Subseasonal Variations of Stable Isotopes in Tropical Andean Precipitation

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ABSTRACT
The tropical Andes of southern Peru and northern Bolivia have several major
mountain summits suitable for ice core paleoclimatic investigations. However,
incomplete understanding of the controls on the isotopic (δD, δ¹⁸O) composi-
tion of precipitation and a paucity of field observations in this region continue
to limit ice-core based paleoclimate reconstructions. This study examines four
years of daily observations of δD and δ¹⁸O in precipitation from a citizen sci-
entist network on the northeastern margin of the Altiplano, to identify controls
on the subseasonal spatiotemporal variability in δ¹⁸O during the wet season
(November to April). These data provide new insights into modern δ¹⁸O vari-
ability at high spatial and temporal scales. We identify a regionally coherent
subseasonal signal in precipitation δ¹⁸O featuring alternating periods of high
and low δ¹⁸O of 9- to 27-day duration. This signal reflects variability in precip-
itation delivery driven by synoptic conditions, and closely relates to variations
in the strength of the South American Low Level Jet and moisture availability
over the study area. The annual layer of snowpack on the Quelccaya Ice Cap
observed in the subsequent dry season retains this subseasonal signal, allowing
the development of a snow-pit age model based on precipitation δ¹⁸O mea-
surements, and demonstrating how synoptic variability is transmitted from the
atmosphere to mountaintop snowpacks along the Altiplano’s eastern margin.
This result improves our understanding of the hydrometeorological processes
governing δ¹⁸O and δD in tropical Andean precipitation, and has implications
for improving paleoclimate reconstructions from tropical Andean ice cores
and other paleoclimate records.
1. Introduction

With mean summit elevations in excess of 4,000 m (Garreaud, 2009), the tropical Andes form a significant barrier to atmospheric flow and play a large role in modulating the weather and climate of western South America (Insel et al. 2010). In southern Peru and northern Bolivia (12-18° S) the Andes reach their widest point, the Altiplano: a high plateau with a NW-SE orientation, bordered by cordilleras on both eastern and western flanks. This study focuses on two heavily glacierized ranges on the northeastern margin of the Altiplano, the Cordillera Vilcanota and Cordillera Real (Fig. 1). The Cordillera Vilcanota is home to the Quelccaya Ice Cap, the world’s largest tropical glacier and the site of one of the most important ice core records in the tropical Andes. Ice cores extracted from Quelccaya in 1983 and again in 2003 are annually resolvable for up to 1,500 years and have been used to reconstruct past changes in Pacific sea surface temperatures, migration of the Intertropical Convergence Zone and conditions during the Little Ice Age (Thompson et al. 2013). In the Cordillera Real, an ice core extracted from Nevado Illimani in 1999 dates back 18,000 years and contains information about tropical Andean climate changes that occurred during the transition from the Last Glacial Stage to the Holocene (Ramirez et al. 2003). Long-term climate records derived from ice cores such as these can improve understanding of climate dynamics and enable the evaluation of possible hydroclimatic responses to past- and modern-day climate forcings. In the tropical Andes, these records are particularly valuable because of their potential to record the El Niño Southern Oscillation (ENSO) - the leading mode of South American climate variability, and because they provide historical climate data in an otherwise data-sparse region (Vimeux et al. 2009). However, significant uncertainties remain in the interpretation of the climate signals recorded in these ice cores (Vimeux et al. 2009), and this data archive is now in danger of
being lost forever due to rapid ablation (Rabatel et al. 2013; Vuille et al. 2017; Thompson et al. 2017).

The key uncertainty addressed in the present study relates to the interpretation of stable water isotopes preserved in tropical Andean ice cores. Stable water isotopes in ice cores are applied in paleoclimate reconstructions based on the assumption that they are closely related to the isotopic composition of precipitation at the time, after taking into account possible post-depositional modification. In polar regions, a robust relationship between surface temperature and the $\text{^{18}O}$ concentration in precipitation allows for reconstructions of historical temperature. Thompson (2000) suggested that a similar relationship might be used to make temperature reconstructions from tropical ice cores. However, in tropical regions, different meteorological regimes and moisture transport processes complicate this relationship, and it not supported by observations. In fact, observations in the tropics suggest a stronger relationship between the isotopic composition of precipitation and precipitation amount, known as the ‘amount effect’ (Dansgaard 1964; Rozanski et al. 1993; Vimeux et al. 2005). Nonetheless, the isotope record from Quelccaya has been incorporated into multi-proxy global temperature reconstructions (Mann and Jones 2003). Hoffmann et al. (2003) show that coherent isotopic profiles from four Andean glaciers, including Quelccaya, appear to relate to precipitation variability over the Amazon basin, however, understanding of the dynamics of this relationship remains limited because of the complexity of the hydrological cycle in the Andes (Vimeux et al. 2009). To fully understand the hydroclimate signal preserved by isotopic tracers in ice cores, a better understanding of modern-day, location specific, controls on the isotopic composition of precipitation, and how this signal is retained in snow and ice, is needed.

In addition, with recent advances in ice core sampling technology, there are emerging opportunities to analyze new cores at an unprecedented resolution. For example, laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS) enables sampling of major and trace elements at
µm resolution (e.g. Mayewski et al. 2014; Sneed et al. 2015), and a novel sampling technique allows mm resolution slices of ice to be analyzed for isotopic composition (Mariusz Potocki, personal communication, 2018). These innovations create the potential to generate paleoclimate reconstructions of subseasonal climate variability from tropical Andean ice cores, allowing the assessment of the timing and precursors to abrupt climate changes, and for the reconstruction of leading modes of subseasonal climate variability at both continental scale (such as the South American See-Saw; González and Vera 2009), and global scale (such as the Madden-Julian Oscillation; Madden and Julian 1971). There are ongoing projects to extract new cores from both of the cordilleras discussed in this study that will be sampled using this new technology. To be able to interpret these data, it is essential to understand the controls on modern-day subseasonal spatiotemporal variability of precipitation isotopes, and whether or not these signals are preserved in the ice.

