A Reconstruction of Precipitation and Hydrologic Variability on the Peruvian and Bolivian Altiplano During the Late Quaternary

by

James Andrew Nunnery

Earth and Ocean Sciences
Duke University

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Approved:

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Paul A. Baker, Supervisor

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Bruce H. Corliss

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Gary S. Dwyer

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Sherilyn C. Fritz

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Avner Vengosh

Dissertation submitted in partial fulfillment of
the requirements for the degree of Doctor of Philosophy
in Earth and Ocean Sciences
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ABSTRACT

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Abstract

The Peruvian/Bolivian Altiplano is an important hydrologic system for paleoclimate reconstruction because it is unique in its ability to record climate variability associated with the near-continental scale South American summer monsoon (SASM), which is responsible for much of the precipitation over the Amazon basin and the southern subtropics. Over long timescales moisture on the Altiplano fluctuates in intensity due to changes in precessional insolation forcing as well as teleconnections to decadal-to-millennial scale abrupt temperature shifts in the Northern hemisphere Atlantic. These long-term changes in moisture transport to the Altiplano have been observed in multiple paleoclimate records, including drill core records and paleo-lake level records, as apparent advances and retreats of large lakes in the terminal basin occupied by the Salar de Uyuni and the Salar de Coipasa.

Presented here are the results from three studies that utilize different methods to create a refined reconstruction of paleohydrology, as well as paleoclimate, on the Altiplano. A major goal of this research is a more detailed understanding of millennial scale climate variability as it relates to insolation changes and abrupt warming and cooling in the north Atlantic. The first study discusses the creation of a paleohydrologic profile to reconstruct north-south hydrological history using previously reported lake core sediment records the northern and southern basins of the Altiplano, and a new 14

m core from the Salar de Coipasa representing the last ~45 ka. The second study uses a terrestrial hydrology model to simulate lake level changes through time given changes in precipitation and temperature. The third study uses strontium isotopic measurements of carbonates and halites in a 220-m core from the Salar de Uyuni to determine how source waters to the southern basin have changed through time.

The paleohydrologic profile in the first study is constructed using records from three major basins within the Altiplano: Lake Titicaca in the north, and Salar de Coipasa and Salar de Uyuni in the south. The new continuous sediment core from Salar de Coipasa indicates a lake that has fluctuated between deep and shallow phases for the last 45 ka. Lacking sufficient calcium carbonate, we instead take advantage of the general correlation between $\delta^{18}O$ and $\delta^{13}C$ in closed basin lakes to approximate water balance using $\delta^{13}C$ from organic carbon. This reconstruction is validated with diatom paleoecological records. The isotope measurements and diatom records indicate that from 45-36 ka Coipasa was moderately deep, consistent with paleoshoreline evidence of paleolake Minchin (46-36 ka). From 36-26 ka a shallow lake <10 m deep occupied the Coipasa basin. During the LGM (26-21 ka) the lake varied from moderate to shallow and during the Holocene (< 10 ka) the lake evolved from a shallow lake to a salt flat.

The hydrologic model in the second study was run through many scenarios including increases in precipitation, decreases in temperature, and combinations of the two. During the LGM southern Altiplano lakes fluctuated between 3,660 – 3,700 masl.

Model results suggest that during this period basin wide precipitation increased up to 250 mm/yr compared to modern values dependent on a temperature decrease of 5 °C relative to modern values. To create a lake at elevation 3,760 masl consistent with the highest paleolake phase (Tauca, ~16 ka) the model requires an increase of 350 mm/yr compared to modern values dependent on a 5 °C decrease in temperature (relative to modern values). An increase in temperature alone of 2 °C above modern values causes Lake Titicaca water level to decrease ~30 m, creating a closed basin lake. Results indicate that Lake Titicaca outflow is necessary to sustain large lakes in the southern basin, providing ~40-60% of total input via the Rio Desaguadero.

Analysis of a 220 m core from the Salar de Uyuni suggests periods of alternating wet and dry phases (indicated by alternating mud and salt units respectively) at the salar. Evident in the record is a transition at ~60 ka from sediments consistent with dry conditions ("playa lakes") to sediments consistent with deep lakes ("great lakes"). It has been shown that rivers and lakes in the Bolivian and Peruvian Altiplano display a range of Sr isotopic ratios that can be connected to the lithologies of specific drainage basins. Measurements of Sr ratios of the alternating halites and carbonate sediments are used to determine when paleolakes in the Salar were supplied by flow from the northern and central basins of the Altiplano, and when they were more a product of increased precipitation in the Uyuni basin. The results from Sr isotope analysis suggest that prior to ~60 ka the primary source of Sr to the Uyuni was local runoff and direct precipitation.

Following the state change from the "play lakes" phase to the "great lakes" phase Sr isotope measurements suggest a significant influence from more radiogenic waters originating in the central and northern Altiplano basins. The reason for this state change is attributed to a combination of a general increase in precipitation following the onset of the MIS-4 (~70 ka) glacial period and downcutting of the Laka Jahuira hydrologic divide, which connects Lago Poopó in the central basin to the Salar de Coipasa.

This approach of reconstructing hydrology using the combination of multiple paleolake records, hydrological modeling, and isotopic tracers allows for a better understanding of how precipitation and temperature changes affect the advance and retreat of large lakes on the Altiplano, and ultimately a more accurate understanding of how decadal-to-millennial forcings influence the climate of the subtropical Andes.

Dedication

This dissertation is dedicated to my wife, Danielle Nunnery, who gives me love and support no matter the weather, and who always reminds me of the truly beautiful and important parts of life.

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Acknowledgements

I first and foremost wish to acknowledge my committee members, Professors Paul Baker, Sherilyn Fritz, Bruce Corliss, Gary Dwyer, and Avner Vengosh for their advice and helpful critiques throughout this process. I am especially grateful to Paul Baker, my advisor, for giving me the opportunity to work on the Bolivian Altiplano, and for pushing me to become a better researcher and writer. I would also like to acknowledge Catherine Rigsby and Gonzalo and Alavaro Mollericon whose support and expertise on the Bolivian Altiplano made fieldwork possible. The fieldwork for this project was funded in part by the Geological Society of America. I thank Dr. Jonathan Karr of the Duke University Environmental Stable Isotope Laboratory for processing samples for carbon and nitrogen isotope analyses, and Nathaniel Warner, Laura Ruhl, Brittany Merola, and Alissa White for their assistance with various laboratory work, from instruction on preparation methods to processing samples for Sr analysis on the TIMS. I am very grateful to Michael Coe at the Woods Hole Research Center for his support with the THMB model and advice on setting up a hydrologic model of the Altiplano. I also thank my fellow graduate students in the Baker Lab, Alana Belcon and Trevor Nace, for advice, moral support, and friendship. Finally, the work that went into this thesis would not have been possible without the love and support of my family. I am forever grateful to my wife, Danielle, my parents, Dr. Alice Nunnery and Dr. James

Nunnery, and my in-laws, Mary Jane Larsen and Ray Larsen. I am also grateful to Bill Taylor for teaching me the value of doing work that is challenging, just for the sake of being challenged.

Introduction

Temperature in the tropical Andes increased by 0.11°C/decade during the period 1939-1998), and in the next century temperature is predicted to rise by another 1-2°C in the high Andes (> 2,500 meters above sea level)(Vuille et al., 2003). The result of this increase has already been observed with the rapid retreat of high altitude glaciers, as rainy season precipitation is unable to keep up with an ever-lengthening melting season (Ramiraz et al., 2001; Francou et al., 2003; Kaser et al., 2005; Coudrain et al., 2005). As temperature increases in the coming century there will also be loss of stored water as lakes contract, which will in turn deprive surrounding regions of the benefit of lake effect precipitation, eventually leading to aridification. This has important implications for the inhabitants of these regions, which rely heavily on glacial meltwater and mountain lakes to supplement rainfall during the dry season (Bradley et al., 2006). The Altiplano of Peru and Bolivia is one region that may be particularly sensitive to the effects of temperature increase, as there is a large population that relies on water from Lake Titicaca, which straddles the border of Peru and Bolivia, for hydropower, agriculture, and consumptive use. Paleolake sediment records of Lake Titicaca water level (e.g. Wirrmann et al., 1988 and 1982; Mourguiart et al., 1995; Wirrmann and Mourguiart, 1995; Abbott et al., 1997; Seltzer et al., 1998; Cross et al., 2000; Baucom and Rigsby, 1999; Baker et al., 2001a; Cross et al., 2001; Rigsby et al., 2005) have shown that during the mid-Holocene, temperature increase and precipitation decrease likely caused

Lake Titicaca to fall below its outlet, cutting off southward flow to the Rio Desaguadero and significantly decreasing transport of water to the south. According to Núñez et al. (2002) the human response to this aridification was to abandon the region. While abandonment of the Altiplano in the next century is unlikely the prospect for Lake Titicaca falling below its overflow due to enhanced evaporation is very real. To gain a better understanding of what impact a 1-2°C temperature increase will have on the hydrology of the Altiplano, as well as the communities within, it is helpful to reconstruct past hydrology and determine how previous hydrologic variability related to climate variability. The main motivation of this research is to investigate how hydrology has changed during previous glacial and interglacial periods and determine how hydrology of the Altiplano is likely to respond to future predicted climate change.

Extensive research of basin geomorphology and analysis of lacustrine sediment cores from regions throughout the Altiplano indicates the rise and fall of Lake Titicaca in the north, and advance and retreat of multiple large lakes in the central and southern regions during the late Quaternary. Methods utilized to investigate changing hydrology include: analysis of paleoshoreline deposits (e.g. Servant and Fontes, 1978; Placzek et al., 2006); analysis of sediment cores from Lake Titicaca (e.g. Wirrmann et al., 1982 and 1988; Mourguiart et al., 1995; Abbott et al., 1997; Cross et al., 2000; Baker et al., 2001a, 2005; Fritz et al., 2007, 2010, 2012), the Rio Desaguadero valley (Baucom and Rigsby, 1999; Rigsby et al., 2005), and the Coipasa and Uyuni salt flats in the southern region (e.g.

Risacher and Fritz, 2000; Baker et al., 2001b; Fritz et al., 2004); and seismic surveys (e.g. Seltzer et al., 1999; D'Agostino et al., 2002). There have also been several hydrologic modeling studies constructed to provide quantitative constraints on paleo-temperature and paleo-precipitation during the paleolake stages (e.g. Kessler, 1984; Hastenrath and Kutzbach, 1985; Blodgett et al., 1997; Condom et al., 2004; Blard et al, 2009). The mechanisms that determine how large hydrologic basins, such as the Altiplano, develop through time are complex. The amount of research that has gone into understanding how the Altiplano has evolved is vast, but necessary to develop a refined history of the dynamics of the region. No single lacustrine core record fully indicates how individual basins throughout the Altiplano have responded to changes in precipitation and temperature, just as no paleoshoreline can yield a continuous record of lake deposition. To reconstruct a more complete history of a large internally draining basin, such as the Altiplano, it is necessary to draw inferences about hydrology and climate from geochemical and lithological records throughout the region. The main objectives of this study are:

1. To utilize paleolake records to reconstruct hydrology throughout the Peruvian and Bolivian Altiplano during the last glacial period.
2. To simulate hydrologic variability using a hydrologic model to determine what changes to precipitation and temperature are necessary to maintain lake levels observed in paleolake records.

3. To determine, using paleolake records and results of hydrologic modeling, how changes in basin morphology and climate (both local and global) affected the rise and fall of large lakes in the southern basin of the Altiplano.
4. To use the information gathered from these paleohydrologic studies to infer how predicted changes in global climate might affect hydrology of the Altiplano in the future.

Chapter 1 examines paleolake records from the Bolivian Altiplano to present a refined history of lake level variability for the last ~50,000 years. A new sediment core recovered from beneath the Salar de Coipasa, a large salt flat in the Bolivian Andes, indicates continuous presence of a lake that has fluctuated between deep and shallow phases for the last 45,000 years. Lacking sufficient calcium carbonate for analysis, I instead take advantage of the general correlation between $\delta^{18}O$ and $\delta^{13}C$ in closed basin lakes (Talbot 1990; Baker et al., 2009) to deduce water balance using $\delta^{13}C$ from organic carbon. This reconstruction is validated with diatom paleoecological records.

In chapter 2, I discuss simulations using a terrestrial hydrology model in order to determine the amount of precipitation and temperature change necessary to create and sustain large lakes in the basins of Lago Poopó, the Salar de Coipasa, and the Salar de Uyuni in the southern Altiplano. By comparing model results to paleoclimate records from throughout the region I am attempting to determine more precisely the climate conditions during lake advance and retreat, and ultimately to gain a better

understanding of how decadal-to-millennial forcing influences the climate of the tropical Andes. The model is run through many scenarios with different combinations of precipitation and temperature.

In chapter 3, I discuss the mechanisms involved in a state change observed in a sediment core from the Salar de Uyuni from a "playa lakes" phase to a "great lakes" phase. At Salar de Uyuni, large paleolakes characterized wetter and/or colder periods (signaled in the salar by mud deposition) and alternated with shallow saltpans characteristic of drier and/or warmer periods (signaled in the salar by deposition of gypsum and halite). Previously published work indicates that the only long-lived lakes in the basin occurred during the past 60,000 years. Strontium (Sr) isotopic data, on both salt and lacustrine mud units in the salar drill core, are used to determine changes to source waters. In this study I also investigate whether this state change was caused by changes in basin morphology, the result of a trend toward wetter conditions throughout the tropical Andes, or a combination of both.

Each of these studies utilizes a specific set of tools to determine how hydrology on the Altiplano changed through time, and how those changes relate to regional and global scale climate variability. The isotopic analysis of a sediment core from the Salar de Coipasa, compared with an earlier record from the Salar de Uyuni, provides more insight into the smaller scale hydrologic dynamics between the Salar de Coipasa and the Salar de Uyuni. Combined with paleolake records from Lake Titicaca and the Salar de

Uyuni, and paleoshoreline records from the central and southern basins, the record from Coipasa also enables the development of a latitudinal paleohydrologic profile of the Altiplano, which allows for a better understanding of how regional and global climate has affected the basin as a whole. Using a terrestrial hydrologic model I am able to determine several possible climate change scenarios necessary to create the various hydrologic states inferred from the paleolake records. And finally, the use of Sr isotopic compositions on carbonates and halites from the 220-meter core from the Salar de Uyuni enables the investigation of the cause of an apparent state change indicated in core lithology. The combination of all of the methods discussed above allows for a more realistic reconstruction of basin paleohydrology, and a better understanding of how the basin responds to regional and global climate variability. The results from these experiments also provide more insight into how the hydrology of the Altiplano will respond in the future as temperatures continue to increase.

1. Paleohydrologic reconstruction of large lakes on the Bolivian southern Altiplano based on a sediment core from the Salar de Coipasa

1.1 Introduction

The Altiplano is a large internally drained basin located in the tropical Andes of Peru and Bolivia. It is home to Lake Titicaca, the world's largest high-altitude lake, and the Salar de Uyuni and the Salar de Coipasa; the former is the world's largest salt flat (Figure 1). The basin has experienced dramatic long-term changes in hydrology, from the expansion of glaciers and large lakes to extreme aridity. Early scientific studies in the basin reported old lacustrine deposits and coral-like benches several meters above the plain of the Salar de Uyuni and the Salar de Coipasa in the south, as well as high shorelines around Lake Titicaca; all these were considered evidence of ancient lakes much larger than those of today (e.g. Forbes 1861, Agassiz 1875, Musters 1877, Minchin 1882, Pompecki 1905, Bowman 1914). Considerable research has explored how long-term hydrologic changes of the Altiplano were influenced by local and global climate processes (Baker et al., 2001a, 2001b; Fritz et al., 2007, Fritz et al., 2010; Sylvestre et al., 1999; Mourguiart et al., 1995; Servant et al., 1977; Bills et al., 1994; Placzek et al., 2006, 2010; Risacher and Fritz, 1991; Rigsby et al., 2005). The purpose of this study is to present a refined history of hydrologic variability on the Altiplano through new high-

resolution geochemical and biotic analyses of a sediment core from Salar de Coipasa and the integration of these new data with prior studies.

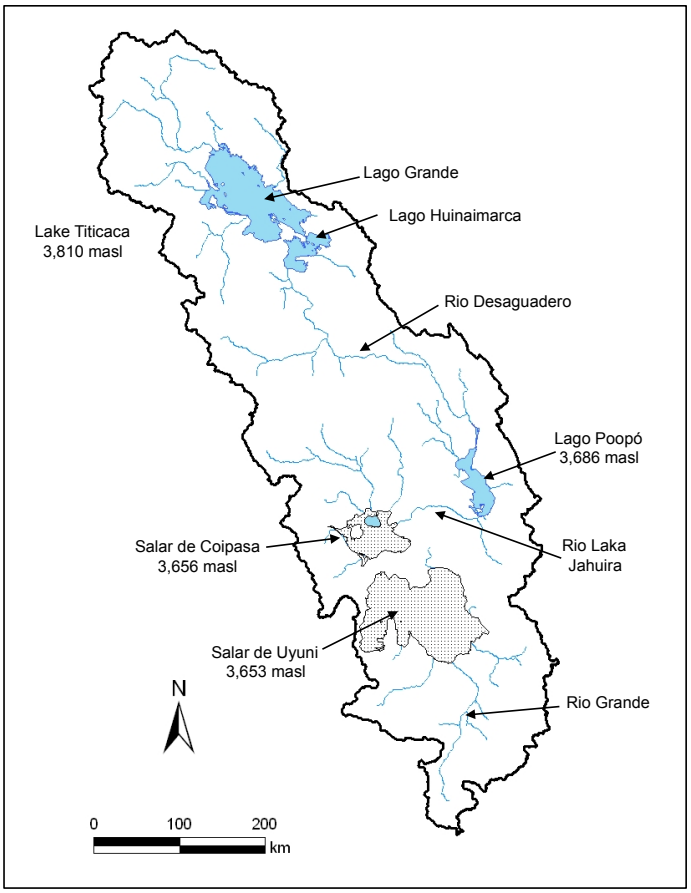

Figure 1: Map of the Peruvian/Bolivian Altiplano showing locations of major hydrologic features.

Evidence for significant hydrologic change through time is present throughout the basin. In the north, sediments of apparently lacustrine origin suggest at least three periods of significant lake expansion - designated as Lake Mataro (3,950 masl), Lake

Cabana (3,900 masl), and Lake Ballivian (3,860 masl); the ages of these have not been firmly established. However, based on fossil analysis of each unit (Pompecki 1905; Hoffstetter et al., 1971) and K-Ar dating of volcanic ash beneath the oldest of these deposits (Lavenu 1984, 1986), the lakes may be of early to mid-Pleistocene age and are attributed to glacial intervals within this time period (Hoffstetter et al., 1971; Ahlfeld and Brasnia 1960; Servant and Fontes 1978; Servant 1977). If the northern and southern Altiplano were connected hydrologically during intervals of lake expansion, the largest of these early lake stages (Lake Mataro) would have created a lake of area 75,000 km^2 with a maximum depth of 420 m. Thus far, there is no evidence in the southern basin for any such paleolake. There are at least three possible reasons for this: 1) The sediment units attributed to lakes Mataro, Cabana, and Ballivian are not truly lacustrine and therefore do not represent an interval of lake expansion; 2) the northern and southern sub-basins were not hydrologically connected during these intervals; or 3) evidence for these lake phases in the southern basin has been lost through erosion.

On the southern Altiplano, paleoshoreline deposits from around the basin (Ahlfeld 1972; Servant and Fontes 1978 and 1984; Wirrmann and Mourguiart 1995; Bills 1994), sediment cores from the Salar de Uyuni (Fornari et al., 2001; Baker et al., 2001; Fritz et al., 2004), and drill cores from the Rio Desaguadero valley (Rigsby et al., 2005) indicate two major phases of lake advance younger than the Ballivian stage in the north: Lake Minchin (36,000-46,000 cal yr BP) and Lake Tauca (15,000-26,000 cal yr BP)(Baker et

al., 2001). Shoreline deposits and core records indicate that Lake Minchin was sub-divided into three distinct lacustrine intervals interrupted by dry periods, with shoreline elevations that vary from approximately 3,700-3,712 masl and a possible maximum of 3,730 masl (Wirrmann and Mourguiart 1995). Radiocarbon dating of paleoshorelines deposited during the Tauca phase suggests a deeper (~110 m depth) lake than Minchin, with surface water reaching a maximum of 3,760 masl (Bills 1994; Wirrmann and Mourguiart 1995).

