2014

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Fornace, Kyrstin L.; Hughen, Konrad A.; Shanahan, Timothy M.; Fritz, Sherilyn C.; Baker, Paul A.; and Sylva, Sean P., "A 60,000-year Record of Hydrologic Variability in the Central Andes from the Hydrogen Isotopic Composition of Leaf Waxes in Lake Titicaca Sediments" (2014). *Papers in the Earth and Atmospheric Sciences*. 449.
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Published in *Earth and Planetary Science Letters* 408 (2014), pp. 263–271; doi:10.1016/j.epsl.2014.10.024
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Submitted April 29, 2014; revised October 6, 2014; accepted October 18, 2014; published online October 30, 2014.

Supporting information for this article is available following the references.

A 60,000-year Record of Hydrologic Variability in the Central Andes from the Hydrogen Isotopic Composition of Leaf Waxes in Lake Titicaca Sediments

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Abstract

A record of the hydrogen isotopic composition of terrestrial leaf waxes (δD_{wax}) in sediment cores from Lake Titicaca provides new insight into the precipitation history of the Central Andes and controls of South American Summer Monsoon (SASM) variability since the last glacial period. Comparison of the δD_{wax} record with a 19-kyr δD record from the nearby Illimani ice core supports the

interpretation that precipitation δD is the primary control on δD_{wax} with a lesser but significant role for local evapotranspiration and other secondary influences on δD_{wax} . The Titicaca δD_{wax} record confirms overall wetter conditions in the Central Andes during the last glacial period relative to a drier Holocene. During the last deglaciation, abrupt δD_{wax} shifts correspond to millennial-scale events observed in the high-latitude North Atlantic, with dry conditions corresponding to the Bølling–Allerød and early Holocene periods and wetter conditions during late glacial and Younger Dryas intervals. We observe a trend of increasing monsoonal precipitation from the early to the late Holocene, consistent with summer insolation forcing of the SASM, but similar hydrologic variability on precessional timescales is not apparent during the last glacial period. Overall, this study demonstrates the relative importance of high-latitude versus tropical forcing as a dominant control on glacial SASM precipitation variability.

Keywords: Lake Titicaca, South American Summer Monsoon, stable hydrogen isotopes, leaf wax, last glacial period, Holocene

1. Introduction

Aided by the availability of different paleoclimate archives, the Central Andes have been the target of many studies of late Quaternary tropical South American climate. For example, sediment cores from the Lake Titicaca basin record changes in water balance (precipitation minus evaporation) and lake level on timescales from glacial–interglacial to millennial (Baker et al., 2001b; Fritz et al., 2007, 2010). Since the majority of regional precipitation falls during austral summer, the Central Andes are particularly sensitive to variability in the South American Summer Monsoon (SASM), a continental-scale circulation that produces much of summer precipitation throughout southern tropical and subtropical South America (Zhou and Lau, 1998). While studies of modern SASM precipitation have identified many controls on interannual to interdecadal SASM variability, including Pacific sea surface temperature (SST)/El Niño Southern Oscillation (ENSO) (Garreaud et al., 2009), Atlantic SST variability (Chiessi et al., 2009), and land surface processes (Collini et al., 2008), SASM behavior on longer timescales is not as well constrained. Characterizing the controls on SASM dynamics on millennial to interglacial–glacial timescales is thus critical to improving our understanding of past and future hydrologic variability throughout much of South America.

Proxies that record the isotopic composition of past precipitation in the Andes, from archives such as ice cores, speleothems, and lacustrine calcite deposits, have proven to be particularly powerful tools for reconstructing past monsoon variability on geological timescales (e.g., Ramirez et al., 2003; Bird et al., 2011; Kanner et al., 2012). Although the interpretation of Andean stable isotopic records has been the subject of much debate (e.g., Broecker, 1997; Hoffmann, 2003; Pierrehumbert, 1999; Thompson et al., 1995), a number of recent modeling and observational studies clearly indicate a strong relationship between precipitation isotopes and the amount of precipitation along the SASM trajectory (Vuille et al., 2003; Vuille and Werner, 2005; Vimeux et al., 2005; Insel et al., 2013). Andean ice cores have yielded insight into the influence of global glacial cycles on regional hydrologic variability, with wetter conditions associated with Last Glacial Maximum (LGM) boundary conditions (Ramirez et al., 2003). In addition, changes in SASM precipitation have been

linked to high-latitude millennial-scale variability, including Dansgaard–Oeschger cycles and Heinrich events during the last glacial period (Kanner et al., 2012), as well as the Younger Dryas and Bølling–Allerød intervals during the last deglaciation (Thompson et al., 1995, 1998). A number of Andean lacustrine calcite $\delta^{18}\text{O}$ records have pointed to the magnitude of local summer insolation as an important control of SASM intensity during the Holocene (Baker et al., 2009; Bird et al., 2011; Seltzer et al., 2000), consistent with southern Brazilian speleothems (Cruz et al., 2006; Wang et al., 2007) that show a strong influence of insolation on SASM precipitation over the entire last glacial cycle. In contrast, speleothem records from the Andes and western Amazonia do not vary strongly on precessional timescales during the last glacial period (MIS 2–4) (Kanner et al., 2012; Mosblech et al., 2012; Cheng et al., 2013).

