Key Points:
- A new 13-year-long data set of precipitation δ¹⁸O and δ²H measurements from Anchorage, Alaska
- Local surface air temperature explains approximately 30 percent of variability in δ¹⁸O and δ²H values
- Temporal variability reflects atmospheric circulation-driven changes in trajectories of precipitating air masses and local topographic effects

Supporting Information:
- Supporting Information S1
- Data Set S1

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Synoptic and Mesoscale Mechanisms Drive Winter Precipitation δ¹⁸O/δ²H in South-Central Alaska

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Abstract
Measurements of oxygen and hydrogen stable isotopes in precipitation (δ¹⁸O_p and δ²H_p) provide a valuable tool for understanding modern hydrological processes and the empirical foundation for interpreting paleo-isotope archives. However, long-term data sets of modern δ¹⁸O_p and δ²H_p in southern Alaska are entirely absent, thus limiting our insight and application of regionally defined climate-isotope relationships in this proxy-rich region. We present and utilize a 13-year-long record of event-based δ¹⁸O_p and δ²H_p data from Anchorage, Alaska (2005–2018, n = 332), to determine the mechanisms controlling precipitation isotopes. Local surface air temperature explains ~30% of variability in the δ¹⁸O_p data with a temperature-δ¹⁸O slope of 0.31%/°C, indicating that δ¹⁸O_p archives may not be suitable paleo-thermometers in this region. Instead, back-trajectory modeling reveals how winter δ¹⁸O_p/δ²H_p reflects synoptic and mesoscale processes in atmospheric circulation that drive changes in the passage of air masses with different moisture sources, transport, and rainout histories. Specifically, meridional systems—with either northerly flow from the Arctic or southerly flow from the Gulf of Alaska—have relatively low δ¹⁸O_p/δ²H_p due to progressive cooling and removal of precipitation as it condenses with altitude over Alaska’s southern mountain ranges. To the contrary, zonally derived moisture from either the North Pacific and/or Bering Sea retains relatively high δ¹⁸O_p/δ²H_p values. These new data contribute a better understanding of the modern Alaska water isotope cycle and provide an empirical basis for interpreting paleo-isotope archives in context of regional atmospheric circulation.

1. Introduction
Measurements of oxygen and hydrogen stable isotopes in precipitation (δ¹⁸O_p and δ²H_p) provide a valuable and increasingly applied tool for tracing hydrological processes through space and time. These include processes occurring in the modern water cycle (Aggarwal et al., 2016; Klein et al., 2015; Puntag et al., 2016), in atmospheric circulation models (Jouzel et al., 1994; Yoshimura et al., 2008), and those recorded in natural precipitation archives such as ice cores, tree rings, and lake sediments (Dansgaard et al., 1993; McCarroll & Loader, 2004; van Hardenbroek et al., 2018).

The utility of isotopic tracers reflects the distinct and well-defined fractionation coefficients and diffusivities of water molecules as they undergo phase transitions in nature (Craig, 1961). These fundamental attributes lead to broadly predictable spatial-temporal patterns in global δ¹⁸O_p and δ²H_p that reflect a multitude of factors, with studies traditionally focusing on the global temperature-isotope relationship (Dansgaard, 1964). However, more recent work highlights how several natural processes such as variations in precipitation amount, altitude, relative humidity, moisture source, latitude, and continentality can also influence precipitation isotopologues (Gat, 1996; Liu et al., 2010; Rozanski et al., 1993). As a result, δ¹⁸O_p and δ²H_p values at any given site will represent an integrated signal of source region conditions, transport pathways, and rainout histories. Hence, our ability to accurately trace and monitor hydrologic changes through time is dependent on a solid understanding of these modern processes at a range of spatial and temporal scales (Sjostrom & Welker, 2009).

Here we focus on the Gulf of Alaska region where ice-covered mountains, forested ecosystems, and marine estuaries and fjords are highly sensitive to and dependent on the availability of seasonal and perennial water resources (Beamer et al., 2017). A growing number of studies across Alaska and the Yukon utilize stable isotope data to trace these hydrological processes through time (Anderson et al., 2005; Bailey et al., 2014, ...
2018; Fisher et al., 2008; Klein et al., 2016), and collectively, these records are important for placing contemporary changes in the context of long-term trends and natural variability. However, our understanding of the physical processes controlling modern and paleo-δ¹⁸O_p and δ²H_p variability in this region is currently underdeveloped due to the sparse observational network of meteoric δ¹⁸O_p and δ²H_p measurements; data are presently confined to six locations in the Global Network of Isotopes in Precipitation (IAEA/WMO, 2018; Figure 1a) and a handful of discrete sampling locations (Lachniet et al., 2016; Putman et al., 2017). Furthermore, existing data typically represent isotopic measurements of monthly composited precipitation samples (IAEA/WMO, 2018), thereby precluding any ability to identify process-based mechanisms that influence individual precipitation events. While some efforts to delineate spatial patterns of δ¹⁸O_p and δ²H_p have been carried out using measured and/or modeled data (Anderson et al., 2016; Bailey et al., 2015; Berkelhammer et al., 2012; Field et al., 2010; Putman et al., 2017), long-term empirical data from Alaska are severely lacking in comparison to the contiguous United States (Akers et al., 2017; Liu et al., 2010; Vachon et al., 2010a, 2010b; Welker, 2000, 2012).

