{‰}$	Picarro CRDS analyzer	LSCE
Snow surface samples	Day 191 to Day 194 of 2011	^{17}O -excess every 12 h	7 samples	6 ppm	IRMS using a flourination line	LSCE
Snow surface samples	Day 143 to Day 214 of 2012	δD_s , $\delta^{18}O_s$ every 24 h (Day 143–165) every 12 h (Day 166–214)	122 samples	$\delta^{18}O_s = 0.1\text{‰}$ $\delta D_s = 1.0\text{‰}$	Picarro CRDS analyzer	CIC

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Table 2. Overview of the distribution of the observed $\delta^{18}\text{O}$ and d-excess together with the slope and correlation for d-excess vs. $\delta^{18}\text{O}$ and $\delta^{18}\text{O}$ vs. air temperature for the atmospheric water vapor, precipitation samples, and snow surface samples.

Sample type	Year	Measurement type	Minimum	Maximum	Average	Slope d-excess vs. $\delta^{18}\text{O}$	<i>R</i> d-excess vs. $\delta^{18}\text{O}$	Slope $\delta^{18}\text{O}$ vs. Air Temp.	<i>R</i> $\delta^{18}\text{O}$ vs. Air Temp.
Atmospheric water vapor	2011	$\delta^{18}\text{O}$	−44	−32	−37	−1.14	−0.76	0.81	0.77
		d-excess	13	34	22				
	2012	$\delta^{18}\text{O}$	−53	−25	−38	−1.03	−0.86	0.85	0.76
		d-excess	10	46	23				
Precipitation samples	2011	$\delta^{18}\text{O}$	−35	−18	−26	−0.47	−0.57	0.64	0.24
		d-excess	2	20	11				
Snow surface samples	2011	$\delta^{18}\text{O}$	−33	−21	−25	−0.31	−0.27	–	0.17
		d-excess	7	19	12				
	2012	$\delta^{18}\text{O}$	−36	−15	−24	−0.44	−0.53	0.31	0.32
		d-excess	3	24	13				

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Table 3. Mean summer surface snow isotopic composition from observations and the calculation of equilibrium with surface water vapor at surface air temperature.

Summer	Observed mean values	Equilibrium with water (Majoube, 1971)	Equilibrium with ice (Majoube, 1970; Merlivat and Nief, 1967)	Equilibrium with ice (Ellehoj et al., 2013)
2011				
$\delta^{18}\text{O}_s$:	-25	-25 ‰	-21 ‰	-22 ‰
d-excess _s :	12	16 ‰	1 ‰	10 ‰
2012				
$\delta^{18}\text{O}_s$:	-24	~ -26 ‰	-22 ‰	-23 ‰
d-excess _s :	13	~ 16 ‰	1 ‰	10 ‰

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Table 4. Changes in snow surface and water vapor isotopic composition in-between precipitation events. The value is calculated by comparing the mean over the specific days.

Day begin	Day end	Change in snow surface $\delta^{18}\text{O}_s$ [‰]	Change in water vapor $\delta^{18}\text{O}_v$ [‰]	Change in snow surface d-excess _s [‰]	Change in water vapor d-excess _v [‰]
2011					
191	194	~ +7.0	~ +8.0	~ -6	~ -10
199	200	~ -1.3	~ -1.3	-	-
204	207	~ +2.4	~ +5.0	~ -5	-
211	214	~ -0.5	~ -3.5	~ +3	-
2012					
153	155	~ -2	~ -2	~ +2	~ +2
160	163/164	~ -4	~ -4	~ +10	~ +7
163/164	166	~ +5	~ +12	~ -13	~ -13
180	183	~ -4	~ -8.5	~ +8	~ +7
196	201	~ -4	~ -8	~ +7	~ +6

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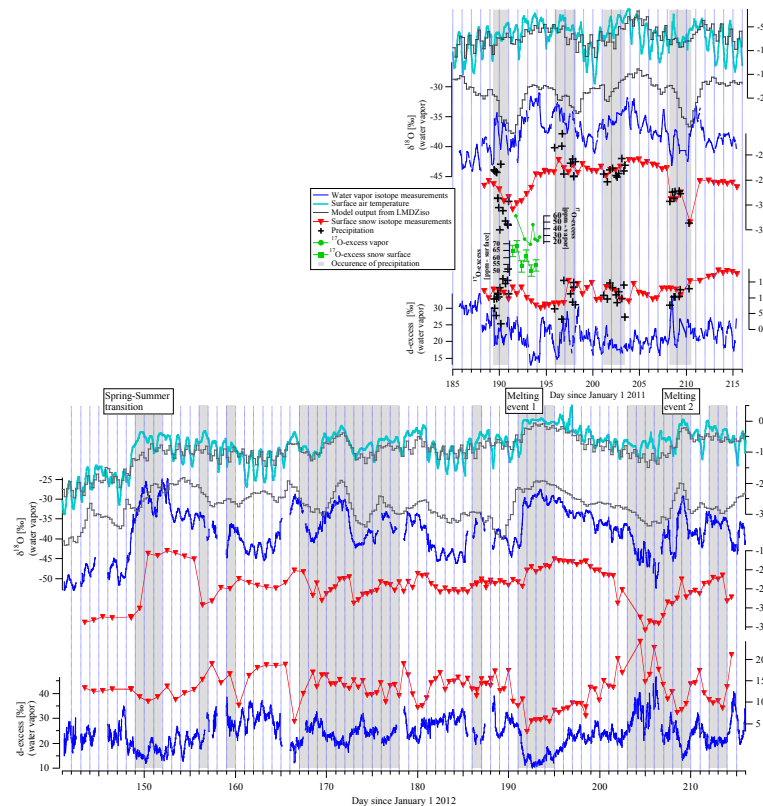


