Authors' response to reviewer comments

Journal: Atmospheric Chemistry and Physics
Manuscript #: acp-2016-539
Title: Annual variation in event-scale precipitation δ^2H at Barrow, AK reflects vapor source region
Authors: Annie Putman, Xiahong Feng, Leslie J. Sonder, and Eric S. Posmentier
Date: Dec, 27, 2016

Dear Dr. Thomas Röckmann,

We are pleased to resubmit for publication the revised version of MS acp-2016-539, 'Annual variation in event-scale precipitation δ^2H at Barrow, AK reflects vapor source region'. We appreciate the constructive criticism of the reviewers, and feel that the resulting paper is stronger as a result of their input. A detailed, point-by-point response to J.L. Bonne and Anonymous Reviewer #2 are followed by a list of relevant changes, and a marked up manuscript. Note that line numbers in this document refer to the final manuscript.

The authors feel that the work detailed in the following document represents a complete response to the concerns of the reviewers. We appreciate your time and consideration as you evaluate our revised manuscript and we look forward to hearing from you in the new year.

Kind Regards,

Annie Putman

Interactive comment on "Annual variation in precipitation δ 2H reflects vapor source region at Barrow, AK" by Annie L. Putman et al.

J.-L. Bonne (Referee) jean-louis.bonne@awi.de

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General comments

This paper presents a new dataset of isotopic composition of precipitation sampled in the Barrow, AK, Arctic station, together with an innovative method to analyze and interpret its seasonal and event time scales variations. The authors propose interesting tools to use the Lagrangian atmospheric backtrajectory model for a quantitative and statistical evaluation of the observed isotopic variations. They conclude that the seasonal variations of water isotopic values are partly due to migration of the moisture origins. They focus on the influence of three parameters which are shown to explain a large part of the observed variations of the isotopic composition: the cooling along atmospheric transport, the dew point at the moisture source and the presence of mountains along the transport. The authors made a good effort to provide a rich interpretation of their observations. The manuscript is well organized and provides necessary tables and figures. There are, however, several issues regarding discussion of the results. I recommend accepting the article after the authors address the points listed below.

The authors thank J.L. Bonne for his helpful and insightful comments. Substantial effort was put toward fully addressing the critiques. Line numbers provided correspond to the final document.

Specific comments

Modeling:

The moisture source modeling used in this paper relies on strong assumptions. However, potential errors caused by these assumptions are poorly pointed out. A companion paper describing the method is currently under review and might contain these information. As this paper is not yet readable, one would need a summary of these information and eventually more details in the method description or in supplementary material, in particular concerning the points addressed below. Contrary to Sodemann et al. 2008b method, the moisture source modeling used here does not take into account variations of the specific humidity in the air parcels along the trajectory. Processes such as the lost moisture through precipitation or reevaporation of already condensed droplets along transport are not taken into account, but could have a strong impact on the isotopic composition. Can you give more details on the potential errors inherent to this moisture sources modeling? Also concerning the moisture sources modeling, moisture uptakes are assimilated to air masses sinking into the planetary boundary layer (PBL) above the ocean surface. Nothing is written about the potential presence of sea ice above the ocean in the region where the PBL is reached, which could however have a strong influence on the evaporation. Do you also take into account the sea ice cover in the region were the air parcels sink into the planetary boundary layer? For example: the moisture sources for the winter events are originating from a very wide range of latitudes. If most sources are originating from the south, some sources are coming from high latitudes, up to 85◦N (see winter sources latitudes on Figure 2). Can we really expect strong evaporation in those regions, over a potentially closed ocean? Have you checked the presence of sea ice in the moisture sources regions for this type of events?

The reviewer has a point, although we still consider our approach adequate for our purpose. We point out that assessment of "the potential errors inherent to this moisture sources modeling" is difficult for any model because true observations of moisture sources for a given event do not exist. Any assessment would be model dependent. Admittedly, the method of tracking the air parcels' moisture evolution through time, such as the one used by Sodemann et al., (2008a), is a more sophisticated way of identifying the source near the PBL than the method presented in this paper, but the model by Sodemann et al. (2008a) is not designed for the event scale moisture tracking in that 1) their parcels do not always start within precipitating clouds, and 2) the vertical distribution of parcels does not reflect condensation rates (precipitation events). Arguably, it would be ideal to combine the two methods. However, we think that for our purpose, which is to characterize the moisture source regions for observed and measured precipitation events, the initial starting points of parcels are most important, because the height of these parcels often primary dictate the source area, in our experience.

Details about modeling that have been added include enhanced discussion of the validity of our 'moisture source' decisions, in particular, our decision to use the last interaction with the PBL as the vapor source. We feel that this choice is justified, though less precise than Sodemann et al (2008a) because the dominance of turbulent transport relative to advective transport within the PBL.

As well, the sea ice presence and concentration is recorded. Air parcels were allowed to sink over sea ice, but were only considered to be a vapor source if the sea ice concentration was < 96% in order to allow for the presence of leads to contribute vapor to the PBL.

p.4, l. 7-14:
'Relative to previous studies that tracked vapor change in an air parcel along the trajectory (e.g., Sodemann et al. (2008a)), we adopted a simpler procedure that assumes vapor in the air parcel is well represented by the air at the latest interaction with the PBL. This assumption is justified because mass movement in the PBL is dominated by vertical turbulence relative to horizontal advection. Figure 1 shows endpoints of all trajectories that sank into the PBL. However, only trajectories that ended over water with < 96% sea ice cover were used for calculations; parcels that sank where there was less than 96% sea ice cover were used for calculations.'

Interpretation of results:
Concerning the interpretation of results at the seasonal scale, the seasonal variations described in the article are mostly the result of the relative preponderance of different types of synoptic scale events across the seasons. The intra-seasonal variability of the different events is often on the same order of magnitude than the variations of seasonal averages, which is too rarely pointed by the author. The clarity of the explanations might benefit from a more stronger distinction of the synoptic scale and seasonal scale variations.
This is a very good point, and useful to understand when interpreting monthly, seasonal and interannual variability. Language to clarify the similarity in magnitude of the event to event and seasonal variability has been added.

P. 7 L. 34:
'… though the inter-event variability in both variables can be as large as the seasonal variability'

P. 12 L.28-29
'…exhibited interannual, annual, and substantial inter-event variability.'

P.12 L. 33
'However, substantial intra-season variability occurred in both source and δ^2H, indicating scatter in the seasonal relationship'

Caption of Fig 3
'Of the three timescales, annual variability shows the greatest amplitude, though variability among events is also substantial.'

Caption of Fig 4
'For both datasets, the variability exhibited among events is of the same the order of magnitude as the seasonal variability.'

Technical corrections

Abstract: The abstract is quite long and could be more concise.
The following sentences at the beginning and end of the abstract were removed.

Removed 'Interpretation of variability in precipitation stable isotopic ratios often relies exclusively on empirical relationships to meteorological variables (e.g., temperature) at the precipitation site. Because of the difficulty of unambiguously determining the vapor source region(s), relatively fewer studies consider evaporation and transport conditions. Increasing accessibility of Lagrangian air parcel tracking programs now allows for an integrated look at the relationship between the precipitation isotope ratios and the evolution of moist air masses. In this study, 70 precipitation events occurring…'

Removed: We expect isotopes to respond similarly for longer-term climate-induced changes to the mean position of meridional circulation features, and expect that the most of the variation in isotopes measured in ice cores and other long term records are driven by changes in circulation, instead of fluctuations in local temperature.

P.1, L.1 to 5: The first three sentences of the abstract could rather be at the beginning of the introduction, as they don't describe the work presented in this article but general situation of research in the domain.

The content of these sentences is covered in the introduction, so they were removed from the abstract, which helps to reduce the length of the abstract. They were replaced with the following:

P.1 L. 1-2
"In this study, precipitation isotopic variations are linked to conditions at the moisture source region, along the transport path, and at the site of precipitation. Seventy precipitation events…"

P.1, L.8: "occurred" > "occuring"
This was changed in the text.
P.1 L. 5

Methods:
P.3, L.13-14: There might be an effect of sublimation of snow which could influence the isotopic composition of water, particularly for sunny periods, even within 24 hours. Did you make some experiments to test the evolution of fresh snow on your sampling site?
The reviewer is correct that this is possible. However, for much of the season when Barrow receives snow, there is little sun to drive sublimation, and the Arctic tends to be quite cloudy. There is also little evidence of sublimation from the data distribution along the meteoric water line. Nonetheless, clarification was added in the referenced section.

P.3 L.13-16
'…and often as soon as snow ended. Though it is possible that snow may have been altered by sublimation before collection, we assume that the degree of alteration of surface snow was minimal relative to the amount of snow gathered. Furthermore, the frequent cloudiness and darkness of Barrow mean that for most events, sunlight-driven sublimation was insignificant.'

P.3, L.14: At which temperature were the samples stored, and how long?
Clarification was added to the text. Samples were stored at less than 5 C, shipped every 3 months, and analyzed within 6 months.

P.3 L.16-19
'Liquid samples were stored in tightly sealed 30mL Nalgene bottles below 5 °C and shipped in batches every three months to the Stable Isotope Laboratory at Dartmouth College. When not in transit, samples were refrigerated. Samples were analyzed within six months of collection.'

P.4, L.1-4: Considering that a moisture source is corresponding to an air parcel sinking into the PBL is a strong assumption. More justifications of this method would be expected. If this method is described in Putman et al. (2015), add a reference here.
The assumption has been more clearly stated and justified in the text. Also see our response, at the beginning, to "Specific comments: Modeling"

P.4, L. 7-14
'Relative to previous studies that tracked vapor change in an air parcel along the trajectory (e.g., Sodemann et al. (2008a)), we adopted a simpler procedure that assumes vapor in the air parcel is well represented by the air at the latest interaction with the PBL. This assumption is justified because mass movement in the PBL is dominated by vertical turbulence relative to horizontal advection. Figure 1 shows endpoints of all trajectories that sank into the PBL. However, only trajectories that ended over water with < 96% sea ice cover were used for calculations; parcels that sank where there was less than 96% sea ice cover were used for calculations.'

P.4, L11-12: By "the most temporally homogeneous three-hour time window", do you mean homogeneity in the precipitation amount or in the meteorological records? Do you have particular criteria to define the preference for the middle of the event? Were the event times defined automatically or manually?
This section has been clarified. Homogeneity is in reference to the radar returns, which give us an idea of precipitation intensity. Event times were selected manually, based on multiple streams of evidence: radar returns, sampling records and surface analysis maps.

P.4 L.15-23
'Back trajectory analysis was performed for dates when precipitation was collected. The starting times for the back trajectories corresponded to times of maximum precipitation intensity, based on a combination of sampling records, surface analysis maps of Alaska available through the National Center for environmental Prediction, and the returns of the millimeter wavelength cloud radar (MMCR) (Johnson and Jensen, 1996; Bharadwaj et al., 2011). Greater Doppler vertical velocities, reflectivities, and spectral widths from the MMCR broadly indicated more intense precipitation. Because the gridded meteorological files 20 used for tracing the back trajectories had three-hour resolution, the chosen starting time represented average conditions over a three-hour period. If precipitation lasted for more than three hours, the most intense three hour time window was selected. If the precipitation was of approximately uniform intensity, the most temporally homogeneous three-hour time window was selected, with preference for time windows where precipitation occurred over the duration of the three hours.'

P5., L.5: "The same was done for an array...": explicit that this is to calculate Qsat,z and define Qsat,z.
This section was rewritten for clarity, and to reflect the change in the equations used.

P.5 L. 14-25
'To determine Qsat;z, we start from the dry adiabatic lapse rate, (-9.8 °C km-1). From this we determine the temperature Tz at altitude z, starting with the 2 meter temperature T2m.
The saturation vapor pressure at elevation z, esat;z is then

$$e_{sat;z} = 0.6113 \exp[5423\ (1/T_0 - T/T_z)] \quad (2)$$

where T0 = 273.15K (Stull, 2015). We may then write the saturation specific humidity, Qsat;z, as

$$Q_z = 0.622\ e_{sat;z}\ h_z / P_z \quad (3)$$

where hz, the relative humidity at height z, is assumed to equal 1 (air is vapor-saturated) and the pressure at height z, Pz, is

$$P_z = 1013.25\ [1-(2{:}25577\ e\text{-}5)\ z]^{5.25588} \quad (4)$$

Calculating the 2m specific humidity, Q2m, is simply a special case of the general calculation: we use the 2m temperature T2m, fractional relative humidity h2m, and pressure P2m from reanalysis in Equations 2 and 3, rather than using the dry adiabatic lapse rate, h = 1, and Equation 4, respectively.

Finally, we find the elevation where Q2m equals Qsat;z. The temperature at this elevation is T_{LCL}.'

P.5, L.5-6: Explicit hz, Tz, Pz: fractional relative humidity, temperature and pressure at elevation z.
This suggestion was incorporated into the newly written section discussed in the previous point.

P.5 L. 14-25

P.5, L.25: Are mtn values assigned manually or automatically? If automatic, then explicit the criteria.
Mtn was assigned manually based on maps of trajectory results.

P.6 L. 18-19
'The value of mtn was assigned manually based on the general pattern of transport observed in the trajectory plots.'

Results and discussion
P. 6, L.6-15: This is a very qualitative description of Figure 1. The mean latitude of moisture sources could be introduced before and used to give quantitative aspects to this description. This description focuses on the seasonal averages of the moisture sources, but Figure 2 shows a very strong variability at the event time scale, which can be of a larger order of magnitude than the variations of the seasonal average for the mean latitude of the moisture source. For example, some events in winter have moisture sources located as north as in summer, or even further north. The normalisation of the maps from Figure 1 can also give an impression of wider or more local moisture sources depending on the total number of events and the difference between each event. Is this description of

moisture sources regions still valid for absolute values without normalization to the number of events, or for individual events instead of the average of all events?
The authors agree that there is substantial variation among events, even within a given season. This follows the response to 'interpretation of results', above.

P. 7 L. 34:
'… though the inter-event variability in both variables can be as large as the seasonal variability'

Caption of Fig 3
'Of the three timescales, annual variability shows the greatest amplitude, though variability among events is also substantial.'

Caption of Fig 4
'The spline fits have R_2 values of 0.60 and 0.19 for the d_2H and V_{Lat} respectively. For both datasets, the variability exhibited among events is of the same the order of magnitude as the seasonal variability.'

P. 6, L. 30-32: Not clear if the last sentence refers to Feng et al. (2007).
Modified sentence.

P.7 L.19
'There is evidence for prior millenium-scale shifts in the southern extent of the polar circulation cell (Feng et al., 2007).'

P. 7, L. 6-7: This sentence is really affirmative, whereas Figure 4 shows a very strong dispersion, particularly for the averaged VLAT. This affirmation should be tempered and a statistical evaluation of the spline fits and there correlations should be given, as well as the standard deviations of the data series. The seasonal scale might not be the better scale to look at.
This is true, in particular for the V_{lat} variable. The variance captured by the spline fits has been added to the text, and the text has been adjusted to better describe the similarity in magnitude between among-event variability and mean seasonal variability.

P. 7 L. 30-34
'Figure 3 also shows the interannual, seasonal and event-scale variability captured by the dataset where the spline captures 65% of the annual and interannual variance. The average annual cycle of the precipitation δ_2H is strong; the spline fit explains 60% of variance in the data. The mean latitude of the vapor source exhibits a weak seasonal pattern, where the spline explains 19% of the variance. The seasonal cycles of δ_2H and vapor source latitude are in phase, as shown in Figure 4,though the inter-event variability in both variables can be as large as the seasonal variability'

Caption of Fig 3
'The spline fit, which highlights seasonal variations, explains 65% of variance in the data with a root mean squared error of 39.7‰. Of the three timescales, annual variability shows the greatest amplitude, though variability among events is also substantial.'

Caption of Fig 4
'The spline fits have R_2 values of 0.60 and 0.19 for the d_2H and V_{Lat} respectively. For both datasets, the variability exhibited among events is of the same the order of magnitude as the seasonal variability.'

P. 8, L.3: How did you choose the temperatures from 10C to -15C in you theoretical cooling experiments? What would be the effect on the slopes of a variation of these temperatures on the order of magnitude of the observed variations?
The temperature range encompasses the temperature change experienced by most trajectories. Making the warmest temp even warmer would yield slightly shallower slopes, and making colder temperatures even colder would yield steeper slopes. The coldest average final temperatures in our dataset are substantially below -15C, though the majority of each trajectory occurs within the 10 to -15C temperature range, and all events except one begin in the selected range.

P. 8, L. 18: Rather write “more than 20C” instead of “> 20C”.
Sentence was deleted when discussion was updated.

P. 9, L. 1: “amount” instead of “amounts”?
Sentence was deleted when discussion was updated.

P.9, L.5: How was the 7C criteria chosen? Is it close to the median of the distribution of ΔTcool?
The 7C criterion was chosen to preserve the statistical power of the short trajectories while preserving the strong relationship of δ2H to Td. Though we have presented the results as categorical, this is for simplicity. It is likely that this feature of isotope systematics is actually continuous.

P.10 L.22-24
‘The breakpoint of 7 °C was chosen by testing different breakpoints and finding one that maximized the statistical power of the short trajectory regression while preserving the strong relationship between δ_2H and T_d.’

P. 9, L.12: Insert a reference to figure 6 to show the repartition of small and large ΔTcool across seasons.
Added sentence:

P.10 L.21-22
‘Table 2 summarizes the results and Figure 6 shows the standard deviation of σT_d by category’

P. 10, L.6: This is not directly about precipitation d-excess but can be of interest: some studies of water vapour d-excess in Arctic regions have depicted a partial conservation of the source d-excess signal under certain atmospheric transport conditions, with relations between observed d-excess and moisture source relative humidity.

Bonne, J.-L., Masson-Delmotte, V., Cattani, O., Delmotte, M., Risi, C., Sodemann, H., and Steen-Larsen, H. C.: The isotopic composition of water vapour and precipitation in Ivittuut, southern Greenland, Atmos. Chem. Phys., 14, 4419-4439, doi:10.5194/acp- 14-4419-2014, 2014.

Bonne, J.-L., et al. (2015), The summer 2012 Greenland heat wave: In situ and remote sensing observations of water vapor isotopic composition during an atmospheric river event, J. Geophys. Res. Atmos., 120, 2970–2989, doi:10.1002/2014JD022602.

Steen-Larsen, H. C., A. E. Sveinbjörnsdottir, Th. Jonsson, F. Ritter, J.-L. Bonne, V. Masson-Delmotte, H. Sodemann, T. Blunier, D. Dahl-Jensen, and B. M. Vinther (2015), Moisture sources and synoptic to seasonal variability of North Atlantic water vapor isotopic composition, J. Geophys. Res. Atmos., 120, 5757–5774, doi:10.1002/ 2015JD023234.
Interesting work, thank you for the citations. The discussion of d-excess has been updated to include these publications.

P.11 L.31-32
‘While studies indicate that d in vapor contains vapor source information (Steen-Larsen et al., 2014; Bonne et al., 2015; Steen-Larsen et al., 2015)…’

P.12 L. 9-11
‘This value is consistent with the -0.4 to -0.6‰ ‰-1 range reported in the literature for vapor (Uemura et al., 2008; Pfahl and Wernli; Bonne et al.; 2014).’

Conclusions
P. 10, L. 29-31: This conclusion on the origins of moisture is valid for the average of the seasonal moisture sources, but should be tempered by pointing out the event to event variation of the moisture sources.
This caveat was added to the conclusions.

P.12 L.33

'However, substantial intra-season variability occurred in both source and δ2H, indicating scatter in the seasonal relationship.'

References
P.13, L. 32-36: Logically, the two papers numbering should be inverted (2008a and 2008b).
Changed.

Tables and figures
Table 1 and 2: The legends do not clearly describe the contents of the tables. Why are different intercepts given for each variable in Table 2 and only one value in Table 1, if the only difference between the two tables are the division of all samples in two groups?
The captions of the tables have been updated to explain that Table 1 contains the results from a single multivariable regression, while Table 2 contains the results from 3 simple linear regressions.

Caption of Tab 1
'Variation in δ2H is explained by a multiple linear regression (R2 = 0.54) of air parcel cooling during transport (ΔTcool), moisture source conditions (Td) and orographic obstacles in vapor transport path (mtn).'

Caption of Tab 2
'Three simple linear regressions against δ2H where β is the regression coefficient and S.E. is the standard error'

Figure 7: Parenthesis not closed in right y-axis label.
Fixed.

Figure 3 and 7: It would be more readable with x-axis ticks corresponding to the beginning of the years instead of the beginning of each December.
Fixed.

Review of
"Annual variation in precipitation δ^2H reflects vapour source region at Barrow, AK"
by A. L. Putman et al.
Paper published in ACPD on 11 August 2016

1 General Comments

This paper presents an interesting dataset of the event-scale δ_2H and deuterium excess signature of precipitation from northern Alaska. The authors use a very simple back-trajectory-based analysis of the transport and moisture source conditions which they summarise in 3 main characteristics to interpret their data. These are 1) the moisture source dew point temperature at 2m, 2) the total cooling between the lifted condensation level at the moisture source and the precipitation level in the cloud at the measurement site (arrival temperature) and 3) whether the air parcels that are transported to the measurement site across the Brooks and/or the Alaskan ranges. I recommend publication of this overall well-written manuscript, but I have four major concerns that should be addressed beforehand as well as a many specific comments listed below:

The authors thank the reviewer for the useful points and ideas. We have considered the suggestions, addressed the questions and revised the paper accordingly, and we hope that the revisions are satisfactory. Line numbers provided refer to the final manuscript.

1 **Moisture source identification and particularly the implicit assumptions made:**
see specific comments 3-7.