There is some controversy over the ages of paleolake phases, especially for those outcrops that are considered to represent Lake Minchin. Based on radiocarbon dates of shells found in outcropping sediments, Servant and Fontes (1978) estimated the age of Minchin at 30,000-32,000 cal yr BP. Rondeau (1990) used U/Th dating of algal bioherms that bracketed the Minchin interval between 30,000-73,000 cal yr BP. More recent radiocarbon dating of subsurface sediments from the Salar de Uyuni placed Minchin between 36,000-46,000 cal yr BP (Baker et al., 2001; Fritz et al., 2004). Yet Placzek et al. (2006) described the exposures typically attributed to paleolake Minchin as three completely separate lake phases. They reject previous radiocarbon dates as too young, and instead used a U/Th chronology of tufa shoreline deposits, re-designating the three phases Ouki (96,000-125,000 cal yr BP), Salinas (80,000-90,000 cal yr BP), and Inca Huasi (45,000-47,000 cal yr BP). These estimations of the timing of large lake expansion on the Altiplano have major implications for interpretations of the role of global climate

dynamics in the South American tropics. Dating discrepancies of the magnitude discussed above have the effect of shifting the correlation of a lake phase relative to various major climate forcings, drastically influencing interpretations of regional dynamics. Therefore, it is critical to have a clear record of the timing of lake advance and retreat in order to develop a more comprehensive understanding of how local climate variability may be tied to global-scale climate variability.

Here we report on a sediment core collected from the southern Altiplano, north of the Salar de Uyuni, at the Salar de Coipasa. We reconstruct lake variability and depth using stable isotope and diatom analyses. Coupled with a hydrologic model, which calculates precipitation amount based on changes in lake volume (Nunnery, 2012), this record allows for a more complete reconstruction of precipitation and hydrologic variability on the Altiplano.

1.2 Site Description

1.2.1 Geology

The Salar de Coipasa (68°8'W, 19°23'S), at an elevation of 3,656 masl, is the second largest salt flat (2,500 km^2) on the central Bolivian Altiplano. The Altiplano is an 187,000 km^2 intermontaine internally-drained basin, located between the Cordillera Oriental (east) and the Cordillera Occidental (west) in the tropical and sub-tropical Peruvian and Bolivian Andes. The Cordillera Oriental is made up of ranges of N-S and

NW-SE trending mountains with maximum altitudes up to 6,000 masl and is primarily composed of intensely folded and faulted Paleozoic sediments, intrusive rocks, and very young ignimbrites (Risacher and Fritz, 1990; Montes de Oca, 1997). The Cordillera Occidental is primarily composed of Cenozoic volcanic rocks. A basement of Paleozoic sediments is overlain by fill of Cretaceous and Tertiary continental sediments (Risacher and Fritz, 1990).The Salar de Coipasa lies northwest of the Salar de Uyuni, and makes up part of the terminal basin of the Altiplano. Based on hydrological modeling, the two salars are hydrologically connected during periods of increased effective moisture (precipitation minus evaporation, P-E) (Blodgett et al., 1997). Geology surrounding the Salar de Coipasa consists of Late Tertiary to Quaternary age alluvial, volcanic, and evaporitic deposits (Banks et al., 2004; Risacher and Fritz, 1991; Clapperton, 1993). The salar is capped by a salt crust that has a thickness of 2.5 meters near its center and tapers to <0.5m at the salar margin (Risacher and Fritz, 1991). The salt is underlain by low permeability lacustrine sediments that consist of detrital material, gypsum, calcite, organic matter, and authigenic clay minerals containing interstitial brines (Ericksen et al., 1978; Risacher and Fritz, 1991; Sylvestre et al., 1998).

1.2.2 Climate and hydrology

Modern climate on the Altiplano is characterized by arid to semi-arid conditions, with rainfall ranging from 800-1,000 mm/yr in the north near Lake Titicaca, to <100

mm/yr in the south in the vicinity of the Salar de Uyuni. The majority of precipitation falls during the austral summer from December to March, with a long dry season lasting from April to November. Annual average precipitation at the Salar de Coipasa is 100-200 mm/yr. The major inflow to Coipasa is the Rio Lauca, with an annual discharge of 0.14×10^9 m^3 (Servicio Nacional de Meterologia Hidrologia de Bolivia), which today creates a shallow lake near the mouth of the river. During the rainy season the salar floods.

During phases of increased effective moisture Lake Titicaca rises above its outlet, at 3,704 masl, and overflows via the Rio Desaguadero into the central basin of the Altiplano, providing increased riverine input to the downstream Lago Poopó. Higher local runoff combined with enhanced flow from the north causes Lago Poopó to fill until surface water elevation is at approximately 3,700 masl. At this elevation, lake waters overflow, via the Laka Jahuira, westward to the Salar de Coipasa. The surface of Salar de Coipasa is only ~4 m above that of the Salar de Uyuni (Risacher and Fritz 1991). Based on topographic data, a surface water elevation of ~3,678 masl is sufficient to connect Uyuni and Coipasa, at which point the Poopó, Coipasa, and Uyuni basins form a series of lakes connected by broad straits (Figure 2). Once surface water rises to the level of Lago Poopó's outlet (3,700 masl), the three basins combine to form a single large lake.

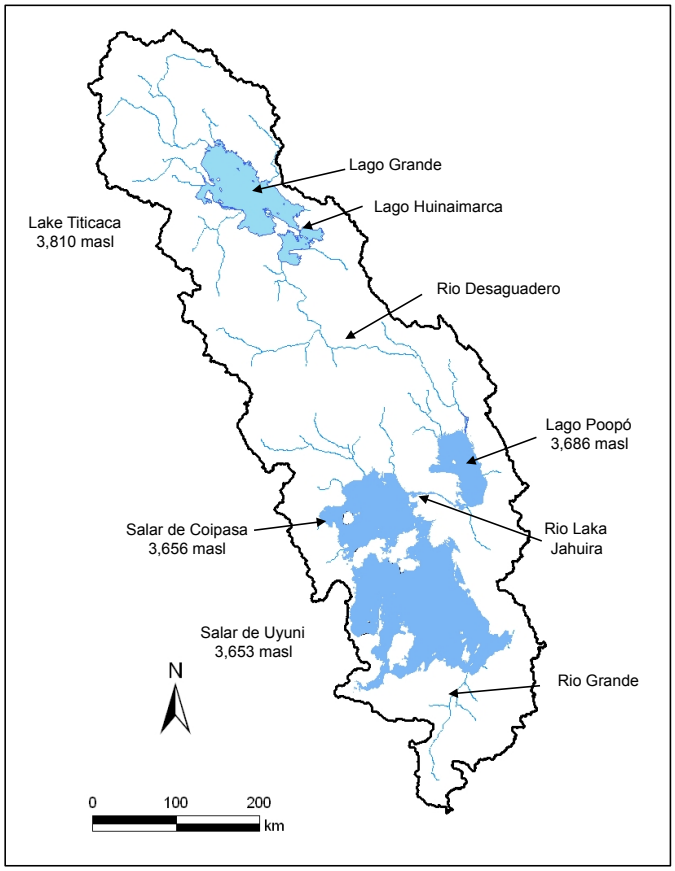

Figure 2: During periods of increased effective moisture the southern Altiplano basins of Poopó, Coipasa, and Uyuni connect via straits. The shaded regions in the southern basins represent lake levels at 3,700 masl.

A long drill core record from the Salar de Uyuni dates wet and dry phases (as shown by the alternation of mud and salt deposits respectively) on the southern Altiplano (Baker et al., 2001; Fritz et al., 2004). According to this record the Salar de Uyuni was wet during the Last Glacial Maximum (LGM). Given the low elevational gradient between the two basins, the reconstructed history of lake expansion from the

Salar de Coipasa should coincide with that in Salar de Uyuni, assuming no significant differences between modern and LGM hydrology.

1.2.3 Previous paleoclimate work

Many paleoclimate records, including those from lake and ice cores (e.g. Baker et al., 2001 and 2005; Thompson et al., 2000) speleothems (e.g. Cruz et al., 2005 and 2006; Wang et al., 2004, 2006, 2007) and paleo-oceanographic cores (e.g. Arz et al., 1998; Jennerjahn et al., 2002), provide evidence that during periods of cold North Atlantic sea surface temperature (SST), precipitation on the Peruvian/Bolivian Altiplano increased (Baker et al., 2009, 2001 a and b, and 2005; Cross et al., 2000; Fritz et al., 2007; Ekdahl et al., 2008; Servant and Fontes, 1978; Sylvestre et al., 1999; Seltzer et al., 2000; Rigsby et al., 2005). This is apparent in records spanning the last glacial period (20,000-60,000 Cal yr BP), including sediment records of alternating wet and dry conditions in the salars (Baker et al., 2001 a and b; Fritz et al., 2004) and speleothem records indicating rainfall variability (Cruz et al., 2005 and 2006; Wang et al., 2004, 2006, 2007) that are in phase with orbital-paced insolation maxima and minima. These records also indicate that there may have been large increases in precipitation connected to decadal-to-millennial scale fluctuations in northern hemisphere cold events, such as the Younger Dryas, Holocene Dansgaard-Oeschger (D-O) oscillations, Heinrich 1 (H1), and the Bølling-Allerød (B-A) warming event (e.g. Baker et al., 2001, Fritz et al. 2010).

Paleoclimate data for the Salar de Coipasa is limited, as most research has focused on the sediment and brine chemistry of Salar de Uyuni (Ericksen et al., 1978; Risacher and Fritz, 1991 and 2000) and on southern Altiplano paleolake formation (Blodgett et al., 1997; Cross et al., 2002; Placzek et al., 2006). Sylvestre et al. (1998) reported on diatom analysis of a 5-meter long core recovered from the southern part of the Salar de Coipasa, which provided a continuous sediment record for the interval 17,000-21,000 ^{14}C yr BP.. Results from this study suggest that a shallow lake continuously covering the Salar de Coipasa during the LGM, with a high degree of variability in lake level and salinity throughout the record .

Hydrologic-modeling studies of the Altiplano (Blodgett et al., 1997, Cross et al., 2002; Hastenrath and Kutzbach, 1984; Condom et al., 2004; Blard et al., 2009) conclude that the Salar de Coipasa is connected to the Salar de Uyuni during wet periods through increases in precipitation, decreases in evaporation, or a combination of the two. Evidence from paleoshoreline deposits supports connection of the two basins (e.g. Ahlfeld 1972; Servant and Fontes 1978 and 1984; Wirrmann and Mourguiart 1995; Bills 1994). Therefore, lacustrine sediments beneath the salt crust at the Salar de Coipasa should be well correlated with lacustrine intervals evident in the stratigraphic record from the Salar de Uyuni.

1.3 Methods

1.3.1 Core collection

In July 2010, we collected an ~14 meter long core from beneath the salt at the Salar de Coipasa using a Russian peat corer. Each section was recovered from the same borehole, because of the difficulty of penetrating the upper salt crust. We first broke through 30 cm of halite salt cap using a hand pick, retaining a portion of the cap for analysis. We collected mud from directly beneath the crust and began hand-coring sediments to a depth of 94 cm, at which point we encountered a 16 cm thick salt bed. Below this was ~13 meters of continuous sediment. Throughout the coring procedure we re-entered the same borehole, collecting sixteen 0.25-1 meter drives (SC_0 through SC_16) for a total of 13.84 meters. The cores were photographed and described on site before being wrapped for shipment and storage. More detailed core description was carried out in the laboaratory at Duke University.

1.3.2 Chronology

Accelerator mass spectrometry (AMS) radiocarbon measurements were made on 17 bulk organic samples and 1 macrophyte sample at the National Ocean Sciences Accelerator Mass Spectrometry Facility (NOSAMS) at Woods Hole Oceanographic Institute (Table 1). Dates were calibrated using Fairbanks0102 (Fairbanks et al., 2005) for all samples. Ages are corrected for stable isotope fractionation but are not adjusted for

reservoir effect, which for this region is considered to be insignificant (Fritz et al., 2004; Baker et al., 2001, and Sylvestre et al., 1999) and, in any event, is unknown. ^{14}C age determinations indicate the 13.84 meters of sediment core represent approximately 42,000 calendar years before present (cal yr BP). An age model was constructed using polynomial equations derived from the ^{14}C age determinations (Figure 3).

Table 1: Radiocarbon ages were calibrated using Fairbanks0107 calibration curve (Fairbanks et al., 2005)

Sample number	Sample depth (m)	$\delta^{13}C$ Source	Radiocarbon age ^{14}C yr BP	Age error	Calibrated age cal yr BP	1 std dev
SC0-30	0.30	-17.32	925	15	847	38
SC1-10	0.40	-16.67	2450	15	2530	98
SC2-8	0.85	-16.6	2610	15	2740	7
SC3-110	1.10	-20.99	4700	25	5415	65
SC4-46	1.66	-24.95	10400	50	12269	117
SC5-50	2.70	-20.41	16350	110	19468	121
SC6-86	4.06	-20.2	13450	75	15658	138
SC7-86	5.06	-18.93	17500	130	20694	175
SC8-50	5.70	-18.26	18800	150	22391	148
SC9-50	6.70	-21.1	20300	190	24208	209
SC10-50	7.70	-20.29	20800	200	24836	304
SC11-50	8.70	-18.64	20500	200	24429	252
SC11-60M	8.80	-11.62	20400	180	24311	206
SC12-50	9.70	-19.3	28500	550	33887	591
SC13-50	10.70	-25	36600	2500	41722	2272
SC15-50	12.70	-19.82	35200	2100	40459	1962
SC16-98	13.82	-20.33	37800	2900	42745	2810

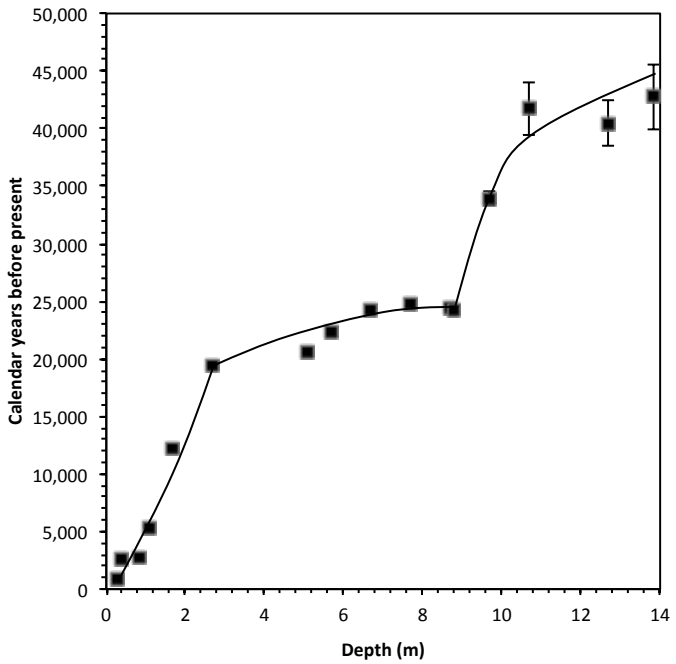

Figure 3: Relationship between core depth and calendar age as established by accelerator mass spectrometry ^{14}C analysis of organic carbon

1.3.3 $\delta^{13}C$ and $\delta^{15}N$ isotopes

Samples were analyzed for $\delta^{13}C$ and $\delta^{15}N$ isotope fractions and % carbon and nitrogen at the Duke environmental stable isotope lab. The core was sub-sampled at 2-cm resolution. Each sample was dried at 60°C, powdered, homogenized, and weighed into silver cups, and transferred into a desiccator containing a beaker of 12N HCl (Schubert and Nielsen 2000). The samples were allowed to acid fume for at least 72 hours to remove any sedimentary carbonate. Samples were dried, transferred into pressed

5x9mm tin caps, and combusted at 1,020°C in a Carlo-Erba NA1500 elemental analyzer. Purified, chromatographically separated N_2 and CO_2 gases were delivered to a ThermoFinnigan Delta+XL continuous flow isotope ratio mass spectrometer via a ThermoFinnigan Conflo III interface. Reproducibility for both isotopes is approximately 0.1 per mil at 1 s.d.

1.3.4 %Calcium

Weight % calcium was determined by digesting powdered samples in 1 N hydrochloric acid and analyzing the leachate for Ca^{2+} by flame atomic absorption spectroscopy (AAS) (Perkin Elmer 5000).

1.3.5 Diatom analysis

Diatom samples were analyzed at the University of Nebraska-Lincoln at 8-cm resolution throughout the length of the core. Samples were processed by treatment with 10% HCl to dissolve carbonates, followed by a cold hydrogen-peroxide treatment to oxidize organic matter. The prepared samples were rinsed repeatedly to remove oxidation by-products. Slurries of the prepared sample were dried onto coverslips, and the coverslips were mounted onto slides with Naphrax. Diatoms were counted at 1000x magnification on a Zeiss Axioskop 2 microscope with differential interference contrast. Where possible, at least 300 diatom valves were identified and counted. In samples

where diatom concentration was low, at least one whole slide was scanned for diatom enumeration. Diatom abundance is expressed as a percentage relative to the total number of counted valves.