Reaching a better understanding of past SASM variability is limited by the small number of continuous isotopic records from the Andes that extend into the last glacial period. Andean ice core records only extend as far back as the LGM (Ramirez et al., 2003; Thompson et al., 1995, 1998), while other extant Andean isotopic records are confined to the Holocene (e.g., Bird et al., 2011) or within the last glacial period (Kanner et al., 2012). Longer records are particularly valuable in light of new speleothem records from the western Amazon that provide important information on SASM activity upstream of the Andes over the last glacial–interglacial cycle (Mosblech et al., 2012; Cheng et al., 2013). The sediment record of Lake Titicaca (LT), continuous over the past four glacial cycles, is a valuable archive of regional climate variability, but traditional stable isotope archives (e.g., authigenic carbonate) are not present in much of the sediment record (Fritz et al., 2007). Compound-specific isotope analysis of biomarkers in LT sediments thus represents a new and advantageous approach to the study of past Andean climate. In particular, the hydrogen isotopic composition (δD) of leaf wax compounds produced by vascular plants has been shown to be highly correlated to precipitation δD (e.g., Sachse et al., 2012), allowing reconstruction of SASM precipitation variability when carbonate proxies are absent. Here we present a continuous record of regional hydrologic variability over the past 60,000 years from leaf wax δD in LT sediments.

2. Study site

The South American Altiplano is a semi-arid, internally drained plateau situated between the western and eastern cordilleras of the Central Andes (fig. 1). Located on the Peruvian/Bolivian border on the northern Altiplano, Lake Titicaca (3810 m asl) is a large, freshwater lake comprised of two basins: the larger and deeper Lago Grande (7131 km²) and the smaller shallow Lago Huiñaimarca (1428 km²) connected by the Straits of Tiquina. The watershed of the Lago Grande basin consists of six major rivers (total catchment area: 52,800 km²). Modern natural vegetation surrounding the lake largely consists of grasses, shrubs, and herbs and is classified as puna vegetation (Paduano et al., 2003).

Figure 1. Map of South American tropics showing location of Lake Titicaca and locations of ice core (triangles), speleothem (squares), and sediment (circles) records discussed in text. **CB:** Cariaco Basin (Hughen et al., 2004), **SC:** Santiago cave (Mosblech et al., 2012), **NAR/ELC:** Cueva del Diamante/El Condor (Cheng et al., 2013), **HIC:** Huascarán ice core (Thompson et al., 1995), **LP:** Laguna Pumacocha (Bird et al., 2011), **LJ:** Lake Junin (Seltzer et al., 2000), **PC:** Pacupahuain cave (Kanner et al., 2012), **IIC:** Illimani ice core (Ramirez et al., 2003), **SIC:** Sajama ice core (Thompson et al., 1998), **CS:** Santana cave (Cruz et al., 2006), **BC:** Botuverá cave (Cruz et al., 2006, 2007; Wang et al., 2007). Also shown are general features of the South American Summer Monsoon (SASM) including northeast trade winds, connections to the marine Intertropical Convergence Zone (ITCZ) and South Atlantic Convergence Zone (SACZ), and a representation of the general SASM moisture trajectory during austral summer. Shaded area indicates extent of Andes (area with elevation ≥ 2000 m), while dashed line indicates approximate extent of Altiplano.

Present annual precipitation ranges from >700 mm in regions near Lake Titicaca to <200 mm on the southern Altiplano (Garreaud et al., 2003). The majority of annual precipitation (50–80%) on the Altiplano occurs during the austral summer and is associated with the SASM (Zhou and Lau, 1998). During the mature phase of the SASM (DJF), strengthened easterly trade winds increase moisture transport from the tropical Atlantic into the Amazon Basin toward the tropical Andes. The development of a continental-scale low-pressure system (Chaco Low) centered on the subtropical plains steers low-level flow southeastward along the eastern Andean flank, carrying moisture from the Amazon Basin to the subtropics (Garreaud et al., 2009). An upper-level anticyclone, known as the Bolivian High, develops over the Altiplano in response to latent heat release by Amazonian and Andean precipitation (Lenters and Cook, 1997). Upper-level easterly flow associated with the north

branch of the anticyclone allows upslope moisture transport from lowland Amazonia to the high Andes (Garreaud, 1999). During the peak months of the SASM (December–February), convective activity is concentrated over the southern Amazon Basin and Central Andes (Vuille and Werner, 2005). During the austral autumn (March–May), the SASM weakens and the region of maximum precipitation over continental South America retreats to the northern tropics. On the Altiplano, prevailing westerly flow from the arid west Andean slope results in dry conditions throughout much of the remainder of the year (May to October) (Garreaud et al., 2003).