To address this knowledge gap, we present a 13-year-long event-based precipitation isotope data set from Anchorage in south-central Alaska. This time series of discrete δ¹⁸O_p and δ²H_p measurements contributes critical information for understanding (1) the diurnal, monthly, seasonal, and long-term trends in south-central Alaska δ¹⁸O_p and δ²H_p values; (2) the relative role of temperature versus local (i.e., mesoscale) and distal (i.e., synoptic-scale) controls on south-central Alaska δ¹⁸O_p and δ²H_p; (3) how isotopic variability recorded in past hydrological archives can be used to reconstruct past changes in regional climate and atmospheric circulation; and (4) the complex atmospheric processes that are critical for accurately constraining climate forecast models (e.g., Steen-Larsen et al., 2017).

2. Study Region

Anchorage (61°N, 149°W) is a coastal city located at the northern end of Cook Inlet in south-central Alaska (Figure 1). The city is flanked by the Chugach Mountains that rise abruptly to >3,000 m above sea level (asl) and, together with the Kenai Peninsula, inhibit the inflow of moist maritime air masses from the Gulf of Alaska (Figure 1). For instance, mean annual precipitation in Anchorage is typically ~15% of that in Whittier on the windward side of the coastal range 80 km away (Figure 1b; NOAA, 2018). Mean annual total precipitation (rain and snow) in Anchorage is ~2,300 mm, and approximately 80% falls between November and March (Figure 2) (1954–2017; NOAA, 2018). Precipitation during the winter months is typically frozen in form though rainfall does occur (Figure 2). Mean annual surface air temperature (SAT) is 2.3 °C, with
mean winter (December–February) and summer (June–August) values −4.2 and +13.8 °C, respectively (1954–2017; NOAA, 2018).

More broadly, “winter” conditions in Anchorage typically last from mid-October through March when air temperatures are below 0 °C and streams and lakes are frozen. Sea ice in Cook Inlet also starts to form in late October and reaches maximum extent and thickness by mid-February to early March (Figure 1b), with ice-out in early to mid-April (NSIDC, 2018). The dynamic climate, marine, and tidal conditions in the inlet typically prevent a continuous intact ice surface, and instead much of the cover is brash ice thickened by compaction, rafting, and ridging (Mulherin et al., 2001).

Regionally, the strong seasonal climate cycle is driven by the Aleutian Low pressure cell: a semi-permanent feature that builds over the North Pacific during late fall and steers storms and abundant precipitation into the Gulf of Alaska (Rodionov et al., 2007). In summer, the Pacific subtropical high expands over the northeast Pacific and brings westerly flow to the region, while the jet stream and its associated polar front account for latitudinal summer temperature gradients over Alaska (Mock et al., 1998). Persistent and long-term changes in these atmospheric patterns are well expressed in climate oscillations such as the Pacific Decadal Oscillation (PDO) and North Pacific Index (NPI; Mantua et al., 1997; Trenberth & Hurrell, 1994).

3. Methods

3.1. Precipitation Collection and Isotope Analysis

Precipitation samples were opportunistically collected from 332 precipitation events (rain and snow) in Anchorage between November 2005 and January 2018 (i.e., not every precipitation event during the 2005 to 2018 period was sampled). The sampling site (Tideview Station) is situated <500 m inland from the shore of Cook Inlet at 30 m asl. Precipitation was collected using a standard cylindrical gauge (CoCoRaHS, 2018) mounted to a wooden deck ~4 m above ground level, thereby eliminating the contribution of windblown snow from previous events. Samples were collected at the end of each precipitation event and typically within 1–3 hr to minimize evaporation or sublimation. Following rainfall, sample water was directly poured into 20-ml screw-cap vials and sealed. During winter, fresh snow samples were collected following an event, sealed in plastic zip lock bags, and thawed at room temperature before transferring into vials. Air temperature, precipitation amount, and time of day were recorded when feasible. Following sample collection, the gauge was emptied, dried, and reinstalled. All samples were directly transported to the Stable Isotope Laboratory at the University of Alaska Anchorage where 2 ml of each sample was pipetted into septa-capped glass vials. Samples were then refrigerated at 4 °C until isotopic analysis, which typically occurred within 6 weeks of sample collection. The remaining sample was archived frozen for 1 year and discarded.

Stable isotopes ($^{16}$O/$^{18}$O and $^1$H/$^2$H) were measured using a Picarro Cavity Ring-Down Spectrometer fitted with an auto-sampler. Each sample was analyzed six times, and reanalysis of the sample occurred if the
standard deviation was greater than 0.3‰ for $\delta^{18}O$ and/or 3.0‰ for $\delta^2H$ or if the internal standard for the run differed from the accepted value by greater than $\pm0.2\%$ or $2.0\%$, respectively. Analytical precision for all samples is $\pm0.2\%$ for $\delta^{18}O$ and $\pm2.0\%$ for $\delta^2H$. All results are reported as parts per thousand difference (per mil; ‰) relative to Vienna-Standard Mean Ocean Water. The secondary parameter deuterium excess ($d$-excess) is calculated as $d$-excess $= \delta D - 8 \times \delta^{18}O$ (Craig, 1961). To calculate monthly isotope averages, calibrated $\delta^{18}O$ and $\delta^2H$ values are reduced to precipitation amount-weighted values using the equation:

$$\text{precipitation weighted } \delta^{18}O/\delta^2H = \frac{\sum_{i=1}^{n} P_i \delta_i}{\sum_{i=1}^{n} P_i},$$

where $P_i$ and $\delta_i$ denote the amount (mm) of precipitation per event and its measured isotope composition, respectively, and $n$ represents the number of precipitation events.