1.4 Results

1.4.1 Stratigraphy

The Salar de Coipasa is topped by a salt cap that is as much as 2.5 meters thick at some locations (Risacher and Fritz, 1991). At the core site the crust is ~30 cm thick and consists of coarse halite crystals (Figure 4). Just below the crust, from 30-55 cm, lies a bed of dark brown mud with large salt crystals. From 55-77 cm we have no core recovery. From 77-94 cm sediments consist of dark to light brown mud with large halite crystals. A second hard salt bed, also consisting of coarse crystalline halite, is present below the mud between 94-110 cm. From 110-192 cm dark to light brown mud with fewer halite crystals is present, which becomes increasingly compacted toward the bottom of the unit. From 192-482 cm the core alternates between green, brown, and black muds with laminations averaging between <1 to 2 mm in thickness. For the intervals 337-400 cm and 420-482 cm we have no core recovery. Between 482-513 cm there is a dark brown mud bed with no laminations present. Sediments from 513-878 cm consist of green, gray, and black muds with 1-2mm laminations. There are several ash

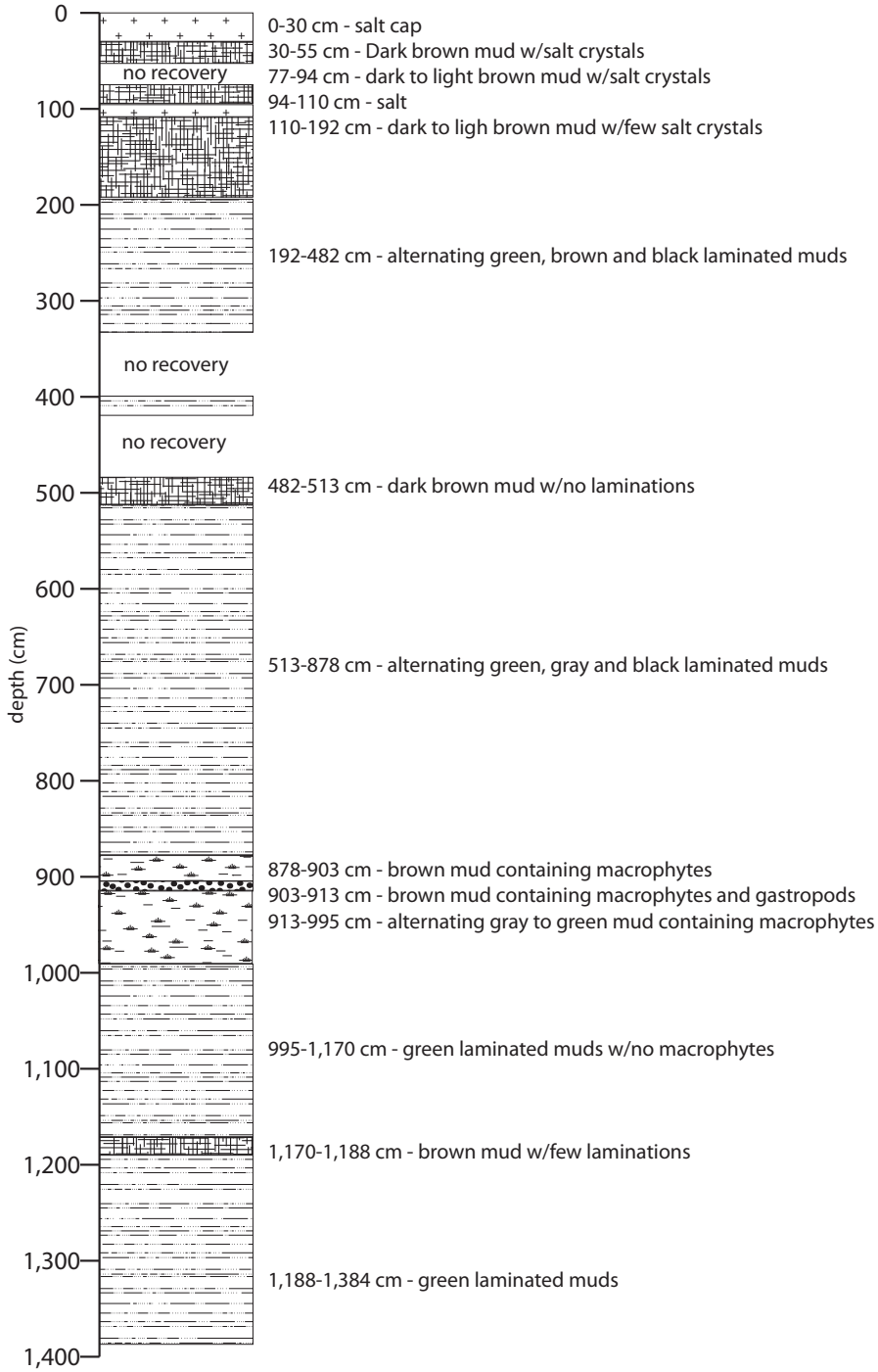

Figure 4: Sediment core log for the Salar de Coipasa. Total depth 13.84 meters below the surface of the salt cap.

lamina in this interval ranging in thickness from 1 mm to 1 cm located at the following core depths: 595 cm, 600 cm, 603 cm, 611.5 cm, 612 cm, 613 cm, 616.5 cm, 619 cm, and 765 cm. Between 878-903 cm there is brown laminated mud bed containing macrophytes. Immediately below this bed, between 903-913, is a unit of brown mud containing macrophytes and gastropods. Under this unit, from 913-995 cm, are alternating gray to green mud beds containing macrophytes. From 995-1,170 cm green laminated muds, containing no macrophytes, continue with the appearance of small (0.5-1 mm) tan beds throughout: 1,037 cm, 1,042 cm 1,044 cm, 1,049 cm, 1,062 cm, 1,086 cm, 1,091 cm, 1,124 cm, and 1,140 cm. The sediment units in the remainder of the core consist of brown muds with few laminations, between 1,170-1,188 cm, and green laminated muds, from 1,188 cm to the base of the core at 1,384 cm.

1.4.2 Chronology

Reconstruction of the sedimentation history of the Salar de Coipasa is based upon ^{14}C radiocarbon dating of 17 bulk organic sediment samples and 1 macrophyte sample. During the earliest interval in the core, between 34,500-45,000 cal yr BP, sedimentation rates were high with a maximum rate of 0.46 cm/yr (Figure 3). From 25,000-34,500 cal yr BP sedimentation slowed significantly with rates ranging from 0.009-0.015 cm/yr. At 25,000 cal yr BP sediment deposition rose to a high rate of almost 1

cm/yr followed by a steady decline over the next 5,700 years to a minimum of 0.062 cm/yr at 19,300 cal yr BP. Between 19,300-1,000 cal yr BP sedimentation rates slowed, depositing between 0.005-0.018 cm/yr. Based on ^{14}C age dating of muds from directly beneath the salt crust, there has been a period of desiccation at the Salar de Coipasa for the last ~1,000 cal yr BP.

1.4.3 $\delta^{13}C$ Isotopes

$\delta^{13}C$ values in the Salar de Coipasa core range from -3.3‰ to -24.7‰ with a mean value of -18.38‰ (Figure 5)(Appendix A) A cross-plot of $\delta^{13}C$ vs. C/N (Figure 6) suggests the core is composed primarily of algal carbon, which generally has $\delta^{13}C$ values between -15‰ and -25‰ (Cross et al., 2000; Rowe et al., 2002).

$\delta^{13}C$ values are most variable between 8 and 14 m (~24-45 Cal kyr BP), as compared to units higher in the core. Within this interval there are two distinct intervals that exhibit markedly enriched $\delta^{13}C$ values: 8-10 m (~24-35 Cal kyr BP) and 10.5-11.5 m (36.5-39 Cal kyr BP). Samples taken between 8-10 m correspond to mud beds containing rooted macrophytes, which generally have more positive $\delta^{13}C$ (>-15‰) indicating shallow water environments. A macrophyte sampled at 8.8 m depth (~24,300 cal yr BP) had a $\delta^{13}C$ value of -11.62‰. Based on ^{14}C age dating, this unit is contemporaneous with

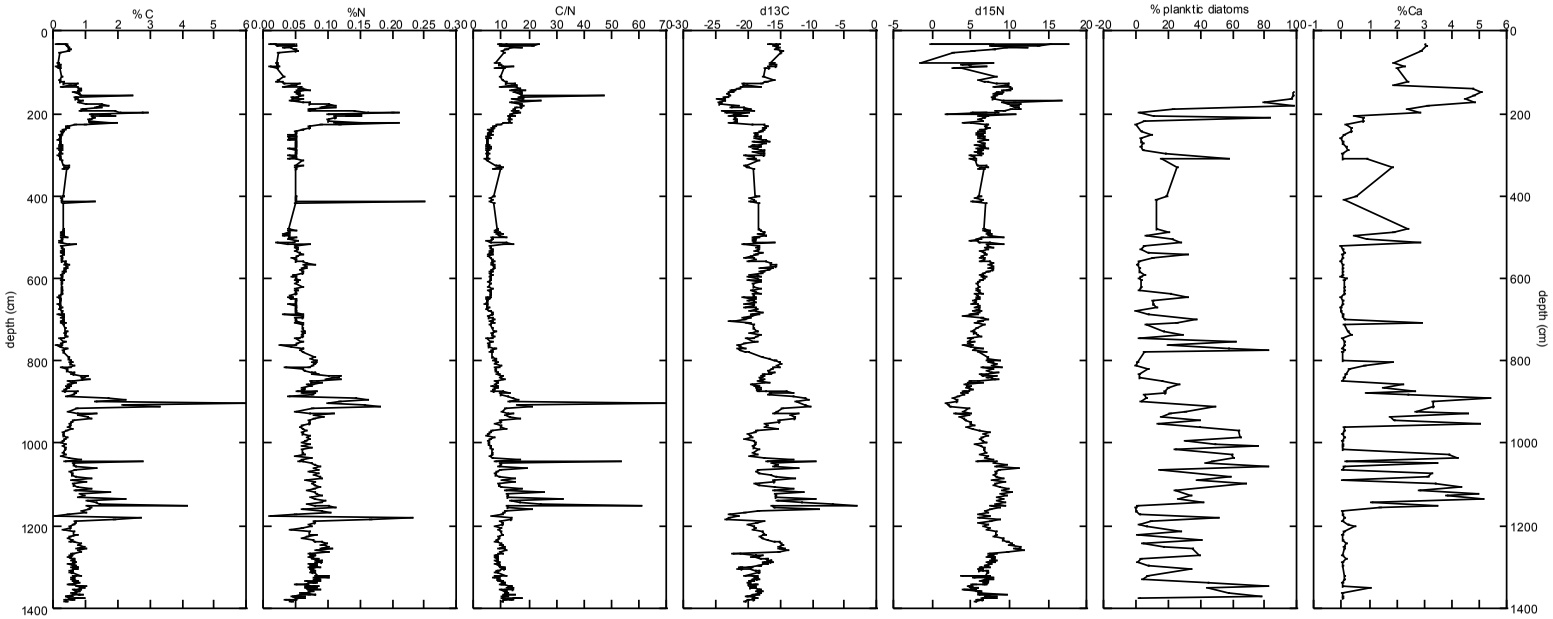

Figure 5: Geochemical measurements in organic muds from the Salar de Coipasa versus depth including: % total organic carbon, % total nitrogen, carbon versus nitrogen ratio (C/N), $\delta^{13}C$, $\delta^{15}N$, % planktic diatoms and % calcium.

a period of salt deposition in the Salar de Uyuni (Baker et al., 2001; Fritz et al., 2004). Samples taken between 10.5-11.5 m (36.5-39 Cal kyr BP) do not correspond with macrophyte-bearing mud. More positive $\delta^{13}C$ in this interval is likely due to lower algal production and increased degassing of CO_2, both products of a lake becoming less voluminous and more saline. $\delta^{13}C$ values for samples taken between 2.5-8 m (17.9-24 Cal kyr BP) are notably less variable than during the preceding interval, ranging between -21‰ and -18‰.

From 1.8-2.25 m (11.5-15.5 Cal kyr BP) there is a decrease of ~5‰ toward more negative $\delta^{13}C$ values, which fall generally between -19‰ and -24‰. At ~1.8 m slightly more positive $\delta^{13}C$, values (-19‰ to -20‰) persist for a brief interval (1.2-1.8 m) before $\delta^{13}C$ once again shifts to more negative values with a minimum of -24.7‰. This interval of more negative $\delta^{13}C$ is coincident with an increase in % planktic diatoms of almost 100%, suggesting a deep lake. For the uppermost section of the core (0-1.2 m) $\delta^{13}C$ values stabilize between ~ -15 and -17‰. Sample resolution seems to be sufficient to capture the real variability in the lake record. There are no single point maximum or minimum $\delta^{13}C$ values (except for macrophyte bearing samples).

1.4.4 $\delta^{15}N$ Isotopes

Nitrogen isotopes in the core range from 0-17.6‰ with an average of 6.83‰ (Figure 6)(Appendix A). There appears to be some positive correlation with $\delta^{13}C$ values for the intervals 2.5-8 m (17.9-24 Cal kyr BP) and 10-14 m (35-45 Cal kyr BP), which are each units of laminated sediments. $\delta^{15}N$ in sediments for the intervals 0-2.5 m (non-laminated mud, 0-17.9 Cal kyr BP) and 8-10 m (macrophyte bearing mud, 24-35 Cal kyr BP) appear to have a negative correlation with $\delta^{13}C$. $\delta^{15}N$ values between 1-2.5 m are high (~10‰), coincident with an excursion to more negative $\delta^{13}C$ and increased % planktic diatoms, %Ca, C/N, TOC, and TN. Shifts in $\delta^{15}N$ in paleolimnological records can have several causes, including changes related to rates of nitrogen fixation by algae, nitrification or denitrification of the lake, ammonia volatilization, or change in nitrogen source (terrigenous vs. marine) (Talbot 2001).

1.4.5 C/N

The majority of C/N values in the Coipasa core fall between 5-10 (Figure 6)(Appendix A). This is a range similar to that found in phytoplankton (C/N 6-7), which are relatively N rich (Talbot 2001). Two C/N spikes >50 between 10-12 m (35-40 Cal kyr BP) are the apparent result of an increase in TOC, and coincide with an increase in %Ca and more positive $\delta^{13}C$ values, possible indicators of a decrease in lake volume. A C/N

spike of 68 at 9 m (26.8 Cal kyr BP) is likely due to an increase of organic carbon from macrophyte-bearing muds found at this depth. Between 1-2.5 m (6.1-17.9 Cal kyr BP) there is a unit of non-laminated muds with elevated C/N values that is similar in structure to rises in % planktic diatoms, %Ca, TN, TOC, and $\delta^{15}N$ within the same interval. These elevated values of C/N generally fall between 10-17, with one spike of 47

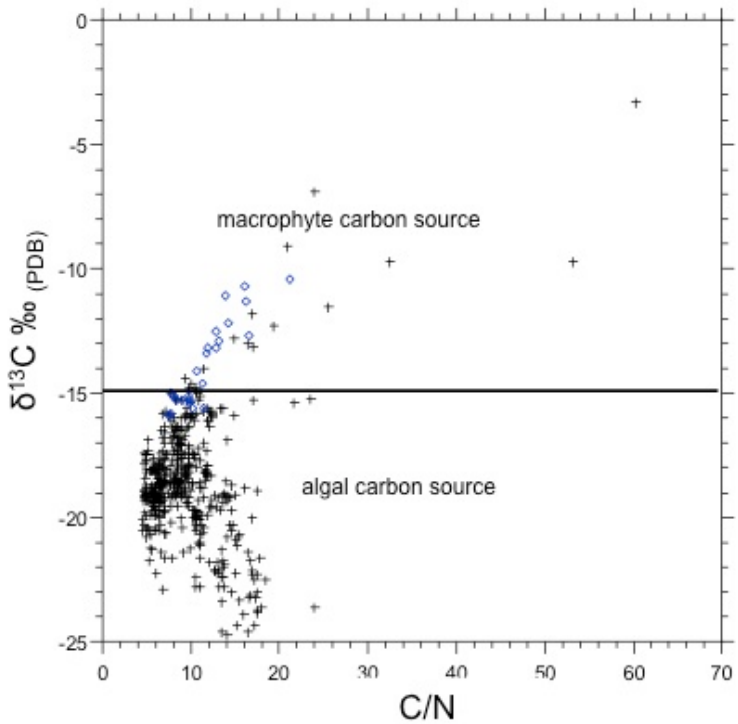

Figure 6: Cross-plot of C/N versus $\delta^{13}C$ of organic carbon samples, and several lacustrine macrophytes from the Salar de Coipasa. The majority of samples are of algal origin ($\delta^{13}C$ < -15 ‰), and are designated by cross markers. Samples within the unit of mud containing rooted macrophytes (800-1,000 cm) are designated by diamond markers.

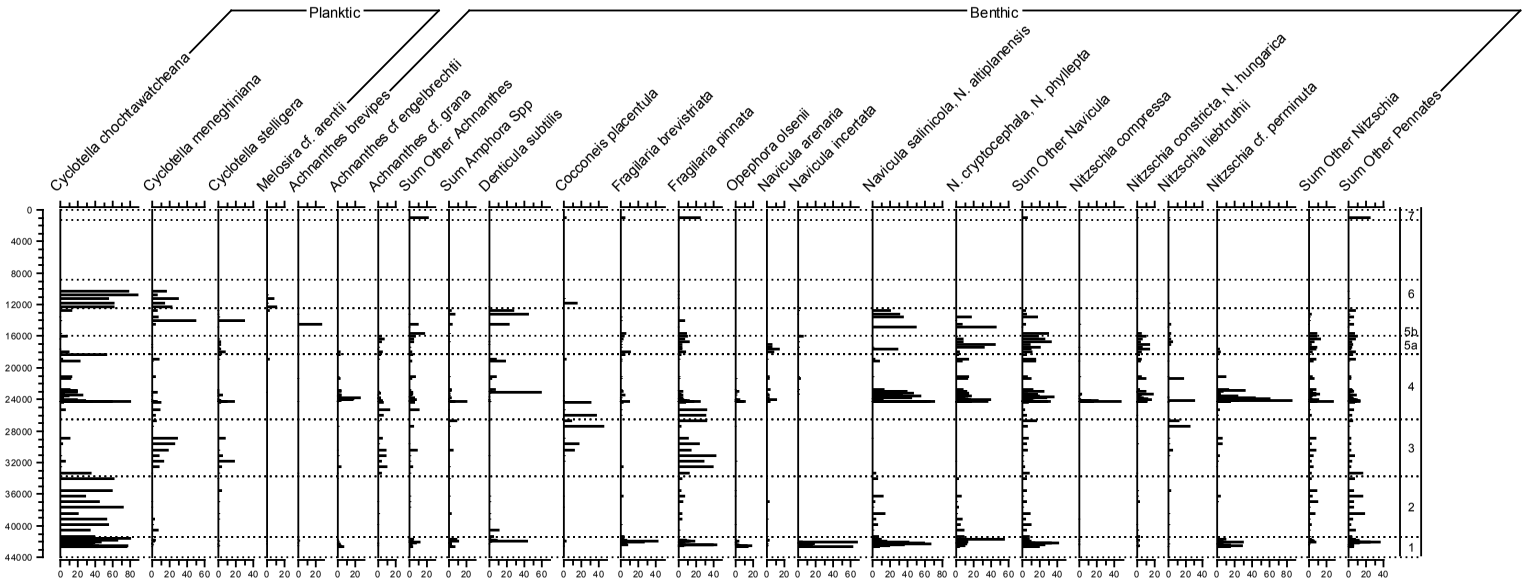

Figure 7: Stratigraphy (percent abundance) of the common (> 5%) diatom taxa in the Salar de Coipasa core. Note that the mud unit between zone 6 and 7 was barren of diatoms.

at 1.6 m (9.8 Cal kyr BP). The coincidence of increased C/N with a shift toward more negative $\delta^{13}C$ values from 1-2.5 m (6.1-17.9 Cal kyr BP) suggests increased algal contributions to carbon sedimentation, as terrestrial carbon would have a more positive $\delta^{13}C$ value. This assumption is bolstered by the large increase of planktic diatoms to nearly 100% within this unit, indicating a relatively deep and fresh lake.

1.4.6 Diatoms

The diatom flora of Salar de Coipasa consists of a mixture of planktic and benthic species, characteristic of variable depth and salinity (Figure 7). The most common planktic taxon is *Cyclotella chochtawatcheana*, which occurs in abundance in deep stably stratified saline lakes (Fritz et al. 1993), brackish estuaries, and in nearshore marine systems (Prasad 1990). Also present in moderate abundance is *Cyclotella meneghiniana*, which grows in lakes of varied depth and salinity but is common in shallow systems. *Cyclotella stelligera* is characteristic of fresh to hyposaline (<10 g L-1) waters and occurs in lakes that are of shallow to moderate depth, as well as in nearshore regions of large deep lakes (Tapia et al. 1993). In the planktic-dominated intervals of the Salar de Coipasa core, we interpret high relative abundance of *C. chochtawatcheana* to represent a deep chemically stratified lake. Salinity is likely lower in the intervals with a high abundance of *C stelligera*.

The attached diatom flora of Salar de Coipasa is diverse. It includes tychoplanktic species, such as *Fragilaria brevistriata* and *F. pinnata*, which can be entrained into the plankton by mixing. The attached benthic species include several taxa characteristic of higher salinity, such as *Denticula subtilis* and *Achnanthes brevipes*; *Cocconeis placentula* which grows on macrophytes; and high representation by species in the genera *Navicula* and *Nitzschia*.

Diatoms are absent from most of the Holocene samples in the upper part of the core, except for the uppermost sample (diatom zone 7), which dates from ~1000 cal BP. The barren samples are those with high to moderate concentrations of halite. Immediately below this, the high representation of *C. choctawatcheana* and the low abundance of benthic diatoms suggest that from ~9000 to 13000 cal yr BP (zone 6) a deep chemically stratified lake occupied the basin. Benthic diatoms dominated most of the period from 13,000 to 18,000 cal yr BP (zone 5), which suggests generally shallow conditions, including some intervals of higher salinity when the relative abundance of *Denticula subtilis* was high. One sample within zone 5, dated at 16,000 cal yr BP, is distinctive in the high percentages of the planktic species *C. stelligera* and *C. meneghiniana*. This suggests a moderately deep freshwater lake and is correlative with the paleolake Tauca phase in the basin. During the LGM (~25000 – 18,000 cal yr BP, zone 4), variation in the relative abundance of planktic and benthic taxa suggests that lake

level fluctuated from moderately deep in the *C. chochtawatcheana*-dominated zones to shallower in samples with higher relative abundance of benthic diatoms. Samples are absent from ~30,000 to 25,000 cal yr BP, which is an interval of very slow sedimentation when the basin was likely a very shallow lake. From 30,000-34,000 cal yr BP (zone 3), the mixture of *C. meneghiniana*, *F. pinnata*, and benthic taxa indicate a moderately shallow system. A deep lake dominated by *C. chochtawatcheana* and with relatively low benthic diatom abundance characterized the basin prior from 34,000 to 42,000 cal yr BP (zone 2) with slightly shallower conditions in the basal sediments (zone 1), including a short very shallow interval about cal 42,500 yr BP.