3. Methods

3.1. *Sediment core retrieval*

Sediment cores from Lake Titicaca were raised in a 2001 International Continental Drilling Program expedition using the GLAD 800 drilling platform and coring system. Sediment samples analyzed here were taken from core LT01-2B, a 136 m core recovered from the Lago Grande basin at a water depth of 228 m (15.8533°S, 69.1404°W). The core was generally sampled at 2 cm resolution during periods of low sediment accumulation (Holocene) and at 10 cm resolution during periods of higher sedimentation rates (last glacial period). In total, 175 samples were analyzed for δD at an average resolution of ~300–400 yr from 61–3.5 kyr. Sample ages were assigned with the LT01-2B age model developed by Fritz et al. (2007) based on 18 bulk organic radiocarbon ages and U/Th ages of aragonite deposits, as well as tuning glacial terminations to the Vostok CO₂ record from Antarctica. Conversion from radiocarbon to calendar years was updated to the IntCal13 calibration (Table S1; Reimer et al., 2013). For the period from 3.5 ka to the present, 11 sediment samples were taken from a box core (NE98 4BXB) raised from the Lago Grande basin at a water depth of 145 m (16.1376°S, 69.1545°W) in 1998. Samples ages were assigned from an age model developed by Tapia et al. (2003) based on 4 bulk organic radiocarbon ages. Calendar ages were also updated to reflect calibration from IntCal13.

3.2. *Sample preparation*

Freeze-dried and finely ground sediment samples (~3–10g) were extracted with a 9:1 dichloromethane:methanol mixture using an accelerated solvent extractor (ASE 200, Dionex). The resulting total lipid extract was dried over Na₂SO₄ and then separated into neutral and acidic fractions by flash column chromatography over aminopropyl-functionalized silica gel (Supelclean LC-NH₂, Sigma Aldrich) using 9:1 dichloromethane:acetone and 2% formic acid in dichloromethane as eluents, respectively. All columns were run on a RapidTrace SPE Workstation. The acidic fraction was transesterified using a 9:1 mixture of anhydrous methanol and acetyl chloride (70°C, 12 h). Fatty acid methyl esters (FAMES) were further purified on a second LC-NH₂ column using hexane as eluent. FAMES were identified by comparison of gas chromatograph/flame ionization detector (GC/FID, Agilent 6850) retention times to those of authenticated external FAME standards.

3.3. Compound-specific hydrogen isotope analysis

The δD values of individual FAMES were measured on a Thermo Scientific DeltaV^{Plus} isotope ratio mass spectrometer (IRMS) coupled to an Agilent 6980 GC via a pyrolysis interface (GC-TC) operated at 1440°C. Each sample was run in duplicate, with the exception of 12 samples distributed throughout the core that were concentration-limited and measured only once. Data were processed using Isodat 3.0 software (Thermo Scientific). The H_3^+ factor (Sessions et al., 2001) was measured daily and ranged from 1.7–2.0 ppm/mV over the course of the measurement period. Peaks of a propane reference gas were inserted at several points before and after analytes during each run and used as internal calibration standards. Instrument variability was accounted for by routinely injecting an external standard mix containing 8 fatty acid methyl and ethyl esters of known δD (F8 mixture, A. Schimmelmann, Indiana University) and adjusting the reference propane δD value to minimize the average offset between the known and measured δD values of the F8 compounds. Average 1σ precision of sample replicates was 3.1‰. Estimated average accuracy is $\pm 4.5\%$ based on daily measurements of the F8 mixture. In order to correct for the δD of methyl hydrogen added during transesterification, a phthalic acid standard of known isotopic composition (A. Schimmelmann, Indiana University) was methylated in parallel to samples using the same batch of anhydrous methanol. The δD of the resulting dimethyl phthalate was measured and used to calculate the δD of the methyl hydrogen by mass balance. Values are reported using standard delta notation (δD) as per mil (‰) deviations from Vienna standard mean ocean water (VSMOW).

4. Results and discussion

Measured δD values for n -C₂₈ and n -C₃₀ FAMES over the past 60,000 years are shown in figure 2. These data can also be found online in the NOAA NCDC Paleoclimate database. The δD values for both compounds show similar trends over this time period and are highly correlated ($r^2 = 0.82$, $p < 0.0001$). In interpreting the record, we focus on n -C₂₈ FAME δD (heretofore referred to as δD_{wax}) as the more concentrated compound with a more robust signal during δD measurement. Lake Titicaca δD_{wax} has a large dynamic range over the past 60 kyr, with values ranging from -236 to -114% . Glacial (60–20 ka) values are on the whole more depleted than Holocene (11.5–0 ka) values; average glacial δD_{wax} is -207% , while average Holocene δD_{wax} is -152% . Within the last glacial period, values range from -174 to -236% and show a high degree of variability on millennial timescales.