In response to these current limitations and opportunities in the analysis of tropical Andean ice cores, the goals of this study are to: 1) to identify subseasonal signals of isotopes in precipitation on the northeastern flank of the Altiplano, and their spatiotemporal characteristics, 2) to determine the dominant meteorological controls on these signals, and 3) to investigate how they are retained in the annual layer snowpack on tropical Andean glaciers. This work will be applicable in the analysis and interpretation of subseasonal isotope signals from new and existing ice cores in this region, and in isotope hydrology studies.

2. Background

a. Overview of isotope systematics

The two most common naturally occurring stable water isotopologues $^1\text{H}_2^1\text{H}_2^1\text{O}$ and $^2\text{H}_2^1\text{O}$ have a lower diffusivity and a higher saturation vapor pressure than the lighter and more common
\(^1\text{H}_2\text{O}\), resulting in isotopic fractionation during evaporation and condensation processes. For this reason, the isotopic composition of precipitation is an important hydrological tracer that encodes information about the source of the precipitation and its evaporation and condensation history. Water isotope composition is generally stated in parts per mille (‰) as relative deviations with respect to a reference value, in this case the Vienna Standard Mean Ocean Water (VSMOW), equation (1) (Gonfiantini 1978).

\[
\delta = \left( \frac{R_x}{R_{\text{VSMOW}}} - 1 \right) \times 10^3
\]  

(1)

\(R_x\) is the heavy-to-light isotope ratio of species \(x\) and \(R_{\text{VSMOW}}\) is the standard heavy-to-light isotope ratio of that species. Hereafter, \(\delta^{18}\text{O}\) refers to the relative deviation of \(^{18}\text{O}\) and \(\delta\text{D}\) to the relative deviation of \(^2\text{H}\). The amount of isotopic fractionation that occurs during evaporation and condensation depends on the initial isotopic composition of the reservoir as well as local temperature, relative humidity (for evaporative processes), and whether the process is taking place under equilibrium conditions (Friedman et al. 1962; Dansgaard 1964; Risi et al. 2008). Therefore, the dominant control on the isotopic composition of precipitation is often location specific.

b. Weather and climate of the northeastern Altiplano

In order to understand the controls on the isotopic composition of precipitation in the Cordillera Vilcanota and Cordillera Real it is therefore essential to understand the local and regional meteorology. These ranges experience distinct wet (Nov-Apr) and dry (May-Oct) seasons. Water vapor that precipitates over the ranges primarily originates over the Atlantic Ocean and undergoes recycling due to convective precipitation and evapotranspiration over the Amazon (Grootes et al. 1989). Precipitation is typically associated with upstream air trajectories from the NNW and weak (approximately 1 m s\(^{-1}\)) low-level moisture advection (Vimeux et al. 2005; Insel et al. 2013;
Perry et al. 2014). These trajectories are broadly parallel with the South American Low Level Jet (SALLJ, Fig. 1), a southeast-flowing low-level air current that reaches maximum velocity close to the 850 hPa level east of the Andes in Bolivia (Vera et al. 2006). An active SALLJ enhances moist advection up the east Andean slope, particularly along NE-SW oriented valleys (Junquas et al. 2017), to the Altiplano where moisture influx from the Amazon basin is the limiting factor to precipitation (Garreaud 1999; Garreaud et al. 2003). During the wet season, precipitation is episodic, with alternating periods of wet and dry conditions lasting 5-20 days (Lenters and Cook 1997; Garreaud 1999, 2000). Despite heterogeneous terrain, these episodes are regionally coherent (Garreaud 2000; Perry et al. 2014; Hurley et al. 2015) and appear to relate to a number of different forcing mechanisms including propagating extratropical cyclones, changes in upper level zonal flow, and cold air incursions (CAIs) (Lenters and Cook 1999). On a diurnal scale, precipitation during the wet season is typically bimodal, featuring precipitation peaks around local midnight and in the afternoon around 16:00 LST (Perry et al. 2014, 2017; Chavez and Takahashi 2017; Junquas et al. 2017). The late afternoon events are convective in nature and result from instability due to daytime heating of the lower atmosphere. The meteorological forcing associated with the nocturnal peak is an area of current research, but observations suggest that these events are regionally coherent and stratiform in structure (Perry et al. 2017; Endries et al. 2018), indicating a greater degree of organization.

c. Current understanding of the controls on isotopes in tropical Andean precipitation

On both interannual and seasonal timescales, $\delta^{18}O$ in tropical precipitation is observed to correlate with changes in precipitation amount, referred to as the ‘amount effect’ (Dansgaard 1964; Rozanski et al. 1993). However, the nature of this amount effect and its physical mechanism remains a topic of ongoing research. One interpretation of the physical mechanism for the amount
effect is that the successive isotopic depletion of precipitation results from a feedback mechanism, whereby the initial isotopic composition of the vapor depends on the isotopic composition of precipitation from the previous storm, reintroduced into the atmosphere by re-evaporation (Risi et al. 2008; Field et al. 2010). More frequent and more intense convective events will thus cause progressive depletion of low-level moisture. Another interpretation is that this effect relates to the degree of organization of storms (Kurita 2013). In the tropical Andes, several studies find the relationship between $\delta^{18}$O and local precipitation amount to be weak (Vimeux et al. 2005; Insel et al. 2013; Fiorella et al. 2015), with some studies finding more robust relationships with upwind precipitation amounts (e.g. Vuille and Werner 2005; Vimeux et al. 2005; Fiorella et al. 2015).