The authors argue that the method employed in this paper is adequate for our purpose. Please see our responses for comments 3-7 for the full discussion.

2 **Choice of the parameters that explain the variance of the isotope signature of precipitation in Barrow:**
For me the choice of the parameters that were used to explain the precipitation isotope signal in Barrow seems random. It makes sense to look at moisture source and transport conditions but in my opinion there is no reason for completely neglecting the local conditions. Particularly at Barrow, the precipitation phase (liquid or snow) probably plays an important role for the end isotope composition of the precipitation event as it determines whether there is isotopic exchange (for rain drops, see specific comment 2) or not (for snowfall) with the local vapour. Also precipitation intensity plays an important role. The authors have some detailed information about the precipitation structure from their radar data and could use this to try to further understand the local processes. If this is done in an other paper, then this should be clearly stated. Also I do not fully support the choice of the variable T_d as representative for the moisture source conditions (see specific comment 12).

This is a good point, and one that was considered by the authors before settling on the variables reported. Indeed, half the variance in δ2H cannot be explained by the 3 variables chosen! The main reason that other variables (including but not limited to precipitation phase, sub-cloud dryness, precipitation intensity, evaporation below the cloud base, supersaturation in the cloud, and storm event type) were not included is because, the statistical power of the limited number of events we were able to consider is not sufficiently high to go after each of those potentially very important variables. When such variables were included in the analysis, they did not explain any more variance. This may be because the isotopic responses to them are not related to δ2H variations, or are related but not sufficiently above noise. For example, sub-cloud dryness may be important for some but not all events, dryness may occur during both high and low δ2H events, but the power or the size of the signal may be limited. Nevertheless, to respond to this point, we added a paragraph at the end of Section 3.2 that includes a list of variables potentially contributing to the 46% of the unexplained variance in δ2H.

P. 11 L. 13-20
'The three chosen variables explain just over half (54%) the variance of δ2H. This is not surprising, considering that many other mechanisms can also influence the δ2H of the vapor and precipitation. These mechanisms include (but are not limited to) condensation temperature, supersaturation in the mixed phase cloud, sub-cloud dryness, phase of precipitation, precipitation intensity, evapotranspiration of land sources, and the amount of sea ice at the vapor

source. The effects of several of these factors, including condensation temperature, sub-cloud dryness, sea ice concentration at the vapor source, and phase of precipitation (rain vs. snow), were tested as additional explanatory variables in the multiple regression, but yielded statistically insignificant results with little to no additional variance explained. Clearly, compared with the three chosen variables, the effects of these variables are relatively minor, such that the statistical power is not sufficient to reveal their significance.'

3 **Expansion of the northern polar circulation cell and its link to moisture source location**
The link between the event-based moisture source location of precipitation and the polar circulation cell is described in a very qualitative way. A link between the weather systems driving the moisture transport at the event timescale leading to precipitation at Barrow and the more climatological description of the polar circulation is not obvious and not trivial to make. The formulations used throughout the paper should be more careful and kept as hypotheses.
The authors acknowledge that the relationship between vapor source and circulation is not simple, and the link between the annual and longer timescales is a possibility, not a certainty. However, the work does substantiate the idea that isotope values measured in ice cores may reflect changes in circulation patterns as well as local temperature, which is how they are often interpreted. The phrasing of these statements has been re-formulated in all discussion to suggest hypothetical as opposed to likely links.

P. 7 L. 20-25
'Aspects of the link between seasonal variability in general circulation and seasonal vapor source cycling may be generalizable to interannual and even millennial timescales. This is relevant to modern changes in the hydrologic cycle as Marvel and Bonfils (2013) suggest that a poleward displacement of circulation cells is already occurring due to recent climate change. Additionally, changes in the isotopic composition of precipitation resulting from systematic vapor source migrations associated with changing climate may allow for interpretation of long-term isotopic records in terms of changes in atmospheric circulation, including but not limited to the precipitation site temperature.'

P. 13 L. 14-16
'The mechanisms identified, most notably the north-south migration of the vapor source region in phase with expansion and contraction of the Polar circulation cell, may also operate on times scales longer than that of our study, and may be a source of variation in isotopes measured in ice cores, pedogenic carbonates, and speleothems.'

4 **Critical discussion of results in view of the existing literature**:
in particular see specific comments 24 and 27.
More discussion has been added to Sections 3.2 and 3.3. In 3.2, which discusses the influence of vapor source on measured precipitation isotopes, greater clarification of the simple Rayleigh model used to contextualize our results has been added, as well as a comprehensive sources of error paragraph, and an expanded discussion of the utility of Td in characterizing the source. In section 3.3, the d-excess results are discussed in greater depth in light of the suggested papers. In particular, we have added discussion of the relationship between local water vapor and evaporation conditions. The new or revised section are in order as listed below.

P. 8 L. 25-29
'Because Rayleigh distillation is considered the main source of spatial variation in δ2H, comparison with the sensitivities calculated from a simple Rayleigh model contextualize our result. In such a model, a saturated air parcel with specified temperature and vapor δ2H is cooled iteratively in 1°C steps. At each temperature step, the condensation amount, remaining vapor, precipitation δ2H and vapor δ2H are calculated. No re-evaporation or non-equilibrium conditions are considered.'

P. 11 L. 13-20
'The three chosen variables explain just over half (54%) the variance of δ2H. This is not surprising, considering that many other mechanisms can also influence the δ2H of the vapor and precipitation. These mechanisms include (but are not limited to) condensation temperature, supersaturation in the mixed phase cloud, sub-cloud dryness, phase of precipitation, precipitation intensity, evapotranspiration of land sources, and the amount of sea ice at the vapor source. The effects of several of these factors, including condensation temperature, sub-cloud dryness, sea ice concentration at the vapor source, and phase of precipitation (rain vs. snow), were tested as additional explanatory variables in the multiple regression, but yielded statistically insignificant results with little to no additional variance

explained. Clearly, compared with the three chosen variables, the effects of these variables are relatively minor, such that the statistical power is not sufficient to reveal their significance.'

P. 9-10 L. 7-35, 1-9
'We prefer Td to the classical variables Tss and h for determining isotopic evaporative fluxes. This choice is based on our understanding that the meteorological variable Td characterizes the bulk vapor content and isotopic ratio of the marine PBL, independent of the vapor temperature. When advected to the free troposphere, it is this vapor that will form precipitation. Additionally, through equilibrium fractionation Td also determines the isotopic ratio of the first condensate at the LCL, where Rayleigh distillation begins.

Within the marine PBL, several inter-related factors/processes are at work to determine the starting point of a Rayleigh trajectory. The first is the isotopic flux of evaporation from the sea surface. Most studies estimate this flux using the classic model by Craig and Gordon (1965). In that model, three variables control the evaporative flux: the sea surface temperature, Tss, δ2H above the laminar layer, and the humidity hss above the laminar layer (e.g., at 2 m), defined relative to Tss. Though hss is not a measured quantity nor one that is normally modeled, it is determined by Td above the laminar layer and Tss. Hence isotopic fluxes can be determined with the classical model using Tss, and Td and _2H above the laminar layer as input variables. From a physical point of view, Tss determines the amount of equilibrium fractionation at the water-air interface. Td and vapor δ2H, as well as Tss, control kinetic fractionation as vapor diffuses across the laminar layer. It should be noted that when Tss is large, Td tends to be large as well, as a result of their change with latitude and season. Td and δ2H are also correlated, which will be discussed below. Therefore, all three variables controlling the evaporative flux, Tss, and hss and δ 2H above the laminar layer, are associated directly or indirectly with Td, making Td a good indicator of evaporation conditions.

The second process is convergence. At a moisture source location, low level air is moist due to evaporation near the sea surface. Convergence and uplift transports low-level moist air into the free troposphere where it mixes with dry, isotopically depleted air descending from surrounding regions resulting in strong humidity and temperature gradients near the sea surface (below 2 m). In contrast, the specific humidity and isotopic ratios in the bulk of the PBL above 2m are relatively constant, resulting from the relative contributions of vertical transport of moist low-level and descending air (Fan, 2016). Td and δ2H at 2m both reflect the outcome of this mixing process, and so it follows that they are positively correlated.

The third process is condensation at the LCL. The temperature of the air mass, which equals or is very slightly less than the local dew point, determines the amount of isotopic fractionation and thus the isotopic ratio of the first condensate. It is this isotopic composition that defines the beginning of the Rayleigh part of the trajectory. Only Td;2m, not Tss nor h2m, is directly associated with the condensation temperature at the LCL (which differs only slightly from Td;2m due to the pressure difference between 2 m and the LCL and its effect on saturation specific humidity).

Since all three processes before Rayleigh distillation are either directly or indirectly related to Td, we consider Td a better indicator for the source conditions than either Tss or h. It is difficult, however, to theoretically assess the sensitivity of precipitation δ2H to variations in source Td, because this would require quantification of the theoretical relationship of Td to δ2H through each of the three processes and perhaps their combinations. We here report the first empirical sensitivity of 3.23‰ °C-1 (Table 1) for δ2H relative to Td. At the sea surface, for Tss between 0 and 25 °C, equilibrium fractionation as a function of temperature yields sensitivities between 1.1-1.6‰°C-1 (Majoube, 1971). However, a large part of this fractionation may be offset by condensation at the LCL. Consequently, the observed sensitivity probably reflects primarily the fraction of vapor contributed by dry, isotopically depleted descending air that converges within the PBL. Mixing with the dry air causes a decrease in Td, which affects the δ2H of the PBL in two ways: 1) making the PBL air dry and isotopically depleted, and 2) isotopically depleting the evaporative flux by enhancing kinetic fractionation (an effect of low relative humidity). Both mechanisms produce a positive association between δ2H and Td, consistent with the sign of our observed partial coefficient (Table 1).'

P. 12 L 9-11
'This value is consistent with the -0.4 to -0.6‰ %-1 range reported in the literature for vapor (Uemura et al., 2008; Pfahl and Wernli; Bonne et al.,2014).'

P. 12 L. 15-20
'This is an interesting result with respect to the utility of Td, a measurable quantity, and is consistent with our earlier argument that Td is strongly related to hss. Both variables provide a better representation of source conditions than Tss and/or h2m. A low value of hss or Td corresponds to a strong influence of descending dry air within the PBL, which enhances kinetic isotopic fractionation and produces a high value of d. This mechanism explains the negative correlation between d and Td, and is expected for the relationship between d and hss. Alternatively, the vapor in descending air may have a high value of d (Fan, 2016), or both mechanisms may contribute to this result.'

2 Specific comments

1. p. 1, title: It would be nice to include in the title the fact that it is event-scale precipitation samples that the authors analyse in this paper. Something like: "Annual variation in event-scale precipitation δ^2H reflects vapour source region at Barrow, AK". Also Barrow, AK could be replaced by northern Alaska.
Title changed as suggested.

New title:
'Annual variation in event-scale precipitation δ2H at Barrow, AK reflects vapor source region'

2. p. 18-23: The local conditions during cloud formation and during precipitation also play an important role for the isotope composition of precipitation. For rainfall for example below cloud effects (evaporation and exchange with ambient vapour) can have a strong impact on the isotope composition of precipitation (20-40h for δ^2H, see Pfahl et al. (2012), Aemisegger et al. (2015)).
This is absolutely true, and we did experiment with including condensation temperature, precipitation type, and sub-cloud humidity in our regressions. The regression presented was the best model in terms of simplicity and variance explained by different parameters. One reason why these local factors may not have been significant influences to our dataset is because of event-to-event variability. e.g., in one case enrichment may be due to sub-cloud evaporation, but in another it may be due to condensation temperature, and within our dataset we did not have the statistical power to disentangle these competing mechanisms. Also see our response to General Comments 2).

P. 11 L. 13-20
'The three chosen variables explain just over half (54%) the variance of δ2H. This is not surprising, considering that many other mechanisms can also influence the δ2H of the vapor and precipitation. These mechanisms include (but are not limited to) condensation temperature, supersaturation in the mixed phase cloud, sub-cloud dryness, phase of precipitation, precipitation intensity, evapotranspiration of land sources, and the amount of sea ice at the vapor source. The effects of several of these factors, including condensation temperature, sub-cloud dryness, sea ice concentration at the vapor source, and phase of precipitation (rain vs. snow), were tested as additional explanatory variables in the multiple regression, but yielded statistically insignificant results with little to no additional variance explained. Clearly, compared with the three chosen variables, the effects of these variables are relatively minor, such that the statistical power is not sufficient to reveal their significance.'

3. p. 3, L. 29: The reanalysis dataset (wind fields) that is used for the trajectory calculation should be mentioned here as well as its horizontal resolution.
The information has been added.

P. 4 L. 3
'…using 1° resolution meteorological data from the Global Data Assimilation System (GDAS).'

4. p. 4, L. 2: What do the authors mean with "The first time"? Is the time reference forward or backward? Does that mean the first time when following the trajectory back from the arrival point? And does that mean that one trajectory can have only 1 associated moisture source? This would be a very strong assumption about the moisture source location. Uptakes of moisture can happen all along an air parcel's trajectory (see Sodemann et al. (2008)) and they can sometimes be linked to surface evaporation even though they are not in the boundary layer (PBL), particularly over land. If for each trajectory only the latest passage in the PBL before arrival at the measurement site is considered then this means that the authors assume very strong mixing. This would imply that the air parcel basically looses all its previous humidity by mixing out and takes up only humidity that has just been evaporated at

this location. The isotope signature of the air parcel thus is fully determined by the freshly evaporated water. This strong assumption has to be explicitly stated.
The 'first time' is in reference to back trajectories; wording in the manuscript has been updated for clarity. Yes, each trajectory has one associated vapor source. Though the method described in Sodemann (2008) is a substantially more sophisticated way of identifying the vapor source, it is not necessary in our work for three reasons. 1) For a given parcel, the spatial range over which the parcel moves up and down across the PBL, is small compared to the region covered by 1000 total parcels of an event. The latter is primarily dictated by the vertical distribution of the initial parcels' altitude 2) Our work analyzes the influence of marine source areas on precipitation at Barrow, AK. Marine surface conditions are relatively homogeneous, which point strengthens the argument in 1). 3) Averaging at the precipitation site of condensate of 1000 trajectories from a wide spatial distribution of source locations implicitly accounts for mixing of moisture from distributed source locations. It is thus in effect equivalent to the more sophisticated Sodemann et al. (2008) model which used on average only 2.6 trajectories per column of air per time window, even if each of those trajectories combined source points at its inception.

P. 4 L. 6-10
'The vapor source location was defined as the place where the back trajectory of the air parcel sank into the planetary boundary layer (PBL). Relative to previous studies that tracked vapor change in an air parcel along the trajectory (e.g., Sodemann et al. (2008a)), we adopted a simpler procedure that assumes vapor in the air parcel is well represented by the air at the latest interaction with the PBL. This assumption is justified because mass movement in the PBL is dominated by vertical turbulence relative to horizontal advection.'

5. p. 4, L. 4: The authors say that air parcels that sank below the PBL over land were ignored? Why then do they find a lot of moisture sources over continental Alaska in Figure 1? This is confusing.
The difference between data used for statistics and data used in figure is confusing, and text has been added to clarify the distinction.

P. 4 L. 11-14
'Figure 1 shows endpoints of all trajectories that sank into the PBL. However, only trajectories that ended over water with < 96% sea ice cover were used for calculations; parcels that sank where there was less than 96% sea ice cover were used for calculations. Parcels that never sank into the PBL or those that sank into the PBL over land or ice-covered ocean were ignored. Ocean-originating air parcels comprised about 71% of all trajectories.'

Fig 1, caption
'The figure indicates that some air parcels originate over land, but these were not included in calculations.'

6. p. 4, L. 4: Were 71% of all trajectories ignored or kept for the analysis?
Changed for clarity.

P. 4 L. 14
'Ocean-originating air parcels comprised about 71% of all trajectories.'

7. p. 4, L. 5-12: For me it is not entirely clear how the trajectory starting dates were chosen. Why do the authors choose only a three hours period instead of the whole precipitation event? Why are the individual dates not weighted by the locally measured precipitation intensity to take into account that when the precipitation intensity is higher the trajectories of that date contribute more to the isotope signal? What means the "most homogeneous" three-hour time window? And why with preference to the "middle" of the event? The selection criteria should be more oriented to the quantitative contribution of moisture to precipitation in my opinion.
The three hour time period was chosen so that each event was treated the same. During analysis, we selected a few events to test the sensitivity of the vapor source to the selected time. In these cases, multiple 3 hour windows were analyzed for a single event, and typically showed very similar results. Thus, we concluded that choosing one three hour window was representative of the whole event. Selection of the specific three hour time period was not automated or quantitative. The qualitative selection used returns from the MMCR and KAZR. Higher Doppler vertical velocity and reflectivity indicate increased precipitation intensity. These criteria, in conjunction with information from surface analysis maps were used to determine the start and end times. The three hour window

reflects the constraints of the reanalysis, i.e. the temporal resolution is three hours. The text has been clarified with respect to these questions.

P. 4 L. 15-23
'Back trajectory analysis was performed for dates when precipitation was collected. The starting times for the back trajectories corresponded to times of maximum precipitation intensity, based on a combination of sampling records, surface analysis maps of Alaska available through the National Center for Environmental Prediction, and the returns of the millimeter wavelength cloud radar (MMCR) (Johnson and Jensen, 1996; Bharadwaj et al., 2011). Greater Doppler vertical velocities, reflectivites, and spectral widths from the MMCR broadly indicated more intense precipitation. Because the gridded meteorological files used for tracing the back trajectories had three-hour resolution, the chosen starting time represented average conditions over a three-hour period. If precipitation lasted for more than three hours, the most intense three hour time window was selected. If the precipitation was of approximately uniform intensity, the most temporally homogeneous three-hour time window was selected, with preference for time windows where precipitation occurred over the duration of the three hours.'

8. p. 4, L. 13: Does "where" mean the starting altitude? The method that is shortly described in this paragraph sounds original and the idea is interesting but it assumes that the reanalysis dataset's wind field and precipitation rate profile are equivalent with the true fields. The reanalysis data error particularly with respect to the representation of small and microscale processes are ignored. Starting trajectories from different locations around the measurement site would allow to take into account the uncertainty arising from the reanalysis data
Yes, 'where' means altitude. The authors agree that accurate wind directions are very important for accurate back trajectories. However, if winds in reanalysis are incorrect, incorporating a wider area will not make them more correct. Furthermore, because the resolution of the reanalysis data is 1x1 degree, using multiple locations may cover a wide region, hundreds of kilometers in size. Such a wide spatial scale could be less representative of local or small-region precipitation events, though it could be helpful for large-region precipitation events. We are not convinced that looking over a larger spatial location would improve the vapor source estimation. However, sending a large number of air parcels (~1000), as we have done, helps to deal with the wind issue. Finally, wind errors in the reanalysis are a potential source of error for any Lagrangian back trajectory study, and are not unique to our study. Therefore, we hope people who are reading studies using Lagrangian back trajectories are in general cautious with this type of reanalysis product.

P. 4 L. 24-25
'The method for selecting the altitudes where the air parcels began their back trajectories is described in full in Putman (2013).'

9. p. 4, L. 25: How did the authors calculate Td and are the average moisture source conditions computed as an arithmetic mean without taking into account the evaporative contribution to the air parcel's humidity at the different source locations?
The description of the calculation of Td is later in Methods section of the manuscript (Section 2.3.2). Because of the way the condensation profile is divided, it is assumed that each parcel contributes an equal amount of vapor to the final precipitating cloud, so we did not weight them.

P. 6 L. 6-9
'We approximate T_d using

$$T_d = [1/T_0 - 1.844*10^{-4} \ln(e_{sat;2m}\, h_{2m}\, 0.6113]^{-1} \quad (5)$$

(Stull, 2015) with saturation vapor pressure, $e_{sat;2m}$, from Equation 2 and the 2m air temperature, T_{2m}, and relative humidity h_{2m} from reanalysis data.'

P. 4 L. 27-29
'The precipitation rate profile was differentiated with respect to height, yielding the condensation rate profile (g m-3 s-1) and then subdivided into the aforementioned 1000 air parcels so as to ensure that each parcel contained an equal fraction of total precipitation.'

10. p. 4, L. 30: The authors should make clear that their T_{cool} is only an estimate of the total cooling that the air parcel has experienced. The same remark for the possibility of multiple moisture sources for one air parcel (see specific comment 4) is valid for cooling and precipitation along an air parcel trajectory. A trajectory can produce rain all along its path and can go through several cycles of cooling and warming. The total cooling would be obtained by integrating the temperature changes along the trajectory.
The reviewer is correct, the cooling indicated by ΔTcool is the net cooling, not the integral of cycles of warming and cooling an air parcel may have experienced along its trajectory. This is a simplification. This has been made explicit by additions and revisions to the text.

P. 5 L 8-10
'An estimate of air parcel cooling that produced condensation, ΔTcool, is a bulk metric quantifying the magnitude of Rayleigh distillation along the trajectory (Sodemann et al., 2008a). This approach simplifies the integration of cycles of warming and cooling that may occur along a trajectory to a net reduction in temperature.'

11. p. 5, L. 3-9: this way of computing T_{LCL} is confusing for me. Where does Eq. 2 come from? See Bolton (1980) and Lawrence (2005).
The equations have been changed to those in Stull (2015) and the description of the calculation has been updated. The previous equation was a linearized approximation that introduced minor, if not insignificant, discrepancy. To be more precise, the equations, calculations, and text have been updated to be consistent with Stull (2015), though the results and discussion require no change. Equations 2, 3, and 5 are affected.