### 3.2. Air Parcel Trajectory Modeling

The air mass history of each Anchorage precipitation event ($n = 332$) was determined using 240-hr (10-day) back-trajectories. Back-trajectories were computed using the Hybrid Single-Particle Lagrangian Integrated Trajectory (HYSPLIT) model (Rolph et al., 2017; Stein et al., 2015) coupled with the 0.5° three-dimensional Global Data Assimilation System (GDAS) data set (NCEP, 2018). The model uses gridded hourly meteorological data ($T$, temperature; $Z$, height; $U$ [zonal] and $V$ [meridional] horizontal wind components; and surface pressure, $P_S$) to calculate air parcel back-trajectories from the precipitation collection site. Each back-trajectory was initiated at a height of 1,500 m above ground level and traced back 240 hr before the precipitation event. A 1,500-m starting height was chosen to reduce atmospheric flow attenuation by surface friction while remaining important for moisture entrainment and transport (Fiorella et al., 2015). Seven environmental parameters were extracted from the GDAS data set at hourly intervals along each back-trajectory: latitude, longitude, pressure, air temperature, relative humidity, precipitation (mm/hr), and altitude (m asl). To determine the density of trajectories in space and time, hourly air parcel positions ($x,y$ coordinates) were binned on a $2^\circ \times 2^\circ$ grid and used to calculate the percentage of trajectories passing through each fixed grid cell.

Sensitivity experiments were performed using starting heights of 500, 1,000, and 2,000 m and resulted in similar trajectories to those at 1,500 m (not shown). To assess how well the numerical fields estimate the true flow field in space and time, comparative tests ($n = 20$) were performed using two alternative wind fields: the NCEP North American Regional Reanalysis (NARR) 32-km-resolution data set (Mesinger et al., 2006) and the NCEP/NCAR REANALYSIS data set (Kalnay et al., 1996). Air parcels were found to follow similar trajectories independent of the reanalysis data set used in the model. The GDAS reanalysis package was ultimately selected for our analyses because (1) it integrates the shortest possible time step (1 hr) when calculating trajectories, thereby minimizing any integration error; by comparison, NARR and REANALYSIS data sets are available in 3- and 6-hr intervals, respectively, and (2) it offers highly resolved regional coverage, thereby enabling the model to assimilate Alaska’s complex topographic features. Lastly, 240-hr forward trajectories were computed from the endpoint of a random subset of trajectories ($n = 20$) and were found to terminate at or close to our Anchorage sampling site, indicating that the trajectories are numerically accurate.

### 3.3. Meteorological and Oceanographic Data

Daily climate data spanning the 2005 to 2018 collection period were acquired from the Anchorage Ted Stevens International Airport meteorological station and are used for our analyses (data coverage: 1952–present; NOAA, 2018). This was necessary due to substantial gaps in retrieving on-site temperature and precipitation measurements. The airport is located adjacent to Cook Inlet at 30 m asl, 10 km from our precipitation sampling site. Data measurements include maximum and minimum SAT, snow and rainfall amounts, and 5-s wind speed and wind direction. Mean daily SATs were calculated from daily temperature extremes.

Daily areal sea ice extent data for Cook Inlet were obtained from the Multisensor Analyzed Sea Ice Extent data set available online from the National Snow and Ice Data Center (NSIDC, 2018). Data include sea ice initiation and ice-out dates, as well as maximum and minimum sea ice coverages (km$^2$). Lastly, NARR data
were used to construct composite analyses of synoptic to mesoscale conditions during Anchorage precipitation events (Mesinger et al., 2006). NARR comprises data with 32-km horizontal grid spacing (0.3°) on 45 vertical levels and are available at 3-hr intervals, and we use composites of the daily mean (i.e., average of the 0z–21z 3-hr data).

4. Results and Discussion

4.1. Anchorage Precipitation Isotopes

The 2005 to 2018 time series of Anchorage precipitation δ¹⁸O, δ²H, and d-excess values are presented in Figure 2 (data available in supporting information Data Set S1). Isotope values range from −5.7‰ to −36.8‰ for δ¹⁸O, and −48.7‰ to −294.4‰ for δ²H (Figure 2a). D-excess values range between −34.2‰ and 31.6‰. Mean annual precipitation-weighted values are −16.5‰ for δ¹⁸O, and −130.4‰ for δ²H.

Collectively, these data define a local meteoric water line for Anchorage and south-central Alaska where δ²H = 7.22 δ¹⁸O − 11.02 (r² = 0.94, n = 332; Figure 2b). The slope and intercept of the local meteoric water line are slightly lower than the Global Meteoric Water Line (GMWL) of δ²H = 8 δ¹⁸O + 10 (Craig, 1961). Cool season precipitation events (October through March) typically plot on the GMWL, while warm season precipitation (April through September) plots slightly below the GMWL (Figure 2b). This may indicate secondary evaporation and/or post-condensational exchange of raindrops during the warmer season (i.e., subcloud processes; Dansgaard, 1964). Conversely, in the cold season, when >90% of precipitation falls as snow (and other solid forms; NOAA, 2018), precipitation will not equilibrate with ambient moisture by continuous exchange during the descent from cloud base to the ground (Dansgaard, 1964).