1.5 Discussion

1.5.1 Hydrologic balance and transfer functions for lake-level reconstruction

Many studies show a positive correlation between $\delta^{18}O$ and $\delta^{13}C$ in closed-basin lakes (e.g. Talbot 1990, Li and Ku 1997), including two lakes on the northern Altiplano Lago Umayo (Figure 8 [Baker et al., 2009]) and Lago Junin (Abbott et al., 2003). Covariance of oxygen and carbon isotopes can be attributed to lake volume change, vapor exchange, evaporation, and productivity (e.g. Li and Ku 1997). The most important aspect of each of these factors may be the change in lake volume, which

affects both isotopic ratios in various ways. For example, increased input of CO_2 poor and $\delta^{13}C$ and $\delta^{18}O$-depleted freshwater to a lake can result in lower values of $\delta^{13}C$ and $\delta^{18}O$. Change in lake volume also affects vapor exchange rates: As lake volume increases rapidly, lower lake surface area to volume ratios decrease the air-water exchange, resulting in more negative values of both $\delta^{18}O_{H2O}$ and $\delta^{13}C_{HCO3}$. A decrease in lake volume has the opposite effect. Higher evaporation rates enrich $\delta^{18}O$ in the lake (Talbot, 1990).

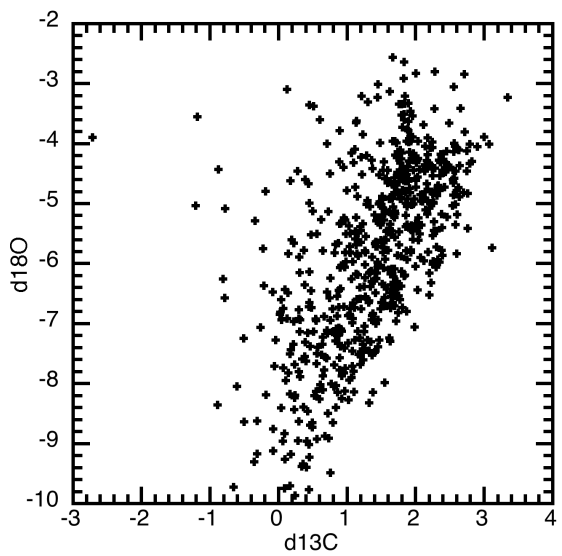

Figure 8: Correlation between $\delta^{13}C$ and $\delta^{18}O$ in closed basin lakes, as demonstrated by samples from Lago Umayo, Peru (Baker et al., 2009). We use this relationship to estimate water balance for the Salar de Coipasa during the last 45 Cal kyr BP. See text for more detail.

For $\delta^{13}C$ two factors are important: 1) Strong evaporation raises P_{CO2} increasing $\delta^{13}C$ of dissolved CO_3^-, 2) In shallow lakes mixing between deep and shallow waters may be enhanced, increasing the flux of nutrients to the surface and increasing productivity (Li and Ku 1996). Higher productivity in surface waters leads to enriched $\delta^{13}C$ values of dissolved HCO_3- in surface waters, where most $CaCO_3$ is produced. Therefore, in surface waters, high productivity leads to high $\delta^{13}C$ of carbonate.

For the Salar de Coipasa we take advantage of the relationship between $\delta^{13}C$ and $\delta^{18}O$ discussed above to use $\delta^{13}C$ of organic carbon as an approximation of water balance. Using a simple equation to determine the change in $\delta^{13}C$ over time, we determine whether there was a positive or negative change in water balance:

$$\frac{\Delta\delta^{13}C}{\Delta t}$$

Based on the results from this calculation (Figure 9), the most active changes in lake volume occurred between 22,000 and 25,000 cal yr BP, contemporaneous with the LGM, showing both positive and negative values, indicating a period of significant variability. Another period of high fluctuation occurs prior to ~35,000 cal yr BP to the base of the record. Each of these intervals corresponds with the previously reported periods of lake advance on the southern Altiplano of Minchin and early Tauca during the LGM (e.g. Servant and Fontes 1978; Rondeau et al., 1990; Baker et al., 2001; Fritz et al., 2004).

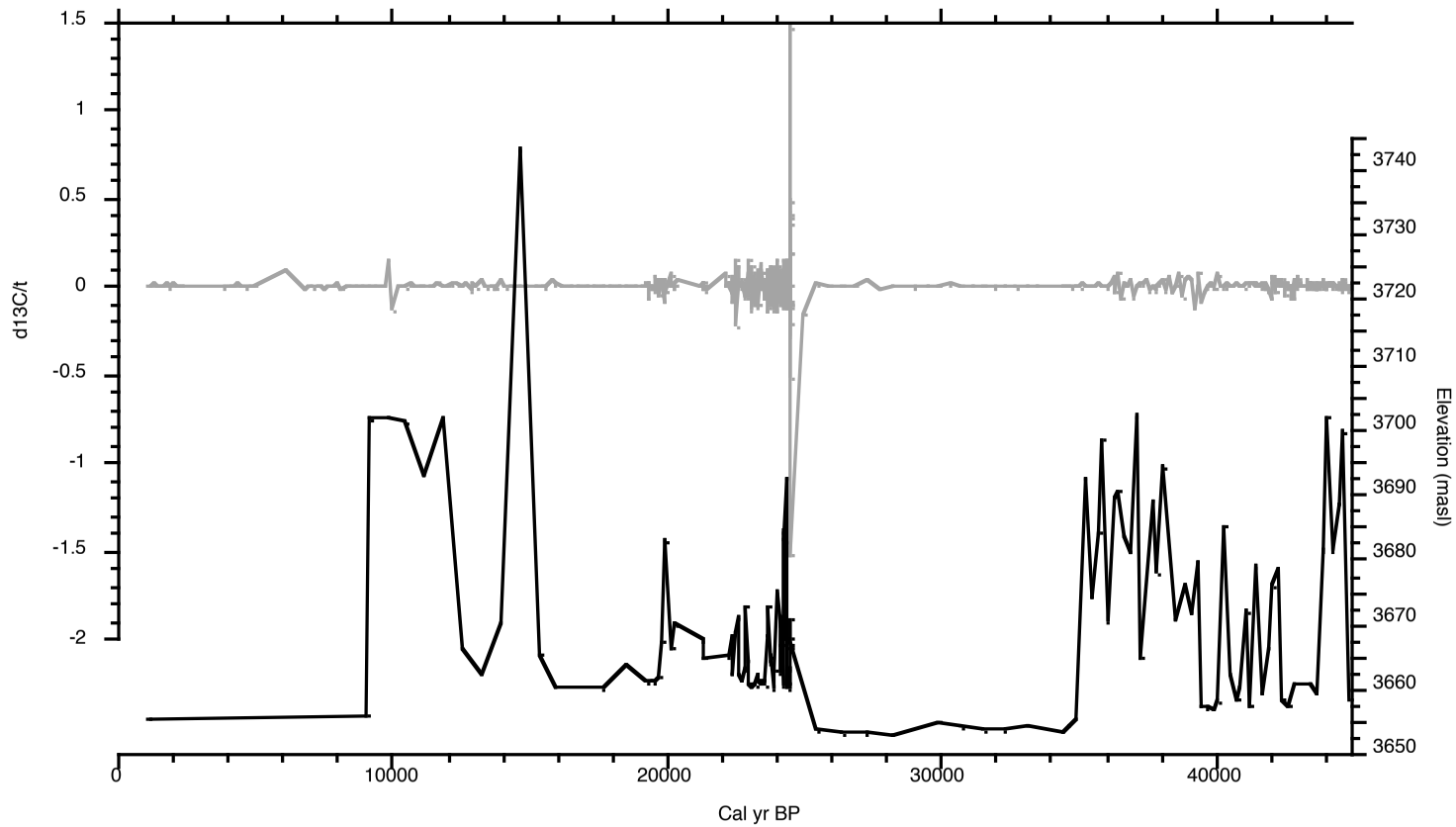

Figure 9: $\Delta\delta^{13}C/\Delta t$ versus time (cal yr BP) indicating fluctuation in lake volume (upper line). A reconstruction of lake level, using % planktic diatoms (lower line), is plotted against time.

Change in $\delta^{13}C$ values is not a direct proxy for lake volume. In order to reconstruct lake level throughout the period represented by the Coipasa sediment core record, we use a transfer function to convert the percentage of planktic diatoms to lake depth. For Coipasa there is no modern lake present at the core site, so we use ^{14}C age determined lake-level elevations measured in prior studies in the basin (e.g. Sylvestre et al., 1999; Servant et al., 1978; Bills et al., 2004; Placzek et al, 2006) to model lake level changes through time. We match these dated elevations with the % planktic diatom values in the Coipasa core for core samples of equivalent age. This correspondence is then used to create transfer functions for each lake phase for the last 45,000 Cal yr BP. The general transfer function is as follows:

$$\text{Elevation} = (P_i / \Delta P_{total}) * \Delta \text{ elev}$$

Where P_i is the value to be rescaled, P_{total} is the range of % planktic diatoms (100%), and Δ elev is the total span of lake level change over the given interval. A new approximation of lake elevation history from the % planktic diatom values is shown in Figure 9.

1.5.2 Salar de Coipasa lake level history

From 45,000-39,000 Cal yr BP proxy data indicate a deep lake with moderate variation in surface elevation throughout the interval (Figure 10). Based on previous

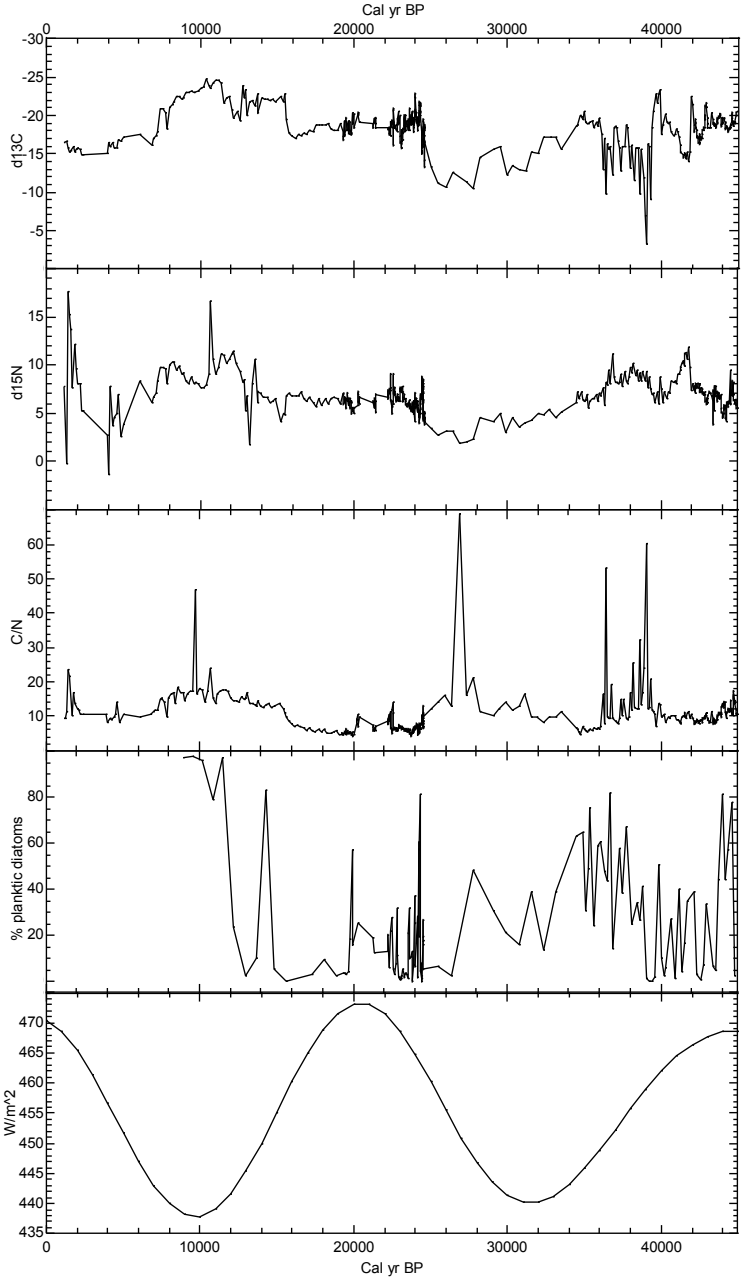

Figure 10: Changes in multiple proxy records of the Salar de Coipasa core displayed versus calendar age: (a) $\delta^{13}C$, (b) $\delta^{15}N$, and (c) C/N or organic muds; (d) percentage planktic diatoms and (e) calculated January insolation at 15°S (Berger, 1978).

lake sediment core records (Baker et al., 2001; Fritz et al., 2004; Sylvestre et al., 1998) and ^{14}C age dated paleoshoreline deposits (e.g. Sylvestre et al., 1999 and Rondeau, 1990), sediments in this interval represent the Lake Minchin phase on the southern Altiplano. Paleoshorelines found around the terminal basin suggest a lake between 3,657 and 3,712 masl. In order to have a lake of this magnitude, precipitation was likely increased 200-350 mm/yr above modern annual average values basin wide, dependent on a 2-5°C temperature depression from modern values (Nunnery, 2012).

At ~39,000 Cal yr BP, $\delta^{13}C$ values quickly jump from <-20‰ to >-15‰ within a very short interval of time, likely the result of a rapid draw down of lake level. High C/N values coincident with the enriched $\delta^{13}C$ suggest a shift toward a higher percentage of terrestrial carbon input, which is consistent with a decreased lake volume.

From 39,000 to 34,000 Cal yr BP the increasing percentage of deep-water planktic diatom species suggest that even though this interval was initially very shallow there was a gradual increase in lake volume. Isotope data appear to agree with this assessment as $\delta^{13}C$ values range from as high as -5‰ early in the interval to as low as -20 ‰ later on. The macrophyte bearing muds have $\delta^{13}C$ and δ15N values similar to those associated with macrophytes or more terrestrial based plants. This period coincides with an insolation minimum, which has been implicated as a driver for decreased moisture transport to the Altiplano (e.g. Baker et al., 2001a, 2001b; Fritz et al., 2007, Fritz

et al., 2010; Mourguiart et al., 1995; Servant et al., 1977; Bills et al., 1994; Seltzer et al., 1999; Rigsby et al., 2005).

From 34,000-26,000 Cal yr BP, Salar de Coipasa core shows an increase in benthic and shallow water planktic taxa coupled with the presence of rooted macrophytes, which suggests that a shallow lake occupied the Coipasa basin throughout this period. Rooted macrophytes are a clear indication of very shallow conditions. The majority of $\delta^{13}C$ values within this interval are more positive (>-15‰), while $\delta^{15}N$ values decrease to <5‰. Both carbon and nitrogen isotope values fall within ranges normally attributed to macrophytes and terrestrial sources (e.g. Cross et al., 2001; Rowe et al., 2002). The Uyuni core shows two short-lived lacustrine units centered at ~34,000 and ~29,000 Cal yr BP, each immediately followed by salt units, suggesting that desiccation occurred in the terminal basin even while a shallow lake was present at Coipasa. In order to maintain this hydrologic balance, precipitation was higher than modern values by 25-50 mm/yr basin wide, with temperatures likely only modestly lower than modern annual average (Nunnery, 2012).

During the LGM (26,000-21,000 Cal yr BP) records of $\delta^{13}C$, $\delta^{15}N$, TOC, and TN are relatively stable; suggesting that while lake levels may have been variable during this interval a lake consistently occupied the basin of Coipasa. The diatom species composition and % planktic diatoms suggest a lake of moderate depth with some

variability within the interval. A spike in δ13C, δ15N, and % planktic diatoms at ~24,000 cal yr BP is likely the result of initial increase in volume at the onset of the LGM lake phase. Paleoshoreline data indicate that a lake of elevation 3,657-3,700 masl occupied the southern Altiplano during the LGM (e.g. Sylvestre et al., 1999; Placzek et al., 2006). Based on hydrologic modeling (Nunnery, 2012), precipitation was elevated 100-250 mm/yr basin wide compared with modern annual averages, dependent on a temperature decrease of 2-5°C from modern values.

For the period 21,000-18,000 cal yr BP insolation is at or near a maximum, which is coincident with a sharp increase in % planktic diatoms and slightly more negative δ13C values. Loss of sediment during coring for the intervals 21,000-20,500 cal yr BP and 22,500-22,000 cal yr BP make an assessment of the true nature of this interval difficult.

From 18,000 to 13,000 cal yr BP, laminated sediments, dominated for most of this period by benthic diatoms, suggest generally shallow conditions and high salinity. A sharp increase in % planktic diatoms at ~15,000-16,000 Cal yr BP is coincident with paleoshoreline evidence of the highest lake stand of the last glacial cycle, designated Lake Tauca, which reached a maximum elevation of 3760 masl (110 m deep). It also coincides with the northern hemisphere Heinrich 1 event (H1), which may have influenced precipitation on the Altiplano through teleconnections that intensified the

SASM (Hastenrath, 1991; Xie and Carton 2004; Garreado and Aceituno, 2000; Garreaud et al, 2002; Zhou and Lau, 1997, Baker et al. 200x). According to hydrologic modeling, precipitation may have been 350-500 mm/yr higher than modern annual average (dependent on a 2-5°C temperature depression from modern values).

Between ~14,000-13,000 cal yr BP low planktic diatom % and a slight increase in δ13C values suggest a rapid decrease in lake level coincident with the Bølling-Allerød (BA) warming event (~15-13 ka) in the northern hemisphere. However, between 13,000-12,000 cal yr BP there is 5‰ increase in δ13C suggesting a decrease in lake level that appears to be coincident with the middle of the northern hemisphere Younger Dryas (YD) cooling event (12.9-11.6 ka). This period is evident in the Uyuni record as a return to a wet lake phase and as more negative δ18O in the Botuvera speleothem record (Cruz et al., 2005, 2006; Wang et al., 2004, 2006 and 2007) indicative of higher precipitation. The δ13C values in this interval are still very much below what would be considered dry conditions (>-15‰), the increase may be less an indication of lake level drop and more an indication of apparent decrease in phytoplankton production due to dilution from higher terrestrial runoff during a period of increased precipitation.

From 12,000-8,500 cal yr BP, leading into the Holocene, the Coipasa core is characterized by deep-water diatoms and δ13C and δ15N values associated with increased lake volume. The Salar de Uyuni transitioned from perennial lake to

perennial shallow saline lake (Fritz et al., 2004; Lowenstein, unpublished) quickly at the onset of the Holocene, whereas Coipasa remained wet and moderately deep until ~8,500 cal yr BP, possibly because of continued input of water from Lake Titicaca overflow, which remained high until ~8,500 cal yr BP, as evident in the $CaCO_3$ curve spanning this interval (Figure 11).

For the remainder of the record (8,000-1,000 Cal yr BP) the Salar de Coipasa seems to have been very shallow or dry. Muds in this interval are not laminated, and diatoms are absent with the exception of a single sample dated ~1,000 cal yr BP. There are two intervals of salt deposition: 6,000-4,800 cal yr BP and <1,000 cal yr BP. . The former interval is consistent with widespread evidence for a dry mid Holocene, as seen in the LT and Uyuni records (Baker et al. 2001a, b) and in the Botuvera speleothem record (Cruz et al., 2005, 2006; Wang et al., 2004, 2006 and 2007). Yet after 4500 cal yr BP, Lake Titicaca rose, whereas lake levels remained shallow in Salar de Coipasa, and Salar de Uyuni was dry (Baker et al. 2001).