Insel et al. (2013), related central Andean $\delta^{18}$O variability to the strength of the SALLJ, which plays an active role in transporting moisture to the study area (Junquas et al. 2017). Although there is clearly a relationship between isotopic composition and precipitation amount, a limitation shared by each of the tropical Andean isotope studies discussed above is that they use monthly precipitation samples that are insufficient to capture isotopic variability associated with the episodic 5- to 20-day wet and dry periods. In contrast, Vimeux et al. (2011) use one year of event-based precipitation samples from the Zongo valley in the Cordillera Real and identified subseasonal $\delta$D variability that is consistent between observation sites despite large differences in elevation and local precipitation amounts. These variations appear to coincide with variations in the position of the South American Convergence Zone (SACZ: a band of convection in eastern Brazil associated with the South American Summer Monsoon, Fig. 1), supporting the findings of an earlier modeling study (Sturm et al. 2007). However, because these data come from a single valley, this result cannot be generalized for the broader region. The present study builds on the findings of Vimeux et al. (2011) by using daily precipitation samples collected from a broader area along the northeastern edge of the Altiplano over multi-year period.
3. Data and Methods

a. Precipitation Isotope Data

Citizen scientist observers, recruited for the project in the Cordillera Vilcanota and Cordillera Real study regions (Fig. 2), measured daily precipitation amount and collected samples for isotopic analysis. Sampling began at two sites in 2013, with others added in 2016 and 2017 (see table inset in Fig. 2). Each site was equipped with a 10 cm diameter measuring cylinder rain gauge with a funnel to limit evaporation. Trained observers took daily liquid-equivalent precipitation measurements at 07 LST (12 UTC in Peru, 11 UTC in Bolivia) using established protocols (Cifelli et al. 2005), and when there was more than 0.5 mm precipitation, a sample was directly transferred and sealed in an 8 mm glass vial for isotopic analysis. In the event of snow, the funnel was removed from the gauge and the snow was melted in a basin of warm water before sampling. Samples were retrieved annually and the University of Arkansas Stable Isotope Laboratory analyzed them for isotopic composition ($\delta^D$, $\delta^{18}O$) using a thermal conversion elemental analyzer (TC/EA) coupled to a continuous flow isotope ratio mass spectrometer (CF-IRMS) (see Gehre et al. 2004). The data was normalized using USGS 48, UASIL L and UASOL R standards (Nelson 2000) and the approximate uncertainty in laboratory results is $\pm 1 \, ^{\circ}\!\!\!\!\!\!_{\circ}^\circ$ for $\delta^D$ and $\pm 0.5 \, ^{\circ}\!\!\!\!\!\!_{\circ}^\circ$ for $\delta^{18}O$. This complete dataset is available online (Appalachian State University 2018). The relationship between $\delta^D$ and $\delta^{18}O$ is described by the deuterium excess parameter (d-excess = $\delta^D - 8 \, \delta^{18}O$) (Dansgaard 1964). The lower diffusivity of $^1H_2^{18}O$ compared to $^2H_2^{16}O$ means that $\delta^{18}O$ is relatively more sensitive to kinetic effects than $\delta^D$, resulting in changes in d-excess. d-excess can therefore be used to detect, and to some extent diagnose, non-equilibrium processes and help to delineate the factors that contribute to differences in the isotopic composition of precipitation. Small or negative d-excess can result from below-cloud evaporation, and higher d-excess values can result from non-equilibrium condensation.
during the growth of ice crystals (Jouzel and Merlivat 1984) or continental moisture recycling (Gat and Matsui 1991). The combined measurement uncertainties in $\delta D$ and $\delta^{18}O$ (above) result in a measurement uncertainty of $\pm 4.1$ $\%$ for $d$-excess. Tests performed during fieldwork indicated that samples could become enriched in heavy isotopologues after small afternoon precipitation events, due to evaporation in a rain gauge prior to collection. Precipitation samples potentially modified in this manner are identifiable by low ($< 0$ $\%$) $d$-excess values, and were excluded from analysis in the present study.

b. Isotope Dataset Processing

To assess the spatial variability of $\delta^{18}O$, each site is compared pairwise by correlating $\delta^{18}O$ measurements on common days with measured precipitation (Table 1.) To assess temporal variability in $\delta^{18}O$ a 15-day moving average of $\delta^{18}O$ is calculated. A 15-day window is chosen because it is short enough to capture subseasonal fluctuations in $\delta^{18}O$ but long enough to fill the majority of data gaps associated with zero-precipitation dates. The seasonal depletion cycle (90-day moving average) is subtracted from the 15-day moving average each year to identify subseasonal $\delta^{18}O$ anomalies.

We define positive (negative) $\delta^{18}O$ anomalies as periods during the de-trended subseasonal $\delta^{18}O$ signal is above the 75th percentile (below the 25th percentile) for at least five consecutive days. All $\delta^{18}O$ averages referred to in this study are weighted by precipitation amount.

c. Relationship with Synoptic Circulations

To examine associations with synoptic circulations, we use data from ERA-Interim, a global atmospheric reanalysis model with 80 km horizontal resolution (Dee et al. 2011). We compare the subseasonal $\delta^{18}O$ signals and selected variables from ERA-Interim averaged over the region shown in Fig. 2 (12.0-18.5 ° S, 65.0-74.0 ° W). This region is selected to encompass all observer sites,
it should be noted that a disadvantage of this region is that it includes parts of the Amazon basin and coastal Altiplano with highly dissimilar climatological regimes, these areas are not considered independently in this study. The following variables from ERA-Interim are tested at both 500 hPa (near surface) and 250hPa (upper level): relative humidity (r), zonal wind (u), meridional wind (v), geopotential height (z) and temperature (t). For each variable significantly correlated (p < 0.01) with the subseasonal $\delta^{18}$O signal for at least three wet seasons, spatial difference plots between the positive and negative $\delta^{18}$O anomalies provide further insight into the relationship between isotopes and synoptic-scale meteorological conditions. Additionally, the possible impact of CAIs on the subseasonal $\delta^{18}$O signal is assessed. CAI event dates are determined using data from ERA-Interim and the method described in Garreaud (2000); when the sea level pressure in a 5 x 5° grid box centered at 25° S, 57.5° W is greater than 1015 hPa ,and the sea level pressure tendency is within the top 10% of the seasonal frequency distribution.

d. Signal Preservation in the Annual Layer Snow-pack

The propensity of deposited snow to retain the subseasonal isotopic signal observed in precipitation is assessed using isotope samples from a 2.16 m deep snowpit at 5670 m above sea level (ASL) on Quelccaya, sampled on 16 July 2017 (location shown in Fig. 1). A vertical profile created using samples of 8 mm of liquid water equivalent (LWE) were collected until a visible dust horizon and ice layer demarcated the base of the annual layer. Samples were allowed melt in sealed plastic bags (with air removed) and then transferred into glass vials to be processed as described in section 3a.