P.5 L. 14-25
'To determine Qsat;z, we start from the dry adiabatic lapse rate, (-9.8 °C km-1). From this we determine the temperature Tz at altitude z, starting with the 2 meter temperature T2m.
The saturation vapor pressure at elevation z, esat;z is then

$$e_{sat;z} = 0.6113 \exp[5423 (1/T_0 - T/T_z)] \quad (2)$$

where T0 = 273.15K (Stull, 2015). We may then write the saturation specific humidity, Qsat;z, as

$$Q_z = 0.622\, e_{sat;z}\, h_z / P_z \quad (3)$$

where hz, the relative humidity at height z, is assumed to equal 1 (air is vapor-saturated) and the pressure at height z, Pz, is

$$P_z = 1013.25\, [1-(2{:}25577\ e\text{-}5)\ z]^{5.25588} \quad (4)$$

Calculating the 2m specific humidity, Q2m, is simply a special case of the general calculation: we use the 2m temperature T2m, fractional relative humidity h2m, and pressure P2m from reanalysis in Equations 2 and 3, rather than using the dry adiabatic lapse rate, h = 1, and Equation 4, respectively.

Finally, we find the elevation where Q2m equals Qsat;z. The temperature at this elevation is T_{LCL}.'

12. p. 5, L. 12-16: The idea to use T_d as a summary variable for both relative humidity with respect to sea surface temperature ($h_{2m\ SST}$) and SST-effects seems not justified to me from a physical point of view. The influence of SST on T_d is only indirect and a strong coupling of the ocean surface conditions with near-surface air characteristics is not necessarily given particularly at the event timescale. From a theoretical perspective and for all isotope-enabled numerical modelling experiments it is the Craig-Gordon model and thus the other two variables that are used to determine *d* of the fresh evaporate. So I am not convinced that it is sensible to introduce a third variable that does not contain more information than the specific humidity at 2 m. Furthermore, it should be made clear in the manuscript that it is not the 2m relative humidity that is important for the non-equilibrium fractionation part during surface evaporation but the humidity gradient towards the surface which is represented by the relative humidity at 2m with respect to sea surface temperature ($h_{2m\ SST}$). The authors should make a stronger case for why they use T_d rather than the classical variables. Also the sentence "T_d depends on the specific humidity of saturated air at the sea surface and on the amount of dry air from aloft that has subsided and mixed into low altitude air" is a confusing statement.

The use of Td and related discussion in this paper reflects that our group has done a substantial amount of work to model and understand isotopic variations in the marine boundary layer (manuscripts in preparation), and our understanding continues to improve. We realize that our discussion about Td in the earlier version was not clear, and it is valid for this reviewer to solicit further explanation. For clarity for the reader, we have completely rewritten section 3.2 that pertains to Td, and we hope the new discussion in the revised version is clearer.

In short, the idea of using Td is to indicate the moisture conditions within the PBL, as this is the moisture that forms the first condensate. This is different from the evaporative flux predicted by the Craig-Gordon model. In addition, the Craig-Gordon model does not consider effects of convection on vapor isotopic ratios in the PBL. However, convection is an important process that 1) transports PBL air to the free troposphere, and 2) brings dry air from aloft to the PBL. The boundary layer air is therefore a mixture of evaporated vapor from the ocean surface, and the dry air from aloft. The extent of this mixing within the PBL is reflected by (2m) dew point, Td.

Td is also useful because it is directly related to relative humidity with respect to the sea surface temperature (h2m, SST), moreso than is the 2 m relative humidity. Indeed, when h2m, SST was used in the multiple regression instead of Td, it was a significant predictor of δ2H. However, in both variance explained and AIC, the multiple regression that incorporated Td performed better. Both because it performs better in the multiple regression, and because it is a measurable quantity, we prefer Td to h2m,SST and have retained it in the paper.

P. 9-10 L. 7-35, 1-9
'We prefer Td to the classical variables Tss and h for determining isotopic evaporative fluxes. This choice is based on our understanding that the meteorological variable Td characterizes the bulk vapor content and isotopic ratio of the marine PBL, independent of the vapor temperature. When advected to the free troposphere, it is this vapor that will form precipitation. Additionally, through equilibrium fractionation Td also determines the isotopic ratio of the first condensate at the LCL, where Rayleigh distillation begins.

Within the marine PBL, several inter-related factors/processes are at work to determine the starting point of a Rayleigh trajectory. The first is the isotopic flux of evaporation from the sea surface. Most studies estimate this flux using the classic model by Craig and Gordon (1965). In that model, three variables control the evaporative flux: the sea surface temperature, Tss, δ2H above the laminar layer, and the humidity hss above the laminar layer (e.g., at 2 m), defined relative to Tss. Though hss is not a measured quantity nor one that is normally modeled, it is determined by Td above the laminar layer and Tss. Hence isotopic fluxes can be determined with the classical model using Tss, and Td and _2H above the laminar layer as input variables. From a physical point of view, Tss determines the amount of equilibrium fractionation at the water-air interface. Td and vapor δ2H, as well as Tss, control kinetic fractionation as vapor diffuses across the laminar layer. It should be noted that when Tss is large, Td tends to be large as well, as a result of their change with latitude and season. Td and δ2H are also correlated, which will be discussed below. Therefore, all three variables controlling the evaporative flux, Tss, and hss and δ 2H above the laminar layer, are associated directly or indirectly with Td, making Td a good indicator of evaporation conditions.

The second process is convergence. At a moisture source location, low level air is moist due to evaporation near the sea surface. Convergence and uplift transports low-level moist air into the free troposphere where it mixes with dry, isotopically depleted air descending from surrounding regions resulting in strong humidity and temperature gradients near the sea surface (below 2 m). In contrast, the specific humidity and isotopic ratios in the bulk of the PBL above 2m are relatively constant, resulting from the relative contributions of vertical transport of moist low-level and descending air (Fan, 2016). Td and δ2H at 2m both reflect the outcome of this mixing process, and so it follows that they are positively correlated.

The third process is condensation at the LCL. The temperature of the air mass, which equals or is very slightly less than the local dew point, determines the amount of isotopic fractionation and thus the isotopic ratio of the first condensate. It is this isotopic composition that defines the beginning of the Rayleigh part of the trajectory. Only Td;2m, not Tss nor h2m, is directly associated with the condensation temperature at the LCL (which differs only slightly from Td;2m due to the pressure difference between 2 m and the LCL and its effect on saturation specific humidity).

Since all three processes before Rayleigh distillation are either directly or indirectly related to Td, we consider Td a better indicator for the source conditions than either Tss or h. It is difficult, however, to theoretically assess the

sensitivity of precipitation δ2H to variations in source Td, because this would require quantification of the theoretical relationship of Td to δ2H through each of the three processes and perhaps their combinations. We here report the first empirical sensitivity of 3.23‰ °C-1 (Table 1) for δ2H relative to Td. At the sea surface, for Tss between 0 and 25 °C, equilibrium fractionation as a function of temperature yields sensitivities between 1.1-1.6‰°C-1 (Majoube, 1971). However, a large part of this fractionation may be offset by condensation at the LCL. Consequently, the observed sensitivity probably reflects primarily the fraction of vapor contributed by dry, isotopically depleted descending air that converges within the PBL. Mixing with the dry air causes a decrease in Td, which affects the δ2H of the PBL in two ways: 1) making the PBL air dry and isotopically depleted, and 2) isotopically depleting the evaporative flux by enhancing kinetic fractionation (an effect of low relative humidity). Both mechanisms produce a positive association between δ2H and Td, consistent with the sign of our observed partial coefficient (Table 1).'

13. p. 5, L. 17: Where does Eq. 3 come from? What is the impact of the simplification involved, the authors should add a chapter reference to Stull (2015). Why did they not use Stull (2015), Equation 4.15a or b or extract directly T_d from the reanalysis dataset?
Calculation was updated to eqn. 4.15b in Stull (2015). See response to 11.

14. p. 5, L. 23: How was *mtn* defined? Using an objective criterion or subjectively by looking at the trajectory plots?
It was determined manually by examining the trajectory plots.

P. 6 L. 17-18
'The value of mtn was assigned manually based on the general pattern of transport observed in the trajectory plots.'

15. p. 6, L. 2: remove parentheses.
Parenthesis removed.

P. 6 L. 20-22
'In this section we discuss the vapor source annual cycle and statistical relationships between the isotopic composition of precipitation, vapor source region, and the variables ΔTcool, Td, and mtn, that characterize the relationship of vapor source and transport to the isotope values measured at Barrow, AK.'

16. p. 6, L. 10-11: It would be useful to add the geographical names in one of the panels in Figure 1.
The authors appreciate the suggestion, but have chosen not to incorporate it because more text would make the figure too busy and obscure the data presented in the plot.

17. p. 6, L. 16-20: Is it really the variation in the moisture source latitude that is relevant or the mean transport distance? I am not convinced about the role of Figure 2. Also see major comment 3.
Because most vapor transport is from mid-latitudes to high latitudes, latitude is actually relevant for this site. Yes, distance might be another reasonable metric to investigate. However, latitude was chosen because latitude covaries with evaporation conditions, so it's more physically useful than distance.

Figure 2 is useful as it shows the relationship of latitude to vapor source and distillation. A similar figure could have been made for distance, but the outcomes would be very similar, as in our case distance variation is roughly the same as that of latitude.

18. p. 6, L. 21-32: For me this relatively long paragraph is a general discussion of the possible link between polar atmospheric circulation and the location of vapour sources and not a result from this study. Either the link with the findings in this paper should be illustrated more clearly or this section should be strongly shortened or even left out. See also my general comment 3: the link between the different timescales that are involved here is not trivial to make at this stage, a more open formulation should be chosen here.
The link between seasonal changes to general circulation and seasonal change in vapor source makes sense because general circulation is the background pattern from which weather events deviate. This paragraph only links the seasonality of vapor source with the seasonality of circulation patterns- nothing over longer timescales. However, the language has been updated to be less causal.

P. 7 L.20-25

'Aspects of the link between seasonal variability in general circulation and seasonal vapor source cycling may be generalizable to interannual and even millennial timescales. This is relevant to modern changes in the hydrologic cycle as Marvel and Bonfils (2013) suggest that a poleward displacement of circulation cells is already occurring due to recent climate change. Additionally, changes in the isotopic composition of precipitation resulting from systematic vapor source migrations associated with changing climate may allow for interpretation of long-term isotopic records in terms of changes in atmospheric circulation, including but not limited to the precipitation site temperature.'

19. p. 6, L. 24-26: In Europe several studies found that during summer the regional moisture recycling and the contribution from continental evaporation is much more important than in winter (see Sodemann and Zubler (2010) and Aemisegger et al. (2014)). Even though on p.4 L.3 the authors say that "only trajectories that sank into the PBL over the ocean" a substantial contribution of evaporation from continental Alaska is found in Spring but also in the other seasons in Figure 1. This possible contribution of continental evaporation should also be discussed as its moisture source isotope signature is different than the one from ocean evaporation.
Local evapotranspiration during spring and summer is likely an important vapor source. However, given the heterogeneity of the event conditions and sources, we did not have the statistical power to pull evapotranspiration out relative to the other factors. Nonetheless, in the discussion, evapotranspiration is added as a potential mechanism of seasonal change as it likely does contribute to the δ2H measured in precipitation. Furthermore, it has been included in the sources of error discussion at the end of Section 3.2.

P. 7 L. 11-17
'The migration of the mean latitude of the vapor source region can be tied to the seasonal cycling of solar insolation in the northern hemisphere via two mechanisms. Decreased solar insolation during winter drives expansion of the northern Polar circulation cell, which increases sea ice cover, and cold temperatures and snow cover prevent evapotranspiration. Both sea ice cover which diminishes the vapor contributions of the Arctic Ocean, and inhibited evapotranspiration allow for enhanced 15 representation of southerly vapor sources. Increased summer insolation drives poleward contraction of the circulation cell, diminishes sea ice coverage, and warmer temperatures favors evapotranspiration such that the average vapor source area migrates north.'

P. 11 L. 13-20
'The three chosen variables explain just over half (54%) the variance of δ2H. This is not surprising, considering that many other mechanisms can also influence the δ2H of the vapor and precipitation. These mechanisms include (but are not limited to) condensation temperature, supersaturation in the mixed phase cloud, sub-cloud dryness, phase of precipitation, precipitation intensity, evapotranspiration of land sources, and the amount of sea ice at the vapor source. The effects of several of these factors, including condensation temperature, sub-cloud dryness, sea ice concentration at the vapor source, and phase of precipitation (rain vs. snow), were tested as additional explanatory variables in the multiple regression, but yielded statistically insignificant results with little to no additional variance explained. Clearly, compared with the three chosen variables, the effects of these variables are relatively minor, such that the statistical power is not sufficient to reveal their significance.'

20. p. 6, L. 3: Add mid- to high latitudes here, other studies could be cited as well (e.g. Bonne et al. (2014))
The phrase is changed, and citation added.

P. 7 L. 29-30
'…that follows the well-established annual cycle for mid- and high latitudes (Feng et al., 2009; Bonne 30 et al., 2014)'

21. p. 7, L. 10-11: References to figures are confusing.
The references to figures were removed.

P. 8 L. 2-4
'…1) the temperature difference between vapor source region and precipitation site, quantified by air parcel cooling ΔTcool, 2) the moisture source conditions, quantified in this work by Td, and 3) the mean air parcel transport path.'

22. p. 7, L. 13: Do the authors mean the regression slopes? It would be useful to add the units of the slopes in all tables. Also in Table 3 it would be useful to add the explanation on what β and S.E. are.

Changed the text to regression slopes. β and S.E. are described in the text that refers to Table 3. The units, originally just of the variable, are now the units of the slope in all tables. This is reflected in the column label.

P. 8 L. 5-6
'Table 1 contains the partial regression slopes (β), p-values, and the unique variance explained by each variable.'
Caption of Tab 1.
'Values of β are the partial coefficients of the regression and S.E. is the standard error.'

Caption of Tab 2
'…where β is the regression coefficient and S.E. is the standard error'

Caption of Tab 3
'β is the regression coefficient and S.E. is the standard error.'

23. p. 7, L. 27: Here and elsewhere the references should be listed chronologically.
Checked references throughout manuscript and reordered where necessary.

24. p. 7, L. 34 - p. 8, L. 10: Here more detailed explanations on the theoretical cooling/Rayleigh experiment are needed to be able to follow. Also the sensitivity range of δ^2H to the diagnosed cooling should be put into context and compared to literature values.
The model is explained in greater detail. The simple model is meant to contextualize the numbers.

P.8 L. 27-29
'In such a model, a saturated air parcel with specified temperature and vapor δ2H is cooled iteratively in 1°C steps. At each temperature step, the condensation amount, remaining vapor, precipitation δ2H and vapor δ2H are calculated. No re-evaporation or non-equilibrium conditions are considered.'

25. p. 7, L. 21: Table 1: do the regression slopes from Table 1 result from multiple linear regression?
Yes, this is stated in the text though it has been added to the Table caption now. The regressions in Tables 2 and 3 are now clearly indicated as simple linear regressions.

P. 8 L. 4-6
'A linear combination of ΔTcool, Td, and mtn statistically represents the event-scale variation in δ2H with an R2 5 value of 0.54 ($p < 0.001$). Table 1 contains the partial regression slopes (β), pvalues, and the unique variance explained by each variable.'

P. 8 L. 24-25
'Our multiple regression yields…'

Caption Tab 1
'Variation in δ2H is explained by a multiple linear regression…'

Caption Tab 2
'Three simple linear regressions…'

Caption Tab 3
'Explaining deuterium excess (d) using simple regressions against…'

26. p. 9, L. 23: "within storm" is a confusing term here as it suggests that the precipitation is due to the passage of a cyclone, which is not always the case. I would suggest using "intra-event" instead.
Changed to intra-event.

P. 11 L.6
'…the distribution of intra-event…'

27. p. 10, L. 17: I am surprised at the *d*–*h* slope which is not at all in agreement (opposite sign and different order of magnitude) with other literature values (*d*-0.6h%$_{-1}$ to -0.32h%$_{-1}$, though a difference with literature values is that h_{2m} is used and not $h_{2m\ SST}$). This mismatch should be explained and the relevant literature should be cited (Pfahl and Wernli, 2008; Steen-Larsen et al., 2014; Aemisegger et al., 2014). Also the *d*-SST regression slope is of opposite sign to what we would expect from the Craig-Gordon model.
We too were also surprised at the outcome with respect to h2m and SST. This comment prompted us to calculate the statistics again with respect to h2m,SST. The results were significant ($p < 0.001$), and in the range of the values described above:- 0.39+/- 0.067 h%-1, with 34% variance explained. This has been added to the paper in this section as the primary result and contextualized by the values reported in the suggested citations.

The non-significant relationship to h2m and SST supports our prior argument (Section 3.2, point 12) that the isotope flux predicted by Craig-Gordon is only one process controlling the vapor properties in the PBL and does not indicate the bulk conditions within the PBL. Thus, a statistical non-relationship with variables that control isotopic flux is reasonable. The statistically significant results with h2m, SST and Td support our argument that the isotope value reflects the bulk conditions of the PBL, which are the result of a combination of the evaporative flux and mixing with drier air from aloft. This is discussed further in point 28, below.

P. 12 L. 9-11
'…it is significantly predicted by h_{ss} ($p < 0.001$, $R_2 = 0.34$), with a slope of -0.4‰%-1. This value is consistent with the -0.4 to -0.6‰ %-1 range reported in the literature for vapor (Uemura et al., 2008; Pfahl and Wernli; Bonne et al.,2014)'

P. 9-10 L. 7-35, 1-9 (see point 12)

28. p. 10, L. 20: What is the theoretical expectation for the sign of the correlation between *d* and T_d? This should be explained in more detail. I do not agree with the statement made here, I would expect a negative *d*-T_d slope from theory since the physical relation between relative humidity and T_d should generally lead to a positive correlation between the latter two (see e.g. Lawrence (2005)).
Yes, we agree that we should expect a negative Td-d slope. The authors thank the reviewer for pointing this out. We now discuss this negative relationship in the context of two processes, 1) dry air (low Td) causing larger kinetic fractionation and higher d, and 2) the descending air may have high d values. Both processes, independently and together yield negative association between d and Td. The new information about the relationship to h2m, SST from point 27 supports our argument that Td and h2m, SST are related, and both are more representative of the PBL conditions than are Tss and h.

P. 12 L. 15-20
'…with a negative slope (-0.53‰°C-1). This is an interesting result with respect to the utility of T_d, a measurable quantity, and is consistent with our earlier argument that T_d is strongly related to h_{ss}. Both variables provide a better representation of source conditions than T_{ss} and/or h_{2m}. A low value of h_{ss} or T_d corresponds to a strong influence of descending dry air within the PBL, which enhances kinetic isotopic fractionation and produces a high value of d. This mechanism explains the negative correlation between d and T_d, and is expected for the relationship between d and h_{ss}. Alternatively, the vapor in descending air may have a high value of d (Fan, 2016), or both mechanisms may contribute to this result.'

29. Figures 3 and 4: more details are needed on the used spline fits. Also the strong inter-event variability that is sometimes of similar amplitude as the seasonal cycle should be discussed.
Details on the spline fits have been added to the figure caption, and the similarity in amplitude among seasonal and event variability is noted for both datasets in various pertinent locations in the paper.

P. 7 L. 30-34
'…where the spline captures 65% of the annual and interannual variance. The average annual cycle of the precipitation δ_2H is strong; the spline fit explains 60% of variance in the data. The mean latitude of the vapor source exhibits a weak seasonal pattern, where the spline explains 19% of the variance. The seasonal cycles of $_2H$ and vapor source latitude are in phase, as shown in Figure 4, though the inter-event variability in both variables can be as large as the seasonal variability.'

P. 12 L.28-29
'…exhibited interannual, annual, and substantial inter-event variability.'

P.12 L. 33
'However, substantial intra-season variability occurred in both source and δ_2H, indicating scatter in the seasonal relationship'

Caption Fig 3
'The spline fit, which highlights seasonal variations, explains 65% of variance in the data with a root mean squared error of 39.7‰. Of the three timescales, annual variability shows the greatest amplitude, though variability among events is also substantial.'

Caption Fig 4
'The spline fits have R_2 values of 0.60 and 0.19 for the δ_2H and V_{Lat} respectively. For both datasets, the variability exhibited among events is of the same the order of magnitude as the seasonal variability.'

30. Figure 5: the role of this Figure is unclear to me, it is only referenced once and not further discussed in the text. Either this Figure should be better embedded in the text or it should be left out. If it is kept: is this figure an average over all events?
Yes, this figure shows the average value of δ2H of precipitation coming from a specific vapor source, which indicates that certain regions tend to be vapor sources for precipitation events that are either more or less enriched than average for that time of year. Mountains along the trajectory appear to be the mechanism at work in producing the spatial structure.

P. 8 L. 14-16
*'*As demonstrated by Figure 5, the presence of mountains along the vapor transport path will deplete the isotope ratio of the precipitation relative to a uniform altitude transport, all other meteorological conditions being equivalent.'

References

Aemisegger, F., S. Pfahl, H. Sodemann, I. Lehner, S. I. Seneviratne, and H. Wernli, 2014: Deuterium excess as a proxy for continental moisture recycling and plant transpiration. *Atmos. Chem. Phys.*, **14 (8)**, 4029–4054, doi:10.5194/acp-14-4029-2014.

Aemisegger, F., J. K. Spiegel, S. Pfahl, H. Sodemann, W. Eugster, and H. Wernli, 2015: Isotope meteorology of cold front passages: A case study combining observations and modeling. *Geophysical Research Letters*, **42 (13)**, 2015GL063 988, doi:10.1002/2015GL063988.

Bolton, D., 1980: The Computation of Equivalent Potential Temperature. *Monthly Weather Review*, **108 (7)**, 1046–1053, doi:10.1175/1520-0493(1980)108¡1046:TCOEPT¿2.0.CO;2.