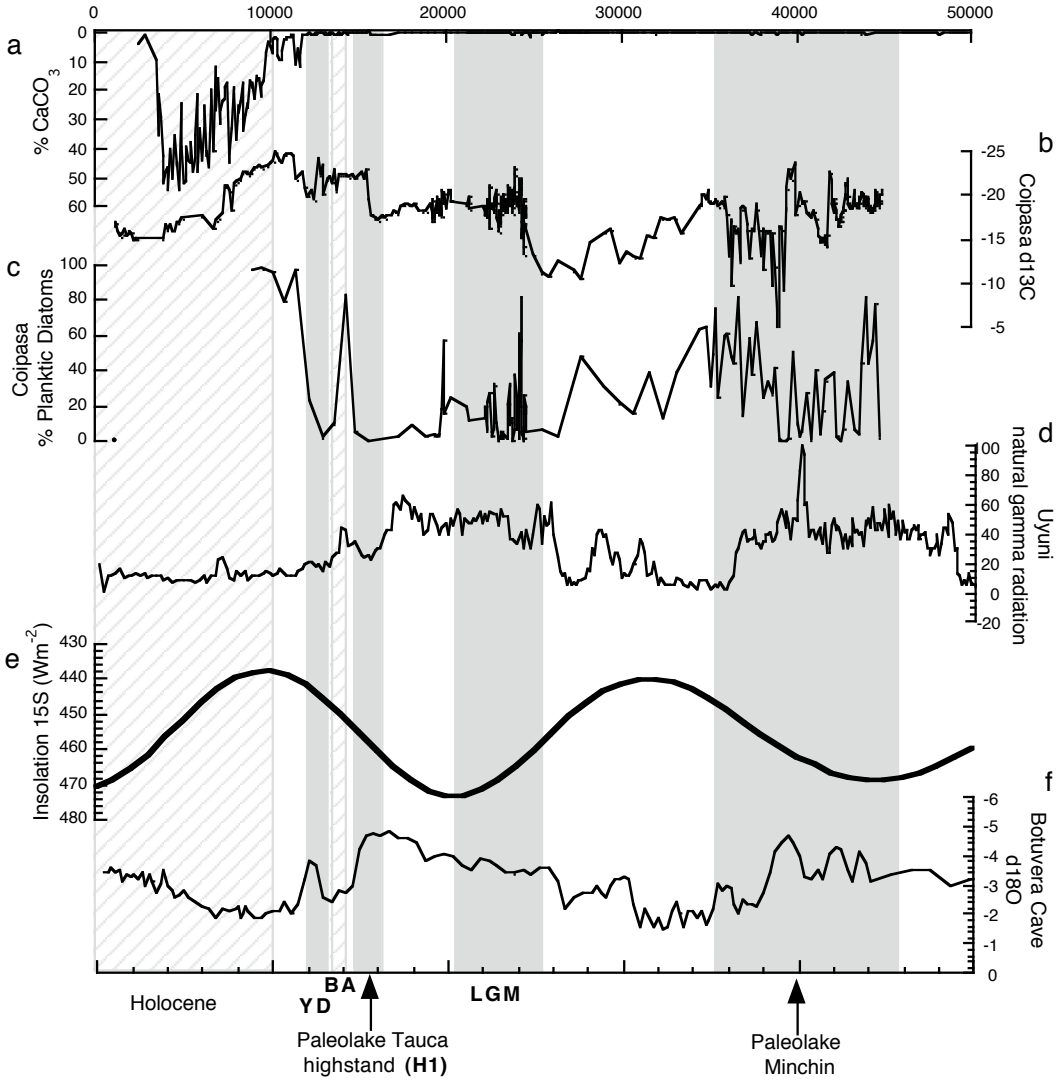

Figure 11: Multiple records from important basins of the Altiplano plotted versus time (0-45,000 years) compared to other paleoclimate time series: (a) CaCO3 of Lake Titicaca sediments, (b) $\delta^{13}C$ and (c) % planktic diatoms from the Salar de Coipasa, (d) January insolation at 15°S (Berger, 1978), (e) natural gamma radiation of the Salar de Uyuni (Baker et al., 2001), and (f) Botuvera Cave speleothem $\delta^{18}O$ record of precipitation (Cruz et al., 2005).

1.5.3 Implications for Altiplano hydrology

The most interesting result from our analysis of the Salar de Coipasa core is the discovery of apparent variability in the hydrologic synchronicity between the northern, central, and southern basins of the Altiplano. During the Minchin and Tauca paleolake stages evidence of deep to moderately deep lakes in the Coipasa (this study) and Uyuni basins (Baker et al., 2001b; Fritz et al., 2004; Lowenstein, unpublished manuscript) is consistent with indications of a deep Lake Titicaca (Fritz et al., 2007), suggesting a possible connection between the northern, central, and southern basins. Likewise, between 34-24 ka during a phase of near desiccation at Uyuni and very shallow lake levels at Coipasa, Lake Titicaca water levels were highly variable (Fritz et al., 2007). The mid-Holocene shows indications of coeval drying throughout the Altiplano, with evidence for closed basin conditions at Lake Titicaca (Fritz et al., 2007) coincident with periods of salt deposition in the southern basins (Baker et al., 2001b; Fritz et al., 2004). However, during the late Holocene Lake Titicaca was apparently deep and overflowing while the Salar de Uyuni and the Salar de Coipasa were desiccated, indicating an asynchronous relationship between the northern and southern basin. There is a similar, if less pronounced, variability in the hydrologic synchronization between the Salar de Coipasa and the Salar de Uyuni. The most noticeable differences in hydrologic histories between these two basins are during the interval 34-24 ka, characterized by continuous

deposition of mud in the Coipasa core and alternating salt/mud in the Uyuni core, and in the early Holocene (10-8 ka), during which the Salar de Uyuni apparently desiccated quickly (Figure 11).

Originally, the Salar de Coipasa core was collected in hopes of getting a high-resolution lacustrine record of the terminal basin for the last glacial and the deglacial interval, assuming that during wet periods the salars of Uyuni and Coipasa function as one lake, and that units of mud in one will be represented in the other. The obvious difference in sedimentation history between these two basins, which should be hydrologically connected with even modest increases in effective moisture, suggests that assumptions regarding basin hydrology may be oversimplified in prior studies of Altiplano paleolakes that assume that a lake stage in either Lago Poopó, the Salar de Coipasa, or the Salar de Uyuni necessitates an equal elevation paleolake stage for the entire southern Altiplano basin. Instead, we suggest that each of the sub-basins of the central and southern Altiplano have individual lake histories that may or may not coincide, and therefore should be evaluated separately.

A controversial point on Altiplano paleolake history has been the timing of the so-called Minchin lake phase (designated Inca Huasi, Salinas, and Ouki by Placzek et al. [2006]). Much of the uncertainty about the timing of this lake phase is apparently due to discrepancies in ^{14}C and U/Th dates from tufas and algal bioherms (Rondeau 1990;

Rouchy et al., 1996; Placzek et al., 2006), especially for samples taken from the Poopó basin. The samples from the Poopó basin were found to have ^{14}C age determinations of ~32,000-45,000 ^{14}C yr BP, while U/Th age determinations for the same samples fell in the range of 100,000-125,000 yr BP (Placzek et al., 2006). The discrepancy is attributed to contamination of carbon sources in the dated material (Rondeau 1990; Rouchy et al., 1996; Clapperton et al., 1997; Placzek et al, 2006). Shoreline deposits attributed to paleolake Ouki are contemporaneous with intervals of salt deposition at the Salar de Uyuni. Otherwise, U/Th dated shorelines in the Salar de Uyuni and Salar de Coipasa basins (Placzek, et al., 2006) are in reasonable agreement with a long core record of lake history from the Salar de Uyuni (Baker et al., 2001; Fritz et al., 2004): paleoshorelines at 45,760 yr BP (3,661 masl), 46,330 yr BP (3,657 masl), and 47,160 yr BP (3,657 masl) correspond to paleolake Minchin (46,000-36,000 yr BP); paleoshorelines at 80,460 yr BP (3657), 86, 320 yr BP (3,660 masl), 88,010 yr BP (3,658 masl), 90,210 yr BP (3,665 masl), and 96,020 (3,660 masl) correspond to perennial lake deposits in the Salar de Uyuni core (Fritz et al., 2004). Therefore, the largest discrepancy lies in the comparison of U/Th dated tufa deposits in the Poopó basin to ^{14}C dated mud units recovered from the Salar de Coipasa and ^{14}C and U/Th dated mud units recovered from the Salar de Uyuni.

A problem with using paleoshoreline deposits from the Poopó basin to infer lake level for the entire southern Altiplano is that there is the implied assumption that a

given lake elevation at Lago Poopó is representative of all lakes in the southern Altiplano, thus that the basins are at equilibrium with each other. However we suggest that the three basins function independently, as demonstrated by comparison of cores from the Salar de Coipasa (Sylvestre et al., 1998; and this study) and the Salar de Uyuni (Baker et al., 2001; Fritz et al., 2004). There is absent from the Coipasa and Uyuni basin paleoshoreline records any evidence for the presence of a lake at the Salar de Coipasa or the Salar de Uyuni contemporaneous with lakes in the Poopó basin suggested by Ouki U/Th ages (>100,000 Cal yr BP [Placzek et al., 2006]), which could be the result of incomplete data coverage in the Coipasa/Uyuni basin for this specific interval. However, the absence of paleoshorelines for the Ouki lake stage in the Coipasa and Uyuni basins may be the result of an alternate hydrology, in which the central and southern Altiplano were not as easily connected as they are today.

Presently, the Rio Desaguadero transports overflow from Lake Titicaca to the southern Altiplano, which has historically provide a significant source of water for expansion in the down-gradient basins of the central Altiplano. Possibly of equal importance is the connection between Lago Poopó and the Salar de Coipasa, known as the Laka Jahuira. The region between Poopó and Coipasa generally lies above 3,720 masl, with the exception of the Laka Jahuira valley (as narrow as 1.2 km across in some

locations). To the east and west of this rise, the Laka Jahuira resembles a wide river flood plain, but in between it is an incised channel (Figure 12). It is possible that this

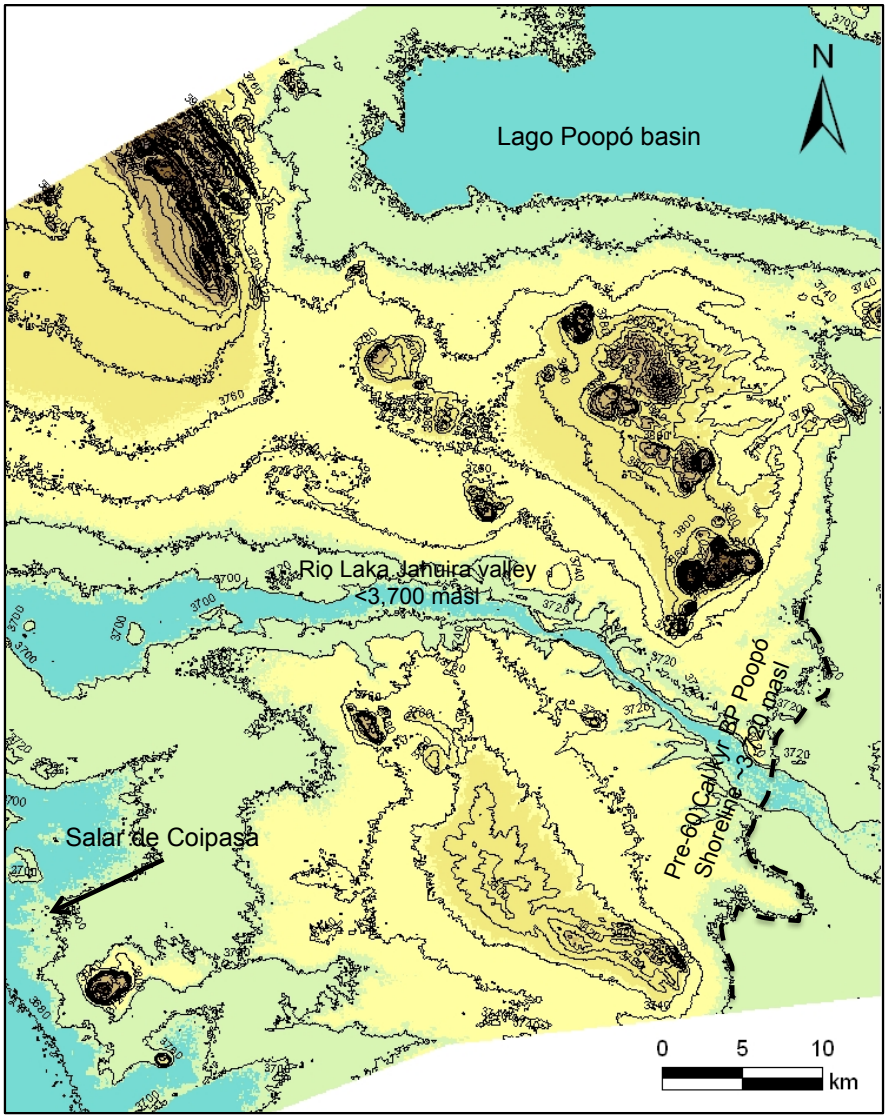

Figure 12: Topographic map of the region between Lago Poopó (east) and the Salar de Coipasa (west) known as the Rio Laka Jahuira. The river lies at <3,700 masl and, during periods of increased lake level, flows westward towards the Salar de Coipasa. The dotted line along the eastern edge of the map indicates a possible breach in the boundary of Lago Poopó.

channel formed during one of the recent high lake intervals (i.e. during the Minchin or Tauca paleolake phases). A conspicuous topographic feature at ~3,720 masl lines the southwestern edge of the Poopó basin, and the river appears to have cut through this feature. Prior to the incision of this feature, lakes in the Poopó basin could have risen to at least 3,720 masl without any overflow to the west.

Recent studies on a sediment core from the Salar de Uyuni (Baker et al., 2001b; Fritz et al., 2004) exploring the history of lake levels on the southern Altiplano, suggested that prior to ~60,000 years before present (yr BP) the southern basin was more arid than through the interval spanning the last 60,000 yr BP. Lacustrine sediment cores from Lake Titicaca (Baker et al., 2001b, Cross et al., 2001; Fritz et al., 2007, 2010, and 2012), as well as cores recovered from the Salar de Atacama in neighboring Chile (Bobst et al., 2001; Lowenstein et al., 2003), show coeval indications of relatively low effective moisture in partial support of the above hypothesis. We postulate that prior to 60,000 Cal yr BP the Rio Laka Jahuira was not present in its current configuration and that the Poopó and Coipasa basins did not connect, except when water elevation rose above ~3,720 masl. An increase of 20 meters represents a significant difference in hydrology, with important implications for our understanding of paleolake connectedness in the central and southern basin. The advance of earlier lakes at Coipasa and Uyuni would, for the most part, be governed by precipitation increase and flow from the Rio Lauca

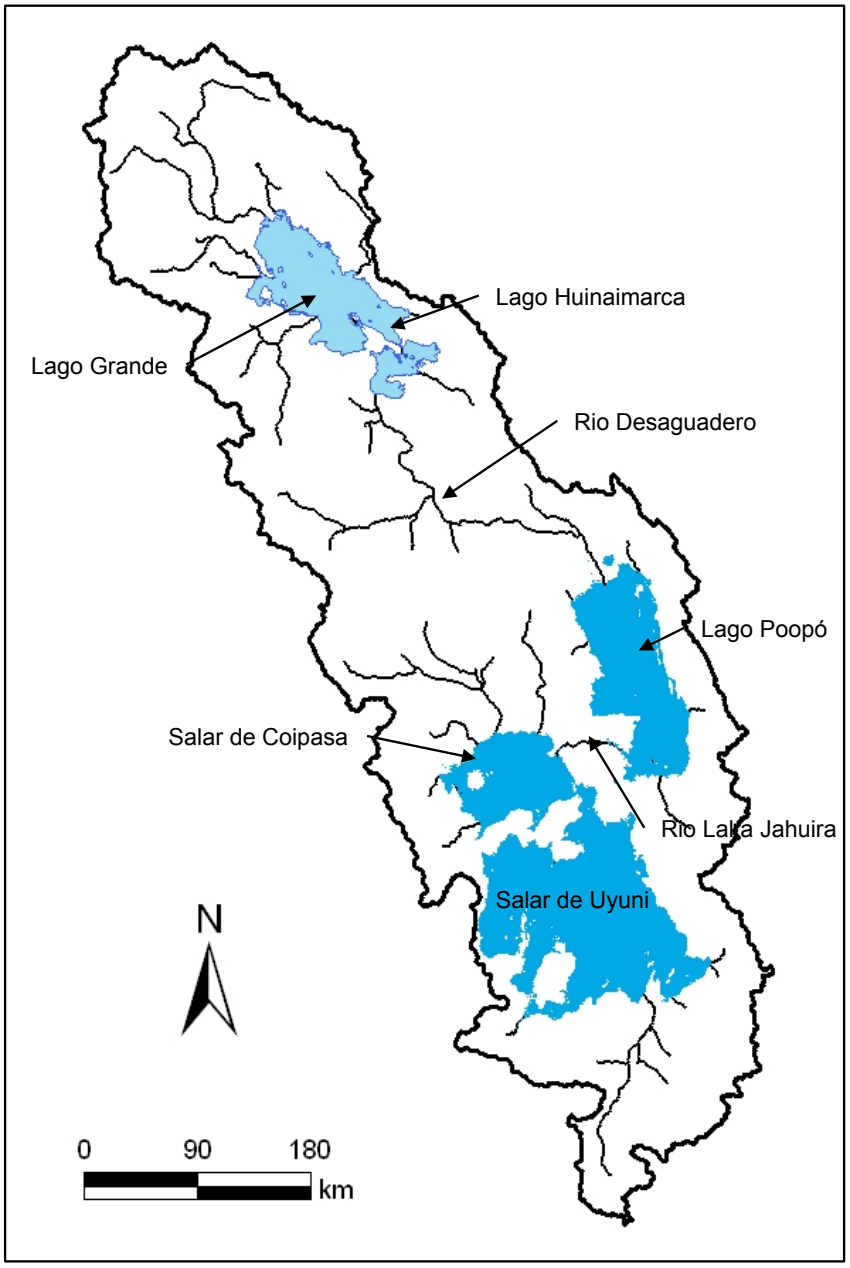

Figure 13: Revised hydrology of the southern Altiplano, with the Rio Laka Jahuira at 3,720 masl. In order to connect Lago Poopó to the Salar de Coipasa would have to be inundated by flooding >3,720 masl, in a lake phase such as Tauca. In the scenario depicted here Lago Poopó is at an elevation of 3,720 masl, while the Salar de Coipasa and the Salar de Uyuni maintain only shallow lakes (<3,700 masl).

and Rio Grande, without the augmentation of flow from Lake Titicaca via the Rio Desaguadero. Model studies show that without flow from the northern Altiplano, via the Rio Desaguadero, only very shallow lakes are possible, which is what is observed in paleoshoreline records prior to the Tauca phase. Another possible scenario is that Lago Poopó volume fluctuated pre-Tauca between 3,668-3,728 masl, while Uyuni and Coipasa fluctuated between 3,656-3,670 masl (Figure 13). At ~15,000 Cal yr BP the Tauca phase began and the Laka Jahuira was flooded to an elevation 3,740-3,760 masl. After subsidence the channel was cut between the basins and now remains below 3,700 masl. In this case, the connection between Poopó and Coipasa as is found in modern times may have been established < 20,000 Cal yr BP. In order to test this hypothesis, further investigation of the Lago Poopó basin, through sediment core retrieval, and more detailed analysis of the Rio Laka Jahuira valley is required.