Anchorage precipitation isotope values exhibit pronounced seasonality that follows the well-established annual cycle for middle and high latitudes (Figure 3a; Rozanski et al., 1993; Welker, 2000). Isotope values are typically higher in the warm season and lower in the cool season, with mean winter (December–February) and summer (June–August) δ¹⁸O values of −19.6‰ and −14.4‰, respectively (Table 1). This seasonal isotope cycle parallels the strong seasonal temperature and precipitation cycle in Anchorage (Figure 2). However, there is no significant correlation between δ¹⁸O and recorded precipitation amount (r² = 0.01, p = 0.18, n = 332), and local SAT explains only ~30% of variability in the δ¹⁸O data (r² = 0.33, p < 0.01, n = 332) with a temperature-δ¹⁸O slope of 0.31‰/°C (Figure S1). This regression is similar to precipitation data from northern Alaska that exhibit a temperature-δ¹⁸O slope of 0.35‰/°C (Klein et al., 2015). Based on these observations, we propose a regional temperature-δ¹⁸O relationship of −0.33‰/°C to be more appropriate for mainland Alaska than the global spatial average of 0.67‰/°C (Dansgaard, 1964).

There are no significant long-term trends in δ¹⁸O, δ²H, or d-excess values (2005–2018; Figure 2a). Long-term interannual seasonal variability exhibits a minor lowering trend in mean precipitation-weighted δ¹⁸O in all seasons, but these trends are not significant (Figure S2). Conversely, long-term interannual mean d-excess values exhibit a slight increasing trend during winter, spring, and summer, while fall d-excess values exhibit long-term decreasing trends (2005–2018), though again neither trend is significant (p > 0.05; Figure S2).

Table 1

<table>
<thead>
<tr>
<th>Month</th>
<th>δ¹⁸O (%)</th>
<th>δ²H (%)</th>
<th>d-excess (%)</th>
<th>n</th>
</tr>
</thead>
<tbody>
<tr>
<td>January</td>
<td>−20.6</td>
<td>−154.0</td>
<td>10.7</td>
<td>26</td>
</tr>
<tr>
<td>February</td>
<td>−16.7</td>
<td>−124.3</td>
<td>9.5</td>
<td>29</td>
</tr>
<tr>
<td>March</td>
<td>−17.5</td>
<td>−129.2</td>
<td>10.7</td>
<td>17</td>
</tr>
<tr>
<td>April</td>
<td>−18.9</td>
<td>−147.5</td>
<td>3.6</td>
<td>14</td>
</tr>
<tr>
<td>May</td>
<td>−15.3</td>
<td>−126.1</td>
<td>−3.5</td>
<td>15</td>
</tr>
<tr>
<td>June</td>
<td>−16.0</td>
<td>−129.0</td>
<td>−0.6</td>
<td>21</td>
</tr>
<tr>
<td>July</td>
<td>−13.7</td>
<td>−112.2</td>
<td>−2.9</td>
<td>53</td>
</tr>
<tr>
<td>August</td>
<td>−14.3</td>
<td>−116.5</td>
<td>−2.4</td>
<td>47</td>
</tr>
<tr>
<td>September</td>
<td>−17.3</td>
<td>−140.3</td>
<td>−2.1</td>
<td>35</td>
</tr>
<tr>
<td>October</td>
<td>−15.6</td>
<td>−116.3</td>
<td>8.6</td>
<td>29</td>
</tr>
<tr>
<td>November</td>
<td>−22.1</td>
<td>−167.4</td>
<td>9.6</td>
<td>22</td>
</tr>
<tr>
<td>December</td>
<td>−23.0</td>
<td>−173.5</td>
<td>10.9</td>
<td>24</td>
</tr>
<tr>
<td>Annual</td>
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<td>−130.4</td>
<td>1.8</td>
<td>332</td>
</tr>
<tr>
<td>December–February</td>
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<td>−147.6</td>
<td>9.1</td>
<td>79</td>
</tr>
<tr>
<td>March–May</td>
<td>−16.6</td>
<td>−129.8</td>
<td>3.1</td>
<td>46</td>
</tr>
<tr>
<td>June–August</td>
<td>−14.4</td>
<td>−118.4</td>
<td>−2.8</td>
<td>121</td>
</tr>
<tr>
<td>September–November</td>
<td>−18.5</td>
<td>−143.9</td>
<td>4.0</td>
<td>86</td>
</tr>
</tbody>
</table>

*Mean precipitation-weighted isotope values.
Across all local environmental data sets, the most significant correlation is between mean winter $\delta^{18}O_p$ and maximum winter sea ice extent (km$^2$) in Cook Inlet, which are positively correlated (i.e., higher $\delta^{18}O_p$ with more sea ice; $r^2 = 0.43$, $p < 0.05$; Figure S3). Conversely, there is a weak negative correlation between Anchorage winter SAT and maximum Cook Inlet sea ice extent (i.e., more sea ice with colder temperatures; $r^2 = 0.24$, $p < 0.05$), as might be expected. An inverse relationship between $\delta^{18}O_p$ and sea ice might be expected from traditional isotope-temperature effects, that is, decreasing $\delta^{18}O_p$ with temperature (Dansgaard, 1964), rather than the observed positive relation. Hence, collectively these results indicate that diurnal and/or seasonal changes in SAT only partially account for isotope variability in Anchorage and that additional factors are of equal or greater importance.

4.2. Precipitation Isotopes and Local Wind Direction

To better understand the seasonal and temperature fractionation effects at the local scale, we partition the Anchorage isotope data into seasonal “Isoroses”—wind rose style summations that depict event-based $\delta^{18}O_p$ as a function of daily cardinal wind direction (Figure 4). We assume wind direction measurements recorded at Anchorage airport are representative of those at our nearby precipitation collection site.