Fig. 1. The observed water vapor isotope signal during the 2011 (top panel) and 2012 (bottom panel) field season at NEEM (shown in blue) together with the concomitant variability in the snow surface isotopes (shown in red). Air temperature from local GC-Net station shown in cyan. Model outputs from LMDZiso shown with solid dark grey line. Occurrence of precipitation indicated by vertical grey columns. Precipitation samples collected in 2011 on sub-event shown with black crosses in top panel.

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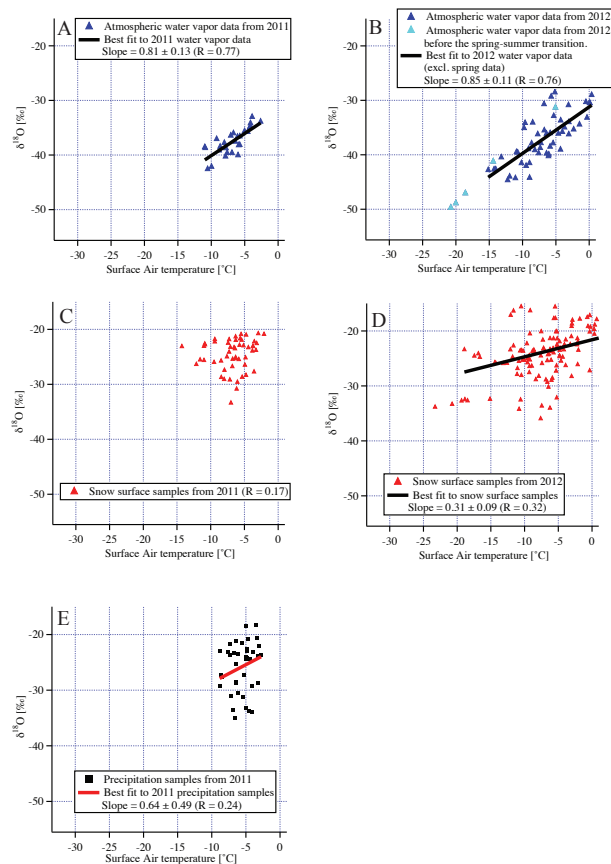


Fig. 2. The $\delta^{18}\text{O}$ vs. surface air temperature for the observed mean daily water vapor isotopes (**A:** 2011, **B:** 2012), Snow surface samples (**C:** 2011, **D:** 2012), and precipitation samples (**E:** 2011).

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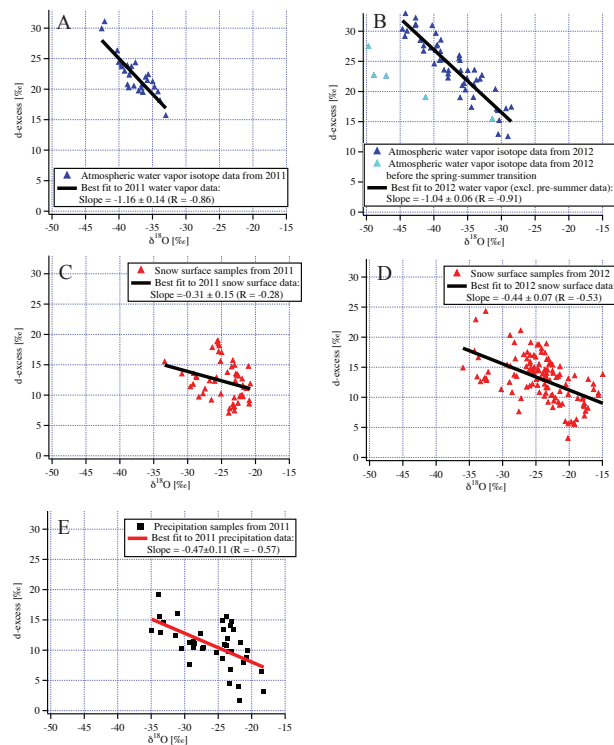


Fig. 3. The d -excess vs. $\delta^{18}\text{O}$ for the observed mean daily water vapor isotopes (**A**: 2011, **B**: 2012), Snow surface samples (**C**: 2011, **D**: 2012), and precipitation samples (**E**: 2011).