Figure 2. Time series of Titicaca C₂₈ (green) and C₃₀ (gray) *n*-alkanoic acid δD over the past 60 kyr. Darker colors indicate samples from LT01-2B core, and lighter colors indicate samples from NE98-4BXB core. Error clouds indicate 1σ precision of replicate sample measurements. For samples with no replicates, an error of 4.5‰ was assigned based on average uncertainty of external standards. Triangles denote age points from radiocarbon dating of sediment organic matter (Table S1) (Tapia et al., 2003; Fritz et al., 2007). We note that despite the strong correlation between *n*-C₂₈ and *n*-C₃₀ ($r^2 = 0.82$, $p < 0.0001$), there are some discrepancies between the δD records of the two compounds, possibly due to different vegetation contributions to each chain length. For interpretation of past precipitation variability, we opted to focus on *n*-C₂₈ δD as the more concentrated compound and generally more robust signal during δD measurement.

The transition from the last glacial period into the Holocene is marked by a series of large and abrupt increases in δD_{wax} . LT δD_{wax} shows an initial rise ca. 16 ka, followed by a reversal from ~13.5–12.5 ka and a further rise at ~12–11 ka into the Holocene. The LT pattern closely resembles that observed in the δD of ice in the nearby Illimani ice core (Ramirez et al., 2003) (see fig. 3), as well as the pattern of the last deglaciation in the NGRIP Greenland ice core $\delta^{18}O$ (Svensson et al., 2008) (see fig. 5). Enriched δD_{wax} values from ~15.5–13.8 ka likely correspond to the Bølling–Allerød interstadial (BA) while the following period of depleted δD_{wax} values coincides with Younger Dryas stadial (YD). We note that there is some age offset between the period of depleted δD_{wax} values and the YD interval in the NGRIP record, but the offset is likely due to uncertainties within the LT01-2B age model since several well-dated speleothem records of SASM precipitation show δD minima in phase with the YD (see fig. 5). During the Holocene, δD_{wax} ranges from –183 to –114‰.

Values decrease from the early Holocene into the present, suggesting an inverse relationship between austral summer insolation and δD_{wax} . However, there is no apparent variability on orbital timescales during the last glacial period.

Figure 3. (Left) Comparison of Titicaca δD_{wax} and Illimani ice core δD over past 20 kyr. Note difference in scales between left and right axes. (Right) Regression between Titicaca δD_{wax} and corresponding Illimani δD data at interpolated time points.

4.1. Interpretation of the LT δD_{wax} record

4.1.1. Controls on Altiplano meteoric water δD

The isotopic composition of modern precipitation (δD_p) in the Central Andes is determined by a combination of several different climatic variables. SASM precipitation that falls on the Altiplano is ultimately derived from Atlantic Ocean moisture that has been transported across the Amazon Basin and lifted over the eastern cordillera of the Andes. Altiplano δD_p thus integrates upstream effects including rainout and moisture recycling over the Amazon Basin, altitude effects associated with upslope moisture transport, and local effects such as the intensity of precipitation (“amount effect”) (Dansgaard, 1964). Changing moisture trajectories can also distinctively impact precipitation isotopic ratios (e.g., Insel et al., 2013). Results from both observational and modeling studies suggest that the dominant control on modern interannual Andean δD_p variability is Rayleigh-type fractionation during rainout along the trajectory of moist air masses across the Amazon Basin up to the Andes (e.g., Rozanski et al., 1993; Hoffmann et al., 2003; Vimeux et al., 2005; Vuille and Werner, 2005). Increased rainout along this trajectory, as well as increased local precipitation in the Andes, results in more depleted δD_p , consistent with the continentality and amount effects that dominate δD_p variability throughout most of tropical South America (Vuille et al., 2003). Accordingly, the intensity of the SASM is significantly negatively correlated with $\delta^{18}\text{O}$ (and correspondingly δD) of modern precipitation in the Central Andes, as well as throughout much of southern tropical South America (Vuille and Werner, 2005).

On glacial–interglacial timescales, there may be additional factors that influence δD_p at LT. First, the effects of changing ice volume caused changes in the isotopic composition of the original moisture source for the SASM (the equatorial Atlantic Ocean). However, given

that mean ocean isotopic composition was approximately 8‰ more enriched in δD during the LGM (i.e., $\delta^{18}O$ enrichment of 1.0‰ scaled by meteoric water line) (Schrage et al., 2002), this correction is very small compared to the scale of LT δD_{wax} changes. Second, an increased temperature gradient between high and low altitude during the last glacial period could have resulted in an “enhanced” altitude effect and more depleted δD_p at high altitude. However, Vimeux et al. (2005) conclude based on modeling simulations that this effect was likely relatively minor ($\sim 10\%$) compared to the dominant hydrologic signal in the Illimani ice core δD record. Finally, changing moisture trajectories and changing rates of evapotranspiration in the Amazon forest may have influenced Altiplano δD_p , but it is difficult to constrain either process. Despite these caveats, we interpret past δD_p variability on the Altiplano as a reasonable indicator of regional SASM precipitation intensity that integrates the effects of rainout in both the Amazon Basin and Central Andes.