1.6 Conclusions

We present a reconstruction of hydrologic variability on the southern Altiplano for the last 45,000 years using the carbon isotopic composition of organic carbon and diatom stratigraphy of a continuous sediment core recovered from the Salar de Coipasa, located in the southern tropical Andes. Our most significant finding is that the Salar de Coipasa and the Salar de Uyuni, long assumed to have been at equilibrium during wet

periods, have apparently different paleolake histories. The most notable example is the interval 26,000-36,000 Cal yr BP, in which the Coipasa core shows evidence of a shallow lake coincident with phases of near desiccation in the Salar de Uyuni sediment record. This leads to the conclusion that the three basins making up the central and southern Altiplano (Lago Poopó, the Salar de Coipasa, and the Salar de Uyuni) must all have separate lake histories. Large discrepancies between dates of lake phases >40,000 Cal yr BP, inferred from paleoshoreline and sediment core records (especially for the Lake Minchin phase [designated Salinas and Ouki from Placzek et al. (2006)]), have been attributed to inaccuracies in dating techniques or contamination of sampling material. Based on studies suggesting that a significant change in hydrology occurred on the southern Altiplano sometime prior to 60,000 years ago (Baker et al., 2001; Bobst et al., 2001; D'Agostino et al., 2002; Lowenstein et al., 2003; Groves et al., 2004; Fritz et al., 2004; Lowenstein, unpublished), we propose that the possibility exists for a lake to have existed in the Poopó basin >100,000 Cal yr BP up to an elevation of 3,720 masl, without significantly affecting the southern basin of the southern Altiplano. In order for this to be true the Rio Laka Jahuira, connecting Lago Poopó to the terminal basin, must have been either non-existent, or situated at an elevation of ~3,720 masl, which seems possible given the topography of the region between Poopó and Coipasa.

We have an incomplete record of lake level history on the southern Altiplano, especially in the Lago Poopó basin and the Rio Laka Jahuira. Collection of a sediment core from Lago Poopó and more detailed analysis of paleoshoreline deposits along the Rio Laka Jahuira valley would help to test the hypothesis that hydrology between the Poopó, Coipasa, and Uyuni basins was significantly different prior to ~60,000 Cal yr BP

2. Reconstructing hydrologic variation on the Peruvian/Bolivian Altiplano for the past ~50,000 years using a terrestrial hydrologic model

2.1 Introduction

Hydrologic variability on the Altiplano of Peru and Bolivia is recorded in historical and paleoclimate records that document the changing water level of Lake Titicaca, Lago Poopó, Salar de Coipasa, Salar de Uyuni and the large paleolakes of the terminal basin. Lake-level changes in these basins are attributed to fluctuations in effective moisture (precipitation minus evaporation) due to a combination of regional and global-scale climate forcing (e.g. Baker et al., 2001 a and b). Many studies have documented high paleo-shoreline and lacustrine sediments in the central and southern Altiplano surrounding Lago Poopó and the salars of Coipasa and Uyuni (e.g. Forbes 1861, Agassiz 1875, Musters 1877, Minchin 1882, Pompecki 1905, Bowman 1914; Servant and Fontes, 1978; Rondeau et al., 1990; Risacher and Fritz, 1991; Wirrmann and Mourguiart, 1993; Clapperton 1993; Bills et al., 1994; Servant et al., 1995; Sylvestre et al, 1999; Fornari et al., 2001; Baker et al., 2001b; Fritz et al., 2004; Rigsby et al., 2005; Placzek et al., 2006). The elevation of the paleoshorelines suggests that during wet periods the Poopó, Coipasa and Uyuni basins filled and sometimes may have been connected as a

single lake. The highest paleoshorelines (3,760 meters above sea level [masl]), which is a product of the largest of all of the paleolakes, has been named Lake Tauca and radiocarbon dated to 16,000 calendar years before present (cal yr BP)(Bills et al., 1994; Sylvestre et al., 1999; Placzek et al., 2006).

Prior studies have explored the contribution of glacial meltwater to the formation of these lakes (e.g. Servant and Fontes, 1978), and the majority have concluded that the meltwater contribution is minimal and that precipitation and evaporation variation are the most important factors in producing lake-level variation on the Altiplano (Kessler, 1984; Hastenrath and Kutzbach, 1985; Servant et al., 1995, Blodgett et al., 1997; Cross et al., 2001; Condom et al., 2004; Blard et al., 2009). However, the amount of precipitation increase and temperature decrease necessary to sustain large lakes on the southern Altiplano is unclear.

Previous studies using energy and water balance models have been applied to paleoclimate records to provide quantitative constraints on paleo-temperature and paleo-precipitation (e.g. Kessler et al., 1984; Hastenrath and Kutzbach, 1985; Blodgett et al., 1997; Cross et al., 2001; Condom et al., 2004; Blard et al., 2009). The study presented here utilizes a terrestrial hydrology model (THMB [Coe, 1998]) that fully integrates basin and stream geometries (based on a digital elevation map) to determine flow and storage of surface water, allowing a more accurate simulation of full basin dynamics. Among

the important features represented in this model are Lake Titicaca bathymetry and Rio Desaguadero valley geometry.

Using THMB several hydrologic simulations are conducted with variable precipitation amount and variable temperature to determine the conditions necessary to create and sustain large lakes in the Poopó, Coipasa and Uyuni basins. We also explore the importance of Lake Titicaca overflow as an input to southern Altiplano lakes. By comparing model results to sediment core records we attempt to determine more precisely the climate conditions during lake formation and retreat, and ultimately, to gain a better understanding of how decadal-to-millennial forcings have influenced the climate of the subtropical Andes. This study is focused on the last 60,000 years because this is an interval that has been shown in paleolake records to include several deep lakes on the southern Altiplano (e.g. Servant and Fontes, 1978; Bills et al., 1994; Baker et al., 2001b).

2.2 Background/Setting

The Altiplano (Figure 14) is an 187,000 km^2 intermontaine internally drained basin, located between the Cordilleras Oriental and Occidental in the Andes of Bolivia and Peru. The northern Altiplano is occupied by Lake Titicaca, which is the world's highest major lake and the largest lake by volume in South America. In wetter times

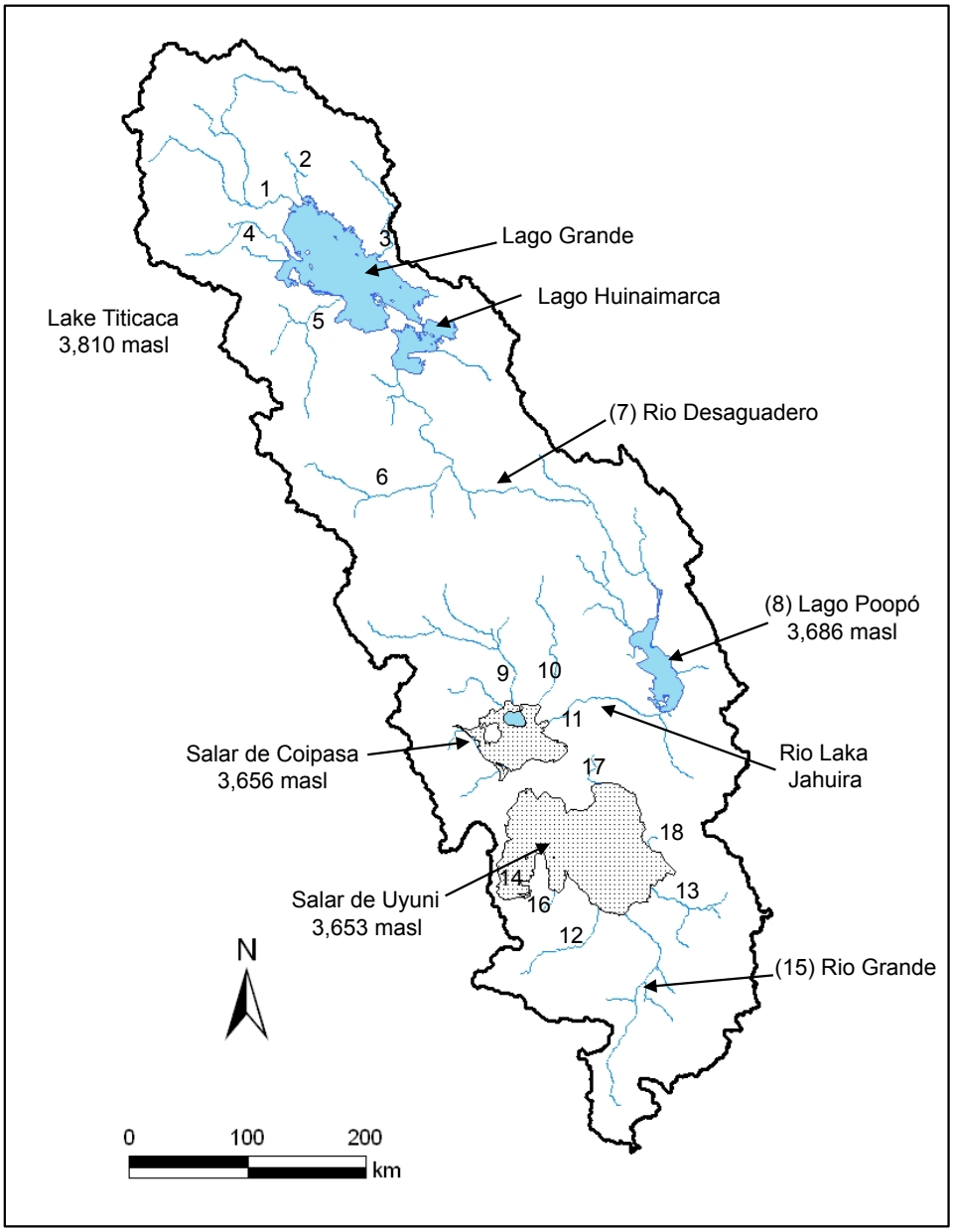

Figure 14: Map of the Peruvian/Bolivian Altiplano showing locations of major hydrologic features. Numbers indicate locations of river discharge for individual watersheds listed in Table 1

Lake Titicaca overflows via the Rio Desaguadero, contributing water to the shallow (2.15 m maximum depth in 2007) saline (25 g/l in 1983) Lago Poopó, which lies in the central Altiplano. Two large salars (salt flats) lie in the southern Altiplano: the Salar de Uyuni and the Salar de Coipasa. The Salar de Uyuni is the largest salar in the world and in wetter times is the terminal basin for the Altiplano.

Lake Titicaca is made up of two sub-basins, Lago Grande and Lago Huinaimarca. Lago Grande is larger (7,131 km^2 in 2009) and deeper (125 m mean depth in 2009), whereas Lago Huinaimarca is smaller (1,428 km^2 in 2009) and shallower (mean depth 9 m in 2009) and makes up ~16% of the area of Lake Titicaca. Two main hydrologic divides affect flow in the Lake Titicaca basin. The narrow (<1 km) and shallow (25 m) Estrecho (straits) de Tiquina divides Lago Grande and Lago Huinaimarca. In addition, a shallow and wide bedrock ledge regulates the natural outflow from Lago Huinaimarca to the Rio Desaguadero. Since 2001 this outlet has been damned, and floodgates and downstream dredged channels control outflow (Autoridad Binacional Autonoma del Sistema Hidrico TDPS [ALT]). However, prior to 2001, natural outflow via the Rio Desaguadero occurred only when lake level in Lago Huinaimarca was equal to or exceeded 3,804 masl. During dry periods when lake level dropped below the Rio Desaguadero outlet level, Lake Titicaca was a closed basin. During the period 1968 to 1987, outflow of Lake Titicaca via the Desaguadero was estimated at 37.5 m^3/s,

comprising 9% of the total water loss, with evaporative loss making up the majority (historically ~91-99%, Roche et al., 1992).

Connecting Lake Titicaca in the northern Altiplano to the central Altiplano is the Rio Desaguadero, which is a transitional fluvial system displaying various morphologies characteristic of marshes, meandering rivers, and braided streams (Baucom and Rigsby, 1999). During periods of increased precipitation and local runoff the northern part of the river basin merges with the southernmost part of Lake Titicaca (Baucom and Rigsby, 1999).

Lago Poopó (3,686 masl) drains an area of 55,000 km^2 in the central Altiplano. In addition to Rio Desaguadero input, inflow to the lake is contributed by precipitation (~20%) and seasonal inflow from regional rivers (~6%). The level of Lago Poopó varies with interannual changes in water level of Lake Titicaca (Carmouze et al., 1977 and 1978; Guyot et al., 1989; Iltis, 1993; Montes de Oca 2005; Zola and Bengtsson, 2007). Historical records of Lago Poopó indicate a lake-level minimum of 1.6-2.2 m with area varying between 1,500 and 2,500 km^2 (Boulange, 1978; Iltis, 1993). In 1985 the lake reached a maximum depth of 6 meters (area of 3,500 km^2), which corresponded with high lake levels in Lake Titicaca (Iltis, 1993).

The Salar de Coipasa is located on the southern Altiplano at an elevation of 3,656 masl. It is the second largest salt flat in Bolivia (second to Salar de Uyuni) with an area of

~2,500 km^2. It has a surficial halite layer (maximum thickness of 2.5 m) with marshy marginal zones (Eriksen et al., 1978; Risacher and Fritz, 1991). Today, the major inflow to Coipasa is the Rio Lauca, with a mean annual discharge of 0.14×10^9 m^3 (Servicio Nacional de Meterologia Hidrologia de Bolivia). Low-permeability lacustrine sediments consisting of detrital components, gypsum, calcite, organic matter, and authigenic clay minerals containing interstitial brines lie beneath the surface salt crust (Ericksen et al., 1978; Risacher and Fritz, 1991). The Salar de Coipasa undergoes seasonal flooding and drying and maintains a permanent lake offshore of the mouth of the Rio Lauca. Brines have been observed just below the crust of the Salar during the dry season, and in some years the salar does not dry (Risacher and Fritz, 1991).

The Salar de Uyuni, located south of the Salar de Coipasa at an elevation of 3,653 masl, is the world's largest salt flat with an area of ~10,600 km^2 (Risacher and Fritz, 1991). The surficial crust of the Salar de Uyuni is similar to that of Coipasa, however, the crust has a maximum thickness up to 11 m (Eriksen et al., 1978; Rettig et al., 1980; Risacher and Fritz, 1991). The major inflow to the Salar is the Rio Grande, which has an estimated mean annual discharge (from 1945-1959) of 0.4×10^9 m^3 (Montes de Oca, 1997)

2.2.1 Modern hydrology

The Altiplano is characterized by arid to semi-arid conditions, and the majority of precipitation is associated with the South American summer monsoon (SASM) (e.g. Lenters and Cook, 1997 and 1999; Zhou and Lau, 1998) and falls during the austral summer from December to March (Montes de Oca, 1997). Precipitation ranges from 800-1,000 mm/yr in the north, near Lake Titicaca, to 300-400 mm/yr in the central basin, near Lago Poopó, and <100 mm/yr in the terminal basin, in the vicinity of the Salar de Uyuni (Mariaca, 1985). Mean annual temperatures range from 10 °C in the north, 6 °C in the central region, and less than 4 °C in the south with a mean annual variation of ~5 °C. The temperature maximums occur during October-November instead of during the austral summer, due to increased cloud cover during the summer months. Daily temperature fluctuates 15-20 °C (Boulange and Aquiz Jaen, 1981). Variability in precipitation and evapotranspiration (partly affected by changes in temperature) determines the amount of effective moisture (total precipitation minus total evapotranspiration) available to create surface water. On the Altiplano an evapotranspiration gradient from low in the north to high in the south creates conditions for a negative water balance in the southern Altiplano (precipitation [P] < evapotranspiration [E]), while maintaining a positive water balance (P>E, higher effective moisture) in the north (Montes de Octa, 1997; Grove et al., 2003).

Rainfall variability on the Altiplano is primarily controlled by seasonal variation in the strength and position of the Bolivian high, an upper troposphere anticyclonic feature located over the central Andes created by the seasonal heating of the South American continent (Kessler, 1984; Aceituno and Montecino, 1993; Lenters and Cook, 1997 and 1999). Wet periods are associated with a southward shift and intensification of the high, and dry periods are associated with northward migration and weakening of the high (Aceituno and Montecino, 1993). During the summer, low-level northeast trade winds carry moisture across the Amazon basin to the Andes, in a recycling process of precipitation and evaporation. On decadal timescales, correlations between Amazon precipitation and sea-surface temperature (SST) suggest that precipitation over the Amazon, and ultimately the Altiplano, may be related to the SST gradient between the northern and southern equatorial Atlantic (Nobre and Shukla, 1996). On even longer time scales (centennial-to-millennial), changes in Altiplano precipitation, via changes in the gradient between northern and southern equatorial Atlantic SST, may be influenced by global scale changes in climate, such as Heinrich events and Dansgaard-Oeschger cycles (Baker et al., 2001a, 2005, 2009), which create anomalous temperature gradients in the north Atlantic (Carton et al., 1995; Nobre and Shukla, 1996; Black et al., 1999; Haug et al., 2001; Broccoli et al., 2006).

2.2.2 Paleohydrology of the Altiplano

There is an extensive background of research of basin geomorphology and analysis of lacustrine sediment cores from regions throughout the Altiplano indicating the rise and fall of Lake Titicaca in the north, and advance and retreat of multiple large lakes in the central and southern regions during the late Quaternary. In the north at least three periods of significant lake expansion have been inferred from observations of high shoreline deposits above the modern surface of Lake Titicaca - designated as paleolakes Mataro (3,950 masl), Cabana (3,900 masl), and Ballivian (3,860 masl)(Agassiz, 1875; Bowman, 1914 and 1916; Ogilvie, 1922; Moon, 1939; Newell, 1949; Ahlfeld and Brasnia, 1960; Servant and Fontes, 1978). The ages of these paleolakes are not firmly established, though they are considered to be of early to mid-Pleistocene age, based on fossil analysis of each unit (Pompecki 1905; Hoffstetter et al., 1971) and K-Ar dating of volcanic ash beneath the oldest of these deposits (Lavenu 1984, 1986). The timing of the formation of these paleolakes has been suggested to be coincident with glacial stages of the mid-Pleistocene (Servant and Fontes, 1978).

There is no evidence in the southern basin for paleolakes coincident with the early paleolakes in the northern basin. Possible reasons for this are: the sediment units attributed to lakes Mataro, Cabana, and Ballivian are not truly lacustrine and therefore do not represent an interval of lake expansion; the northern and southern sub-basins

were not hydrologically connected during these intervals; or evidence for these lake phases in the southern basin has been lost through erosion.

More recent investigations of northern Altiplano lake level change during the late Quaternary have involved the use of lacustrine sediment cores from the Lake Titicaca basins of Lago Grande and Lago Huinaimarca, as well as outcrops and sediment cores from the Rio Desaguadero Valley. Multiple sediment cores collected from both Lago Grande and Lago Huinaimarca, spanning the last 8,000 ^{14}C yr BP, indicate that during the early Holocene Lago Huinaimarca was relatively low (~15 meters below modern lake surface)(Wirrmann and Olveira Almeida, 1988; Wirrmann et al., 1988 and 1982; Wirrmann et al., 1992. Ybert, 1992; Mourguiart et al., 1995; Wirrmann and Mourguiart, 1995). Subsequent coring in Lago Huinaimarca (Abbott et al., 1997), which produced an extremely well dated record extending to ~3,800 cal yr BP, also concluded that Lago Huinaimarca was lower during the Holocene and became deeper by ~15-20 m at ~3,500 cal yr BP.

Wirrmann et al. (1988) and Mourguiart et al. (1995) presented evidence from a core taken near shore (< 50 m water depth) in Lago Grande suggesting that during the mid-Holocene (~7,700-3,900 ^{14}C yr BP) Lake Titicaca experienced a lake level drop of 50-55 meters. Cross et al. (2000) later determined, based on a newer suite of piston cores from deep-water sites (54-280 m water depth) and high-resolution seismic surveys

(Seltzer et al., 1998), that during the mid-Holocene (> 6,000-3,800 ^{14}C yr BP) Lake Titicaca likely dropped 85-100 meters below the modern surface elevation. Later analysis of high-resolution seismic surveys identifying erosional surfaces at intervals equivalent to mid-Holocene lake level lowering (D'Agostino et al., 2002) supported this conclusion. The changes in Lake Titicaca were attributed to orbitally forced oscillations in insolation, with deep lakes forming during insolation maxima (Cross et al., 2000).