The data demonstrate that northerly winds account for 38% of all sampled precipitation events in Anchorage between December through February (2005–2018), and these events are characterized by the lowest measured $\delta^{18}O_p$ values, between $-37\%$ and $-23\%$ (Figure 4). Conversely, winter precipitation events with the highest $\delta^{18}O_p$ values derive primarily from systems arriving from the south and southeast (26%) and exhibit $\delta^{18}O_p$ values ranging from $-18\%$ to $-10\%$ (Figure 4). In spring, northerly precipitation events are also characterized by relatively low $\delta^{18}O_p$ values, but these events occur less frequently (18%) in the data set. Instead, the prevalence of southwesterly to southeasterly winds increases to 44%, and these are associated with precipitation characterized by the highest spring $\delta^{18}O_p$ values, from $-18\%$ to $-6\%$ (Figure 4).

In summer, southeasterly and southerly winds continue to prevail in 42% and 34% of all events, respectively, and associated isotope values are also relatively high (Figure 4). Of all summer precipitation events, 12% derived from north or northeasterly wind events, and these exhibit $\delta^{18}O_p$ values between $-10\%$ and $-18$.

**Figure 4.** Seasonal relations between Anchorage $\delta^{18}O_p$ and cardinal wind direction (2005–2018). Wind direction data recorded at Ted Stevens International Airport in Anchorage (NOAA, 2018).
In the fall, southeasterly events prevail at 28%, though southerly and northerly winds are also prevalent with near equal occurrence at 24% and 27%, respectively (Figure 4).

4.3. Storm Track Trajectories and \(\delta^{18}O_p\)

The well-defined seasonal relationships between wind direction and \(\delta^{18}O_p\) suggest that both local (i.e., mesoscale) and distal (i.e., synoptic-scale) processes influence isotopic variability in Anchorage precipitation. For example, seasonal variations in \(\delta^{18}O_p\) may be driven by the different precipitation moisture source regions and transport pathways during the winter and summer. To understand these long-range moisture transport processes, back-trajectories for all sampled precipitation events are presented in Figure 5 and categorized by season.

Air parcel tracks span a wide latitudinal range from 22°N to 89°N, though trajectory densities are the highest in the proximal zones of the Gulf of Alaska and Cook Inlet. As all trajectories were modeled for the same duration (240 hr), the longer trajectories indicate faster moving air masses (e.g., December through February), and shorter trajectories indicate slower moving ones (e.g., June through August; Figure 5). Most trajectories (~90%) have an “oceanic” origin at 240 hr rather than “continental.” This typically occurs over the Pacific Ocean, but source regions also include the Bering, Chukchi, and Okhotsk Seas, as well as the Arctic Ocean (Figure 5). This suggests a minimal component of “recycled” continental moisture contributing to precipitation in Anchorage, wherein the term recycled reflects moisture that originates from evapotranspiration over land, in contrast to a maritime moisture source. In all seasons, the eastward propagation of air masses across the North Pacific is evident and can be attributed to the region’s prevailing midlatitude westerlies (Figure 5; Mock et al., 1998).

Winter back-trajectories exhibit the largest variation in both trajectory range and source, spanning an area between 22°N and 86°N (Figure 5). While most winter trajectories originate—at 240 h before the event—in the western North Pacific and track east across the North Pacific into the Gulf of Alaska, ~40% derive from the Arctic and travel south across interior Alaska or the Yukon. During spring (March through May), storms follow similar trajectories as winter and track east across the North Pacific and into Anchorage, though fewer trajectories originate and/or pass over the Arctic Ocean and interior Alaska compared to winter.

Instead, storms typically traverse eastward across the Bering Sea and into Cook Inlet (Figure 5). Storm
trajectories span the smallest latitudinal range during the summer months (June through August) where they track between ~25°N and 68°N. Most of these storms (>95%) originate in the North Pacific and track either directly northeast into the Gulf of Alaska, or they traverse eastward along the Aleutian Islands and/or the Bering Sea (Figure 4). Only ~10% of the summer precipitation events have trajectories that traverse eastward across Alaska compared to ~55% during winter. During the fall (September through November), storm trajectories become longer (i.e., faster) and have a more northerly track, such as those during spring. It is apparent that during the transition into winter when the polar front typically becomes more well defined and positioned further south (Mock et al., 1998), the average trajectory of westerly North Pacific storms is also located further south (Figure 5).

These observed nuances in storm trajectory can begin to account for the seasonal isotopic variability in Anchorage precipitation. For example, winter is characterized by the lowest δ18Op/δ2H and highest d-excess values, as well as the largest proportion of trajectories originating in the Arctic (~40%, Figure 5a). Similar relationships were observed in the northern interior of Alaska (Klein et al., 2015) and central Aleutian Islands (Bailey et al., 2015) where precipitating air masses from the Arctic bring water vapor and precipitation that has relatively low δ18O and high d-excess values. Alternatively, prevailing southerly storm tracks entrain warm 18O-enriched water vapor from lower latitude regions (i.e., the Pacific), and in these cases precipitation was found to have relatively high δ18Op/δ2H and low d-excess values (Bailey et al., 2015; Klein et al., 2015). In winter, sea ice in the Arctic Ocean and Bering Sea may further influence air masses originating in the north by altering the availability of atmospheric moisture (e.g., Klein et al., 2015; Kurita, 2011).