Bonne, J.-L., V. Masson-Delmotte, O. Cattani, M. Delmotte, C. Risi, H. Sodemann, and H. C. Steen-Larsen, 2014: The isotopic composition of water vapour and precipitation in Ivittuut, southern Greenland. *Atmospheric Chemistry and Physics*, **14 (9)**, 4419–4439, doi:10.5194/acp-14-4419-2014.

Lawrence, M. G., 2005: The Relationship between Relative Humidity and the Dewpoint Temperature in Moist Air: A Simple Conversion and Applications. *Bulletin of the American Meteorological Society*, **86 (2)**, 225–233, doi:10.1175/BAMS-86-2-225.

Pfahl, S. and H. Wernli, 2008: Air parcel trajectory analysis of stable isotopes in water vapor in the eastern Mediterranean. *Journal of Geophysical Research: Atmospheres*, **113 (D20)**, D20 104, doi: 10.1029/2008JD009839.

Pfahl, S., H. Wernli, and K. Yoshimura, 2012: The isotopic composition of precipitation from a winter storm – a case study with the limited-area model COSMOiso. *Atmos. Chem. Phys.*, **12 (3)**, 1629–1648, doi:10.5194/acp-12-1629-2012.

Sodemann, H., C. Schwierz, and H. Wernli, 2008: Interannual variability of Greenland winter precipitation sources: Lagrangian moisture diagnostic and North Atlantic Oscillation influence. *Journal of Geophysical Research: Atmospheres*, **113 (D3)**, D03 107, doi:10.1029/2007JD008503.

Sodemann, H. and E. Zubler, 2010: Seasonal and inter-annual variability of the moisture sources for Alpine precipitation during 1995–2002. *International Journal of Climatology*, **30 (7)**, 947–961, doi: 10.1002/joc.1932.

Steen-Larsen, H. C., et al., 2014: Climatic controls on water vapor deuterium excess in the marine boundary layer of the North Atlantic based on 500 days of in situ, continuous measurements. *Atmospheric Chemistry and Physics*, **14 (15)**, 7741–7756, doi:10.5194/acp-14-7741-2014.
Stull, R., 2015: *Practical Meteorology: An Algebra-Based Survey of Atmospheric Science, Chapter 4, Water Vapor*. Univ. of British Columbia.

All of these suggested references have been checked and cited when appropriate.

List of all relevant changes in manuscript
Line numbers refer to the final document, not the markup. Substantial blocks of text that have been removed are noted. Though the text included here may not be entirely new, all have significant additions, or have undergone substantial structural or content change in response to comments from reviewers.

Title:
Added 'event-scale'
Rearranged to 'precipitation δ2H at Barrow, AK reflects vapor source region'

Abstract
Removed 'Interpretation of variability in precipitation stable isotopic ratios often relies exclusively on empirical relationships to meteorological variables (e.g., temperature) at the precipitation site. Because of the difficulty of unambiguously determining the vapor source region(s), relatively fewer studies consider evaporation and transport conditions. Increasing accessibility of Lagrangian air parcel tracking programs now allows for an integrated look at the relationship between the precipitation isotope ratios and the evolution of moist air masses. In this study, 70 precipitation events occurring…'

P. 1 L. 1-2
"In this study, precipitation isotopic variations are linked to conditions at the moisture source region, along the transport path, and at the site of precipitation. Seventy precipitation events…"
P. 1 L. 5
'…occurring…'
Removed: We expect isotopes to respond similarly for longer-term climate-induced changes to the mean position of meridional circulation features, and expect that the most of the variation in isotopes measured in ice cores and other long term records are driven by changes in circulation, instead of fluctuations in local temperature.

P.3 L.13-16
'…and often as soon as snow ended. Though it is possible that snow may have been altered by sublimation before collection, we assume that the degree of alteration of surface snow was minimal relative to the amount of snow gathered. Furthermore, the frequent cloudiness and darkness of Barrow mean that for most events, sunlight-driven sublimation was insignificant.'

P.3 L.16-19
'Liquid samples were stored in tightly sealed 30mL Nalgene bottles below 5 °C and shipped in batches every three months to the Stable Isotope Laboratory at Dartmouth College. When not in transit, samples were refrigerated. Samples were analyzed within six months of collection.'

P. 4 L. 3
'…using 1° resolution meteorological data from the Global Data Assimilation System (GDAS).'

P.4 L. 6-14
'The vapor source location was defined as the place where the back trajectory of the air parcel sank into the planetary boundary layer (PBL). Relative to previous studies that tracked vapor change in an air parcel along the trajectory (e.g., Sodemann et al. (2008a)), we adopted a simpler procedure that assumes vapor in the air parcel is well represented by the air at the latest interaction with the PBL. This assumption is justified because mass movement in the PBL is dominated by vertical turbulence relative to horizontal advection. Figure 1 shows endpoints of all trajectories that sank into the PBL. However, only trajectories that ended over water with < 96% sea ice cover were used for calculations; parcels that sank where there was less than 96% sea ice cover were used for calculations. Ocean-originating air parcels comprised about 71% of all trajectories.'

P.4 L.15-23
'Back trajectory analysis was performed for dates when precipitation was collected. The starting times for the back trajectories corresponded to times of maximum precipitation intensity, based on a combination of sampling records,

surface analysis maps of Alaska available through the National Center for environmental Prediction, and the returns of the millimeter wavelength cloud radar (MMCR) (Johnson and Jensen, 1996; Bharadwaj et al., 2011). Greater Doppler vertical velocities, reflectivities, and spectral widths from the MMCR broadly indicated more intense precipitation. Because the gridded meteorological files 20 used for tracing the back trajectories had three-hour resolution, the chosen starting time represented average conditions over a three-hour period. If precipitation lasted for more than three hours, the most intense three hour time window was selected. If the precipitation was of approximately uniform intensity, the most temporally homogeneous three-hour time window was selected, with preference for time windows where precipitation occurred over the duration of the three hours.'

P. 4 L. 24-25
'The method for selecting the altitudes where the air parcels began their back trajectories is described in full in Putman (2013).'

P. 4 L. 27-29
'The precipitation rate profile was differentiated with respect to height, yielding the condensation rate profile (g m-3 s-1) and then subdivided into the aforementioned 1000 air parcels so as to ensure that each parcel contained an equal fraction of total precipitation.'

P. 5 L 8-10
'An estimate of air parcel cooling that produced condensation, ΔTcool, is a bulk metric quantifying the magnitude of Rayleigh distillation along the trajectory (Sodemann et al., 2008a). This approach simplifies the integration of cycles of warming and cooling that may occur along a trajectory to a net reduction in temperature.'

P.5 L. 14-25
'To determine Qsat;z, we start from the dry adiabatic lapse rate, (-9.8 °C km-1). From this we determine the temperature Tz at altitude z, starting with the 2 meter temperature T2m.
The saturation vapor pressure at elevation z, esat;z is then

$$esat;z = 0.6113 \exp[5423 (1/T_0 - T/T_z)] \quad (2)$$

where T0 = 273.15K (Stull, 2015). We may then write the saturation specific humidity, Qsat;z, as

$$Qz = 0.622\ esat;z\ hz\ /Pz \quad (3)$$

where hz, the relative humidity at height z, is assumed to equal 1 (air is vapor-saturated) and the pressure at height z, Pz, is

$$Pz = 1013.25\ [1-(2.25577\ e\text{-}5)\ z]^{5.25588} \quad (4)$$

Calculating the 2m specific humidity, Q2m, is simply a special case of the general calculation: we use the 2m temperature T2m, fractional relative humidity h2m, and pressure P2m from reanalysis in Equations 2 and 3, rather than using the dry adiabatic lapse rate, h = 1, and Equation 4, respectively.

Finally, we find the elevation where Q2m equals Qsat;z. The temperature at this elevation is T_{LCL}.'

P. 6 L. 6-9
'We approximate Td using

$$T_d = [1/T_0 - 1.844*10^{-4} \ln(e_{sat;2m}\ h_{2m}\ 0.6113]^{-1} \quad (5)$$

(Stull, 2015) with saturation vapor pressure, esat;2m, from Equation 2 and the 2m air temperature, T2m, and relative humidity h2m from reanalysis data.'

P. 6 L. 17-18
'The value of mtn was assigned manually based on the general pattern of transport observed in the trajectory plots.'

P. 6 L. 18-19
'The value of mtn was assigned manually based on the general pattern of transport observed in the trajectory plots.'
P. 6 L. 20-22
'In this section we discuss the vapor source annual cycle and statistical relationships between the isotopic composition of precipitation, vapor source region, and the variables ΔTcool, Td, and mtn, that characterize the relationship of vapor source and transport to the isotope values measured at Barrow, AK.'

P. 7 L. 11-17
'The migration of the mean latitude of the vapor source region can be tied to the seasonal cycling of solar insolation in the northern hemisphere via two mechanisms. Decreased solar insolation during winter drives expansion of the northern Polar circulation cell, which increases sea ice cover, and cold temperatures and snow cover prevent evapotranspiration. Both sea ice cover which diminishes the vapor contributions of the Arctic Ocean, and inhibited evapotranspiration allow for enhanced 15 representation of southerly vapor sources. Increased summer insolation drives poleward contraction of the circulation cell, diminishes sea ice coverage, and warmer temperatures favors evapotranspiration such that the average vapor source area migrates north.'

P.7 L.19
'There is evidence for prior millenium-scale shifts in the southern extent of the polar circulation cell (Feng et al., 2007).'

P. 7 L. 20-25
'Aspects of the link between seasonal variability in general circulation and seasonal vapor source cycling may be generalizable to interannual and even millennial timescales. This is relevant to modern changes in the hydrologic cycle as Marvel and Bonfils (2013) suggest that a poleward displacement of circulation cells is already occurring due to recent climate change. Additionally, changes in the isotopic composition of precipitation resulting from systematic vapor source migrations associated with changing climate may allow for interpretation of long-term isotopic records in terms of changes in atmospheric circulation, including but not limited to the precipitation site temperature.'

P. 7 L. 29-30
'…that follows the well-established annual cycle for mid- and high latitudes (Feng et al., 2009; Bonne
30 et al., 2014)'

P. 7 L. 30-34
'Figure 3 also shows the interannual, seasonal and event-scale variability captured by the dataset where the spline captures 65% of the annual and interannual variance. The average annual cycle of the precipitation δ2H is strong; the spline fit explains 60% of variance in the data. The mean latitude of the vapor source exhibits a weak seasonal pattern, where the spline explains 19% of the variance. The seasonal cycles of δ2H and vapor source latitude are in phase, as shown in Figure 4, though the inter-event variability in both variables can be as large as the seasonal variability'

P. 8 L. 2-4
'…1) the temperature difference between vapor source region and precipitation site, quantified by air parcel cooling ΔTcool, 2) the moisture source conditions, quantified in this work by Td, and 3) the mean air parcel transport path.'

P. 8 L. 5-6
'Table 1 contains the partial regression slopes (β), p-values, and the unique variance explained by each variable.'

P. 8 L. 14-16
'As demonstrated by Figure 5, the presence of mountains along the vapor transport path will deplete the isotope ratio of the precipitation relative to a uniform altitude transport, all other meteorological conditions being equivalent.'

P. 8 L. 24-25
'Our multiple regression yields…'

P. 8 L. 25-29
'Because Rayleigh distillation is considered the main source of spatial variation in δ2H, comparison with the sensitivities calculated from a simple Rayleigh model contextualize our result. In such a model, a saturated air parcel with specified temperature and vapor δ2H is cooled iteratively in 1°C steps. At each temperature step, the condensation amount, remaining vapor, precipitation δ2H and vapor δ2H are calculated. No re-evaporation or non-equilibrium conditions are considered.'

P. 9-10 L. 7-35, 1-9
'We prefer Td to the classical variables Tss and h for determining isotopic evaporative fluxes. This choice is based on our understanding that the meteorological variable Td characterizes the bulk vapor content and isotopic ratio of the marine PBL, independent of the vapor temperature. When advected to the free troposphere, it is this vapor that will form precipitation. Additionally, through equilibrium fractionation Td also determines the isotopic ratio of the first condensate at the LCL, where Rayleigh distillation begins.

Within the marine PBL, several inter-related factors/processes are at work to determine the starting point of a Rayleigh trajectory. The first is the isotopic flux of evaporation from the sea surface. Most studies estimate this flux using the classic model by Craig and Gordon (1965). In that model, three variables control the evaporative flux: the sea surface temperature, Tss, δ2H above the laminar layer, and the humidity hss above the laminar layer (e.g., at 2 m), defined relative to Tss. Though hss is not a measured quantity nor one that is normally modeled, it is determined by Td above the laminar layer and Tss. Hence isotopic fluxes can be determined with the classical model using Tss, and Td and _2H above the laminar layer as input variables. From a physical point of view, Tss determines the amount of equilibrium fractionation at the water-air interface. Td and vapor δ2H, as well as Tss, control kinetic fractionation as vapor diffuses across the laminar layer. It should be noted that when Tss is large, Td tends to be large as well, as a result of their change with latitude and season. Td and δ2H are also correlated, which will be discussed below. Therefore, all three variables controlling the evaporative flux, Tss, and hss and δ 2H above the laminar layer, are associated directly or indirectly with Td, making Td a good indicator of evaporation conditions.

The second process is convergence. At a moisture source location, low level air is moist due to evaporation near the sea surface. Convergence and uplift transports low-level moist air into the free troposphere where it mixes with dry, isotopically depleted air descending from surrounding regions resulting in strong humidity and temperature gradients near the sea surface (below 2 m). In contrast, the specific humidity and isotopic ratios in the bulk of the PBL above 2m are relatively constant, resulting from the relative contributions of vertical transport of moist low-level and descending air (Fan, 2016). Td and δ2H at 2m both reflect the outcome of this mixing process, and so it follows that they are positively correlated.

The third process is condensation at the LCL. The temperature of the air mass, which equals or is very slightly less than the local dew point, determines the amount of isotopic fractionation and thus the isotopic ratio of the first condensate. It is this isotopic composition that defines the beginning of the Rayleigh part of the trajectory. Only Td;2m, not Tss nor h2m, is directly associated with the condensation temperature at the LCL (which differs only slightly from Td;2m due to the pressure difference between 2 m and the LCL and its effect on saturation specific humidity).

Since all three processes before Rayleigh distillation are either directly or indirectly related to Td, we consider Td a better indicator for the source conditions than either Tss or h. It is difficult, however, to theoretically assess the sensitivity of precipitation δ2H to variations in source Td, because this would require quantification of the theoretical relationship of Td to δ2H through each of the three processes and perhaps their combinations. We here report the first empirical sensitivity of 3.23‰ °C-1 (Table 1) for δ2H relative to Td. At the sea surface, for Tss between 0 and 25 °C, equilibrium fractionation as a function of temperature yields sensitivities between 1.1-1.6‰°C-1 (Majoube, 1971). However, a large part of this fractionation may be offset by condensation at the LCL. Consequently, the observed sensitivity probably reflects primarily the fraction of vapor contributed by dry, isotopically depleted descending air that converges within the PBL. Mixing with the dry air causes a decrease in Td, which affects the δ2H of the PBL in two ways: 1) making the PBL air dry and isotopically depleted, and 2) isotopically depleting the evaporative flux by enhancing kinetic fractionation (an effect of low relative humidity). Both mechanisms produce a positive association between δ2H and Td, consistent with the sign of our observed partial coefficient (Table 1).'

Removed:
'This is because source meteorological conditions control the δ2 H of the water evaporated from the ocean surface. Studies typically attribute variations in _2 H values at the source to mean sea surface temperature Tss and mean 2m relative humidity h2m, which affect the magnitudes of equilibrium and kinetic fractionation, respectively (Craig and Gordon, 1965). Assuming that Tss influences 2m air temperature, so that the two temperatures correlate spatially, we may use the 2m dew point (Td) at the vapor source to combine the effects of Tss and h2m . Either high Tss or high h2m results in high Td . Therefore, we expect Td to be positively associated with δ2 H in the original vapor in an airparcel at the vapor source. This δ2 H signal at the source is then carried to the precipitation site. Differences in T2m cause 20°C> of the range we report for Td , whereas h2m contributes 2-4°C.The substantial difference between the vapor source Td for Arctic compared with subtropical sources makes Td a more useful metric for characterizing the vapor source than either Tss or h2m alone.'

P.10 L.22-24
'The breakpoint of 7 °C was chosen by testing different breakpoints and finding one that maximized the statistical power of the short trajectory regression while preserving the strong relationship between δ2H and Td.'

P.10 L.21-22
'Table 2 summarizes the results and Figure 6 shows the standard deviation of σTd by category'

P. 11 L.6
'…the distribution of intra-event…'

P. 11 L. 13-20
'The three chosen variables explain just over half (54%) the variance of δ2H. This is not surprising, considering that many other mechanisms can also influence the δ2H of the vapor and precipitation. These mechanisms include (but are not limited to) condensation temperature, supersaturation in the mixed phase cloud, sub-cloud dryness, phase of precipitation, precipitation intensity, evapotranspiration of land sources, and the amount of sea ice at the vapor source. The effects of several of these factors, including condensation temperature, sub-cloud dryness, sea ice concentration at the vapor source, and phase of precipitation (rain vs. snow), were tested as additional explanatory variables in the multiple regression, but yielded statistically insignificant results with little to no additional variance explained. Clearly, compared with the three chosen variables, the effects of these variables are relatively minor, such that the statistical power is not sufficient to reveal their significance.'

P.11 L.31-32
'While studies indicate that d in vapor contains vapor source information (Steen-Larsen et al., 2014; Bonne et al., 2015; Steen-Larsen et al., 2015)…'

P.12 L. 9-11
'This value is consistent with the -0.4 to -0.6‰ %-1 range reported in the literature for vapor (Uemura et al., 2008; Pfahl and Wernli; Bonne et al.; 2014).'

P. 12 L. 15-20
'This is an interesting result with respect to the utility of Td, a measurable quantity, and is consistent with our earlier argument that Td is strongly related to hss. Both variables provide a better representation of source conditions than Tss and/or h2m. A low value of hss or Td corresponds to a strong influence of descending dry air within the PBL, which enhances kinetic isotopic fractionation and produces a high value of d. This mechanism explains the negative correlation between d and Td, and is expected for the relationship between d and hss. Alternatively, the vapor in descending air may have a high value of d (Fan, 2016), or both mechanisms may contribute to this result.'

P. 12 L.28-29
'…exhibited interannual, annual, and substantial inter-event variability.'

P.12 L. 33

‘However, substantial intra-season variability occurred in both source and δ2H, indicating scatter in the seasonal relationship’

P. 13 L. 14-16

‘The mechanisms identified, most notably the north-south migration of the vapor source region in phase with expansion and contraction of the Polar circulation cell, may also operate on times scales longer than that of our study, and may be a source of variation in isotopes measured in ice cores, pedogenic carbonates, and speleothems.’

Caption of Fig 1

‘The figure indicates that some air parcels originate over land, but these were not included in calculations.’

Caption of Fig 3

*‘*The spline fit, which highlights seasonal variations, explains 65% of variance in the data with a root mean squared error of 39.7‰. Of the three timescales, annual variability shows the greatest amplitude, though variability among events is also substantial.*’*

Caption of Fig 4

*‘*The spline fits have R2 values of 0.60 and 0.19 for the δ2H and VLat respectively. For both datasets, the variability exhibited among events is of the same the order of magnitude as the seasonal variability.*’*

Caption of Tab 1

‘Values of β are the partial coefficients of the regression and S.E. is the standard error.’

*‘*Variation in δ2H is explained by a multiple linear regression (R2 = 0.54) of air parcel cooling during transport (ΔTcool), moisture source conditions (Td) and orographic obstacles in vapor transport path (mtn).’

Caption of Tab 2

‘…where β is the regression coefficient and S.E. is the standard error’

‘Three simple linear regressions against δ2H where β is the regression coefficient and S.E. is the standard error’

Caption of Tab 3

‘Explaining deuterium excess (d) using simple regressions against…’

‘β is the regression coefficient and S.E. is the standard error.’