Rowe et al. (2002) utilized analysis of the composition of organic carbon, C/N, wt $\%C_{org}$, $\%CaCO_3$, and % biogenic silica from deep-water cores (> 50 meters water depth) from Lago Grande to reconstruct changes in Lake Titicaca surface water elevation during the last 20,000 years. Results indicated that between ~20,000-13,500 yr BP the lake was above its outflow (~3,810 masl) and during the Pleistocene-Holocene transition (13,500-7,500 yr BP) lake level regressed to as low as 3,775 masl. At ~ 7,250 yr BP a transgression occurred raising lake level to ~3,798, followed by an ~85 m decrease in lake level at ~6,250 yr BP. Deeper and fresher conditions between 5,000-4,000 yr BP, and relatively high lake level, similar to modern conditions, are seen during the late Holocene. Similar to Cross et al. (2000), Rowe et al. concluded that change to lake level was caused by changes in insolation combined with long-term changes in El Nino southern oscillation variability.

Further analysis of deep-water sediment cores from Lago Grande, extending the lacustrine record to ~25,000 cal yr BP, determined that Lake Titicaca was deep and overflowing throughout the LGM, spanning the period > 25,000 – 13,000 cal yr BP (Baker et al., 2001a). Increased precipitation in the Andes during the LGM is attributed to an intensified South American summer monsoon (SASM) driven by a combination of wet season insolation maxima, increased SST gradient in the equatorial Atlantic enhancing northeastern trade winds, and lower equatorial Atlantic SST creating increased sea-land temperature gradient and enhancing water vapor advection (Baker et al., 2001a).

Analysis of Rio Desaguadero morphology (Baucom and Rigsby, 1999) and river valley sediment cores (Rigsby et al., 2005) showed that during periods of increased precipitation and local runoff the northern part of the river basin merged with the southernmost part of Lake Titicaca. Analysis of sediment cores collected along the 390 km long Rio Desaguadero valley indicate the presence of major paleolakes during the LGM (26,000-21,000 cal yr BP), the late glacial (14,000-12,000 cal yr BP), the early Holocene (10,000-7,9000 cal yr BP), and in the late Holocene (4,5000 cal yr BP to present)(Baucom and Rigsby, 1999; Rigsby et al., 2005), supporting findings of earlier Lake Titicaca sediment core studies that wet periods on the northern Altiplano coincide with periods of cold north Atlantic SST.

Fritz et al. (2010) presented analysis of core sediments for the period ~60,000-20,000 yr BP from Lago Grande showing evidence for millennial scale lake level variability that correlates well with millennial climate variability indicated in the Botuvera Cave isotopic record of southern subtropical Brazil (Cruz et al., 2005; Wang et al., 2004, 2006, 2007a), the Paraiso Cave isotopic record of the eastern Amazon (Wang et al., 2007b), and the speleothem and marine sedimentary records of the Brazilian Nordeste region (Arz et al., 1998; Wang et al., 2004), suggesting that large-amplitude climate variability extended throughout the southern tropics of South America during this interval. Changes between wet and dry conditions throughout the southern tropics of South America are apparently in agreement with millennial scale changes in cold to warm (respectively) SST in the north Atlantic, as indicated by NGRIP glacial isotope records (Fritz et al., 2010).

A 136-meter long sediment core recovered from Lago Grande represents 370,000 years of sediment deposition and reveals changes in hydrology of the late Quaternary (Fritz et al., 2007). Lake level and salinity were inferred from records of $CaCO_3$, $\delta^{13}C$ of organic sediment, and diatom stratigraphy. These analyses showed that periods of regional glacial advance and retreat coincided with global glacial and interglacial stages respectively. Also, deeper and fresher periods at Lake Titicaca correspond with regional

glacial ice advance, while periods of negative water balance were contemporaneous with periods of locally reduced glaciation (Fritz et al., 2007).

On the southern Altiplano, old lacustrine deposits and algal bioherms located in the Lago Poopó basin (central Altiplano) and the basins of the Salar de Uyuni and the Salar de Coipasa (southern Altiplano)(Musters, 1877; Minchin; 1882; Troll, 1927; Moon, 1939; Ahlfeld and Brasnia, 1960; Servant and Fontes 1976; Servant, 1977; Servant and Fontes, 1978), indicate two major phases of lake advance, younger than the Ballivian stage in the north, designated paleolakes Minchin and Tauca. The ages of these paleolakes have been the source of a great deal of controversy.

Servant and Fontes (1978) estimated the age of Minchin at 30,000-32,000 cal yr BP based on radiocarbon dates of shells found in outcropping sediments. Rondeau (1990) used U/Th dating of algal bioherms that bracketed the Minchin interval between 30,000-73,000 cal yr BP. Analysis of a 121 m sediment core from the Salar de Uyuni (Risacher and Fritz, 2000; Fornari et al., 2001) suggested a paleolake Minchin interval of > 31,000 cal yr BP with oldest age estimates (based on U/Th isochron method) of ~68,000-73,000 yr BP. Radiocarbon dating of subsurface sediments from another deep core (220 m) from the Salar de Uyuni placed Minchin between 36,000-46,000 cal yr BP (Baker et al., 2001; Fritz et al., 2004). Yet Placzek et al. (2006) described the exposures typically attributed to paleolake Minchin as three completely separate lake phases, rejecting

previous radiocarbon dates as too young. They used a U/Th chronology of tufa shoreline deposits, re-designating the three phases Ouki (96,000-125,000 cal yr BP), Salinas (80,000-90,000 cal yr BP), and Inca Huasi (45,000-47,000 cal yr BP). The ages of the Salinas and Inca Huasi paleolake phases correlate well with mud intervals of the same age in a core from the Salar de Uyuni (Baker et al., 2001b; Fritz et al., 2004). However, there is no corresponding mud unit in the core for the Ouki phase tufa deposits.

Ouki phase samples were collected exclusively from the Lago Poopó basin. There have been no Ouki aged outcrops located thus far in either the Salar de Coipasa or the Salar de Uyuni basins, suggesting that the central and southern Altiplano were not hydrologically connected during the lake interval that created these deposits

Early radiocarbon measurements of shells found in outcropping sediments in the Lago Poopó basin placed the age of paleolake Tauca at 14,100-11,400 cal yr BP (Servant and Fontes, 1978). More recent dating of algal bioherms place the timing of the highest phase of paleolake Tauca at ~16,000 cal yr BP (Bills et al., 1994; Servant et al., 1995; Rouchy et al., 1996; Sylvestre et al., 1999; Grove et al., 2003; Placzek et al., 2006), while lacustrine sediment cores from the Salar de Uyuni (Risacher and Fritz, 2000; Baker et al., 2001b; Fornari et al., 2001; Fritz et al., 2004) and Salar de Coipasa (Sylvestre et al., 1998; Nunnery, 2012) suggest that the Tauca phase spanned the interval 25,000-16,000 cal yr BP, consistent with the northern hemisphere LGM.

A third relatively shallow (~3,660 masl) and short-lived paleolake, designated paleolake Coipasa, has been identified in late Pleistocene paleoshoreline deposits around the Salar de Uyuni (Rondeau, 1990; Servant et al., 1995; Rouchy et al., 1996; Sylvestre et al., 1999; Fornari et al., 2001; Placzek et al., 2006). The age of this lake has been reported as 13,400-11,500 cal yr BP (Servant et al., 1995) and 12,800-11,400 cal yr BP (Placzek et al., 2006). There is no representation of paleolake Coipasa in a sediment core from the Salar de Uyuni, due to low-resolution recovery of this interval during collection. However, the natural gamma radiation log of the borehole indicates a period of increased precipitation of lacustrine muds centered at ~12,500 cal yr BP (Baker et al., 2001b), consistent with Servant et al. (1985) and Placzek et al. (2006). The timing of paleolake Coipasa is contemporaneous with the Younger Dryas cooling event in the north Atlantic, further supporting the hypothesis that southern tropical Andean precipitation is connected to temperature variability in northern hemisphere (Baker et al., 2001b).

A 220 m sediment core and a ~190 m natural gamma radiation borehole log from the Salar de Uyuni (Baker et al., 2001b; Fritz et al., 2004) provides late Quaternary records of hydrologic variation in the terminal basin of the Altiplano. The sediment core indicates an alternation between wet and dry phases, represented in the record by the respective alternation of mud and salt units. Because mud units have much higher

values of natural gamma radiation than salt deposits, borehole measurements of natural gamma radiation provide a powerful qualitative indication of hydrologic variability through time.

The stratigraphy of the Salar de Uyuni consists of sediments derived from three primary environments: perennial lakes that were fresh to saline, perennial hypersaline lakes, and saline pans and mudflats (Fritz et al., 2004;). Perennial lake muds from fresh to saline intervals are laminated, thin bedded, or massive, with calcite-rich carbonate mud, sulfides, organic matter, volcanic grains (biotite and amphibole), gypsum, diatoms and ostracodes (Fornari et al., 2001; Fritz et al., 2004). Perennial hypersaline lake units are principally composed of white-bedded halite, with associated brown to black laminae and thin beds of gypsum crystals and carbonate mud and pellets. Mudflat and saltpan deposits are characterized by massive mud/halite/gypsum units, which include evaporite mineral dissolution and diagenetic growth of gypsum and halite in the subsurface (Lowenstein and Hardie, 1985; Lowenstein unpublished).

Recently a 13-meter lacustrine sediment core was recovered from beneath the salt cap at the Salar de Coipasa, providing a record of nearly continuous sedimentation for the last 45,000 years (Nunnery, 2012). Hydrologic balance and lake level is inferred from $\delta^{13}C$ isotopic measurements of sedimentary organic carbon and diatom stratigraphy. $\delta^{13}C$ in the Coipasa core is largely derived from lacustrine algae (not terrestrial sources)

and is a proxy for changes in lake volume (Talbot, 1990; Nunnery, 2012). Results from these analyses are in general agreement with other local records and previous studies (e.g. Baker et al., 2001a,b; Fritz et al., 2004; Fritz et al., 2007) and suggest that intervals of increased effective moisture on the Altiplano coincided with cold periods in the northern hemisphere Atlantic.

Presented here are results from modeling of Altiplano hydrology that generate a reconstruction of Altiplano climate history for the last ~50,000 years, encompassing the Minchin (~46-36 ka), Tauca (~16 ka), and Coipasa (~12.5 ka) lake phases. The terrestrial hydrology model utilized in this study allows for simulation of Altiplano hydrology relying on changes to precipitation and temperature, two variables that can reasonably be related to observable measurements such as lake level. Additionally, we explore how possible future increases in global temperature may affect the water level of Lake Titicaca.

2.3 Methods

2.3.1 Model Principle

This study uses a terrestrial hydrology model (THMB) to determine how surface water accumulation and flow on the Altiplano are affected by changes to modern mean climate values of precipitation and temperature. The model is run through several

scenarios of change in precipitation and change in temperature to simulate the observed changes in lake levels derived from paleoclimate records. THMB (formerly HYDRA) (Coe 1998) is designed to simulate the flow and storage of water in terrestrial hydrologic systems over time. The version of THMB used in this study is the same as that used by Li et al. (2007) in which the river directions are not prescribed but are calculated dynamically at each time step based on the water head (land elevation plus the water depth). This version of the model allows for a more dynamic simulation of in-stream and closed-basin lake systems of the Altiplano. For this study THMB depends on four important boundary conditions: topography, precipitation, potential evaporation (PET), and surface runoff.

A digital elevation model (DEM) of the Altiplano (in the region between 12.5°S-24.5°S and 72.5°W-65.5°W) was constructed at 5′x 5′ resolution (approximately 9 km x 9 km) using high-resolution Shuttle Radar Topography Mission (SRTM) data (Figure 15). Topography below the modern surface of Lake Titicaca was constructed using depth soundings from nautical joint Peruvian-Bolivian maps. The depths were digitized using ArcMap software, and gaps between the depth soundings were filled using a spline interpolation method. For Lago Poopó the topography beneath the modern surface of

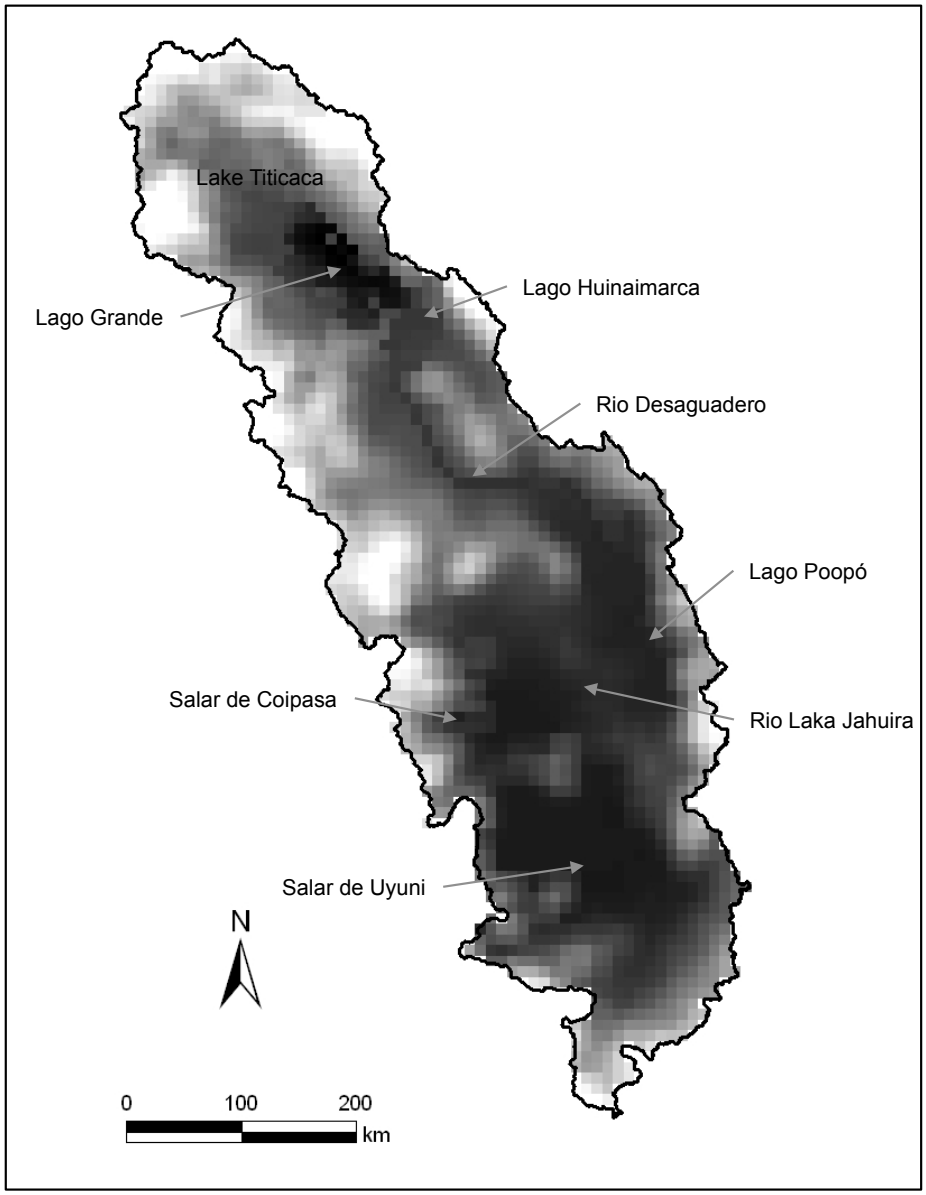

Figure 15: The 5'x5' digital elevation model (DEM) used to model hydrology of the Altiplano. Highest elevations are indicated by white and lowest elevations are indicated by black. THMB uses elevation data from the DEM to determine stream flow and water storage.

the lake is assumed to be generally flat with an average lake bottom elevation of 3,684 masl (Zola and Bengtsson, 2007). Sub-basin watersheds were determined using river inflow points for important rivers throughout the Altiplano. Surface runoff was calculated using gridded precipitation and temperature data for individual sub-watersheds (Table 2). Gridded temperature data are used to calculate potential and actual evaporation over land and water.

Table 2: Attributes of major watersheds of the Peruvian/Bolivian Altiplano. Runoff is calculated using annual mean values for precipitation (P), temperature (T), evapotranspiration (ET), and watershed area. Watersheds associated with smaller streams and local runoff around the Salar de Coipasa and the Salar de Uyuni are designated SC_local and SU_local respectively.

Id	Watershed	Area (km^2)	Avg P (mm/yr)	Avg T (deg C)	ET (mm/yr)	Runoff (m^3/yr)
1	Ramis	1.57E+07	792.26	6.92	530	4.10E+09
2	Huancane	3.62E+06	778.31	8.40	542	8.56E+08
3	Suchas	2.97E+06	900.26	12.56	587	9.29E+08
4	Coata	5.32E+06	633.14	7.43	480	8.14E+08
5	Llave	8.03E+06	507.09	7.54	420	7.00E+08
6	Mauri	1.01E+07	300.42	5.78	276	2.46E+08
7	Desaguadero (south of Mauri)	2.17E+07	679.43	9.60	512	3.64E+09
8	Poopó (includes Marquez)	1.31E+07	312.07	8.78	250	8.15E+08
9	Lauca	1.54E+07	267.58	5.62	250	2.71E+08
10	SC_local_1	3.55E+06	223.13	5.62	216	2.53E+07
11	Laka Jahuira	6.99E+06	123.31	6.01	120	2.32E+07
12	SU_local_1	6.76E+06	253.57	9.21	242	7.83E+07
13	SU_local_2	7.50E+06	208.04	8.80	202	4.53E+07
14	SU_local_3	4.25E+06	144.51	3.23	140	1.92E+07
15	Rio_Grande	1.12E+07	222.00	7.12	216	6.71E+07
16	SU_local_4	2.98E+06	119.87	6.20	118	5.57E+06
17	SU_local_5	5.47E+06	132.89	6.01	130	1.58E+07
18	SU_local_6	1.53E+06	195.78	8.17	188	1.19E+07

2.3.2 Input

Input to the model consists of topography (DEM), mean annual precipitation (P), potential evapotranspiration (PET), and mean annual temperature (T). The model is run as a sensitivity test of changes to these inputs over long timescales (centennial-to-millennial scale), so mean annual values are used, rather than seasonal values, in order to simplify model inputs. Gridded historic precipitation and temperature values, obtained from the National Climatic Data Center Global Historical Climate Network via the Web-based, Water-Budget, Interactive, Modeling Program (WebWIMP) at University of Delaware (Willmott et al., 2009) were used in the calculation of runoff for watersheds within the Altiplano. Precipitation and Temperature are defined as spatially variant. For each watershed average precipitation (P, mm/yr) minus evapotranspiration (ET, mm/yr) is multiplied by watershed area (A, m^2) giving runoff (R, m^3/yr):

$$R = (P - ET) * A \tag{1}$$

All watersheds and calculated runoff is presented in Table 2. The model uses these runoff calculations to prescribe input of major rivers given changes to precipitation and temperature over each corresponding watershed.

ET in equation (1) is calculated using a relation between precipitation and potential evapotranspiration (PET)(Pike, 1964), in which ET is actual evapotranspiration

(mm/yr), P is annual average precipitation (mm/yr), and PET is potential evapotranspiration (mm/yr):

$$ET = \frac{P}{\left[1+\left(\frac{P}{PET}\right)^2\right]^{1/2}} \quad (2)$$

Potential evapotranspiration (PET, mm/yr) is calculated via a modified Hargreaves equation (Hargreaves 1975, Hargreaves and Samni, 1982 and 1985) described by Xu and Singh (2000) and used for similar hydrologic modeling of the Altiplano by Condom et al. (2004) and Blard et al. (2009):

$$PET = \left(0.1 + 0.7\frac{dtrue}{dcs}\right) * Re * \frac{(T+17.8)*0.0145}{595-0.51*T} * 365 \quad (3)$$

d_{true} is the actual day length (hours), d_{cs} is the clear sky day length (assumed to be 12 hours), Re is extraterrestrial radiation (11,794 J/cm^2/day [Blard et al., 2009]), and T is average annual temperature (°C). Values of d_{true} are determined by an empirical equation developed by Condom et al. (2004) that takes into account the increase in cloud cover with increased precipitation (mm/day):

$$d_{true} = 9.18 - 0.008 * P \quad (4)$$

Total input to surface water via rainfall (P-E) is calculated using the potential evapotranspiration equation given in equation 3.