To compare the isotopic profile of the snowpit to isotopic signal observed in precipitation, it is necessary to develop an age model for the snowpit. To approximate an age model, we use daily snow height and precipitation measurements from an automatic weather station on Quelccaya (Appalachian State University, unpublished data), further details of this weather station are available...
in Perry et al. (2017). The first day of continuous wet season accumulation on Quelccaya (the start of the snowpit) is determined using snow height measurements. Then, the ratio of the total precipitation recorded by the Quelccaya pluviometer (between the estimated start date and the snowpit sampling date) to the total liquid water equivalent of the snowpit is used to estimate daily snowpit accumulation. This approach assumes no loss of precipitation from the snowpack on Quelccaya due to ablation or sublimation during the wet season. The age model allows a broad comparison between the isotopic profile of the annual layer snowpit on Quelccaya and the isotopic signal observed in precipitation.

4. Results

a. Overview of Patterns in Precipitation $\delta^{18}O$

All raw precipitation measurements and isotope data are shown in Fig. 3. Because the majority of precipitation (and therefore the majority of sample collection) occurs from November to April, when comparing interannual variability we consider each hydrological year to start and finish in August; thereby capturing the entire wet season. Henceforth, when discussing the year 2017, for example, we refer to 1 August 2016 to 1 August 2017 (See Fig. 3). A breakdown of all data from individual sites in 2017 (when all 10 sites were active) is provided in the supplementary materials.

There is a strong linear relationship between $\delta^{18}O$ and $\delta D$ in all samples and all seasons (Fig. 4); the rest of this analysis therefore focuses on $\delta^{18}O$ for brevity. Each year, there is a gradual depletion of $^{18}O$ in precipitation during the wet season, with minimum $\delta^{18}O$ observed around March, followed by enrichment at the onset of the dry season (Fig. 3). On subseasonal timescales, there is high variability in $\delta^{18}O$ with large amplitude oscillations (>10 ‰) over 1 to 7 days. In addition, throughout the wet season, there are periods of higher and lower $\delta^{18}O$ that last between 6
and 30 days. Subseasonal variations in δ¹⁸O exhibit strong synchronicity at all sites despite large differences in local precipitation amount; this is especially apparent in 2017, where there are data from a larger number of stations that extend along a 500 km transect and range in elevation between 3,350-5,050 m ASL (Fig. 5). The range of δ¹⁸O values in the first three years is -30 ‰ to +5 ‰, however in 2017 there are several samples from multiple sites recording δ¹⁸O < -35 ‰. During the exceptionally strong El Niño event of 2015-2016 (Xue and Kumar 2017), the annually averaged δ¹⁸O in precipitation was 5.6 ‰ higher than the average of the other three years (Fig. 3).

b. d-Excess

For the majority of samples at all sites, the relationship between δD and δ¹⁸O is very close to the Global Meteoric Water Line (GMWL): the globally averaged relationship between δD and δ¹⁸O in meteoric waters that have not undergone surface evaporation (Craig 1961) (Fig. 4). Deviations from the GMWL can occur due to kinetic processes during precipitation (for example, sub-cloud evaporation), kinetic processes along the moisture transport pathway (for example, continental moisture recycling) or from changes in evaporation conditions at the moisture source (Pfahl and Sodemann 2014). Local meteoric water lines for each individual site, and for the dry season (JJA) samples compared to the wet season (DJF) samples are provided in the supplementary materials. Highest d-excess values (> 30 ‰) occur during the peak of the wet season (Fig. 3), and are associated with precipitation that is highly depleted in ¹⁸O. Over several multiday periods, high d-excess is observed at certain stations in the Cordillera Vilcanota (for example Pucarumi and Sallayoc) but not at others that are only 25 km away (i.e. Murmurani, Chillca). This suggests that these high d-excess values may result from local effects, as discussed in Section 5b.
c. Spatial Variability

Despite a 1,700 m difference in elevation between the highest and lowest sites, no relationship is evident between elevation and $\delta^{18}O$ in 2017 (Fig. 5). There is also no evident relationship between local precipitation amount and $\delta^{18}O$ (Fig. 5). The range of mean $\delta^{18}O$ between sites is 4.7 ‰ and standard deviation is 1.0 ‰ suggesting a high degree of agreement between sites despite the differences in elevation, daily precipitation, and horizontal separation. There are strong, statistically significant ($p < 0.01$) correlations between $\delta^{18}O$ measured at every site (Table 1). Particularly noteworthy is the consistency of the correlation coefficients among sites within the Cordillera Vilcanota and between the Cordillera Vilcanota and the Cordillera Real. This result clearly demonstrates that subseasonal variations of isotopes in precipitation along the north-eastern margin on the Altiplano are recording a common signal.

d. Temporal Variability

Since subseasonal variations of $\delta^{18}O$ in precipitation are demonstrated to be coherent across the Cordillera Real and the Cordillera Vilcanota, a region-wide $\delta^{18}O$ mean is calculated for the 2017 wet season by taking the mean 15-day $\delta^{18}O$ weighted by precipitation amount between all sites (Fig. 6a). There are three key modes of temporal variability apparent in these data. The first, characterized by gradual depletion of $^{18}O$ in precipitation during the wet season and enrichment at the onset of the dry season, is the seasonal cycle, well captured by a 90-day moving average (orange line, fig. 6a). The second consists of 6- to 30-day cycles of relatively more-depleted and relatively more-enriched periods that are clearly delineated by the 15-day moving average; this is temporally consistent with the recognized timescale of variations between wet and dry episodes on the Altiplano during the wet season (Lenters and Cook 1997; Garreaud 1999, 2000). The remaining variability relates to short-term ‘storm scale’ oscillations in $\delta^{18}O$. Although these
variations may represent an interesting meteorological signal, they are less consistent between sites, and it is likely that post-depositional processes will smooth out this signal in the snowpack; this study therefore focuses on the subseasonal signal (15-day moving average), which is retained in snowpack into the subsequent dry season. Subtracting the seasonal cycle from the 15-day moving average isolates this subseasonal signal, yielding a time series of subseasonal $\delta^{18}O$ anomalies (Fig. 6b). Due to the lack of precipitation in the dry season, there are only enough data to define the subseasonal $\delta^{18}O$ anomalies between November and May, hence the rest of this study will focus on subseasonal variations of $\delta^{18}O$ during this period. Assuming that the spatial coherency of the $\delta^{18}O$ variations is true in all years, repeating this analysis for 2014-2016 identifies the timing of positive and negative $\delta^{18}O$ anomalies along the NE margin of the Altiplano for each year, despite fewer sampling locations (Fig. 7). This method identifies two or three negative $\delta^{18}O$ anomalies each year lasting between 9 and 24 days. In 2014 and 2016 there is one positive anomaly, with two in 2015 and four in 2017. The positive $\delta^{18}O$ anomalies last between 9 and 27 days. The amplitude of $\delta^{18}O$ anomalies varies year to year, with the largest amplitude anomalies occurring in 2017.