Annual variation in event-scale precipitation ~~δ^2 H reflects vapor source region~~ $\delta^2 H$ at Barrow, AK reflects vapor source region

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Abstract. ~~Interpretation of variability in precipitation stable isotopic ratios often relies exclusively on empirical relationships to meteorological variables (e.g., temperature) at the precipitation site. Because of the difficulty of unambiguously determining the vapor source region(s), relatively fewer studies consider evaporation and transport conditions. Increasing accessibility of Lagrangian air parcel tracking programs now allows for an integrated look at the relationship between the precipitation isotope ratios and the evolution of moist air masses. In this study, 70 precipitation events occurring~~ In this study, precipitation isotopic variations at Barrow, AK, USA are linked to conditions at the moisture source region, along the transport path, and at the precipitation site. Seventy precipitation events between January 2009 and March 2013 ~~at Barrow, AK, USA,~~ were analyzed for ~~δ^2 H~~ $\delta^2 H$ and deuterium excess. For each precipitation event, vapor source regions were identified with the Lagrangian air parcel tracking program, HYSPLIT, in back-cast mode. The results show that the vapor source region migrated annually, with the most distal (proximal) and southerly (northerly) vapor source regions ~~occurred~~ occurring during the winter (summer). This may be ~~linked~~ related to equatorial expansion and poleward contraction of the Polar circulation cell and the extent of Arctic sea ice cover. Annual cycles of vapor source region latitude and ~~δ^2 H~~ $\delta^2 H$ in precipitation were in phase; depleted (enriched) ~~δ^2 H~~ $\delta^2 H$ values were associated with winter (summer) and distal (proximal) vapor source regions. Precipitation ~~δ^2 H~~ $\delta^2 H$ responded to variation in vapor source region as reflected by significant correlations between ~~δ^2 H~~ $\delta^2 H$ with the following three parameters: 1) total cooling between lifted condensation level and precipitating cloud at Barrow, $\Delta\bar{T}_{cool}$, 2) ~~the~~ meteorological conditions at the evaporation site quantified by 2 m dew point, $\bar{T}_d$, and 3) whether the ~~transport~~ vapor transport path crossed the Brooks and/or Alaskan ranges, expressed as a Boolean variable, mtn. These three variables explained 54 % of the variance ($p < 0.001$) in precipitation ~~δ^2 H~~ $\delta^2 H$ with a sensitivity of -3.51 $\pm$ 0.55 ‰ °C^{-1} ($p < 0.001$) to $\Delta\bar{T}_{cool}$, 3.23 $\pm$ 0.83 ‰ °C^{-1} ($p < 0.001$) to T_d, and -32.11 $\pm$ 11.04 ‰ ($p =$ ~~0.0028~~0.0049) depletion when mtn is true. The magnitude of each effect on isotopic composition also varied with vapor source region proximity. For storms with proximal vapor source regions (where $\Delta\bar{T}_{cool}$ <7 °C), $\Delta\bar{T}_{cool}$ explained 3 % of the variance in ~~δ^2 H~~ $\delta^2 H$, $\bar{T}_d$ alone accounted for 43 %, while mtn explained 2 %. For storms with distal vapor sources ($\Delta\bar{T}_{cool}$ >7°C), $\Delta\bar{T}_{cool}$ explained 22 %, $\bar{T}_d$ explained only 1 %, and mtn explained 18 %. The deuterium excess annual cycle lagged by 2-3 months the ~~δ^2 H~~ $\delta^2 H$ cycle, so the direct correlation between the two variables is weak. ~~Neither vapor source~~ Vapor source region relative humidity with respect to the sea surface temperature, ~~nor vapor source relative humidity, nor a linear combination of the two, was a statistically significant predictor of precipitation deuterium excess. Vapor source region $\bar{T}_d$ explained~~ $\bar{h}_{ss}$, explained 34 % of variance in deuterium excess,

(-0.395 $\pm$ 0.067 ‰ % $^{-1}$, p < 0.001). The patterns in our data suggest that on an annual scale, isotopic ratios of precipitation at Barrow may respond to changes in the southerly extent of the Polar circulation cell. We expect isotopes to respond similarly for longer-term climate-induced changes to the mean position of meridional circulation features, and expect that the most of the variation in isotopes measured in ice cores and other long term records are driven by changes in circulation, instead of fluctuations in local temperature, a relationship that may be applicable to interpretation of long term climate change records like ice cores.

1 Introduction

Changes to spatial patterns of water vapor transport and precipitation are an important component of incipient climate change (Santer et al., 2007; Marvel and Bonfils, 2013). The Arctic exhibits a particularly strong hydrologic response, including a notable increase in Arctic precipitation (Min et al., 2008; Bintanja and Selten, 2014; Kopec et al., 2016). Current and future changes in the hydrologic cycle may impact fresh water resources, natural disasters, and earth's radiation balance, due to changes in timing, extent, and duration of snow or cloud cover (Liu et al., 2012).

Like changes in the timing or amount of precipitation, changes in the relative abundance of heavy-isotope substituted water molecules in precipitation (e.g., $^{1}H_{2}^{16}O$ vs. $^{1}H_{2}^{18}O$ and $^{1}H^{2}H^{16}O$) may reflect effects of changing climate on the hydrologic cycle. Historically, researchers have measured the isotopic ratios of precipitation on monthly or longer timescales and attempted to explain the temporalisotopic variations over time, altitude, and latitude variation of the isotopic ratios (Cappa et al., 2003; Rindsberger et al. Empirical analysis has focused on weather and climate conditions at the precipitation site (?)(Dansgaard, 1964). Models developed to understand the spatial and temporal variability of water stable isotopes include evaporation and Rayleigh distillation models (Merlivat and Jouzel, 1979; Jouzel and Merlivat, 1984), models examining the balance of vertical mixing and meridional advection (Hendricks et al., 2000; Noone, 2008), and isotope-enabled general circulation models (GCMs) (e.g., Yoshimura et al., 2008; Dee et al., 2015; Jouzel et al., 1987)(e.g., Jouzel et al., 1987; Yoshimura et al., 2008; Dee et al., 2015).

Variation in condensation temperatures and sub-cloud humidity has been shown to explain substantial variation in the measured isotopic ratios of precipitation (Aemisegger et al., 2015; Stewart, 1975) over short timescales. Until recently, few isotope models have considered meteorological conditions at the vapor source, in part because the evaporation site could not be unambiguously identified. Not knowing the vapor source prevents a comprehensive examination of the full vapor history. Recently developed Lagrangian air parcel tracking programs with quantitative source and trajectory meteorology have enabled estimation of evaporation sites and thus have become a useful tool for interpreting precipitation isotope ratios (Ichiyanagi and Yamanaka, 2005; Strong et al., 2007; Treble et al., 2005; Sodemann et al., 2008a; Good et al., 2014; Wang et al., 201

The objective of this study is to understand how source and trajectory meteorology contribute contributes to event-scale variations in the precipitation isotopic ratios and how such contributions vary over time (e.g., seasonally). To do this, we investigate the isotopic ratios of precipitation from event-scale sampling at Barrow, Alaska, USA. Barrow is one of nine sites that comprise the pan-Arctic Isotopic Investigation of Sea Ice and Precipitation in the Arctic Climate System campaign (iisPACS, (Feng, 2011)) . This work utilizes The work presented here uses intensive observations at Barrow under the Atmospheric Radiation

Monitoring (ARM) program. Specifically, we use millimeter wavelength cloud radar (MMCR) to identify the ~~precipitating clouds'~~ altitude and rate of condensation in the precipitating clouds in order to initialize Lagrangian air parcel tracking. Using direct cloud observations means that the backward trajectories are initiated at the appropriate time and from a distribution of altitudes representative of the actual heights of condensation. Such an initial distribution of air parcels is unique to our study. We distribute air parcels in proportion to the condensation rate, so that for a given event, each air parcel represents an equal ~~amount of the~~ fraction of precipitated water (Putman, 2013). This simplifies calculating the average vapor source, transport, and condensation conditions, which we use to interpret the observed precipitation isotope ratios. Although this research focuses on precipitation data from a single location, the results ~~link circulation to the~~ may indicate a link between atmospheric circulation and precipitation isotope systematics ~~of~~ across the sea-ice-sensitive high latitudes.

2 Methods

Event-scale precipitation samples were collected from 70 precipitation events at Barrow, AK, between January 2009 and April 2013. Below we describe methods for sample collection and measurement of ~~δ^2 H and δ^{18} O~~ δ^2H and $\delta^{18}O$ of precipitation, identification of vapor source regions, and characterization of evaporation and transport conditions using meteorological data from the source regions.

2.1 Sample collection and isotopic analysis

The sampling equipment was installed on a skydeck within the North Slope of Alaska facility of the Atmospheric Radiation Measurement (ARM) program. If the precipitation was rain, a rain funnel was used to collect the sample. If the precipitation was snow, the fresh snow was scooped into a plastic bag from a designated surface on the skydeck. The collection surface was five meters above the ground on the tower, ensuring minimal contribution of windblown snow from previous events. Samples were gathered less than 24 hours after the event ended ~~.~~ and often as soon as snow ended. Though it is possible that snow may have been altered by sublimation before collection, we assume that the degree of alteration of surface snow was minimal relative to the amount of snow gathered. Furthermore, the frequent cloudiness and darkness of Barrow mean that for most events, sunlight-driven sublimation was insignificant. Liquid samples were stored in tightly sealed 30 mL Nalgene bottles below 5 °C and shipped in batches every three months to the Stable Isotope Laboratory at Dartmouth College. When not in transit, samples were refrigerated. Samples were analyzed within six months of collection.

Upon arrival at Dartmouth the samples were prepared for analysis of hydrogen and oxygen isotopic ratios with a Delta Plus XL Isotope Ratio Mass Spectrometer (IRMS). For hydrogen measurements, the IRMS was connected to an HDevice reduction furnace: a reactor tube filled with a volumetric 1:1 mix of 100 mesh and 300 mesh chromium powder and set at 850 °C. One μL of sample was injected into the HDevice, and the water was allowed to react for two minutes in the hot chromium chamber, reducing to hydrogen gas, which was then introduced to the dual inlet system of the mass spectrometer and measured by the IRMS. For oxygen isotope measurements, the IRMS was coupled to a GasBench. A 500 μL aliquot of liquid sample was placed in a vial, flushed with a mixture of 0.3 % CO_2 in helium, and allowed to equilibrate for at least 18 hours at 25 °C. The

isotopic ratios of the CO_2 were measured by the IRMS. For both the oxygen and hydrogen measurements, the measured value was converted to the water-isotope equivalent by calibration with known standards. Isotopic ratios ($^2H/^1H$ and $^{18}O/^{16}O$), are reported in delta notation: the per mil (‰) deviation from the international standard VSMOW on the VSMOW-SLAP scale, defined as $\delta = [\frac{R_{SA} - R_{ST}}{R_{ST}}]$, where ~~$R_{SA\ or\ ST} = \frac{^2[H]}{^1[H]}$ or $\frac{[^{18}O]}{[^{16}O]}$~~ $R_{SA\ or\ ST} = \frac{^2[H]}{^1[H]}$ or $\frac{[^{18}O]}{[^{16}O]}$. SA and ST indicate sample and standard, respectively. The uncertainties of the reported values are within $\pm$ 0.5 and $\pm$ 0.1 ‰ (one standard error) for ~~δ^2 H and δ^{18} O~~ $\delta^2 H$ and $\delta^{18} O$, respectively.

2.2 Back trajectories

Back trajectories were performed ~~using~~ with the air parcel tracking program HYSPLIT ~~(Draxler, 1999; Draxler and Hess, 1997, 1998; Stein~~ 1° resolution meteorological data from the Global Data Assimilation System (GDAS). To obtain a representative ~~view~~ sampling of the vapor source region, the condensing air above Barrow, AK was subdivided into 1000 air parcels, each representing an equal amount of ~~the~~ condensing water. ~~The~~ We refer to the height of each air parcel ~~will be referred to~~ as the 'air parcel arrival height'. Each of the 1000 air parcels was tracked backward in time for 10 days (240 hours). ~~A vapor source was declared the first time that the~~ The vapor source location was defined as the place where the back trajectory of the air parcel sank ~~below the free troposphere~~ into the planetary boundary layer (PBL). ~~Only~~ Relative to previous studies that tracked vapor change in an air parcel along the trajectory (e.g., Sodemann et al. (2008a)), we adopted a simpler procedure that assumes vapor in the air parcel is well represented by the air at the latest interaction with the PBL. This assumption is justified because mass movement in the PBL is dominated by vertical turbulence relative to horizontal advection. Figure 1 shows endpoints of all trajectories that sank into the PBL~~over the ocean were considered~~. However, only trajectories that ended over water with < 96% sea ice cover were used for calculations; parcels that sank where there was less than 96% sea ice cover were used for calculations. Parcels that never sank into the PBL or those that sank ~~below~~ into the PBL over land or ice-covered ocean were ignored. ~~These were~~ Ocean-originating air parcels comprised about 71% of all trajectories~~initiated~~.

Back trajectory analysis was performed for ~~selected dates and times that most closely matched the occurrence of the precipitation events. The dates and times were chosen~~ dates when precipitation was collected. The starting times for the back trajectories corresponded to times of maximum precipitation intensity, based on a combination of sampling records, surface analysis maps of Alaska available through the National Center for Environmental Prediction, and the returns of the millimeter wavelength cloud radar (MMCR) (Johnson and Jensen, 1996; Bharadwaj et al., 2011). ~~On the date of precipitation, back trajectory start times were selected for maximum precipitation intensity as indicated by the MMCR~~ Greater Doppler vertical velocities, reflectivities, and spectral widths from the MMCR broadly indicated more intense precipitation. Because the gridded meteorological files used for tracing the back trajectories had three-hour resolution, the ~~starting time chosen~~ chosen starting time represented average conditions over a three-hour period. If precipitation lasted for more than three hours, the most intense three hour time window was selected. If the precipitation was of approximately uniform intensity, the most temporally homogeneous three-hour time window was selected, with preference for ~~the middle of the event~~time windows where precipitation occurred over the duration of the three hours.

The method for selecting the altitudes where the air parcels began their back trajectories is described in full in Putman (2013). Briefly, returns of the reflectivity and Doppler vertical velocity ~~(Holdridge et al., 1994; Regional Climate Center, 2012; Johnson and Jensen~~ the MMCR were processed with algorithms developed by Zhao and Garrett (2008) to estimate the precipitation rate profile (g m^{-2} s^{-1}) as a function of height. The precipitation rate profile was differentiated with respect to height, yielding the condensation rate profile (g m^{-3} s^{-1}) and then subdivided into the aforementioned 1000 air parcels so as to ensure that each parcel contained an equal fraction of total precipitation.

At both the ~~vapor source region and the~~ air parcel initiation altitude above the precipitation site and the vapor source region, the meteorological data for our analysis came from the Global Data Assimilation System ~~(GDAS)~~ reanalysis gridded dataset. At the condensation site, we extracted from GDAS the air temperature at each height containing an air parcel. At the vapor source, we extracted the 2 m relative humidity and 2 m air temperature. Sea surface temperature data for the deuterium excess analysis came from the NOAA gridded sea surface temperature dataset (NOAA/OAR/ESRL PSD at Boulder Colorado USA, 2013). ~~At the condensation site, we extracted from GDAS the air temperature at each height containing an air parcel.~~

2.3 Calculation of $\Delta\bar{T}_{cool}$, $\bar{T}_d$, ~~$\Delta\bar{T}_{cool}$~~ and mtn

To quantify the relationship between the vapor source region and the isotopic composition of precipitation, we used three physically based metrics: the average amount of cooling during air parcel transport $\Delta\bar{T}_{cool}$, the average dew point ~~(evaporation conditions)~~ at the vapor source region $\bar{T}_d$, which characterizes planetary boundary layer conditions, and the presence or absence of mountains along the transport path, described by the Boolean variable mtn. The first two metrics were calculated from the meteorological data at the vapor source and precipitation site. The third ~~, a Boolean variable,~~ was assigned based on the air parcel trajectory.

~~The portion of the total~~

2.3.1 ΔT_{cool}

An estimate of air parcel cooling that produced condensation, ΔT_{cool}, ~~represents~~ is a bulk metric quantifying the magnitude of Rayleigh distillation along the trajectory (Sodemann et al., 2008a). ~~For a given~~ This approach simplifies the integration of cycles of warming and cooling that may occur along a trajectory to a net reduction in temperature. For each air parcel, we calculate ΔT_{cool} as the difference between the temperature at the air parcel lifted condensation level (LCL) above the source region T_{LCL}, and the condensation temperature, T_c, at the air parcel arrival height extracted from reanalysis above Barrow, AK, i.e.,

$$\Delta T_{cool} = T_{LCL} - T_c \tag{1}$$

~~The temperature at the LCL, T_{LCL}, was determined by finding the altitude where the specific humidity Q, of the surface air parcel equaled the~~ To determine $Q_{sat,z}$, we start from the dry adiabatic lapse rate, (-9.8 °C km^{-1}). From this we determine the temperature T_z at altitude z, starting with the 2 meter temperature T_{2m}.

The saturation vapor pressure at elevation z, $e_{sat,z}$ is then

$$e_{sat,z} = 0.6113e^{[5423(\frac{1}{T_0} - \frac{1}{T_z})]} \quad (2)$$

where T_0 = 273.15 K (Stull, 2015). We may then write the saturation specific humidity ~~Q_{sat}. To calculate~~, $Q_{sat,z}$, as

$$Q_z = 0.622\frac{e_{sat,z}h_z}{P_z} \quad (3)$$

where h_z, the relative humidity at height z, is assumed to equal 1 (air is vapor-saturated) and the pressure at height z, P_z, is

$$P_z = 1013.25[1 - (2.25577 * 10^{-5})z]^{5.25588} \quad (4)$$

Calculating the 2 m specific humidity, Q_{2m}, ~~the~~ is simply a special case of the general calculation: we use the 2 m temperature T_{2m}, fractional relative humidity h_{2m}, and pressure P_{2m} ~~, were combined as in Equation 3. The same was done for an array of elevations z, where h_z was assumed to be 1~~from reanalysis in Equations 2 and 3, ~~T_z was calculated from~~ rather than using the dry adiabatic lapse rate~~(), and P_z was calculated as in~~, h = 1, and Equation 4. ~~Thus T_{LCL} was T_z where $Q_{sat,z}$ equaled~~, respectively.

Finally, we find the elevation where Q_{2m} equals $Q_{sat,z}$. The temperature at this elevation is T_{LCL}.

$$Q = \frac{3270h}{P}2^{\frac{T}{10}}$$

$$P(z) = 1013.25[1 - (2.25577 * 10^{-5})z]^{5.25588}$$

ΔT_{cool} was calculated individually for each of the 1000 ~~trajectories~~ air parcels in an event. We report the mean of all ~~trajectories~~ air parcels that were traced to the marine PBL, $\Delta\bar{T}_{cool}$, as characteristic of the event.

2.3.2 T_d

We used the vapor source 2 m dew point T_d, to represent the ~~evaporation conditions at the~~ conditions of the PBL in the vapor source region~~. T_d depends on the specific humidity of saturated air at the sea surface and on the amount of dry air from aloft that has subsided and mixed into low altitude air. The~~, because the relative proportions of the ~~saturated~~ moist surface air and ~~unsaturated~~ dry subsiding air determine the ~~properties~~ T_d of the marine PBL. ~~This makes~~ The choice of T_d ~~a useful indicator of integrated evaporation conditions at the vapor source.~~ rather than sea surface temperature T_{ss} and relative humidity h_{2m} reflects

our conviction that T_d provides a better representation of conditions within the PBL, from which vapor with its characteristic $\delta^2 H$ will start its' trajectory to the precipitation site (see section 3.2) We approximate T_d ~~with~~ using

$$T_d = [\frac{1}{T_0} - 1.844 * 10^{-4} ln(\frac{e_{sat,2m} h_{2m}}{0.6113}]^{-1} \quad (5)$$

(Stull, 2015) with saturation vapor pressure, $e_{sat,2m}$, from Equation 2 and the 2 m air temperature, T_{2m}, and relative humidity ~~,~~h_{2m} ~~, using the following (?)~~ from reanalysis data.

~~$T_d = T_{2m} + 10 * log_2(h_{2m})$~~

T_d was calculated for the vapor source indicated by each of the ~~1000 trajectories that were~~ trajectories that was traced to the marine PBL~~, and ;~~ the mean of ~~the 1000~~ these T_d values is reported as a single value, $\bar{T}_d$, characteristic of the event.

2.3.3 *mtn*

Vapor originating in the Gulf of Alaska typically must be transported over the Alaska and Brooks Ranges to contribute to precipitation at Barrow, whereas vapor originating anywhere in the Arctic Ocean, Bering Strait, or western North Pacific typically does not encounter major orographic obstacles during its transport to Barrow. The orographic effect on isotope ratios of precipitation was quantified with a Boolean variable, 'mtn'~~:~~, defined as whether (1) or not (0) most air parcels crossed the Alaskan and/or Brooks ranges during transport to Barrow. The value of mtn was assigned manually based on the general pattern of transport ~~for the event, not to individual trajectories~~observed in the trajectory plots.

3 Results and discussion

In this section we discuss the vapor source annual cycle ~~,~~and statistical relationships between the isotopic composition of precipitation, vapor source region, and the variables ~~(~~$\Delta\bar{T}_{cool}$, $\bar{T}_d$, and mtn~~)~~ , that characterize the relationship of vapor source and transport to the isotope values measured at Barrow, AK.

3.1 Vapor source region annual cycle

The vapor source regions for precipitation at Barrow ~~change~~ changed seasonally (Figure 1). Vapor fueling winter (December, January, February) precipitation originated furthest south, typically in the Gulf of Alaska, and for most winter events, trajectories crossed the Alaskan and Brooks Ranges. In spring (March, April, May) the vapor for roughly half the precipitation events came from the North Pacific and traveled over the mountain ranges, as in winter. The vapor for the remaining precipitation events generally came from the southwest of Barrow, from the Bering Strait and Chukchi Sea. Vapor source regions for summer (June, July and August) precipitation were the most northerly, typically the Chukchi Sea or Bering Strait. Synoptic systems moving counterclockwise around the Arctic Ocean characterized summer air parcel transport. In fall (September, October,

November), vapor also came from the Chukchi and Beaufort ~~seas~~Seas, but with air parcel transport from the east to Barrow, the reverse of the spring and summer parcel transport patterns. The Gulf of Alaska provided vapor for a few fall events, with air parcel transport over the Brooks and/or Alaskan mountain ranges, as in winter.

In association with the latitudinal variation in the vapor source region, the temperature difference along the trajectory $\Delta\bar{T}_{cool}$ and vapor source dew point $\bar{T}_d$ also varied (Figure 2). The mean latitude of the vapor source region, $\bar{V}_{Lat}$, and $\Delta\bar{T}_{cool}$ varied inversely, with more cooling being associated with lower $\bar{V}_{Lat}$, i.e. greater meridional transport. For any given season $\bar{T}_d$ was warmer in the south, and cooler in the north. There are also seasonal differences; at any latitude ~~the dew point, $\bar{T}_d$,~~ was warmer in summer and cooler in winter.