4.1.2. Controls on δD_{wax}

Long-chain (C_{24} – C_{32}) *n*-fatty acids with a strong even over odd carbon number preference are produced nearly exclusively by vascular plants as components of leaf epicuticular waxes (Eglinton and Hamilton, 1967). Leaf wax compounds are deposited in lacustrine or marine sediments by eolian or fluvial transport. Though these compounds may be transported over long distances (Conte and Weber, 2002), the steep Andean topography and isolation of the LT basin make any significant contribution of leaf waxes from long-range transport unlikely. The relatively good preservation of long-chain fatty acids in sediments (Meyers, 1997) and very slow rates of isotopic exchange of the alkyl chain hydrogen (Sessions et al., 2004) make these compounds well suited for isotopic reconstructions.

The hydrogen isotopic composition of leaf wax compounds in sediments has previously been demonstrated to reliably track the isotopic composition of plant source water, i.e., meteoric water for terrestrial plants, thus allowing reconstruction of past trends of precipitation isotopes (Hou et al., 2008; Sachse et al., 2004, 2006). However, a number of secondary effects may obfuscate this primary signal. Such effects include soil water evaporation and leaf transpiration, which can cause isotopic enrichment of soil and leaf water, respectively. Furthermore, a number of studies have shown variability in biosynthetic fractionation between source water and leaf waxes among different vegetation growth forms (i.e., trees, shrubs, grasses) (Sachse et al., 2012). Because the magnitudes of these effects in different environments are not well constrained, there is some uncertainty about their combined influence on δD_{wax} in records of past climate.

We assessed the extent to which LT δD_{wax} reflects isotopic variability of precipitation versus local environmental influences by comparing the LT record to the δD record of the Illimani ice core (Ramirez et al., 2003), which provides a more direct record of isotopic variation of precipitation. Nevado Illimani is situated in the eastern cordillera of the Andes approximately 100 km directly east (generally upwind) of LT. Illimani δD should thus provide a good estimate of δD_p variability over the LT basin. Over the past 19 kyr, LT δD_{wax} and linearly interpolated Illimani δD data are well correlated (fig. 3, $r^2 = 0.69$, $p < 0.0001$). This correlation is comparable to that of modern calibration studies of δD_{wax} across spatial climate gradients (Sachse et al., 2012) and shows that LT δD_{wax} variability is closely related to δD_p variability. δD_{wax} may be particularly successful in capturing precipitation isotopic

variability at this site because of the relatively large changes in δD_p over this time period and the isolated and relatively small source area of leaf waxes in the LT watershed, helping to focus the precipitation signal in LT sediments. Although we cannot compare δD_{wax} with ice core data before 19 ka, we assume that the observed correlation between δD_{wax} and δD_p from 19–0 ka remained intact over the duration of our record.

Although LT δD_{wax} and Illimani δD are well correlated, the slope of this relationship is significantly greater than 1. Thus while the isotopic composition of precipitation appears to be the primary control on the structure of the LT δD_{wax} record, it is clear that secondary effects also contribute to this signal. Most notably, the much larger range at LT (102‰) than at Illimani (58‰) over the past 19 kyr suggests the influence of compounding effects on LT δD_{wax} , such as decreased regional precipitation, resulting in a more enriched δD_p , and low local humidity, which would tend to further enrich plant source water by soil water evaporation and/or leaf transpiration. Since such local evapotranspiration feedbacks would be expected to act in step with δD_p changes, this would serve to amplify the δD_{wax} signal and increase its sensitivity to δD_p variability.