e. Relationship with Region-wide Meteorology and Continental Scale Circulation

The spatiotemporal consistency of $\delta^{18}O$ across the observational domain observed in 2017 suggests that regionally coherent meteorological conditions are responsible for the anomalies identified in Section 3d. To identify the key meteorological variables that are associated with the $\delta^{18}O$ anomaly signal we examined five domain-averaged ERA-Interim variables ($z$: geopotential height, $t$: temperature, $u$: zonal wind, $v$: meridional wind and $r$: relative humidity) at two pressure levels (250 hPa and 500 hPa). Correlation coefficients calculated between the $\delta^{18}O$ anomaly time series and each variable processed in the same way (i.e., 15-day moving averages minus the seasonal cycle) are presented in Table 3, which also includes correlations between the $\delta^{18}O$ anomaly series
and the precipitation anomaly series calculated from the mean precipitation between all observer
sites. The strongest correlations are with 500 hPa and 250 hPa relative humidity, mean observed
precipitation, 250 hPa zonal wind and 500 hPa meridional wind (Table 3). These variables were
examined for spatial differences in synoptic conditions between the positive and negative δ¹⁸O
anomalies each year (Fig. 8).

On average, the wet season positive δ¹⁸O anomalies that occurred in 2014-2017 coincide with
significantly reduced relative humidity at 500 hPa and 250 hPa over west-central South America
relative to the negative δ¹⁸O anomalies (Fig. 8a-b). At upper levels (250 hPa), the negative δ¹⁸O
anomalies are typically associated with strengthened easterly winds over the Amazon basin and
increased geopotential heights over southern South America, likely associated with more southward
placement of the Bolivian High (Lenters and Cook 1997, Vuille 1999) (Fig. 8c-d). In the 500
and 850 hPa meridional wind fields, the negative δ¹⁸O anomalies are associated with increased
northerly winds, with the largest differences in 850 hPa meridional wind centered on eastern Bolivia
(Fig. 8f), aligned along the axis of the SALLJ, suggesting that negative (positive) anomalies are
typically associated with stronger (weaker) northerly winds to the east of the Andes.

To assess the possible influence of CAIs on the subseasonal isotope signal, we calculated the
dates of CAIs using the method outlined in Garreaud (2000), and plotted them against the δ¹⁸O
anomaly time series in Fig. 9. Over the four years in this study, there is no consistent response
to CAIs observable in the subseasonal isotope signal, with some CAI dates preceding both high
δ¹⁸O and low δ¹⁸O episodes. However, strong relationships are apparent between the strength of
the SALLJ at Santa Cruz, domain averaged 500 hPa relative humidity, and δ¹⁸O (Fig. 9).

At the continental scale, there is a significant positive correlation between the δ¹⁸O anomaly
signal and detrended 500 hPa relative humidity over the north-eastern Brazil (Fig. 10) for each
year sampled in the present study. Additionally, there are negative correlations between these same variables over NE Bolivia. These results are consistent with the findings of Vimeux et al. (2011).

f. Subseasonal Isotopic Signal Representation in Annual Layer Snowpack

The potential for subseasonal isotopic signals to become useful diagnostics in ice core paleoclimatic investigations requires that the signal be retained in annual layer snowpacks. Based on snow height measurements from an automatic weather station on Quelccaya (Appalachian State University, unpublished data), the period of continuous wet season accumulation on Quelccaya approximately begins on 11 Nov 2016. The subseasonal $\delta^{18}$O signal in precipitation from this date exhibits excellent pattern matching to the $\delta^{18}$O profile from the annual layer snowpit on Quelccaya (Fig. 11). The average snowpit $\delta^{18}$O is 3‰ lower than the average $\delta^{18}$O from precipitation; although this could be attributed to the observed ‘altitude effect’ whereby precipitation falling at higher elevations is typically more depleted (Gonfiantini et al. 2001), it is odd that such a relationship is not observed in the precipitation data. Another possible explanation for this is that the majority of the observer sites receive liquid precipitation, whereas solid precipitation (accumulating on Quelccaya) is typically more depleted in 18O because isotopic exchange with boundary layer moisture is limited, and because the equilibrium fractionation factor is larger between vapor and ice than it is between ice and liquid (Jouzel and Merlivat 1984).

5. Discussion

a. $\delta^{18}$O Characteristics in the Cordilleras Vilcanota and Real

The range of $\delta^{18}$O variations each year in this study is similar to the range of $\delta^{18}$O identified by previous studies in this region (e.g., Gonfiantini et al. 2001; Fiorella et al. 2015) with the exception of some strongly 18O-depleted events ($\delta^{18}$O < -35‰) that occurred during 2017. It is also very
close to the range of values predicted by Grootes et al. (1989). Higher $\delta^{18}$O in precipitation than predicted by Grootes et al. (1989) ($\delta^{18}$O > -4‰) may result from below-cloud evaporation on days when the boundary layer was particularly dry (Risi et al. 2008).

The 9-27 day subseasonal periods of higher and lower $\delta^{18}$O in precipitation superimposed upon the seasonal cycle are consistent with the Zongo Valley study (Vimeux et al. 2011), and with Hurley et al. (2016) who identify similar subseasonal variability in $\delta^{18}$O profiles from eight annual-layer snow pits on Quelccaya. This result suggests that subseasonal $\delta^{18}$O variations are a robust feature of precipitation in the eastern tropical Andes and are retained in annual-layer snowpacks.