The migration of the mean latitude of the vapor source region ~~is tied in two ways to the change in~~ can be tied to the seasonal cycling of solar insolation in the northern hemisphere via two mechanisms. Decreased solar insolation during winter drives expansion of the northern Polar circulation cell~~and~~ , which increases sea ice cover~~. Increased~~ , and cold temperatures and snow cover prevent evapotranspiration. Both sea ice cover which diminishes the vapor contributions of the Arctic Ocean, ~~allowing~~ and inhibited evapotranspiration allow for enhanced representation of southerly vapor sources. Increased summer insolation drives poleward contraction of the circulation cell~~and~~ , diminishes sea ice coverage, and warmer temperatures favors evapotranspiration such that the average vapor source area migrates north. Feng et al. (2009) documented similar vapor source migration over a much larger scale, in association with the annual north-south migration of circulation cells.

~~The latitudinal~~ There is evidence for prior millenium-scale shifts in the southern extent of the ~~Polar circulation cell likely varies over time scales ranging from interannual to millenial, causing changes to vapor sources and the Arctic hydrologic cycle on corresponding time scales.~~ polar circulation cell (Feng et al., 2007). Aspects of the link between seasonal variability in general circulation and seasonal vapor source cycling may be generalizable to interannual and even millennial timescales. This is relevant to modern changes in the hydrologic cycle as Marvel and Bonfils (2013) suggest that a poleward displacement of circulation cells is already occurring due to recent climate change. ~~Southward migration of the polar cell during the last glacial maximum has been documented by Feng et al. (2007). Changes~~ Additionally, changes in the isotopic composition of precipitation resulting from systematic vapor source migrations associated with changing climate may allow for interpretation of long-term isotopic records in terms of changes in atmospheric circulation, including but not limited to the precipitation site temperature.

3.2 The influence of vapor source on precipitation ~~δ^2 H~~$\delta^2 H$

The local meteoric water line (with 95 % confidence intervals) is $\delta^2 H = 7.78(\pm 0.12)\delta^{18}O + 7.18(\pm 2.61)$. Figure 3 shows that the measured ~~δ^2 H~~ $\delta^2 H$ values of the 70 precipitation events fall between -280 ‰ and -50 ‰, with a pattern of summer enrichment and winter depletion that follows the well-established annual cycle for mid- and high latitudes ~~(Feng et al., 2009). The local meteoric water line (with confidence intervals), $\delta^2 H = 7.78(\pm 0.12)\delta^{18}O + 7.18(\pm 2.61)$, is statistically distinguishable from the global meteoric water line ($\delta^2 H = 8\delta^{18}O + 10$).~~ (Feng et al., 2009; Bonne et al., 2014). Figure 3 also shows the interannual, seasonal and event-scale variability captured by the dataset where the spline captures 65% of the annual and interannual variance. The average annual cycle of the precipitation ~~δ^2 H is in phase with the~~ $\delta^2 H$ is strong; the spline fit explains 60% of

variance in the data. The mean latitude of the vapor source ~~,~~ exhibits a weak seasonal pattern, where the spline explains 19% of the variance. The seasonal cycles of $\delta^2 H$ and vapor source latitude are in phase, as shown in Figure 4, though the inter-event variability in both variables can be as large as the seasonal variability.

The phase relationship between ~~δ^2 H~~ $\delta^2 H$ and the north-south migration of the vapor source region occurs because the vapor source region governs three critical metrics that affect the ~~δ^2 H~~ $\delta^2 H$ of precipitation: 1) the temperature difference between vapor source region and precipitation site, quantified by air parcel cooling $\Delta\bar{T}_{cool}$ ~~(Figure 2)~~, 2) the ~~evaporation~~ moisture source conditions, quantified in this work by $\bar{T}_d$ ~~(Figure 2)~~, and 3) the mean air parcel transport path ~~(Figure 5)~~. A linear combination of $\Delta\bar{T}_{cool}$, $\bar{T}_d$, and mtn statistically represents the event-scale variation in ~~δ^2 H~~ $\delta^2 H$ with an R^2 value of ~~0.52~~ 0.54 ($p < 0.001$). Table 1 contains the ~~correlation~~ partial regression slopes (β), p-values, and the unique variance explained by each variable. Below we discuss the physical mechanisms that may explain the influence of each of these metrics on ~~δ^2 H~~ $\delta^2 H$.

In contrast with previous assumptions that local (precipitation site) surface temperature alone is a metric for Rayleigh distillation (e.g., Dansgaard, 1964), our study ~~compares δ^2 H with~~ relates $\delta^2 H$ to $\Delta\bar{T}_{cool}$, $\bar{T}_d$ and mtn. Using these metrics instead of local surface temperature allows us to circumvent two restrictive assumptions. First, we do not assume that ~~δ^2 H~~ $\delta^2 H$ has a spatially and temporally stationary relationship to local temperature. Bowen (2008) demonstrated that this assumption does not hold. Rather, because meridional temperature gradients are an important driver of the isotope-temperature sensitivity (Hendricks et al., 2000), when the meridional temperature gradient fluctuates, a quantity that $\Delta\bar{T}_{cool}$ captures, the sensitivity of ~~δ^2 H~~ $\delta^2 H$ to local temperature also fluctuates. ~~Likewise~~ As demonstrated by Figure 5, the presence of mountains along the vapor transport path will deplete the isotope ratio of the precipitation relative to ~~an over-ocean~~ a uniform altitude transport, all other meteorological conditions being equivalent. The second restriction associated with using local surface temperature as a metric of Rayleigh distillation ~~assumes~~ is the assumption that vapor for all precipitation events comes from a single, homogeneous source. It requires that the ~~δ^2 H~~ $\delta^2 H$ of the water vapor, and thus the initial condensate, is constant in space and time. However, global measurements from the Tropospheric Emissions Spectrometer (Good et al., 2015) indicate that the vapor in the planetary boundary layer over the ocean varies with space and season, confirming previous land and ship measurements ~~(e.g., Steen-Larsen et al., 2014; Kurita, 2011; Uemura et al., 2008)~~(e.g., Uemura et al., 2008; Kurita, 2011; Steen-Larsen et al., 201
Likewise, our results indicate that vapor may come from a heterogeneous source region or variety of source regions (Figure 1) and the initial condensate, based on the evaporation conditions, should be expected to vary. The effect of a meteorologically heterogeneous source region(s) is captured by $\bar{T}_d$.

As expected, $\Delta\bar{T}_{cool}$ accounts for the largest proportion of variance in ~~δ^2 H (23~~ $\delta^2 H$ (28.7%) among the explanatory variables. Our ~~statistical correlation~~ multiple regression yields a sensitivity of ~~for δ^2 H~~ -3.51‰ °C^{-1} for $\delta^2 H$ with respect to $\Delta\bar{T}_{cool}$ (Table 1). Because Rayleigh distillation is considered the main source of spatial variation in ~~δ^2 H~~ $\delta^2 H$, comparison with the sensitivities calculated from a simple Rayleigh model contextualize our result. In such a model, a saturated air parcel with specified temperature and vapor $\delta^2 H$ is cooled iteratively in 1°C steps. At each temperature step, the condensation amount, remaining vapor, precipitation $\delta^2 H$ and vapor $\delta^2 H$ are calculated. No re-evaporation or non-equilibrium conditions are considered. We determined condensation in ~~a moist~~ this air parcel for both adiabatic decompression and isobaric radiative cooling using equilibrium isotope fractionation factors from Majoube (1971). Because the association between precipitation

~~δ^2 H~~ δ^2H and $\Delta\bar{T}_{cool}$ during a Rayleigh process varies (Dansgaard, 1964), the sensitivity range for moist adiabatic cooling from 10 °C to -15 °C, with a lapse rate of -6.5 °C km^{-1}, ranges between -3.46 ‰ °C^{-1} to -5.45 ‰ °C^{-1}, while moist isobaric radiative cooling across the same temperature range yields sensitivities from -5.47 ‰ °C^{-1} to -7.88 ‰ °C^{-1}. The sensitivity exhibited by our data is just ~~below~~ above the low end of the range determined for moist adiabatic cooling and was ~~significantly~~ substantially lower than the range using isobaric cooling. The similarity between our data and the moist adiabatic model results suggest that moist adiabatic cooling was likely the dominant mechanism for precipitation during air parcel transport to Barrow, although scatter in the ~~δ^2 H~~ δ^2H data could also be due to variable contributions of radiative cooling. The relatively low observed sensitivity relative to ~~the~~ both theoretical sensitivities may be explained by additions of vapor to air parcels during poleward meridional transport, ~~following~~ which were not considered by our back trajectory scheme, but are supported by the two-stream isentropic vapor source transport model (Noone, 2008).

Our multiple linear ~~model~~ regression attributes a substantial fraction of the variance in ~~δ^2 H~~ δ^2H to variations in $\bar{T}_d$ (~~17%~~)10.5%, Table 1), which is used to represent the source conditions. We prefer T_d to the classical variables T_{ss} and h for determining isotopic evaporative fluxes. This choice is based on our understanding that the meteorological variable T_d characterizes the bulk vapor content and isotopic ratio of the marine PBL, independent of the vapor temperature. When advected to the free troposphere, it is this vapor that will form precipitation. Additionally, through equilibrium fractionation T_d also determines the isotopic ratio of the first condensate at the LCL, where Rayleigh distillation begins.

Within the marine PBL, several inter-related factors/processes are at work to determine the starting point of a Rayleigh trajectory. The first is the isotopic flux of evaporation from the sea surface. Most studies estimate this flux using the classic model by Craig and Gordon (1965). In that model, three variables control the evaporative flux: the sea surface temperature, T_{ss} , δ^2H above the laminar layer, and the humidity h_{ss} above the laminar layer (e.g., at 2 m), defined relative to T_{ss} . Though h_{ss} is not a measured quantity nor one that is normally modeled, it is determined by T_d above the laminar layer and T_{ss} . Hence isotopic fluxes can be determined with the classical model using T_{ss} , and T_d and δ^2H above the laminar layer as input variables. From a physical point of view, T_{ss} determines the amount of equilibrium fractionation at the water-air interface. T_d and vapor δ^2H , as well as T_{ss} , control kinetic fractionation as vapor diffuses across the laminar layer. It should be noted that when T_{ss} is large, T_d tends to be large as well, as a result of their change with latitude and season. T_d and δ^2H are also correlated, which will be discussed below. ~~This is because source meteorological conditions control the δ^2 H of the water evaporated from the ocean surface. Studies typically attribute variations in δ^2 H values at the source to mean sea surfacetemperature $\bar{T}_{ss}$ and mean~~ Therefore, all three variables controlling the evaporative flux, T_{ss} , and h_{ss} and δ^2H above the laminar layer, are associated directly or indirectly with T_d , making T_d a good indicator of evaporation conditions.

The second process is convergence. At a moisture source location, low level air is moist due to evaporation near the sea surface. Convergence and uplift transports low-level moist air into the free troposphere where it mixes with dry, isotopically depleted air descending from surrounding regions resulting in strong humidity and temperature gradients near the sea surface (below 2 m~~relative humidity $\bar{h}$, which affect the magnitudes of equilibrium and kinetic fractionation, respectively (Craig and Gordon, 1965). Assuming that $\bar{T}_{ss}$ influences~~). In contrast, the specific humidity and isotopic ratios in the bulk of the PBL above 2 m ~~air temperature, so that the two temperatures correlate spatially, we may use the~~ are relatively constant, resulting from the relative

contributions of vertical transport of moist low-level and descending air (Fan, 2016). T_d and $\delta^2 H$ at 2 m ~~dew point ($\bar{T}_d$) at the vapor source to combine the effects of $\bar{T}_{ss}$ and $\bar{h}_{2m}$. Either high $\bar{T}_{ss}$ or high $\bar{h}_{2m}$ results in high $\bar{T}_d$. Therefore, we expect $\bar{T}_d$ to be positively associated with δ^2 H in the original vapor in an airparcel at the vapor source. This δ^2 H signal at the source is then carried to the precipitation site. Differences in $\bar{T}_{2m}$ cause > of the range we report for $\bar{T}_d$, whereas $\bar{h}_{2m}$ contributes. The substantial difference between the vapor source $\bar{T}_d$ for Arctic compared with subtropical sources makes $\bar{T}_d$ a more useful metric for characterizing the vapor source than either $\bar{T}_{ss}$ or $\bar{h}_{2m}$ alone.~~ both reflect the outcome of this mixing process, and so it follows that they are positively correlated.

~~Given the magnitude of variability in $\bar{T}_d$ attributed to temperature,~~ The third process is condensation at the LCL. The temperature of the air mass, which equals or is very slightly less than the local dew point, determines the amount of isotopic fractionation and thus the isotopic ratio of the first condensate. It is this isotopic composition that defines the beginning of the Rayleigh part of the trajectory. Only $T_{d,2m}$, not T_{ss} nor h_{2m}, is directly associated with the condensation temperature at the LCL (which differs only slightly from $T_{d,2m}$ due to the pressure difference between 2 m and the LCL and its effect on saturation specific humidity).

Since all three processes before Rayleigh distillation are either directly or indirectly related to T_d, we consider T_d a better indicator for the ~~relationship of δ^2 H to $\bar{T}_d$ should be comparable to the equilibrium fractionation relationship between δ^2 H and $\bar{T}_{ss}$. For δ^2 H relative to $\bar{T}_d$ we report a sensitivity of (Table 1). In comparison, for $\bar{T}_{ss}$ between~~ source conditions than either T_{ss} or h. It is difficult, however, to theoretically assess the sensitivity of precipitation $\delta^2 H$ to variations in source T_d, because this would require quantification of the theoretical relationship of T_d to $\delta^2 H$ through each of the three processes and perhaps their combinations. We here report the first empirical sensitivity of 3.23‰ °C^{-1} (Table 1) for $\delta^2 H$ relative to T_d. At the sea surface, for T_{ss} between 0 and 25 °C, equilibrium fractionation as a function of temperature yields sensitivities between 1.1-1.6‰°C^{-1} (Majoube, 1971). ~~The sensitivities have the same sign, indicating that warm evaporation temperatures correlate with high vapor isotopic ratios. The sensitivity of δ^2 H to $\bar{T}_d$ is 2 to 4 times that of δ^2 H to $\bar{T}_{ss}$, meaning that equilibrium fractionation accounts for less than half of the sensitivity. Kinetic fractionation associated with variation in $\bar{h}_{2m}$ (e.g., entrainment of upper atmosphere air) may contribute some of the remaining sensitivity. Other factors that influence evaporation, such as wind speed and sea surface roughness (Merlivat and Jouzel, 1979) may have covaried spatially with $\bar{T}_d$ and influenced the reported sensitivity. Partitioning the effects of windspeed and surface roughness requires further investigation.~~ However, a large part of this fractionation may be offset by condensation at the LCL. Consequently, the observed sensitivity probably reflects primarily the fraction of vapor contributed by dry, isotopically depleted descending air that converges within the PBL. Mixing with the dry air causes a decrease in T_d, which affects the $\delta^2 H$ of the PBL in two ways: 1) making the PBL air dry and isotopically depleted, and 2) isotopically depleting the evaporative flux by enhancing kinetic fractionation (an effect of low relative humidity). Both mechanisms produce a positive association between $\delta^2 H$ and T_d, consistent with the sign of our observed partial coefficient (Table 1).

Upon leaving the vapor source region, the isotopic composition of vapor depends on the trajectory taken. To reach Barrow, AK, air parcels originating in the Gulf of Alaska must cross the Alaska and/or Brooks Ranges, whereas air parcels from the Bering Strait or Chukchi Sea do not have to cross high topography. Our work shows that transport across mountain ranges

resulted in significant ~~δ^2 H~~ $\delta^2 H$ depletion in Barrow precipitation. Transport of vapor over mountain ranges occurred more frequently during cold months, when the Gulf of Alaska and North Pacific were the dominant vapor source regions. Since the vapor source location in winter is governed by the expansion of the Polar circulation cell, the projected northward displacement of subtropical highs and the Polar front (Marvel and Bonfils, 2013) in a warming climate may be associated with ~~a reduced amounts of~~ less vapor transported over the Alaskan and/or Brooks ranges during fall, winter and spring. Fewer events traveling over the Alaskan and/or Brooks ranges would correspond to a pronounced enrichment in measured ~~δ^2 H~~ $\delta^2 H$ at Barrow during cold months.

To study the importance of $\bar{T}_d$ and mtn as explanatory variables with respect to cooling during transport ($\Delta\bar{T}_{cool}$), we divided our data into subgroups~~:~~, those with $\Delta\bar{T}_{cool}$ <7 °C ~~and~~ (corresponding to short trajectories), and those with $\Delta\bar{T}_{cool}$ >7 °C (corresponding to long trajectories) and recalculated the statistics. Table 2 summarizes the results and Figure 6 shows the standard deviation of $\bar{T}_d$ by category. The breakpoint of 7 °C was chosen by testing different breakpoints and finding one that maximized the statistical power of the short trajectory regression while preserving the strong relationship between $\delta^2 H$ and $\bar{T}_d$. For the small $\Delta\bar{T}_{cool}$ subgroup, $\bar{T}_d$ explains almost half the variance in ~~δ^2 H~~ $\delta^2 H$ (R^2 = ~~0.48~~0.43), whereas for the large $\Delta\bar{T}_{cool}$ subgroup, $\bar{T}_d$ explains very little variance (R^2 = ~~0.10~~0.007). This difference implies enhanced isotopic modification over long trajectories. In contrast, the ~~δ^2 H~~ $\delta^2 H$ values of the small $\Delta\bar{T}_{cool}$ subgroup are not well explained by the Boolean variable mtn (R^2 = ~~0.05~~0.03), whereas mtn explains ~~a quarter~~ about one-fifth of the variability of the large $\Delta\bar{T}_{cool}$ subgroup (R^2 = ~~0.24). As expected, some variability in each subgroup is explained by~~ 0.18). For the small $\Delta\bar{T}_{cool}$ subgroup, $\Delta\bar{T}_{cool}$ ~~. The small R^2 = 0.02. For the large~~ $\Delta\bar{T}_{cool}$ subgroup, $\Delta\bar{T}_{cool}$ explained a quarter (R^2 = ~~0.26~~0.22) of the variance in ~~δ^2 H, and for the large $\Delta\bar{T}_{cool}$ subgroup R^2 = 0.13~~ $\delta^2 H$.

Because the events with smallest $\Delta\bar{T}_{cool}$ tended to occur in summer, the strong relationship between $\bar{T}_d$ and ~~δ^2 H~~ $\delta^2 H$ indicates that precipitation ~~δ^2 H~~ $\delta^2 H$ in summer predominantly reflects variability in ~~evaporation~~ source conditions. The strong relationship between mtn and the variation in ~~δ^2 H~~ $\delta^2 H$ for large $\Delta\bar{T}_{cool}$ indicates that precipitation ~~δ^2 H~~ $\delta^2 H$ in winter predominantly reflects whether most air parcels crossed the Alaska and/or Brooks mountain ranges. Notably, $\Delta\bar{T}_{cool}$ could significantly predict ~~δ^2 H for both long and short~~ $\delta^2 H$ for long trajectory events, ~~but~~ and it explained less variance than expected, given the emphasis on Rayleigh distillation in ~~isotope hydrology~~explaining spatial variation in precipiation stable isotopes.

Among the simple regressions, almost half the variance in ~~δ^2 H~~ $\delta^2 H$ for events with $\Delta\bar{T}_{cool}$ <7 °C was explained by $\bar{T}_d$. This is a notable result, as the isotope composition of the initial vapor is not emphasized to the same degree as Rayleigh distillation in isotope hydrology. There are two reasons why $\bar{T}_d$ may explain so much variance for short trajectory events. First, storm events with minimal cooling during air parcel transport typically originated close to Barrow in the Arctic Ocean. A smaller vapor source area predicts less variation in T_d among air parcels: a more homogeneous source. We quantify this effect by examining the distribution of ~~within-storm~~ intra-event $\bar{T}_d$ standard deviations ($\sigma\bar{T}_d$) for the short and long trajectory event subsets (Figure 6). Short trajectory ($\Delta\bar{T}_{cool}$ <7 °C) events had a median $\sigma\bar{T}_d$ of 2.79 °C, which was ~~smaller~~ less than the long trajectory ($\Delta\bar{T}_{cool}$ >7 °C) median $\sigma\bar{T}_d$ of 4.68 °C. Less variability among air parcels in the short trajectory subset allowed the among-event relationship of ~~δ^2 H~~ $\delta^2 H$ to $\bar{T}_d$ to emerge. ~~Second~~In addition, some of the variability in measured precipitation ~~δ^2 H~~ $\delta^2 H$ may be caused by processes occurring during transport, such as radiative cooling, air mass mixing, and different

degrees of mountain-induced rainout. The opportunity for these effects to impact the precipitation isotope value increases with increasing transport distance, obscuring the relationship of the precipitation ~~δ^2 H to the δ^2 H~~ $\delta^2 H$ to the $\delta^2 H$ of the initial vapor at the source and therefore to $\bar{T}_d$.

3.3 ~~The influence of vapor source on deuterium excess~~

The three chosen variables explain just over half (54%) the variance of $\delta^2 H$. This is not surprising, considering that many other mechanisms can also influence the $\delta^2 H$ of the vapor and precipitation. These mechanisms include (but are not limited to) condensation temperature, supersaturation in the mixed phase cloud, sub-cloud dryness, phase of precipitation, precipitation intensity, evapotranspiration of land sources, and the amount of sea ice at the vapor source. The effects of several of these factors, including condensation temperature, sub-cloud dryness, sea ice concentration at the vapor source, and phase of precipitation (rain vs. snow), were tested as additional explanatory variables in the multiple regression, but yielded statistically insignificant results with little to no additional variance explained. Clearly, compared with the three chosen variables, the effects of these variables are relatively minor, such that the statistical power is not sufficient to reveal their significance.