Another potential source of variability in the LT δD_{wax} record is vegetation change. However, pollen records from LT sediments do not reveal major vegetation shifts since the last glacial period (Paduano et al., 2003; Hanselman et al., 2011). Vegetation in the LT basin over the past 60,000 years was dominated by puna (grassland) and sub-puna (shrubland) taxa, particularly *Poaceae*. The most significant difference between the last glacial period and the Holocene was an increase in vegetation density in the LT watershed rather than large shifts in vegetation composition (Paduano et al., 2003). However, during the Holocene, a large drop in lake level between ~8 and 4 ka caused a significant retreat in the LT shoreline (Cross et al., 2000; D'Agostino et al., 2002) and likely increased the amount of littoral organic detritus delivered to deepwater sites (Baker et al., 2001b). A change in the relative contributions of purely terrestrial versus emergent aquatic plants residing near the shoreline could thus have significantly affected δD_{wax} . Several studies have shown aquatic plants to have more depleted δD_{wax} compared to purely terrestrial plants, presumably due to accessing different water sources (i.e., lake water/groundwater vs. soil water) and the absence of soil evaporation (Shuman et al., 2006; Douglas et al., 2012). After removing the long-term negative trend through the Holocene, we observe generally more depleted δD_{wax} values during periods of low rather than high lake level (fig. 4). Furthermore, an increased proportion of n -C₂₄ fatty acids relative to total abundance of long-chain fatty acids (n -C₂₄– n -C₃₀) during the mid-Holocene lowstand is consistent with an increased contribution from macrophytes at this time (Gao et al., 2011). We note that Holocene δD_{wax} trends are complicated by the switch between LT01-2B and NE98-4BXB cores at 3.5 ka, but the chain length pattern remains generally consistent across the two cores. Shifts in the sources of leaf waxes to the core site are thus a possible explanation for some of the variability in δD_{wax} during the Holocene, although some large deviations (i.e., at ~8.5ka) do not appear associated with lake level change.

Figure 4. Comparison of Holocene detrended Titicaca δD_{wax} record (A), changes in n -alkanoic chain length abundance (B), and lake level indicators (C). **A.** Titicaca δD_{wax} data from 11–0 ka was fit with a linear regression and the slope subtracted to remove long-term trend. Dark green indicates LT01-2B samples, light green indicates NE98-4BxB samples, and dashed line shows average Holocene value. **B.** Ratio of C_{24} n -alkanoic acid abundance to total long-chain even carbon number fatty acid ($n-C_{24}$ – $n-C_{30}$) abundance. Dark orange indicates LT01-2B samples, light orange indicates NE98-4BxB samples, and dashed line shows average Holocene value. **C.** Black line shows weight percent $CaCO_3$ in Lake Titicaca sediments from Baker et al. (2001b). $CaCO_3$ content is inversely related to lake level due to salinity and pressure effects on $CaCO_3$ precipitation/dissolution. Blue squares indicate lake level estimates from Cross et al. (2000) using sedimentological, geochemical, and biological analyses of Lake Titicaca sediment cores. Solid dark blue line indicates lowstand level estimate from seismic data (D'Agostino et al., 2002). Shaded bar shows correspondence between depleted δD_{wax} values, high $n-C_{24}$ abundance and low lake level during mid-Holocene.

4.2. Remote forcing of SASM variability

4.2.1. Glacial-interglacial variability

The substantial difference between glacial and Holocene δD_{wax} values at Lake Titicaca points to higher precipitation along the SASM trajectory during the last glacial period than during the Holocene. The similar patterns of LT δD_{wax} and Greenland temperature over the last 60 kyr suggest that glacial boundary conditions, e.g., the buildup of large northern continental ice sheets, exerted important control on SASM dynamics (fig. 5). As LT δD_{wax} is influenced by both local and upstream precipitation variability, a number of other tropical South American records help to constrain the geographical extent of glacial precipitation increases. Ample evidence of wetter conditions in the Central Andes and Altiplano during the LGM (Baker et al., 2001a, 2001b) and the previous three glacial periods (Fritz et al., 2004, 2007) clearly indicate significant increases in Central Andean precipitation under glacial conditions. Enhanced Andean rainfall under glacial conditions is also supported by regional climate simulations that suggest a mechanism whereby atmospheric circulation changes induced by cooler tropical Atlantic temperatures could have led to increased moisture transport from the Amazon to the Andes (Vizy and Cook, 2007).

Outside the Andes, there is some evidence of moderate precipitation increases in western Amazonia (Cheng et al., 2013) and southern Brazil (Cruz et al., 2007) throughout the last glacial period. Such widespread changes imply a larger regional mechanism to increase precipitation during the last glacial period. Based on simulations of LGM boundary conditions, Chiang et al. (2003) and Chiang and Bitz (2005) have suggested that the presence of large ice sheets at high northern latitudes could have resulted in a southward shift of the Atlantic ITCZ. The latitudinal position of the Atlantic ITCZ controls the rainy season over northern South America and also affects moisture transport into the Amazon Basin, with a southward shift in mean ITCZ position associated with increased Amazon moisture transport and a more intense SASM on a range of timescales (Vuille et al., 2012). A southward displacement of the ITCZ during the last glacial period together with increased moisture transport to the Andes could thus have caused widespread SASM precipitation increases throughout southern tropical South America. However, greater geographic coverage of records of tropical South American hydrologic variability over the last glacial cycle would be necessary to test this idea.