Subseasonal variability is coherent across a large domain encompassing, at a minimum, all of our observation sites in both the Cordilleras Real and Vilcanota. This has implications for paleoclimate studies because it demonstrates that the isotopic signals in precipitation that are ultimately preserved in glacial ice have common controls. Therefore, under the assumption that circulation patterns are similar to the present day, the isotopic profiles from ice cores extracted from Quelccaya and from Illimani, for example, can be meaningfully compared. Such agreement between ice cores extracted from Quelccaya and Illimani is already observed on interannual time scales (Hoffmann et al. 2003).

b. Synoptic Controls on the Subseasonal Isotope Signal

The excellent spatial matching between the subseasonal $\delta^{18}$O signals at all locations implies that this signal is reflecting synoptic scale rather than local conditions. This finding is in agreement with previous studies that have demonstrated that the local amount effect in this region is weak for precipitation sampled on daily or longer timescales (Vimeux et al. 2011, Fiorella et al. 2015).

However, strong negative correlations between $\delta^{18}$O and regionally averaged precipitation and 500 hPa relative humidity suggest that this variability does reflect region-wide variations in precipitation amount. Additionally, the duration of positive and negative $\delta^{18}$O anomalies closely aligns
with the known duration of dry and wet periods in the NE Altiplano (6-30 days). This result strongly
supports that isotopes in tropical Andean precipitation are reflecting the regional ‘amount effect’ at
subseasonal timescales. Precipitation amount cannot explain all of the observed variability, how-
ever, and the question of what physical processes are controlling the amount effect here remains.
The synoptic setup that results in precipitation anomalies over the study area is not the same for
each positive and negative $\delta^{18}O$ anomaly; therefore interpreting them as an ‘amount effect’ alone is
likely to have large uncertainties. Of more use to improving paleoclimate reconstructions from ice
cores would be to relate the positive and negative $\delta^{18}O$ anomalies to the specific synoptic weather
systems associated with these precipitation anomalies, and the attendant continental-scale modes
of subseasonal variability. Lenters and Cook (1999) describe three synoptic patterns that can
result in positive precipitation anomalies on the Altiplano: a) propagating extratropical cyclones
that result in a strengthening of the SALLJ; b) a westerly shift of the SACZ and anomalous high
pressure over the central Amazon basin forcing a strengthening and southward shift of the Bolivian
High; and c) deep narrow bands of low pressure from the subtropics extending into tropical regions
along the eastern slopes of the Andes associated with CAIs. When the synoptic patterns associated
with individual negative $\delta^{18}O$ anomalies are examined (not shown) each anomaly appears to be
associated with either scenario a or scenario b. A strengthening and southward shift of the Bolivian
high is only observed in scenario b and is insufficient to explain all the $\delta^{18}O$ anomalies. However,
a strengthening of the SALLJ is a robust feature in each negative $\delta^{18}O$ anomaly; this is discussed
further in Section 5c.
Another way of distinguishing the synoptic setup associated with each $\delta^{18}O$ anomaly is to
consider d-excess in more detail. Several samples from 2017 had very high ($>30 \‰$) d-excess that
did not occur at all stations simultaneously and that do not appear to be related to station elevation
or moisture source trajectories. Most of these high d-excess values occur during the third negative
δ¹⁸O anomaly that appears to be associated with scenario a. This δ¹⁸O anomaly had weaker 500 hPa relative humidity anomalies compared to the other negative anomalies (both scenario b) despite being of a similar magnitude. A possible explanation is that this precipitation event was limited in spatial extent; a strong positive precipitation anomaly with limited spatial extent is suggestive of deep convection. Localized high d-excess can occur in the presence of deep convective storms as a result of very high local rain rates (Bony et al. 2008; Galewsky et al. 2016) or kinetic processes during ice crystal formation (Jouzel and Merlivat 1984). Verifying this is beyond the scope of the present study; it is worth noting that Vimeux et al. (2011) observed a similar event in the 2000 season (very low δD with high d-excess).

c. The Role of the South American Low Level Jet

The strengthening of the SALLJ directly to the east of the study site is a robust feature for all of the subseasonal negative δ¹⁸O anomalies in all years 2014-2017 (Fig. 9) despite other differences in the synoptic setups (discussed in section 5b). Conversely, the majority of positive anomalies are associated with a SALLJ that is anomalously weak, absent or reversed (by the passage of a CAI). For the few positive anomalies during which the SALLJ is present, the 500 hPa relative humidity directly to the east of the Altiplano is anomalously low. This result implies that SALLJ plays a key role in enhancing moisture transport from the western Amazon basin to the study region, by strengthening the upslope flow that advects moisture to mountains along northwest-oriented valleys (Marengo et al. 2002; Junquas et al. 2017). When there is sufficient moisture available, this results in the positive relative humidity anomalies that are associated with negative δ¹⁸O anomalies.

In addition, precipitation along the eastern edge of the Altiplano occurs under two key modes: afternoon convection and widespread nocturnal stratiform events (Perry et al. 2014, 2017; Chavez and Takahashi 2017; Endries et al. 2018). More organized precipitating systems with large
stratiform components are associated with decreased $\delta^{18}\text{O}$ in precipitation compared to localized convective storms (Kurita et al. 2011; Aggarwal et al. 2016). The marked difference between the isotopic composition in stratiform and convective precipitation types results from differences in cloud dynamics and hydrometeor formation. In convective storms, boundary layer moisture (more enriched in heavy isotopes) is entrained into the cloud by strong updrafts, whereas in stratiform precipitation, hydrometeors grow by vapor diffusion above the melting layer where moisture is relatively depleted in heavy isotopes, and below cloud subsidence may result in low $\delta^{18}\text{O}$ vapor transport to the sub-cloud layer. See Aggarwal et al. (2016) for a detailed description of this process. The SALLJ is strongest overnight (Garreaud and Wallace 1997; Marengo et al. 2004; Junquas et al. 2017), and Junquas et al. (2017) present evidence from numerical simulations that the upslope flow influenced by the presence of a strong SALLJ persists overnight. It is possible that this enhanced moisture transport combined with the stable nighttime atmosphere may mechanically produce precipitation more stratiform in structure, and hence with lower $\delta^{18}\text{O}$.