4.4. Mesoscale Mechanisms Influencing Extreme Winter δ18Op

Despite the wide range of water vapor sources for Anchorage precipitation, the North Pacific remains the dominant source and storm trajectory pathway in all seasons (Figure 5). To delineate the specific mechanisms controlling δ18Op, we focus on the extreme high and low δ18Op events that occurred in the cold season: an interval when >80% of Anchorage’s annual precipitation falls (i.e., October through March, herein referred to as “winter”; Figure 2). High events are classified as having δ18Op values higher than one standard deviation from the mean winter δ18Op value (greater than ~14.2‰), and low events are one standard deviation lower (less than ~23.9‰).

Composite maps of the individual storm trajectories for these events reveal clear trends among the two scenarios (Figure 6). For high δ18Op winter events (n = 26, range: ~14.2‰ to ~8.5‰), the highest density of trajectories occurs in Cook Inlet (>50%) where there is a prominent southwesterly flow component directly into Anchorage (Figure 6a). For the low δ18Op winter events (n = 26, range: ~36.8‰ to ~18.9‰), a higher density of storms track into the Gulf of Alaska where they are steered northwest into Prince William Sound and pass into Anchorage from the east and northeast (>70%; Figure 6b).

To obtain a semiquantitative estimate of the processes influencing δ18Op along these contrasting trajectories, we analyze the environmental variables extracted along each HYSPLIT back-trajectory started from Anchorage. Specifically, we use a factor analysis with varimax rotation (Klovan & Imbrie, 1971) to explore variance among the latitude, longitude, altitude, pressure, air temperature, precipitation, and relative humidity values extracted at hourly intervals along each 240-hr trajectory (e.g., Ersek et al., 2010) (Figure 7). The analysis was performed separately using all variables extracted at 240, 120, 48, and 24 hr before the precipitation event, for both extreme high and low δ18Op trajectories, to determine which trajectory length explains the highest total variance among the data (i.e., the optimal trajectory length).

The trajectory length accounting for most of the variance among environmental variables is 48 hr in duration and explains 84% and 86% of the total variance in the high and low δ18O events, respectively. By comparison, extracted variables along the 240-hr trajectories account for only 64% and 60%; and focusing on conditions during the final 24 hr explained 70% and 81% of the variance. These data suggest that environmental conditions along the storm trajectory during the 48 hr preceding a precipitation event may be most influential in terms of δ18Op (in Anchorage).

Along the 48-hr trajectories, three stable factors explain 84% of the total variance in the high δ18Op events (Figure 7a). Factor score loadings greater than or equal to 0.5 are dominating factors; that is, these factors characterize the main features of the air mass as it moves along its respective trajectory. Factor 1 explains 48% and is dominated by trajectory altitude (~0.96), pressure (~0.90), and temperature (~0.83; Figure 7a).
Factor 2 has the highest score loadings for latitude (+0.78) and explains 20% of the variance, while Factor 3 accounts for 16% and is dominated by precipitation (+0.94). In the low $\delta^{18}$OP trajectories, the first three factors explain 86% of the variance. Factor 1 explains 43% of the variance and is similarly dominated by altitude ($-0.98$), pressure (+0.93), and temperature (+0.89; Figure 7b). Factor 2 accounts for 24% of the variance and is dominated by latitude ($-0.98$) and longitude (+0.99), while Factor 3 is dominated by precipitation (+0.81) and relative humidity (+0.82), accounting for 19% of the total variance (Figure 7b).

To summarize, the dominant environmental factor along the storm track, in both the high and low $\delta^{18}$OP events, relates to trajectory altitude (i.e., height, temperature, and pressure). In support of this observation, we use regression analysis to explore the relations between $\delta^{18}$OP and all environmental variables extracted along each HYSPLIT modeled back-trajectory; the only significant relationship is found between event $\delta^{18}$OP and the corresponding mean trajectory altitude ($r^2 = 0.27$, $p < 0.05$, $n = 52$). Accordingly, mean trajectory altitude is higher for the low $\delta^{18}$OP events than during the high $\delta^{18}$OP events, with mean altitude values of 1,055 m asl (range: 1–3,911 m asl) and 675 m asl (range: 1–1,981 m asl), respectively. This "altitude effect" is typical in mountainous regions where $\delta^{18}$OP is linked with both altitude and temperature (Dansgaard, 1964). Moist air cools adiabatically and condenses as it ascends the windward side of a mountain, and this progressive removal of precipitation produces an isotopic gradient with altitude. The effect leads to precipitation with higher $\delta^{18}$OP at lower altitudes on the windward side and lower $\delta^{18}$OP at higher altitudes and on the downslope in the lee of the mountain.