Figure 5. Comparison of Titicaca δD_{wax} with Greenland and southern tropical South American isotopic records over the past 60 kyr. **A.** Greenland (NGRIP) ice core $\delta^{18}O$ (Svensson et al., 2008). **B.** Speleothem $\delta^{18}O$ from Santiago cave in western Amazonia (Mosblech et al., 2012). **C.** Speleothem $\delta^{18}O$ from El Condor cave in western Amazonia (Cheng et al., 2013). **D.** Titicaca δD_{wax} . **E.** Speleothem $\delta^{18}O$ from Botuverá cave in southern Brazil (Wang et al., 2007). For South American records, gray lines show mean summer (December–February) insolation for latitude of each study site. Note that insolation axes are reversed (increasing down). Insolation values were calculated with AnalySeries 2.0 (Paillard et al., 1996) using data from Laskar et al. (2004). Shaded bar indicates Younger Dryas (YD) interval in all records.

4.2.2. Millennial-scale variability

The pattern of LT δD_{wax} during the last deglaciation suggests a period of reduced precipitation intensity corresponding to the BA and wetter conditions during the YD, a pattern also suggested by other paleolimnological data from the Altiplano (Baker et al., 2001a, 2001b). A well-dated paleolake sequence further shows that the timing of lake highstands on the central Altiplano coincides with Heinrich Event 1 and the YD, providing evidence of synchronicity between temperature fluctuations in the North Atlantic region and Andean hydrologic variability (Blard et al., 2011). Intervals of depleted $\delta^{18}O$ in the Botuverá speleothem (southern Brazil) and El Condor speleothem (western Amazonia) coincident with the YD (fig. 5) also point to increased precipitation across southern tropical South America due to a more intense SASM at that time (Wang et al., 2007; Cheng et al., 2013).

In contrast, a reconstruction of C3/C4 vegetation from the Cariaco Basin in northern tropical South America shows wetter conditions coincident with the BA followed by a drier YD (Hughen et al., 2004). This observed anti-phase relationship between changes in northern and southern tropical South American precipitation is consistent with meridional shifts of the Atlantic ITCZ during the last deglaciation. Southward ITCZ shifts on millennial timescales during the deglaciation were likely due to remote forcing from high-latitude North Atlantic cooling and an increase in the latitudinal temperature gradient. Southward displacements of the Atlantic ITCZ are consistently reproduced in general circulation model simulations of reduced North Atlantic SST associated with weakened Atlantic meridional overturning circulation (AMOC) (e.g., Vellinga and Wood, 2002; Zhang and Delworth, 2005), thereby linking southward ITCZ migration with cold North Atlantic events and Greenland stadials.

North Atlantic forcing has also been invoked during the last glacial period to explain connections between millennial-scale tropical South American hydrologic variability and high-latitude climate fluctuations (Baker et al., 2001a, 2001b). Cold North Atlantic conditions, associated with cold phases of Dansgaard–Oeschger (D-O) cycles and Heinrich events, coincided with dry conditions in northern South America inferred from a longer record of C3/C4 vegetation shifts from the Cariaco Basin (Drenzek, 2007). In the Central Andes, more depleted $\delta^{18}O$ values in the Pacupahuain cave speleothem record point to a more intense SASM during Heinrich events and most cold phases of D-O cycles (Kanner et al., 2012). Millennial fluctuations of several paleolimnological indicators (including grain size, biogenic silica, bulk organic carbon $\delta^{13}C$, and planktonic/benthic diatom ratio) in the sediments of Lake Titicaca during MIS 3 are consistent with increased precipitation, terrestrial runoff, and shallow water input to deep water sediments coincident with North Atlantic cold events (Fritz et al., 2010). Due to relatively low resolution of the LT δD_{wax} record and uncertainty in the LT01-2B age model during the same time period, we cannot unequivocally correlate δD_{wax} variability to North Atlantic events, but it is likely that Andean hydrologic variability tied to high-latitude forcing contributed to the large fluctuations of LT δD_{wax} between 60 and 20 ka.

4.3. Local forcing of SASM variability

The magnitude of local summer insolation does not show a consistent relationship with δD_{wax} variability at Lake Titicaca over the past 60 kyr (fig. 5). The LT record instead shows

a much stronger relationship between summer insolation and δD_{wax} variability over the last 20 kyr than during the last glacial period (60–20 ka). The negative trend in δD_{wax} from the early to late Holocene is consistent with increasingly wet conditions in response to increasing summer insolation. This Holocene trend has also been seen in a number of isotopic records from the Central Andes, including the Illimani and Huascaran ice cores (Ramirez et al., 2003; Thompson et al., 1995) and carbonate records from Lake Pumacocha (Bird et al., 2011) and Lake Junin (Seltzer et al., 2000), as well as speleothem records from southern Brazil (Cruz et al., 2006; Wang et al., 2007) and western Amazonia (Cheng et al., 2013). As discussed by Bird et al. (2011), the uniformity of Holocene records throughout southern tropical South America suggests an insolation-controlled strengthening of the SASM throughout the Holocene.