3.3 The influence of vapor source on deuterium excess

Deuterium excess (d-excess, or d) of precipitation is often used to investigate ~~conditions at vapor source regions~~ source region conditions such as T_{ss} and h that affect evaporation (Dansgaard, 1964). Empirical studies have linked marine boundary layer vapor deuterium excess ($d = \delta^2 H - \delta^{18} O$) to T_{ss} and h or h_{ss} (Uemura et al., 2008; Kurita, 2011; Steen-Larsen et al., 2014). These results agree qualitatively or semi-quantitatively with theoretical predictions ~~(Merlivat and Jouzel, 1979; Craig and Gordon, 1965)~~(M However, in order for source vapor d values to be preserved in precipitation, d must be conserved through condensation and post-condensation processes. This assumption may not be realistic for several reasons. First, even simple equilibrium Rayleigh distillation does not yield constant d values in precipitation (Dansgaard, 1964). Second, non-equilibrium processes associated with snow formation may substantially alter d (Jouzel and Merlivat, 1984). Third, evaporation or sublimation under the cloud base and/or at the snow surface tends to decrease d (Stichler et al., 2001).

~~Direct~~ While studies indicate that d in vapor contains vapor source information (Steen-Larsen et al., 2014; Bonne et al., 2015; Steen-Lars direct comparison of precipitation d to vapor source conditions via Lagrangian back trajectory vapor source estimation has produced complicated results. For example, Sodemann et al. (2008b) found that while the d of precipitation contains identifiable source information, it 'does not directly translate into the source region $\bar{T}_{ss}$'. In a study of vapor sources for precipitation in Antarctica, Wang et al. (2013) noted that the classical interpretation of measured d would predict that the highest average d found at Dome Argus would correspond to the warmest (most northerly) vapor sources. However, precipitation at Dome Argus was linked to southerly (cooler) vapor sources. The authors suggested the high d value was due to the vapor pressure deficit of dry air blowing off sea ice. Likewise Good et al. (2014) attributed the significant correlation between high d and source relative humidity (h) for precipitation collected at four northeast U.S. locations during Superstorm Sandy to oceanic evaporation into a dry continental air mass that was entrained into the superstorm.

Our study reveals a ~~similarly inconclusive~~ relatively more conclusive relationship between vapor source and event-scale precipitation d~~:~~ , as summarized by four simple regressions against h_{2m} , h_{sst} , T_{ss} , and T_d shown in Table 3. Though d is not significantly predicted by $\bar{h}_{2m}$ (p = ~~0.36) but~~ 0.86) it is significantly predicted by ~~$\bar{T}_{ss}$ (p~~ $\bar{h}_{ss}$ (p < 0.001, R^2 = ~~0.016),~~ 0.34), with a slope of -0.4 ‰% $^{-1}$. This value is consistent with the -0.4 to -0.6‰ $\%^{-1}$ range reported in the literature for vapor (Uemura et al., 2008; Pfahl and Wernli; Bonne et al., 2014). $\bar{T}_{ss}$ is also a significant predictor (p = 0.0023) though the variance explained is 12 % and the sign of the coefficient is negative~~(Table 3), opposite of~~ , opposite to expectations. If d is regressed against both $\bar{T}_{ss}$ and ~~$\bar{h}_{2m}$~~ $\bar{h}_{ss}$, the multiple regression is ~~marginally~~ significant (p ~~= 0.051~~ < 0.001, not shown in Table 3) and explains ~~of variance~~36 % of variance, most of which is due to the strong relationship with h_{ss} . The vapor source region dew point, $\bar{T}_d$, significantly predicts d (p <0.001) and explains a non-trivial portion of the variance (R^2 = 0.24) ~~, although the sign of the regression coefficient is opposite of theoretical expectations.~~

~~Despite the apparent poor~~ with a negative slope (-0.53 ‰°C^{-1}). This is an interesting result with respect to the utility of T_d , a measurable quantity, and is consistent with our earlier argument that T_d is strongly related to h_{ss} . Both variables provide a better representation of source conditions than T_{ss} and/or h_{2m} . A low value of h_{ss} or T_d corresponds to a strong influence of descending dry air within the PBL, which enhances kinetic isotopic fractionation and produces a high value of d . This mechanism explains the negative correlation between d and ~~vapor source conditions, our dataset~~ T_d , and is expected for the relationship between d and h_{ss} . Alternatively, the vapor in descending air may have a high value of d (Fan, 2016), or both mechanisms may contribute to this result.

Our dataset also shows systematic seasonal variations in d . Figure 7 shows that d cycles annually, with the maximum occurring in October or November ~~, and~~ lagging the annual maximum of ~~δ^2 H~~ $\delta^2 H$ by 2-3 months (or $\sim 90°$). This phase relationship explains the lack of linear association between d and $\bar{T}_{ss}$ and $\bar{h}_{2m}$ ~~we report~~because the two latter variables are both in phase with $\delta^2 H$. Systematic seasonal variations in precipitation d occur in the Northern Hemisphere (Feng et al., 2009), particularly in the Arctic ~~(White et al., 1988; Johnsen et al., 1989; Kopec et al., 2016; Kurita, 2011)~~(White et al., 1988; Johnsen et al., 1989; Kurita, 20 These studies suggest that the conditions producing d variation have systematic annual variations in their magnitude and relative importance.

4 Conclusions

The vapor source regions identified by HYSPLIT for storms at Barrow, AK, USA ~~exhibit~~ exhibited interannual, annual, and substantial inter-event variability. ~~Vapor comes~~ On average, vapor came from the North Pacific and Gulf of Alaska, the most southerly vapor source areas, in cold months when the Polar circulation cell ~~extends~~ extended southward. Vapor ~~comes~~ came from the Bering, Chukchi and Beaufort seas, the most northerly sources, in warm months when the Polar cell ~~contracts~~ contracted northward. The cycle of winter depletion and summer enrichment exhibited by the ~~δ^2 H~~ $\delta^2 H$ of the precipitation ~~follows~~ followed the annual changes in the latitude of the vapor source region, as a result of source region controls on evaporation, transport, and condensation conditions. However, substantial intra-season variability occurred in both source and $\delta^2 H$, indicating scatter in the seasonal relationship. A linear combination of the average vapor source region dew point ($\bar{T}_d$, β

= 3.23 ‰ °C^{-1}), average cooling of the air parcels during transport ($\Delta\bar{T}_{cool}$, β = -3.51 ‰ °C^{-1}) and passage of air parcels over mountains or not (mtn, β = -32.11 ‰ when mtn = 1) ~~explains~~ explained 54 % of the event-scale variance in ~~δ^2 H~~$\delta^2 H$. For the subset of events where $\Delta\bar{T}_{cool}$ was < 7 °C (short trajectories), $\bar{T}_d$ alone explained 43 % of the variance in ~~δ^2 H~~$\delta^2 H$. For the subset of events where $\Delta\bar{T}_{cool}$ was > 7 °C (long trajectories), $\bar{T}_d$ did not significantly predict ~~δ^2 H~~$\delta^2 H$, but mtn alone explained 18 % of the variance in ~~δ^2 H~~$\delta^2 H$. Neither the average vapor source relative humidity, $\bar{h}_{2w}$, nor the average vapor source sea surface temperature, $\bar{T}_{ss}$, nor both combined, significantly explained the variations in deuterium excess~~, although a systematic seasonal variation with maximum d in October and minimum d in March was noted~~. The vapor source region ~~dew point explained~~ relative humidity with respect to sea surface temperature, $\bar{h}_{sst}$, explained 34 % of the variance in d, ~~though the sensitivity of d with respect to~~ with the expected negative sensitivity, and the source dew point, $\bar{T}_d$ ~~was negative. Additional~~ , explained a nontrivial proportion of 22%. Our results suggest that T_d is related to h_{ss} , and that both variables are more indicative of PBL conditions that directly affect vapor supplied to the free troposphere than T_{ss} or h_{2w} . Deuterium excess also exhibited a systematic seasonal variation with maximum d in October and minimum d in March; though additional study is needed to ~~understand among event variations in precipitation d~~ identify the mechanism responsible for the annual cycle.

Our study highlights how ~~the~~ variations in stable isotopes of precipitation measured on an event-by-event basis can be interpreted in the context of the vapor source~~when precipitation is measured on an event-by-event basis~~. The mechanisms identified, most notably the north-south migration of the vapor source region in ~~response to~~ phase with expansion and contraction of the Polar circulation cell, ~~are expected to~~ may also operate on times scales longer than that of our study~~e.g., interannual, decadal, millenial, or glacial-interglacial. The associated precipitation isotopic response at a given site to the variability of the vapor source should be recorded in~~ , and may be a source of variation in isotopes measured in ice cores, ~~lake sediments,~~ pedogenic carbonates, and speleothems.

5 Data availability

The processed data used for this research ~~is~~ are available as a supplement to the manuscript. Raw and partially processed results of the back trajectory runs may be obtained from Annie Putman (putmanannie@gmail.com).

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References

Aemisegger, F., Spiegel, J. K., Pfahl, S., Sodemann, H., Eugster, W., and Wernli, H.: Isotope meteorology of cold front passages: A case study combining observations and modeling, Geophys Res Lett, 42, 5652–5660, doi:10.1002/2015GL063988, 2015.

Bharadwaj, N., Widener, K., Nelson, D., Venkatesh, V., Lindenmaier, I., and Johnson, K.: KAZRGE 4-1-2011 to 4-1-2013, 71.323 N 156.609 W: North Slope Alaska (NSA) Central Facility, Barrow AK (C1), 2011.

Bintanja, R. and Selten, F. M.: Future increases in Arctic precipitation linked to local evaporation and sea-ice retreat, Nature, 509, 479–482, doi:10.1038/nature13259, 2014.

Bonne, J.-L., Masson-Delmotte, V., Cattani, O., Delmotte, M., Risi, C., Sodemann, H., and Steen-Larsen, H. C.: The isotopic composition of water vapour and precipitation in Ivittuut, southern Greenland, Atmospheric Chemistry and Physics, 14, 4419–4439, doi:10.5194/acp-14-4419-2014, 2014.

Bonne, J.-L., Steen-Larsen, H. C., Risi, C., Werner, M., Sodemann, H., Lacour, J.-L., Fettweis, X., Cesana, G., Delmotte, M., Cattani, O., Vallelonga, P., Kjær, H. A., Clerbaux, C., Sveinbjörnsdòttir, A. E., and Masson-Delmotte, V.: The summer 2012 Greenland heat wave: In situ and remote sensing observations of water vapor isotopic composition during an atmospheric river event, J Geophys Res - Atmos, 120, 2970–2989, doi:10.1002/2014JD022602, 2015.

Bowen, G.: Spatial analysis of the intra-annual variation of precipitation isotope ratios and its climatological corollaries, J Geophys Res - Atmos, 113, doi:10.1029/2007JD009295, 2008.

Cappa, C. D., Hendricks, M. B., DePaolo, D. J., and Cohen, R. C.: Isotopic fractionation of water during evaporation, J Geophys Res, 108, 4525, doi:10.1029/2003JD003597, 2003.

Craig, H. and Gordon, L. I.: Deuterium and oxygen 18 variations in the ocean and marine atmosphere, in: Stable Isotopes in Oceanographic Studies and Paleotemperatures, edited by Tongiogi, E., pp. 9–130, 1965.

Dansgaard, W.: Stable isotopes in precipitation, Tellus, 16, 436–468, doi:10.1111/j.2153-3490.1964.tb00181.x, 1964.

Dee, S., Noone, D., Buenning, N., Emile-Geay, J., and Zhou, Y.: SPEEDY-IER: A fast atmospheric GCM with water isotope physics, J Geophys Res - Atmos, 120, 73–91, doi:10.1002/2014JD022194, 2015.

Draxler, R. R.: HYSPLIT4 user's guide, Tech. Rep. ERL ARL-230D, NOAA Tech Memo, 1999.

Draxler, R. R. and Hess, G. D.: Description of the HYSPLIT4 modeling system, Tech. Rep. ERL ARL-224, NOAA Tech Memo, 1997.

Draxler, R. R. and Hess, G. D.: An overview of the HYSPLIT4 modeling system of trajectories, dispersion, and deposition., Aus Meteorol Mag, p. 295, 1998.

Fan, N.: Atmospheric control on isotopic composition and d-excess in water vapor over ocean surface, Master's thesis, Dartmouth College, 2016.

Feng, X.: Isotopic Investigation of Sea-ice and Precipitation in the Arctic Climate System, http://www.dartmouth.edu/~iispacs/, 2011.

Feng, X., Reddington, A. L., Faiia, A. M., Posmentier, E. S., Shu, Y., and Xu, X.: The changes in North American atmospheric circulation patterns indicated by wood cellulose, Geology, 35, 163–166, doi:10.1130/G22884A.1, 2007.

Feng, X., Faiia, A. M., and Posmentier, E. S.: Seasonality of isotopes in precipitation: A global perspective, J Geophys Res, 114, D08 116, doi:10.1029/2008JD011279, 2009.

Good, S., Mallia, D. V., Lin, J. C., and Bowen, G. J.: Stable Isotope Analysis of Precipitation Samples Obtained via Crowdsourcing Reveals the Spatiotemporal Evolution of Superstorm Sandy, PloS one, 9, e91 117, doi:10.1371/journal.pone.0091117, 2014.

Good, S., Noone, D., Kurita, N., Benetti, M., and Bowen, G.: D/H isotope ratios in the global hydrologic cycle, Geophys Res Lett, 42, 5042–5050, doi:10.1002/2015GL064117, 2015.

Hendricks, M. B., DePaolo, D. J., and Cohen, R. C.: Space and time variation of δ^{18}O and δD in precipitation: Can paleotemperature be estimated from ice cores?, Global Biogeochem Cy, 14, 851–861, doi:10.1029/1999GB001198, 2000.

Holdridge, D., Kyrouac, J., and Coulter, R.: SONDEWNPN 2009-01-01 to 2013-03-28, 71.323 N 156.609 W: North Slope Alaska (NSA) Central Facility, Barrow AK (C1), 1994.

Ichiyanagi, K. and Yamanaka, M. D.: Interannual variation of stable isotopes in precipitation at Bangkok in response to El Niño Southern Oscillation, Hydrol Process, 19, 3413–3423, doi:10.1002/hyp.5978, 2005.

Johnsen, S. J., Dansgaard, W., and White, J. W. C.: The origin of Arctic precipitation under present and glacial conditions, Tellus B, 41, 452–468, 1989.

Johnson, K. and Jensen, M.: ARSCL1CLOTH 1-1-2009 to 4-1-2013, 71.323 N 156.609 W: North Slope Alaska (NSA) Central Facility, Barrow AK (C1), 1996.

Jouzel, J. and Merlivat, L.: Deuterium and Oxygen-18 in Precipitation: Modeling of the Isotopic Effects During Snow Formation, J Geophys Res, 89, 11 749–11 757, doi:10.1029/JD089iD07p11749, 1984.

Jouzel, J., Russell, G. L., Suozzo, R. J., Koster, R. D., White, J. W. C., and Broecker, W. S.: Simulations of the HDO and $H_2^{18}O$ atmospheric cycles using the NASA GISS general circulation model: The seasonal cycle for present-day conditions, J Geophys Res - Atmos, 92, 14 739–14 760, doi:10.1029/JD092iD12p14739, 1987.

Kopec, B., Feng, X., Michel, F. A., and Posmentier, E.: Influence of sea ice on Arctic precipitation, PNAS, 113, 46–51, doi:10.1073/pnas.1504633113, 2016.

Kurita, N.: Origin of Arctic water vapor during the ice-growth season, Geophys Res Lett, 38, L02 709, doi:10.1029/2010GL046064, 2011.

Liu, J., Curry, J. A., H. Wang, M. S., and Horton, R. M.: Impact of declining Arctic sea ice on winter snowfall, PNAS, doi:10.1073/pnas.1114910109, 2012.

Liu, Z., Bowen, G. J., and Welker, J. M.: Atmospheric circulation is reflected in precipitation isotope gradients over the conterminous United States, J Geophys Res - Atmos, 115, D22 120, doi:10.1029/2010JD014175, 2010.

Majoube, M.: Oxygen-18 and deuterium fractionation between water and steam, J Chim Phys PCB, 68, 1423, 1971.

Marvel, K. and Bonfils, C.: Identifying external influences on global precipitation, PNAS, 110, 19 301–19 306, doi:10.1073/pnas.1314382110, 2013.

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

Min, S. K., Zhang, X., and Zwiers, F.: Human-Induced Arctic Moistening, Science, 320, 518–520, doi:10.1126/science.1153468, 2008.

NOAA/OAR/ESRL PSD at Boulder Colorado USA: NOAA OI SST V2 data, 2013.

Noone, D.: The influence of midlatitude and tropical overturning circulation on the isotopic composition of atmospheric water vapor and Antarctic precipitation, J Geophys Res - Atmos, 113, doi:10.1029/2007JD008892, 2008.

Pfahl, S. and Wernli, H.: Air parcel trajectory analysis of stable isotopes in water vapor in the eastern Mediterranean.

Putman, A.: Tracking the moisture sources of storms at Barrow, Alaska: Seasonal variations and isotopic characteristics, Master's thesis, Dartmouth College, 2013.

Regional Climate Center, W.: Barrow WSO Airport (500546), 2012.

Rindsberger, M., Magaritz, M., Carmi, I., and Gilad, D.: The relation between air mass trajectories and the water isotope composition of rain in the Mediterranean Sea area, Geophys Res Lett, 10, 43–46, doi:10.1029/GL010i001p00043, 1983.

Santer, B. D., Mears, C., Wentz, F. J., Taylor, K. E., Gleckler, P. J., Wigley, T. M. L., Barnett, T. P., Boyle, J. S., Brüggemann, W., Gillett, N. P., Klein, S. A., Meehl, G. A., Nozawa, T., Pierce, D. W., Stott, P. A., Washington, W. M., and Wehner, M. F.: Identification of human-induced changes in atmospheric moisture content, PNAS, 104, 15 248–15 253, doi:10.1073/pnas.0702872104, 2007.

Sodemann, H., Schwierz, C., and Wernli, H.: Interannual variability of Greenland winter precipitation sources: Lagrangian moisture diagnostic and North Atlantic Oscillation influence, J Geophys Res - Atmos, 113, D03 107, doi:10.1029/2007JD008503, 2008a.

Sodemann, H., Masson-Delmotte, V., Schwierz, C., Vinther, B. M., and Wernli, H.: Interannual variability of Greenland winter precipitation sources: 2. Effects of North Atlantic Oscillation variability on stable isotopes in precipitation, J Geophys Res - Atmos, 113, D12 111, doi:10.1029/2007JD009416, 2008b.

Steen-Larsen, H., Sveinbjörnsdòttir, A. E., Peters, A., Masson-Delmotte, V., Guishard, M., Hsiao, G., Jouzel, J., Noone, D., Warren, J., and White, J.: Climatic controls on water vapor deuterium excess in the marine boundary layer of the North Atlantic based on 500 days of in situ, continuous measurements, Atmos Chem Phys, 14, 7741–7756, doi:10.5194/acp-14-7741-2014, 2014.

Steen-Larsen, H. C., Sveinbjörnsdòttir, A. E., Jonsson, T., Ritter, F., Bonne, J.-L., Masson-Delmotte, V., Sodemann, H., Blunier, T., Dahl-Jensen, D., and Vinther, B. M.: Moisture sources and synoptic to seasonal variability of North Atlantic water vapor isotopic composition, J Geophys Res - Atmos, 120, 5757–5774, doi:10.1002/2015JD023234, 2015JD023234, 2015.

Stein, A. F., Draxler, R. R., Rolph, G. D., Stunder, B. J. B., Cohen, M. D., and Ngan, F.: NOAA's HYSPLIT Atmospheric Transport and Dispersion Modeling System, B Am Meteorol Soc, 96, 2059–2077, doi:10.1175/BAMS-D-14-00110.1; 21, 2015.

Stewart, M. K.: Stable Isotope Fractionation Due to Evaporation and Isotopic Exchange of Falling Waterdrops: Applications to Atmospheric Processes and Evaporation of Lakes, J Geophys Res, 80, 1133–1146, doi:10.1029/JC080i009p01133, 1975.

Stichler, W., Schotterer, U., Fröhlich, K., Ginot, P., Kull, C., Gäggeler, H., and Pouyaud, B.: Influence of sublimation on stable isotope records recovered from high-altitude glaciers in the tropical Andes, J Geophys Res - Atmos, 106, 22 613–22 620, doi:10.1029/2001JD900179, 2001.

Strong, M., Sharp, Z. D., and Gutzler, D. S.: Diagnosing moisture transport using D/H ratios of water vapor, Geophys Res Lett, 34, L03 404, doi:10.1029/2006GL028307, 2007.

Stull, R.: Practical Meteorology: An Algebra-based Survey of Atmospheric Science, Dept. of Earth, Ocean & Atmospheric Sciences University of British Columbia, Vancouver, BC, Canada, 3rd edn., http://www.eos.ubc.ca/books/Practical_Meteorology/, 2015.

Treble, P. C., Budd, W. F., Hope, P. K., and Rustomji, P. K.: Synoptic-scale climate patterns associated with rainfall $\delta^{18}O$ in southern Australia, J Hydrol, 302, 270–282, doi:10.1016/j.jhydrol.2004.07.003, 2005.

Uemura, R., Matsui, Y., Yoshimura, K., Motoyama, H., and Yoshida, N.: Evidence of deuterium excess in water vapor as an indicator of ocean surface conditions, J Geophys Res - Atmos, 113, D19 114, doi:10.1029/2008JD010209, 2008.

Wang, Y., Sodemann, H., Hou, S., Masson-Delmotte, V., Jouzel, J., and Pang, H.: Snow accumulation and its moisture origin over Dome Argus, Antarctica, Clim Dynam, 40, 731–742, doi:10.1007/s00382-012-1398-9, 2013.