Hence, Anchorage storm tracks associated with low $\delta^{18}$OP events are found to typically track into Prince William Sound and pass over the Chugach Mountains (>2,000 m) where they will precipitate out (Figure 6b). Alternatively, moist maritime air masses that do not experience significant vertical lifting, either orographically or induced by other natural processes (e.g., lifting in advance of a front, or convection), do not undergo such fractionation processes and will retain relatively high $\delta^{18}$OP values. In the case of high $\delta^{18}$OP events, this precipitation primarily derives from trajectories that funnel northeastward up Cook Inlet and do not encounter any major terrain barriers (Figure 6a). This localized effect and divergence in storm track $\delta^{18}$OP is also manifested at the continental scale. For instance, as Pacific moisture is transported from west to east across continental North America, orographic depletion in $^{18}$O is manifested by up to 20% from coastal California and Oregon to the crest of the Rocky Mountains; whereas moisture flow from the Gulf of Mexico into the American Midwest exhibits comparatively minor depletion due to the absence of high terrain features (Vachon et al., 2010a, 2010b; Welker, 2000; Winnick et al., 2014). These changes in $\delta^{18}$OP with mountain height and temperature are also governed by the microphysical processes through which precipitation is formed. For instance, lower $\delta^{18}$OP with increased altitude may reflect not only decreasing temperature but also the variable contributions and distinct isotopic signatures of riming.
of cloud liquid and vapor deposition onto snow (e.g., Aggarwal et al., 2016; Moore et al., 2016). In the case of the extreme high and low $\delta^{18}$O events, each precipitation sample represents a snow event, with mean snowfall amounts of 36.9 and 43.4 mm for the high and low $\delta^{18}$O events, respectively. However, mean SAT was +7.4 °C warmer during the high $\delta^{18}$O events ($\mu = 0.3 °C$, $sd = 6.2$) compared to the low $\delta^{18}$O events ($\mu = −7.1 °C$, $sd = 4.4$), possibly suggesting snow melt and/or regrowth by vapor deposition with higher $\delta^{18}$O during decent from the cloud base.

The second most dominant factor identified in the analysis relates to trajectory position (i.e., latitude and longitude). For the low $\delta^{18}$O events, the strong negative relation between latitude and longitude primarily reflects the cyclonic motion of winter systems as they are steered into the Gulf of Alaska (Figures 5 and 6; Mock et al., 1998). For instance, the low $\delta^{18}$O events typically track further eastward into the Gulf of Alaska before turning northward (i.e., increasing latitude) and steering west (i.e., decreasing longitude), thereby arriving at Anchorage from the southeast to northeast (Figure 6b). In contrast, longitude is not identified as a dominant variable for the high $\delta^{18}$O events that instead exhibit strong southwesterly flow. Lastly, precipitation is only a tertiary factor in both the high and low $\delta^{18}$O scenarios (Figure 7), indicating that Anchorage $\delta^{18}$O values do not reflect an “amount affect” whereby heavier rainfall results in more isotopic depletion of $^{18}$O (Dansgaard, 1964).

4.5. Synoptic Scale Influences on Anchorage $\delta^{18}$O/$\delta^2$H

Our analyses present a strong case for altitude and temperature effects on Anchorage $\delta^{18}$O along individual storm tracks. To understand what dictates intraseasonal variation in storm trajectory position at the synoptic
scale, we utilize the NCEP NARR data set for a regional estimate of the state of the atmosphere during extreme high and low $\delta^{18}O_P$ events (Mesinger et al., 2006). The terrain used in the model is roughly equivalent to one grid point every 32 km. While the mountains surrounding Anchorage are not sufficiently resolved to allow extraction of detailed local-scale information, the model does allow an assessment of major synoptic features and trends in key meteorological fields.

NARR daily composite maps based on the top 10 highest and lowest winter $\delta^{18}O_P$ events are shown in Figure 8 (Table S1). For the low $\delta^{18}O_P$ events, 850 mb geopotential heights indicate a low pressure centered over the northern Gulf of Alaska and Kenai Peninsula (Figure 8a). The net result is strong winds advected southward across the Bering Sea and into the Gulf of Alaska, where cyclonic flow steers these winds into Anchorage from the north to northeast (Figure 8b). The vapor flux profile indicates how moisture in this synoptic pattern is advected directly into the southeast Alaskan coastline along the eastern Gulf, although relatively high amounts also converge in Prince William Sound (e.g., in Whittier; Figure 8c). The case provides a strong example of the long-range transport mechanisms leading to isotopic depletion of low-level moisture as it passes over the Chugach Mountain barrier. Additionally, although Alaska’s high elevation Brooks and Alaska Ranges typically shelter Anchorage from incursions of northerly Arctic air masses, the synoptic picture for low $\delta^{18}O_P$ events highlights how Arctic air masses can travel considerable distances to southern Alaska via the Bering Strait (Figure 8b).

For the high $\delta^{18}O_P$ winter events, a different synoptic pattern emerges: The 850-mb heights show an elongated low extending from the western Aleutian Islands northeastward into the Bering Sea (Figure 8d). The net result is strong southwesterlies that track along the Aleutian Islands and into the Bering Sea and Cook Inlet (Figure 8e). These synoptic systems move from the west toward southern Alaska and advect high amounts of water vapor directly into Cook Inlet (Figure 8f). The steep terrain gradients in the upper Inlet will force strong updrafts along the slopes and cause precipitation in Anchorage (e.g., Papineau & Holloway, 2011). In contrast to the low $\delta^{18}O_P$ scenario, this moisture does not encounter any major terrain barriers during its passage to Anchorage that would cause fractional removal of
heavy $^{18}$O isotopes, and hence, these events are characterized by precipitation with relatively high $\delta^{18}$Op values.

These new findings support recent efforts by Lachniet et al. (2016) to constrain the climatic and physiographic controls on $\delta^{18}$O/$^2$H values of Alaskan stream waters. In contrast to their observed state-wide lowering isotope trend with distance from the coast, anomalously high stream $\delta^{18}$O/$^2$H values in the Susitna Valley—to the northwest of Anchorage—were hypothesized to reflect the funneling of storms through topographic lows (Lachniet et al., 2016). We find that under certain synoptic conditions, moisture can be advected through Cook Inlet and further inland (Figure 8f), and that these precipitation events will be characterized by relatively high $\delta^{18}$Op values.