Relationships between positive precipitation anomalies and the strength of the SALLJ (e.g., Garreaud 1999; Lenters and Cook 1999; Junquas et al. 2017), and negative $\delta^{18}\text{O}$ anomalies and the strength of the SALLJ (e.g, Vimeux et al. 2011; Insel et al. 2013) have been previously suggested. Our results indicate that the presence and strength of the SALLJ is a key—and probably dominant—control on moisture availability and the type and extent of precipitation during the wet season that ultimately results in subseasonal $\delta^{18}\text{O}$ anomalies that are synchronous and coherent across a wide region. On a broader scale, the strength of the SALLJ relates to continental scale precipitation variability by association with the South American See-Saw dipole (González and Vera 2009) migration of the SACZ, this is discussed further in section 5e.
d. The Role of Cold Air Incursions

Lenters and Cook (1997) propose CAIs as the third mechanism to explain positive precipitation anomalies. More recent studies have found evidence that CAIs may modulate subseasonal variations of $\delta^{18}$O in the snowpack on Quelccaya (Hurley et al. 2015, 2016). However, on the subseasonal timescale that is the focus of the present study, none of the negative $\delta^{18}$O anomalies identified appear to result from CAIs, and we observe no response to CAIs in the 15-day moving average signal of isotope anomalies in precipitation. It is possible that CAIs do influence isotopes in precipitation on shorter timescales; however signals over shorter timescales are not observed in the annual-layer snowpit on Quelccaya (Fig. 11).

e. Relationship with Continental Scale Circulation Patterns

The controls on stable isotopes in precipitation using daily-resolution sampling over an entire season (1999-2000 hydrological year) were examined by Vimeux et al. (2011), who used observations from several sites in the Zongo valley near Huayna Potosí in the Cordillera Real (Fig. 1). They found that intraseasonal variations in $\delta D$ in the Zongo valley are associated with the South American See-Saw, a continental precipitation dipole related to the position of the SACZ, whereby higher (lower) $\delta D$ in precipitation in the Zongo valley coincides with enhanced (reduced) convection over north-eastern Brazil and reduced (enhanced) convection over the subtropical plains. The results of the present study are consistent with these findings and demonstrate that this correspondence is observed over a number of years and applies to a broader region, encompassing both the Cordillera Vilcanota and Cordillera Real. The South American See-Saw has a period of 15 to 20 days (Nogués-Paegle and Mo 1997), consistent with the leading mode of subseasonal variability in $\delta D$ identified by Vimeux et al. (2011), and of the same order of the as the subseasonal oscillations depicted in the present study. In one phase, the SACZ strengthens, precipitation over the subtropics
is reduced and the SALLJ weakens (González and Vera 2009) - associated with higher $\delta^{18}$O/ $\delta$D along the northeastern margin of the Altiplano (Vimeux et al. 2011 and the present study). In the opposite phase, precipitation in the subtropics increases associated with a strengthening of the SALLJ, increased moisture influx from the Amazon basin, and lower $\delta^{18}$O/ $\delta$D. The phase of the South American See-Saw dipole is potentially related to the Madden Julian Oscillation (Nogués-Paegle and Mo 1997; Paegle et al. 2000; Alvarez et al. 2016), implying that subseasonal variations in isotopes preserved in annual-layer snowpits in the Cordilleras Real and Vilcanota might record information about hemispheric tropical climate variability.

f. The 2015-2016 El Niño recorded in isotopes in precipitation

Although not a primary focus of this study, our isotope sampling period includes the strong El Niño year of 2015-16, providing the opportunity to compare isotopic variations between a strong El Niño and non-El Niño years. During 2015-16, there was a severe reduction in snow accumulation on Quelccaya and annually averaged $\delta^{18}$O in the annual-layer snowpack was higher by 5.2 ‰ (Thompson et al. 2017). The results of this study are consistent with this finding: the average $\delta^{18}$O in precipitation during 2016 is 5.6 ‰ higher than the average of the other three years. This establishes that the change in $\delta^{18}$O in the snowpack on Quelccaya resulted from more precipitation that was more enriched in $^{18}$O at the time of deposition rather than post-depositional enrichment. Despite this, we still observe subseasonal variations in $\delta^{18}$O that occur on similar timescale to the other years, suggesting that while the El Niño event affected the baseline isotopic composition of precipitation, as described by previous studies (e.g., Vuille et al. 2003), the processes driving the subseasonal variability are the same.
g. Retention of the Subseasonal Signal in Annual Layer Snowpack

The annual-layer snowpack on Quelccaya deposited in 2017 and measured in the subsequent dry season clearly retains the region-wide 15-day averaged $\delta^{18}$O signal observed in precipitation (Fig. 11). The isotopic profile therefore effectively records subseasonal changes in synoptic conditions and, in particular, changes in the strength of the SALLJ and in moisture availability in the Amazon lowlands directly east of the Altiplano. To apply the results of this study to interpret paleoclimate signals, it is necessary to consider how the processes that act together to generate the signal we observe today might have changed in the past, and how the subseasonal isotopic signal is retained and possibly modified in deeper firn and ice. Molecular diffusion during firnification smooths the isotopic signal over time and, in deep ice, even the seasonal isotopic signal cannot be identified (e.g., Ramirez et al. 2003). However, the isotopic profile from the ice core extracted from Quelccaya in 2003 exhibits annually resolvable resolution for 1,500 years (Thompson et al. 2006). Although subseasonal isotopic signals are diminished before they are interred in glacial ice (e.g., Thompson et al. 2017), some insoluble chemical tracers that fluctuate in response to the same climatic variability may remain in place within annual layering. How much subseasonal variability can be detected in both isotopes and chemical tracers from new ice cores extracted from both cordilleras, that will be sampled using new techniques, remains to be seen.