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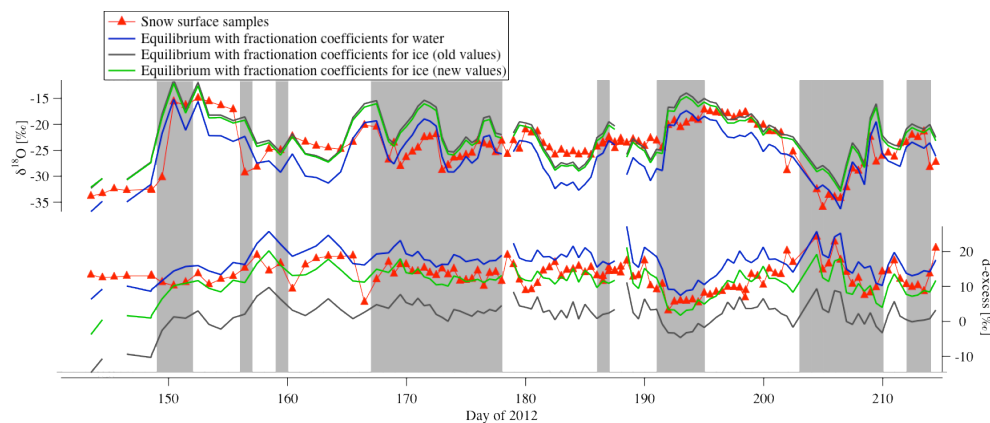


Fig. 4. $\delta^{18}\text{O}$ and d-excess of the condensate calculated based on the air temperature and observed atmospheric water vapor isotopic composition under the assumptions of isotopic equilibrium. Different fractionation coefficients were used assuming liquid water (Majoube, 1971), ice (Majoube, 1970; Merlivat and Nief, 1967), and ice but with new fractionation coefficients estimated by Ellehoj et al. (2013).

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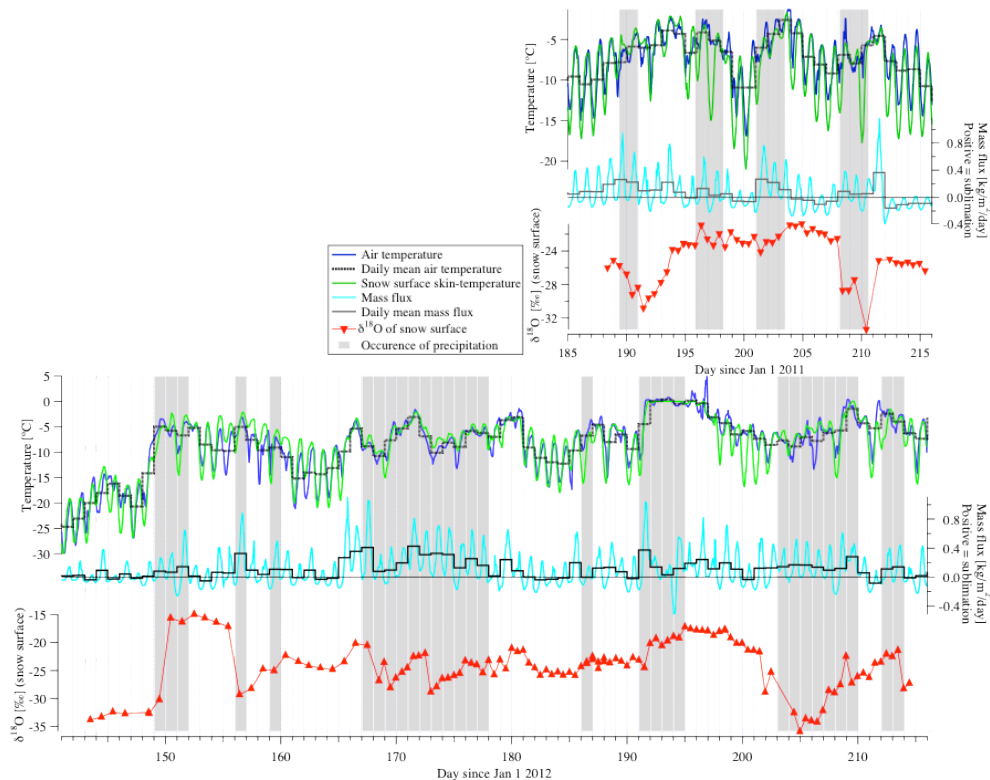


Fig. 5. Top and bottom panel show respectively data from 2011 and 2012. The air temperature and snow surface skin-temperature modeled by CROCUS is shown in respectively blue and green (Daily mean air temperature shown in black). The estimated mass flux from CROCUS with hourly resolution and daily mean values is shown in respectively cyan and black. The variability in the snow surface $\delta^{18}\text{O}_s$ value is shown below.

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