Lastly, these synoptic mechanisms may account for the observed positive correlation between Anchorage $\delta^{18}$Op and maximum sea ice extent in Cook Inlet (Figure S3). For instance, when the prevailing atmospheric flow is zonal (higher $\delta^{18}$Op), southwesterly winds may help contain more sea ice in the Upper Cook Inlet. Conversely during prevailing cyclonic flow (lower $\delta^{18}$Op), the north and northeast trending winds are typically found to move ice down Cook Inlet to areas of melting (Mulherin et al., 2001).

### 4.6. Application to Paleoisotope Studies

Our analyses demonstrate how Anchorage $\delta^{18}$Op values are tightly coupled to local weather changes reflecting the passage of air masses with different transport paths and rainout histories. The prevailing synoptic-scale circulation patterns that drive these events are known to fluctuate on diurnal to interannual and decadal time scales, and these changes are well-expressed in regional climate indices such as the NPI, PDO, and the Pacific-North American pattern (Mantua & Hare, 2002; Trenberth & Hurrell, 1994; Wallace & Gutzler, 1981). Accordingly, several paleoisotope archives from Alaska and the Yukon have been interpreted to reflect atmospheric circulation driven changes in moisture source and/or storm trajectory (Anderson et al., 2005; Bailey et al., 2018; Jones et al., 2014) and in some instances exhibit significant correlations with instrumental climate indices over the past 100 years (Bailey et al., 2015; Klein et al., 2015).

There is no statistically significant relationship between the time series of Anchorage $\delta^{18}$Op, $\delta^2$H, or $d$-excess and either the NPI, Pacific-North American, or PDO between 2005 and 2018. However, our analyses do identify the broad synoptic-scale SLP features that dictate prevailing storm track locations that drive precipitation $\delta^{18}$Op in southern Alaska (Figure 8). This modern understanding of the water isotope cycle offers a previously unavailable empirical basis for interpreting $\delta^{18}$O recorded in local proxy archives of paleoprecipitation, such as in glacial ice, tree-rings, and lake sediments. For instance, higher or lower $\delta^{18}$O in the past may similarly reflect shifts between prevailing westerly or northeasterly storm systems, which in turn may coincide with warmer/wetter or cooler/drier conditions. Evaluating new and existing proxy isotope data in this light may facilitate a better understanding of the climate mechanisms influencing past $\delta^{18}$O variability, as well as elucidate some apparent disparities observed among the existing regional paleoisotope data sets (Kaufman et al., 2016).

Wherein the majority of Alaska paleoisotope studies relies on the regionally sparse Global Network of Isotopes in Precipitation network for modern meteoric $\delta^{18}$O and $\delta^2$H data, an alternative resource is the Online Isotopes in Precipitation Calculator (OIPC; Bowen, 2018). There is strong agreement between our measured mean annual isotope values and OIPC-modeled values for Anchorage (Figure S4); both exhibit a mean annual value of $-16.6‰$ for $\delta^{18}$Op, with a 0.5‰ difference between measured and modeled mean annual $\delta^2$H (Bowen & Revenaugh, 2003; Welker, 2000). Yet on the monthly scale, the OIPC slightly overestimates fall $\delta^{18}$Op values (i.e., higher) and underestimates $\delta^{18}$Op in spring (i.e., lower; Bowen, 2018; Bowen et al., 2005; Figure S4). These disparities likely reflect the inability of the model to capture finer local-scale processes such as the seasonal shifts in storm trajectory and topographic complexities we identify.

### 5. Conclusions

This new event-based $\delta^{18}$Op and $\delta^2$H data set from south-central Alaska fills a critical gap in the North American precipitation isotope network. We find that while the annual isotope cycle follows the well-established annual cycle for middle and high latitudes (Rozanski et al., 1993; Welker, 2000), only $\sim$30% is
explained by local temperature effects. Our observed temperature-δ¹⁸O slope of 0.31‰/°C is consistent with empirical δ¹⁸O data from northern Alaska that exhibit a slope of 0.35‰/°C (Klein et al., 2015). Based on these observations, we propose a regional temperature-δ¹⁸O relationship of −0.33‰/°C to be more appropriate for mainland Alaska than the global spatial average of 0.67‰/°C (Dansgaard, 1964). However, the relatively weak relationship between local SAT and δ¹⁸O/δ¹⁵N means we emphasize caution over the use of south-Alaskan isotopic archives as paleo-temperature records.

Instead, we demonstrate how isotopic variability in south-central Alaska is tightly coupled to synoptic atmospheric circulation processes that drive changes in the passage of air masses with different moisture sources, transport, and rainout histories. Specifically, meridional storm systems approaching Anchorage from either the Gulf of Alaska or the Arctic have relatively low δ¹⁸O and δ¹⁵N values due to the progressive cooling and removal of precipitation as it condenses with altitude over the Chugach Mountain Range; whereas zonally derived moisture from the North Pacific and/or Bering Sea retains relatively high δ¹⁸O and δ¹⁵N values. These new data and analyses contribute a previously unavailable empirical basis for interpreting δ¹⁸O recorded in local and regional precipitation archives. Future studies combining isotopic measurements of water vapor and/or temperature profiles at multiple atmospheric levels would help to reveal the full range of processes contributing to the resulting δ¹⁸O and δ¹⁵N characteristics.

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