6. Conclusions

This study demonstrates that subseasonal variations in $\delta^{18}$O in precipitation in the Cordillera Vilcanota and the Cordillera Real are spatially coherent and reflect variability in synoptic and continental-scale meteorology. Periods of relatively high and low $\delta^{18}$O in precipitation during the wet season superimposed on the seasonal cycle last between 6 and 30 days and are observed over a multi-year period including during a strong El Niño year. These results expand upon a previous
study of subseasonal isotopic variations in the same region (Vimeux et al. 2011), and demonstrate
that the findings in Vimeux et al. (2011) hold for a wider region along the northeastern margin
of the Altiplano and over a multi-year period. Events with low (high) $\delta^{18}$O are associated with
anomalously high (low) mid-tropospheric relative humidity averaged over the study area that is
in turn influenced by propagating extratropical Rossby waves and correlate with variability in the
strength and position of the SACZ. Anomalously low $\delta^{18}$O in precipitation occurs when the SALLJ
is strengthened, and anomalously high $\delta^{18}$O occurs when the SALLJ is weakened or reversed, or
when there is reduced relative humidity in the Amazon basin along the eastern tropical Andes. The
relationship between subseasonal $\delta^{18}$O variability and the continental-scale precipitation dipole
associated with the changes in the position of the SACZ implies that subseasonal $\delta^{18}$O variability
encodes information about continental climatic anomalies and may be related to the Madden
Julian Oscillation; such a relationship is an important area for future research. This subseasonal
variability is retained in the isotopic profile of the annual layer snowpit on Quelccaya, enabling
the development of an accurate snowpit age model, and demonstrating that the snowpit isotopic
profile reflects synoptic and continental-scale meteorological variability. These results suggest
that annual layering in Andean ice cores may contain information about subseasonal variability in
precipitation delivery and associated circulation anomalies, which will improve future efforts to
better understand Andean hydroclimate through ice cores.

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LIST OF TABLES

Table 1. Correlation coefficients between $\delta^{18}$O at each station on days with common precipitation (the number of days used for each correlation is bracketed). All correlations are significant at the 99% confidence interval or greater. . . . . . . 37

Table 2. Correlation coefficients between $\delta^{18}$O anomalies and selected variables from ERA-Interim reanalysis averaged over the region 12.0-18.5° S and 65.0-74.0° W. All variables are 15-day moving averages minus the seasonal cycle. Variables: $r =$ relative humidity, $t =$ temperature, $z =$ geopotential height, $u =$ zonal wind, $v =$ meridional wind. The numbers beside each variable refer to the atmospheric pressure level (hPa). P = mean observed precipitation at all sites. Correlations that are significant at the 99% confidence level are highlighted with asterisks. . . . 38
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<table>
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<tr>
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<th>Parayoc</th>
<th>Phinaya</th>
<th>Chillca</th>
<th>Tuxahuira</th>
<th>Warisata</th>
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<td>0.75 (56)</td>
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<td>0.89 (23)</td>
<td>0.85 (21)</td>
<td>0.90 (21)</td>
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<td>0.61 (68)</td>
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<td>0.71 (24)</td>
<td>-</td>
<td>0.55 (36)</td>
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<td>-0.53*</td>
<td>-0.28*</td>
<td>-0.8*</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

Fig. 1. Study area. Left panel: the locations of the Cordilleras Vilcanota and Real, cities (white stars), and mountains (blue triangles) discussed in this study. Right panel: key features of South American circulation discussed in this study. SACZ = South Atlantic Convergence Zone, SALLJ = South American Low Level Jet.

Fig. 2. Locations of precipitation collection sites. The table inset lists station names, elevations (meters above sea level), and the start date of each record.

Fig. 3. Summary of all daily precipitation (top), $\delta^{18}$O (middle) and $\delta$-excess (bottom) measurements used in this study. Lines connect data collected on consecutive days. Dashed blue lines on the $\delta^{18}$O plots are precipitation amount weighted means between 1-Nov and 1-May each year.

Fig. 4. $\delta^{18}$O and $\delta$D of each daily precipitation sample used in this study in all seasons (2014-2017). Different shapes/colors correspond to different stations (see legend inset). The global meteoric water line (GMWL, solid black) and local meteoric water line (LMWL, dashed black) are plotted for comparison.

Fig. 5. a) 15-day moving average $\delta^{18}$O (weighted by precipitation amount) at each site that collected samples in 2017. b) The same as a), but for measurements of daily precipitation amount.

Fig. 6. a) $\delta^{18}$O of all precipitation samples collected in 2017 (points, see legend inset), 3-day precipitation weighted mean of all samples where more than 2 data points are available (thin black line), 15-day precipitation weighted mean (green line, referred to as the subseasonal signal in this study), and the 90-day precipitation weighted mean (thick orange line, referred to as the seasonal cycle in this study). b) Subseasonal $\delta^{18}$O anomalies during the 2017 wet season (subseasonal signal minus the seasonal cycle). Red shading highlights positive anomalies and blue shading highlights negative anomalies.

Fig. 7. Subseasonal $\delta^{18}$O anomaly plots (as described in Fig. 5b) for each wet season 2014-2017 (a-d respectively).

Fig. 8. Difference plots between the average conditions during all positive anomalies and all negative anomalies identified in Fig. 6 (positive - negative). a) 500 hPa relative humidity, b) 250 hPa relative humidity, c) 250 hPa zonal wind, d) 250 hPa geopotential height, e) 500 hPa meridional wind, and f) 850 hPa meridional wind. Black star is the location of Quelccaya. Stippling denotes differences that are significant at the 99% confidence level.

Fig. 9. Standardized subseasonal $\delta^{18}$O anomaly signal (thick grey line). Standardized subseasonal variations in 500 hPa relative humidity averaged over the region 18.5 to 12 S and 74 to 65 W. Standardized subseasonal variations in 850 hPa meridional wind at Santa Cruz (dashed red line). Dates of Cold Air Incursions are shown by blue circles plotted along y=0. Pearson’s correlation coefficients in the upper right corner are between each variable and the subseasonal $\delta^{18}$O anomaly signal. Asterisks identify correlations that are significant at the 99% confidence level.

Fig. 10. Pearson’s product-moment correlation coefficients between the time-series of $\delta^{18}$O anomalies and time-series of 500 hPa relative humidity anomalies at every grid-point for each year. Black dots highlight areas where the correlation is significant at the 99% confidence level. On each plot the black star shows the location of Quelccaya.
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