White, J. W. C., Johnsen, S. J., and Dansgaard, W.: The origin of Arctic precipitation as deduced from its deuterium excess, Ann Glaciol, 10, 219–220, 1988.

Yoshimura, K., Kanamitsu, M., Noone, D., and Oki, T.: Historical isotope simulation using Reanalysis atmospheric data, J Geophys Res - Atmos, 113, doi:10.1029/2008JD010074, 2008.

Zhao, C. and Garrett, T. J.: Ground-based remote sensing of precipitation in the Arctic, J Geophys Res, 115, 1, doi:10.1029/2007JD009222, 2008.

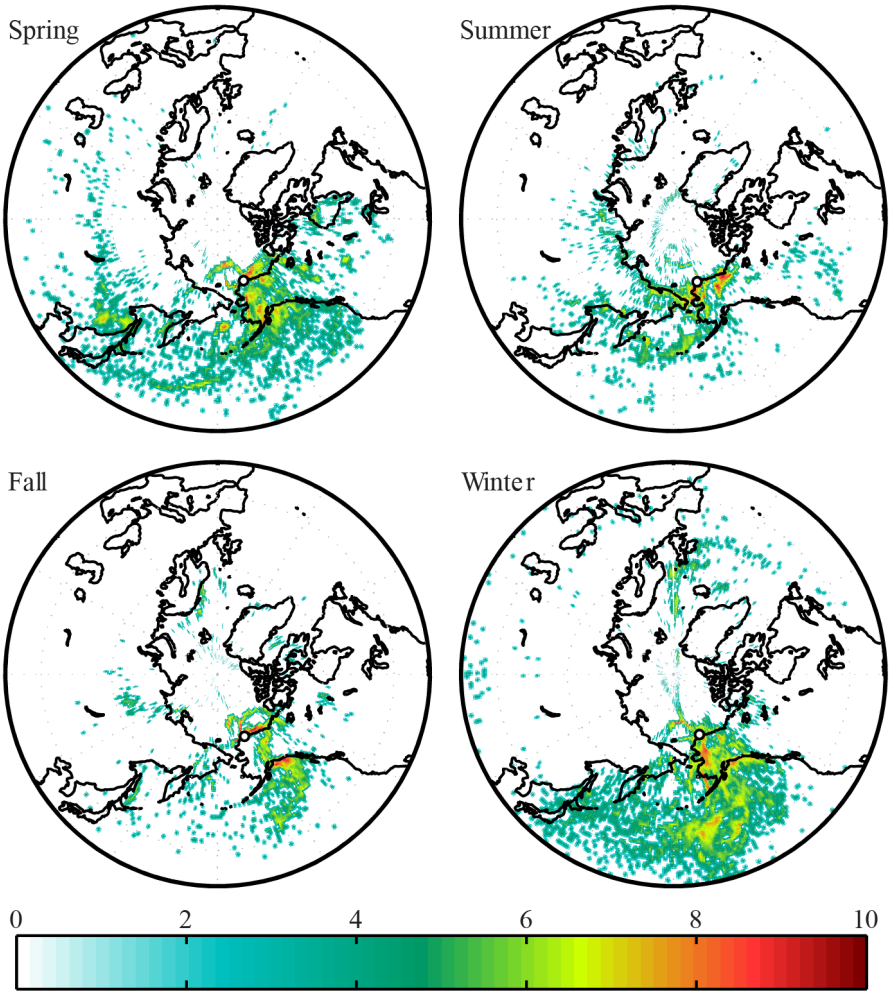

Figure 1. Spatial distribution of the vapor source region by season. Color indicates the relative frequency that a pixel was identified by HYSPLIT as a vapor source. Red indicates the most frequent vapor source for a given season, whereas dark blue indicates few air parcels were traced to that location. Because different numbers of events occurred in each season, each season's color scale is normalized to the total number of air parcels tracked during that season. The figure indicates that some air parcels originate over land, but these were not included in calculations.

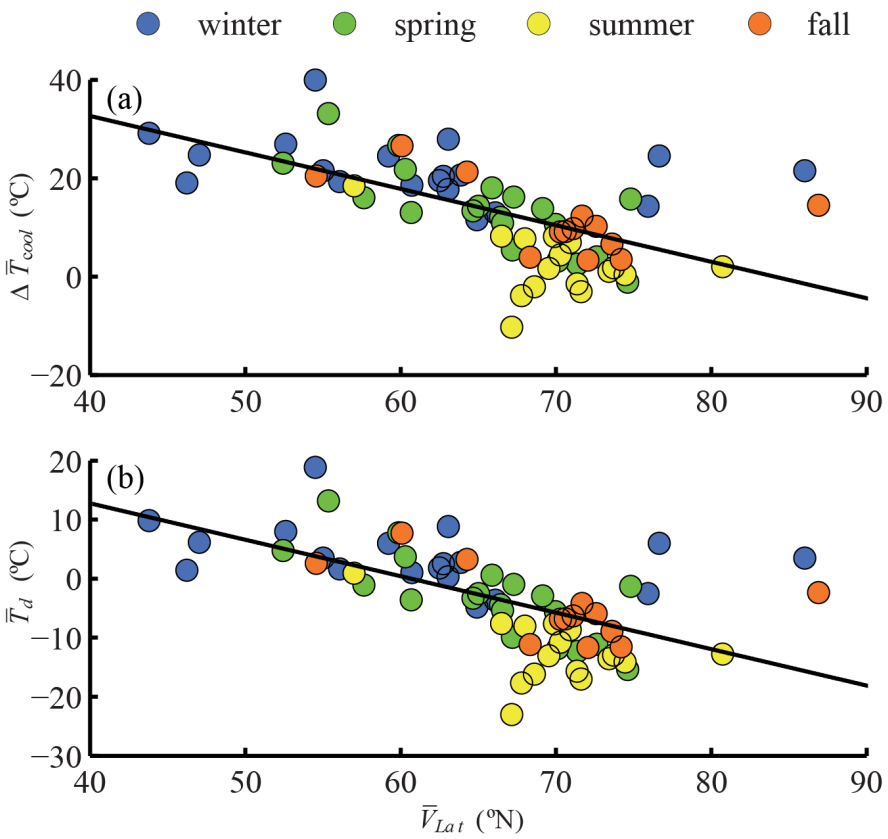

Figure 2. (a) ~~Mean~~ Covarying behavior of mean vapor source region latitude, $\bar{V}_{Lat}$, and mean air parcel cooling during transport, $\Delta\bar{T}_{cool}$~~,~~ ~~covary~~. (b) ~~Mean~~ Covariation of the mean vapor source region latitude, $\bar{V}_{Lat}$, and dew point, $\bar{T}_d$~~, also covary~~. Both $\Delta\bar{T}_{cool}$ and $\bar{T}_d$ influence the ~~δ^2 H~~ $\delta^2 H$ of precipitation at Barrow, AK. ~~Scatter~~ Lines are best-fits; scatter from ~~best fit line~~ them is due, in part, to seasonal variation in latitudinal temperature gradients and vapor source conditions.

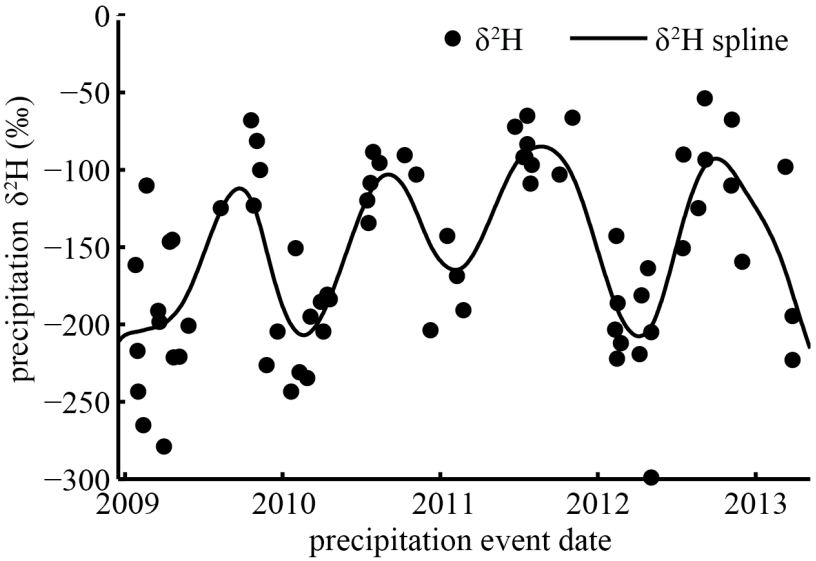

Figure 3. ~~The δ^2 H measured~~ Measured $\delta^2 H$ in precipitation at Barrow, AK, ~~exhibited~~ exhibits variability on interannual, annual, and event time scales. ~~Annual~~ The spline fit, which highlights seasonal variations, explains 65% of variance in the data with a root mean squared error of 39.7‰. Of the three timescales, annual variability ~~is~~ shows the greatest amplitude, ~~with~~ though variability among events is also substantial. Maximum enrichment ~~corresponding~~ corresponds roughly to the warmest months (June, July, August), and maximum depletion ~~corresponding~~ corresponds roughly to the coldest months (December, January, February).

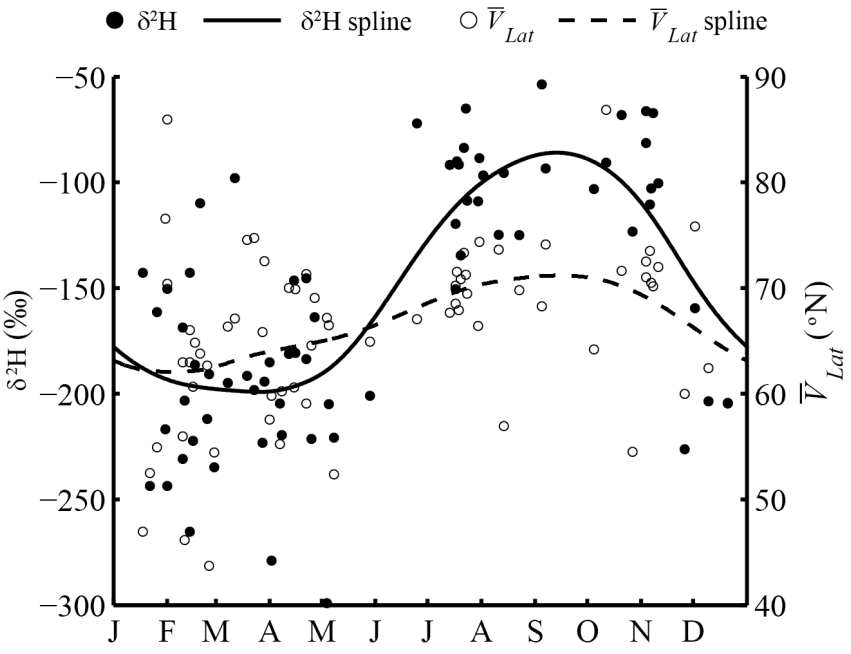

Figure 4. ~~The measured δ^2 H~~ Measured $\delta^2 H$ of Barrow precipitation and ~~the~~ mean latitude of the vapor source both exhibit an annual cycle and are in phase. The circles depict ~~the~~ raw data, while ~~the lines~~ curves are ~~a~~ spline ~~fit~~ fits to the data. The spline fits have R^2 values of 0.60 and 0.19 for the $\delta^2 H$ and $\bar{V}_{Lat}$ respectively. For both datasets, the variability exhibited among events is of the same the order of magnitude as the seasonal variability.

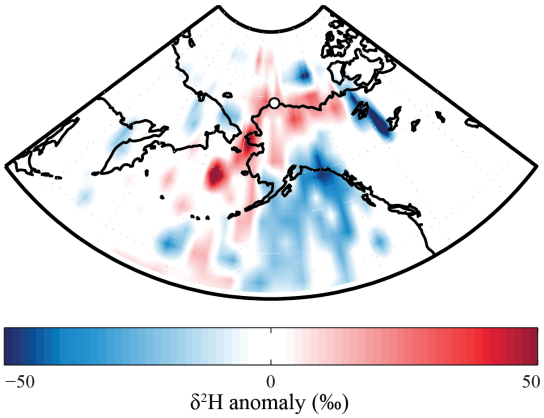

Figure 5. To demonstrate the effect of air parcel transport path, the ~~δ^2 H~~ residual $\delta^2 H$ of Barrow precipitation ~~with seasonal variation removed~~ is plotted at the vapor source. The residual $\delta^2 H$ is determined by subtracting the spline shown in Figure 4 ~~, is plotted at~~ from the ~~vapor source~~$\delta^2 H$ of each precipitation event. The ~~data~~vapor source locations, which ~~are on a~~ have 1 ° by 1 ° ~~spatial scale~~resolution, are smoothed for clarity. Vapor from the Bering Strait or Chukchi Sea tends to produce precipitation that is enriched relative to the average~~for that time of year~~. Likewise, vapor from the Gulf of Alaska tends to produce precipitation that is depleted relative to the average~~for that time of year~~. This variation in vapor source reflects a difference in transport path. Vapor originating from the Gulf of Alaska must rise to cross over the Alaska Range, inducing orographic precipitation and isotopic depletion relative to air masses that do not encounter orographic obstacles.

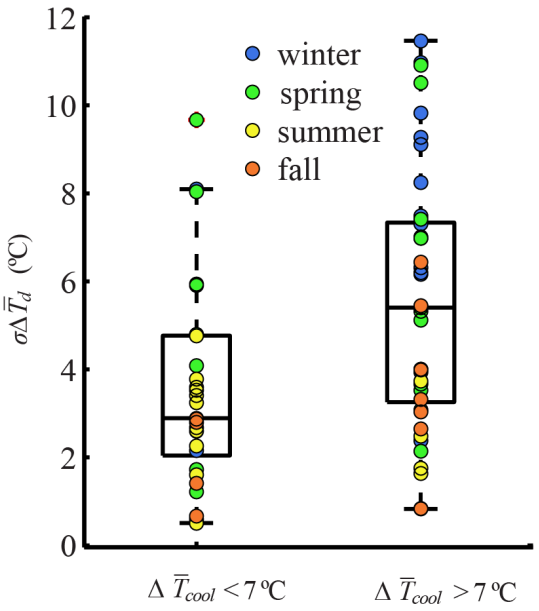

Figure 6. Distribution of standard deviations (σ) of $\bar{T}_d$ for events with $\Delta\bar{T}_{cool}$ <7 °C(short trajectories) and $\Delta\bar{T}_{cool}$ >7 °C (long trajectories). Colors indicate seasons. In general, small $\Delta\bar{T}_{cool}$ was associated with small $\sigma\bar{T}_d$. The variation in standard deviation is related to season, where warmer months tend to have smaller $\sigma\bar{T}_d$ and cooler months tend to have larger $\sigma\bar{T}_d$.

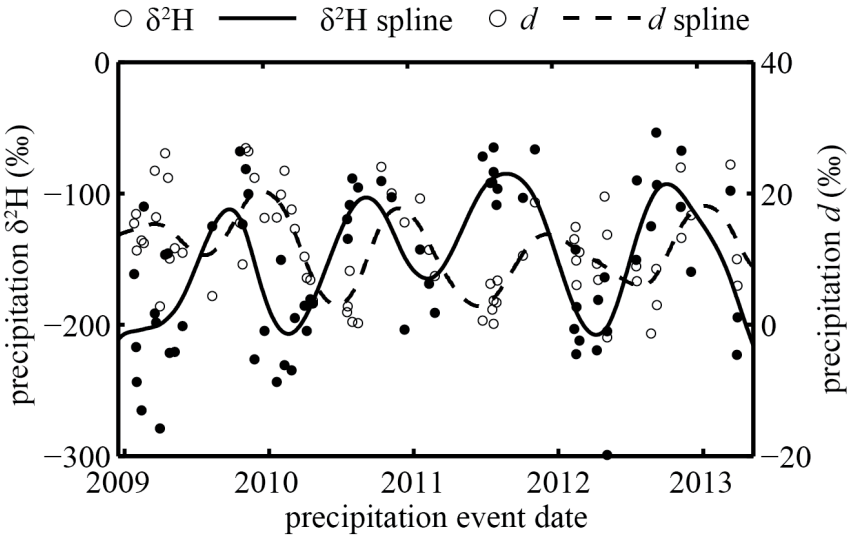

Figure 7. Annual maxima and minima in deuterium excess, *d*, lag those of ~~δ^2 H~~ $\delta^2 H$ by 2 - 3 months, such that the ~~maximum is~~ maxima are in fall and ~~minimum~~ minima in spring.

Table 1. Response variable: ~~δ^2 H, R^2 = 0.52~~$\delta^2 H$. Variation in ~~δ^2 H~~ $\delta^2 H$ is explained by a multiple linear regression (R^2 = 0.54) of air parcel cooling during transport ($\Delta\bar{T}_{cool}$), ~~evaporation~~ moisture source conditions ($\bar{T}_d$) and orographic obstacles in vapor transport path (mtn). ~~The slope~~ Values of β are the ~~correlation~~ partial coefficients of the regression and $S.E.$ is ~~β~~ the standard error. The variance estimate for each explanatory variable is calculated as the square of the semi-partial correlation for that variable with ~~δ^2 H~~$\delta^2 H$. The variances reported do not sum to the total variance explained because the explanatory variables are not perfectly orthogonal.

independent variable (slope units)	β ($\pm$ S.E.)	p-value	variance estimate
intercept	~~-96.9 (8.70~~-95.33 (8.62)	< 0.001	
$\Delta\bar{T}_{cool}$ (~~°C~~‰°C^{-1})	~~-3.25 (0.58~~-3.51 (0.55)	< 0.001	~~0.227~~0.287
$\bar{T}_d$ (~~°C~~‰°C^{-1})	~~3.80 (0.78~~3.23 (0.83)	< 0.001	~~0.171~~ 0.105
mtn (~~boolean~~‰ when mtn = 1)	~~-34.29 (11.05~~-32.11 (11.04)	~~0.0028~~ 0.0049	~~0.069~~0.059

Table 2. ~~Response variable: δ^2 H~~Three simple linear regressions against $\delta^2 H$ where β is the regression coefficient and $S.E.$ is the standard error. ~~Evaporation~~ Source conditions ~~(~~parameterized by $\bar{T}_d$ ~~)~~ explain most variation in ~~δ^2 H~~ $\delta^2 H$ for small $\Delta\bar{T}_{cool}$, while topographic highs below the trajectory (mtn) explain substantial variation for large $\Delta\bar{T}_{cool}$. $\Delta\bar{T}_{cool}$ explains variability significantly only for ~~both subregions, but explains a third of~~ the ~~variance for the small $\Delta\bar{T}_{cool}$~~ long transport subgroup ($\Delta\bar{T}_{cool}$ > 7°C).

	$\Delta\bar{T}_{cool}$ < 7 °C			$\Delta\bar{T}_{cool}$ > 7 °C	
Independent variable (slope units)	β ($\pm$ S.E.)	p-value	R^2	β ($\pm$ S.E.)	p-value
Intercept (‰)	~~-120.4 (8.01~~-111.4 (8.6)	< 0.001		~~-127.0 (21.8~~-115.1 (18.9)	< 0.001
$\Delta\bar{T}_{cool}$ (~~°C~~‰°C^{-1})	~~-5.81 (1.87~~-1.68 (2.1)	~~0.004~~ 0.428	~~0.26~~ 0.03	~~-2.84 (1.16~~-3.44 (0.97)	~~0.019~~ < 0.001
Intercept (‰)	~~-109.1 (7.52~~-104.9 (6.75)	< 0.001		~~-172.6 (9.60~~-176.5 (8.4)	< 0.001
$\bar{T}_d$ (~~°C~~‰°C^{-1})	~~3.9 (0.79~~2.89 (0.74)	< 0.001	~~0.48~~ 0.43	~~3.77 (1.81~~1.04 (1.8)	~~0.038~~ 0.58
Intercept (‰)	~~-124.7 (9.74~~-115.2 (9.1)	< 0.001		~~-140.1 (13.2~~-147.6 (11.9)	< 0.001
mtn (~~boolean~~‰when mtn = 1)	~~-21.5 (18.6~~-16.6 (24.6)	~~0.26~~ 0.51	~~0.05~~ 0.02	~~-62.2 (17.6~~-49.8 (15.5)	~~0.001~~ 0.0025

Table 3. Explaining ~~Deuterium~~ deuterium excess (d) ~~with~~ using simple regressions against various metrics that ~~quantify evaporation~~ characterize source conditions. β is the regression coefficient and $S.E.$ is the standard error. We show results from simple linear regressions with ~~three~~ four different independent variables: evaporation site relative humidity ($\bar{h}_{2m}$), ~~Sea~~ evaporation site relative humidity relative to sea surface temperature ($\bar{h}_{sst}$), sea surface temperature ($\bar{T}_{ss}$), and 2 m dew point ($\bar{T}_d$). ~~Relative humidity and sea surface temperature were explored independently because of correlation between the two.~~

Independent variable (slope units)	β ($\pm$ S.E.)	p-value	R^2
$\bar{h}_{2m}$ (~~‰~~‰$\%^{-1}$)	~~13.8 (15.0~~0.027 (0.157)	~~0.359~~ 0.86	0.0
$\bar{h}_{sst}$ (‰$\%^{-1}$)	-0.395 (0.067)	< 0.001	0.34
$\bar{T}_{ss}$ (~~°C~~‰°C^{-1})	~~-0.87 (0.35~~-1.17 (0.37)	~~0.016~~ 0.0023	~~0.08~~ 0.12
$\bar{T}_d$(~~°C~~‰°C^{-1})	~~-0.53 (0.12~~-0.56 (0.13)	< 0.001	~~0.24~~ 0.22