In contrast, there does not appear to be any significant relationship between LT δD_{wax} and summer insolation from 60 to 20ka, consistent with absent or very weak precessional signals in speleothem $\delta^{18}\text{O}$ records from the Andes (Kanner et al., 2012) and western Amazonia (Mosblech et al., 2012; Cheng et al., 2013) over this interval. Although there is strong evidence of insolation forcing of monsoonal precipitation in southern Brazil throughout the last glacial period (Cruz et al., 2006; Wang et al., 2007), insolation forcing of Andean and Amazonian precipitation intensity was apparently diminished during the same period, presumably due to the competing influence of glacial boundary conditions. The interaction between insolation forcing and glacial boundary conditions thus appears to have been regionally variable, with glacial conditions suppressing the effects of insolation changes in western Amazonia and the Central Andes, while insolation forcing remained dominant in southern Brazil.

5. Conclusion

The hydrogen isotopic composition of terrestrial leaf waxes in Lake Titicaca sediments reveals large amplitude changes in regional precipitation on a range of timescales over the last 60 kyr. The coherence of LT δD_{wax} and the δD record from the nearby Illimani ice core demonstrates that variability in precipitation isotopes is the dominant influence on δD_{wax} at this site, reaffirming the value of δD_{wax} as a paleoclimate proxy. Local environmental factors are also significant, and changes in the extent of evapotranspiration of plant source water likely accentuated δD_{wax} sensitivity to isotopic shifts in precipitation.

Considerably depleted δD_{wax} values suggest wetter conditions during the last glacial period relative to the Holocene along the SASM trajectory from the Amazon Basin to the Central Andes. On millennial timescales, the LT δD_{wax} record is consistent with a number of SASM records that show drier conditions coincident with the Bølling–Allerød interstadial followed by a return to wetter conditions concurrent with the Younger Dryas stadial. The LT record further supports a model of meridional shifts in Atlantic ITCZ position, possibly related to AMOC and North Atlantic SST variability during the last deglaciation. Local insolation forcing appears to have been strong during the Holocene, evidenced by a general negative trend of δD_{wax} concomitant with increasing summer insolation. Insolation control of SASM intensity throughout the last glacial period, however, is not apparent in

the record, suggesting that glacial boundary conditions led to dampening of local insolation forcing. Overall, the LT δD_{wax} record underscores the importance of remote forcing from high latitudes as a control of tropical South American hydrology over the past 60kyr, with cold periods at high northern latitudes generally associated with increases in SASM precipitation.

Acknowledgments – This work was supported by an MIT Presidential Fellowship and a WHOI Ocean and Climate Change Institute fellowship (KLF) and NSF grants DEB-0447281 (KAH), ATM-0502517 (KAH), and EAR-1338694 (PAB and SCF). We thank Eva Niedermeyer and one anonymous reviewer for comments that helped to improve the manuscript.

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Supplemental Material**Table S1.** NE98-4BXB and LT01-2B bulk organic carbon radiocarbon ages (Tapia et al., 2003; Fritz et al., 2007) updated to IntCal13 calibration

Core	Depth (m)	¹⁴ C age (yr BP)	Age Error (yr)	Original Calendar Age (yr BP)	IntCal13 Calendar Age (yr BP)	1σ Age Error (yr)
NE98-4BXB	0	0	0	0	0	0
NE98-4BXB	0.30	1490	45	1366	1368	43
NE98-4BXB	0.60	2370	40	2357	2392	48
NE98-4BXB	0.83	3490	40	3765	3772	57
LT01-2B	0.108	3050	30	3268	3282	75
LT01-2B	0.28	3625	43	3933	3928	56
LT01-2B	1	5988	52	6818	6832	58
LT01-2B	1.83	8601	62	9584	9576	54
LT01-2B	2.9	12,123	81	14217	13970	120
LT01-2B	3.12	14,360	50	17211	17508	89
LT01-2B	6.135	18,790	65	22311	22625	110
LT01-2B	7.21	20,000	150	23210	24065	186
LT01-2B	9.04	20,760	80	24046	25075	162
LT01-2B	11.84	22,310	140	25752	26550	239
LT01-2B	15.04	24,190	160	27821	28217	204
LT01-2B	16.78	25,110	222	28834	29147	256
LT01-2B	16.78	25,310	220	29054	29346	284
LT01-2B	18.04	28,390	180	32444	32301	354
LT01-2B	21.15	33,370	200	37925	37755	427
LT01-2B	21.72	31,230	660	35569	35179	633
LT01-2B	24.16	36,680	270	41142	41324	269
LT01-2B	25.99	37,900	1900	41416	41977	1668

Updated Age-Depth Models

Core	Upper Depth (mblf)	Lower Depth (mblf)	Age-depth equations, x = depth (m)
NE98-4BXB	0	0.83	$4399x - 20$
LT01-2B	0	4.56	$4320x + 2531$
LT01-2B	4.56	16.78	$576x + 19605$
LT01-2B	16.78	38.57	$1486x + 4410$