A control run with values of P, T, ET, and R calculated from modern-day mean annual precipitation and temperature values, is used to validate the model. For the northern Altiplano, which includes the Lake Titicaca watershed, precipitation maxima

varies from 600 to 1,100 mm/yr, air temperature ranges from 7 - 10 °C, and potential evapotranspiration ranges from 1,200 to 1,600 mm/yr. Total runoff from major watersheds contributing to Lake Titicaca is 7.40×10^9 m^3/yr. In the model these conditions allow for the surface of Lake Titicaca to be maintained at approximately 3,810 masl. For the central and southern Altiplano, which includes Lago Poopó, the Salar de Uyuni, and the Salar de Coipasa, precipitation varies from <100 mm/yr to 400 mm/yr, annual average air temperatures range from 4 to 9 °C, and potential evapotranspiration ranges from 1,500-1,700 mm/yr. Given these modern conditions, the model simulates Lago Poopó with surface elevation of 3,686 masl (2 m maximum depth), while the Salars of Coipasa and Uyuni display surficial flooding within a few kilometers of major discharge points (Rio Lauca for Coipasa and Rio Grande for Uyuni).

2.3.3 THMB experiments

Three experiments were conducted, designed to quantitatively reconstruct precipitation and temperature on the Altiplano based on interpretation of paleoclimate records. In each of these experiments, THMB is run through several scenarios with different values of P and T relative to modern values. The modern precipitation gradient from north to south on the Altiplano, defined as the ratio of ΔP (1,000 – 100 mm/yr) to Δdistance (~500 km), is roughly 1.8 mm yr^{-1} km^{-1}. The modern temperature

gradient, with ΔT ranging from 11-4 °C from north to south, is roughly 0.014 °C km^{-1}. Gradients are maintained throughout the various climate simulations by increasing or decreasing P by mm/yr and T by °C, rather than increasing these values by percentage.

Our first experiment uses THMB to test how increases in precipitation and decreases in temperature might affect Altiplano hydrology, simulating colder periods on the Altiplano, such as the LGM (26-21 ka) and the Younger Dryas (11.6-12.9 ka). With increased precipitation we expect to see an increase in runoff, as well as a corresponding increase in cloud cover (based on equation 4), which would reduce evaporation basin wide. Likewise, with a decrease in air temperature we expect a corresponding decrease in evaporation, providing more available water for runoff. Therefore, given any of the climate scenarios examined in this experiment, conditions on the Altiplano will move toward a positive water balance. Northern Altiplano contribution (defined as outflow from Lake Titicaca and runoff from the Rio Mauri watershed) to central and southern Altiplano hydrology is presented as % of total input ($\%N_{input}$), which is defined as the ratio of northern flow via the Rio Desaguadero (N_{input}) to the sum of total basin wide riverine input (N_{input} + central and southern river input [S_{input}] and direct precipitation (P_{direct}):

$$\%N_{input} = \frac{N_{input}}{N_{input} + S_{input} + P_{direct}} \times 100$$

Our second experiment is aimed at examining the importance of Lake Titicaca overflow to the formation of lakes in the Poopó, Coipasa, and Uyuni basins. This is accomplished by running THMB through the same set of climate scenarios that were run in the first experiment (increases in precipitation and decreases in temperature), without overflow from Lake Titicaca. Without contribution from the northern Altiplano, the creation of a positive water balance on the southern Altiplano (i.e. maintenance of standing water in the salar basins) will require significantly higher precipitation amounts than those suggested by previous hydrologic modeling studies (i.e. Kessler, 1983; Hastenrath and Kutzbach, 1984; Blodgett et al., 1997; Condom et al., 2004; Blard et al., 2009). Therefore, this experiment is a test of the feasibility of lake expansion in the southern Altiplano given a precipitation gradient in which the northern basin is dry and contributes minimally to the south.

The third experiment uses THMB to examine the effects of rising temperature or decreasing precipitation amount on Altiplano hydrology. Modern climate on the central and southern Altiplano has a negative water balance, as demonstrated by a shallow saline Lago Poopó and desiccated basins at the Salar de Coipasa and the Salar de Uyuni. Temperature increase or precipitation decrease will intensify that negative water balance in the central and southern regions. Therefore, this experiment focuses on the possible effects of a warmer or drier climate on the hydrology of Lake Titicaca. Such a climate

may correspond to the post interglacial periods (e.g. MIS 5e or mid Holocene) or future climate change. According to the fourth assessment report (AR4) of the Intergovernmental Panel on Climate Change (IPCC)(2007), global temperatures are likely to rise 2-4°C by the end of the 21st century. Climate observations in the Andes show that temperatures have been rising on average by about 0.9-0.20°C per decade (1939-2003) in the highest elevations, with the most significant increases occurring in the last thirty years. This suggests that by the end of the 21st century Altiplano temperatures could rise by ~1-2°C (Vuille and Bradley, 2003). THMB is run through several scenarios with increased temperature and decreased precipitation amounts in order to simulate previous low stands of Lake Titicaca (e.g. Seltzer et al., 1999; Cross et al., 2000; D'Agostino et al., 2002; Fritz et al., 2007), as well as to examine the possibility of future drawdown of Lake Titicaca water levels.

2.4 Results

2.4.1 Experiment 1: Precipitation increase, temperature decrease

THMB was run through several climate scenarios with increases in precipitation between 100-1000 mm/yr and decreases in temperature between 1-10 °C. Results are discussed here for two different lake surface elevations: (1) 3,700 masl, at which elevation Lago Poopó, the Salar de Coipasa, and the Salar de Uyuni reach equilibrium,

and (2) 3,760 masl, which is observed in paleoshoreline records to be the elevation of paleolake Tauca, the deepest of the central and southern Altiplano paleolakes.

With no change in temperature and a ~350 mm/yr increase of precipitation, the Poopó, Coipasa, and Uyuni basins fill to the level of Lago Poopó's outlet (~3,700 masl)(Figure 16). Since increase in precipitation causes a decrease in surface water evaporation in our calculations, due to enhanced cloud cover (equation 4), Lake Titicaca shows a 12% reduction in surface evaporation (~1,300 mm/yr down from as high as 1,600 mm/yr), while the southern Altiplano has a less dramatic drop in surface water evaporation of 2-6% (minimum ~1,500 mm/yr down from ~1,700 mm/yr). In this scenario, annual mean northern Altiplano flow (including overflow from Lake Titicaca and runoff from the Rio Mauri) to the central Altiplano via the Rio Desaguadero is ~1.7×10^{10} m^3/yr, total riverine input to the central and southern Altiplano is ~1.55×10^{10} m^3/yr, and input from direct precipitation is ~1.78×10^7 m^3/yr. Approximately 52% of the total input to the central and southern Altiplano is derived from flow from the northern Altiplano via the Rio Desaguadero.

To maintain a lake at elevation of 3,760 m in the Poopó, Coipasa, and Uyuni basins with no change in temperature, a ~600 mm/yr increase of precipitation is necessary. In this scenario, average precipitation in the central and southern Altiplano is ~850 mm/yr, and surface water evaporation decreases to ~300 mm/yr relative to modern

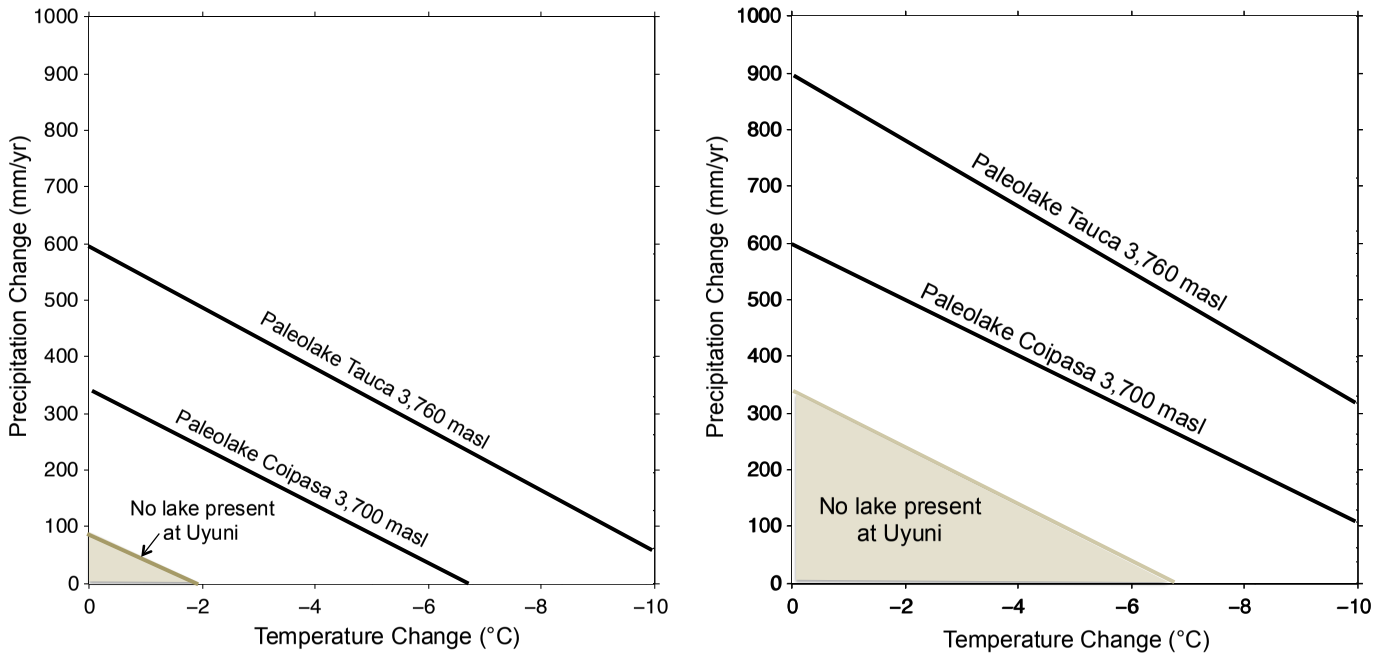

Figure 16: Plots showing results of THMB simulations for the Poopó, Coipasa, and Uyuni basins with various combinations of changes to precipitation (P) and temperature (T). P increase along the y-axis is in mm/yr above modern annual average values and T decrease along the x-axis is in degrees Celsius below modern annual average values. The plot on the left shows results of model simulations with increases in P and T, or a combination of the two, and includes input from Lake Titicaca overflow. The plot on the right also shows the results of model simulations with changes in P and T, but does not include input from Lake Titicaca overflow.

conditions (~1,400 mm/yr). Flow from the northern Altiplano to the south is ~2.2×10^{10} m^3/yr, total riverine input to the central and southern Altiplano is ~3.22×10^{10} m^3/yr, and input from direct precipitation is ~4.38×10^{7} m^3/yr. Input from the north accounts for approximately 41% of the total input to the central and southern Altiplano basins.

2.4.2 Experiment 2: Precipitation increase, temperature decrease, no flow from Lake Titicaca

THMB is run with several climate scenarios that are similar to the previous experiment, with increases in precipitation between 100-1000 mm/yr and decreases in temperature between 1-10 °C, but without input from Lake Titicaca. With no change to temperature and with a precipitation increase of 600 mm/yr above modern annual average, the lake levels in the southern Altiplano basins of Poopó, Coipasa, and Uyuni reach an elevation of 3,700 masl, equivalent to the Lago Poopó outlet (Figure 16). For the area around the Salar de Uyuni, this represents a 300% increase over modern precipitation values and a 70% increase over the precipitation necessary to reach this elevation when overflow from Lake Titicaca is included. With this increase in precipitation calculated evaporation is reduced over southern Altiplano surface water by about 9-11% (minimum of 1,300 mm/yr).

Without overflow from Lake Titicaca and without increased precipitation, a temperature decrease of 12 °C from the modern annual average is necessary to attain the

3,700 masl lake level that connects the southern Altiplano basins. The more reasonable 5 °C temperature decrease (similar to estimated LGM conditions [Hostetler and Mix, 1999]) combined with a precipitation increase of 350 mm/yr is necessary to fill all three basins to the 3,700 masl level. In order to create a lake on the southern Altiplano equivalent to paleolake Tauca (3,760 masl) with changes in precipitation alone and no Lake Titicaca contribution, it is necessary to raise precipitation basin wide by 900 mm/yr above modern annual average. This amounts to an increase in precipitation of 300-450% for watersheds surrounding Lago Poopó and the Salars of Coipasa and Uyuni. With a temperature decrease of 5 °C, the model requires precipitation to increase by 600 mm/yr in order to produce paleolake Tauca with no Lake Titicaca overflow. Under these conditions surface water evaporation over the Salar de Uyuni would decrease from 1,570 mm/yr (modern calculated value) to 1,050 mm/yr.

2.4.3 Experiment 3: Increase in temperature and decrease in precipitation

To simulate how Lake Titicaca might respond to warmer and drier climates, THMB is run with decreases in precipitation between 100-1000 mm/yr and increases in temperature between 1-10 °C. High-resolution seismic reflection surveys of Lake Titicaca (Seltzer et al., 1998; D'Agostino et al., 2002; Cross et al., 2000) indicated that mid-Holocene lake levels were 85-100 m below the modern lake surface, and MIS 5e lake

levels were ~200 m below the modern surface (Fritz et al., 2007). Analysis of trace elements and stable isotopes of ostracod calcite, diatom assemblages, and observation of rooted macrophytes from 85-meter water depth are consistent with the findings of the seismic survey (Seltzer et al., 1999). For the lake to fall to 85-100 meters below modern lake level due solely to temperature change would require an increase of 5-6 °C from modern temperatures (Figure 17).

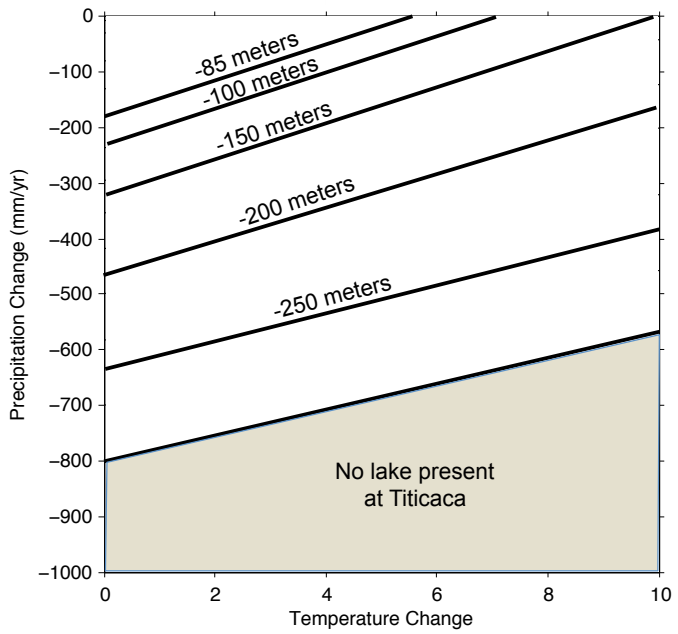

Figure 17: Plot showing results of THMB simulations for Lake Titicaca with various combinations of changes to P and T. P decrease along the y-axis is in mm/yr below modern annual average values and T increase along the x-axis is in degrees Celsius above modern annual average values. A temperature increase of 2 °C is sufficient to cause Lake Titicaca to become a closed basin. With a combined temperature increase of 2 °C and a precipitation decrease of ~150 mm/yr lake level would drop ~85 m.

If lake level fell exclusively due to lower precipitation, an average annual precipitation of 600-900 mm/yr (as much as 200 mm/yr less than modern) is necessary to create the low stand indicated by the seismic study and sediment records.

2.5 Discussion

2.5.1 Comparison with previous modeling studies

Several previous modeling studies have been applied to Altiplano paleoclimate records to provide quantitative constraints on paleo-temperature and paleo-precipitation. Kessler (1984), using a bulk transfer water budget calculation, determined that the conditions necessary to maintain paleolake Tauca were increased precipitation of 30% (~195 mm/yr over Uyuni) with a temperature ~3 °C cooler compared to modern conditions. Hastenrath and Kutzbach (1985) used a heat-budget method and determined that in order to maintain paleolake Tauca, precipitation was 100-300 mm/yr above modern conditions (based on Kessler's temperature estimation of a 3 °C decrease compared to modern values).

Blodgett et al. (1997) constructed an evaporation model based on a combination of the bulk-transfer method (e.g. Kessler) and energy-budget method (e.g. Hastenrath and Kutzbach). The model has an increased level of complexity over prior models, including parameters, such as surface albedo, Bowen ratio (ratio of sensible heat to

latent heat), cloud fraction, surface emissivity, saturation vapor pressure, air pressure, relative humidity, and windspeed. Model results concluded that the paleolake Tauca high stand required precipitation 20% higher than modern at temperatures 5 °C colder than modern.

Condom et al., (2004) developed a reservoir model to determine the difference between the timing of Altiplano climate and lake level change for the large lakes implied by the paleolake records. The model uses an evapotranspiration calculation based on precipitation, temperature, radiative budget, and hours of daylight. The study found that for Lake Tauca to be maintained, precipitation would have to be >330 mm/yr. Blard et al. (2009) combined a glacier model and a lake reservoir model (the later based on the model developed by Condom et al. [2004]) to reconstruct precipitation and temperature during the lake Tauca high stand. The results showed that during the Tauca lake phase air temperatures were 6.5 °C with average precipitation increased by a factor of 1.6-3 times modern mean values.

In these prior studies, input to the Poopó, Coipasa, and Uyuni basins originating from Lake Titicaca was either based on the empirical relationship between measured Rio Desaguadero flow and Lake Titicaca water levels (Condom et al., 2004), calculated based on a water budget method (Kessler, 1983; Blard et al., 2009), or was ignored (Hastenrath and Kutzbach, 1985). Blodgett et al. (1997) included discussion of Lake Titicaca

evaporation during the Minchin and Tauca lake phases, but did not include calculation of the Rio Desaguadero flow or northern contribution to central and southern Altiplano lakes.

The problem with the methods for calculating Lake Titicaca overflow discussed above is that the flow of the Rio Desaguadero is treated as a separately calculated input rather than an integrated part of the lake system. However, the rivers and lakes of the Altiplano are part of a linked hydrologic network. Therefore, to accurately simulate the water budget, water storage in lakes and flow in river valleys should be modeled simultaneously (Coe, 1998 and 2000). An advantage of using THMB to model Altiplano hydrology is that it fully integrates lakes and rivers, based on topography of the entire basin. The Rio Desaguadero is a dynamic feature in the model rather than a prescribed input. This method allows for accurate modeling of the communication between the northern, central, and southern Altiplano, resulting in more realistic water balance calculation. Another important contribution of this study to previous hydrologic modeling of the Altiplano is the inclusion of Lake Titicaca bathymetry, which allows for Lake Titicaca filling and overflow to be more realistically simulated rather than based on simplified water budget calculations (Kessler, 1983; Blard et al., 2009) or empirical measurements related to lake surface elevation (Condom et al., 2004).

2.5.2 Hydrologic history of the Altiplano

A north-south profile of effective moisture variation on the Altiplano through time is constructed using sedimentary records from the northern and southern basins (Figure 18). These records are combined with THMB output to reconstruct the likely paleohydrologic history of the Altiplano.

During the interval 46-36 ka, designated the Lake Minchin phase, lake levels ranged from ~3700-3,720 masl (Clapperton 1993; Servant and Fontes, 1978; Servant et al., 1995; Sylvestre et al., 1999; Placzek et al., 2006). With no change in temperature, precipitation would have been 350-425 mm/yr higher than modern to maintain paleolake Minchin, with the Rio Desaguadero contributing ~40% of total input to the lake. With a decrease in temperature consistent with glacial stage estimates (~5°C [Hostetler and Mix, 1999]) basin wide precipitation may have ranged from 100-200 mm/yr above modern values. Flow from the Rio Desaguadero to the central Altiplano accounts for ~50% of total input to paleolake Minchin.

Output from THMB shows that the Poopó, Coipasa, and Uyuni basins are not at equilibrium unless all three basins have a lake level greater than the elevation of Lago Poopó's overflow (~3,700 masl). When lake levels in the southern basins are below this overflow the central and southern basins are hydrologically disconnected. Equilibrium between the Salar de Uyuni and the Salar de Coipasa is maintained when both have

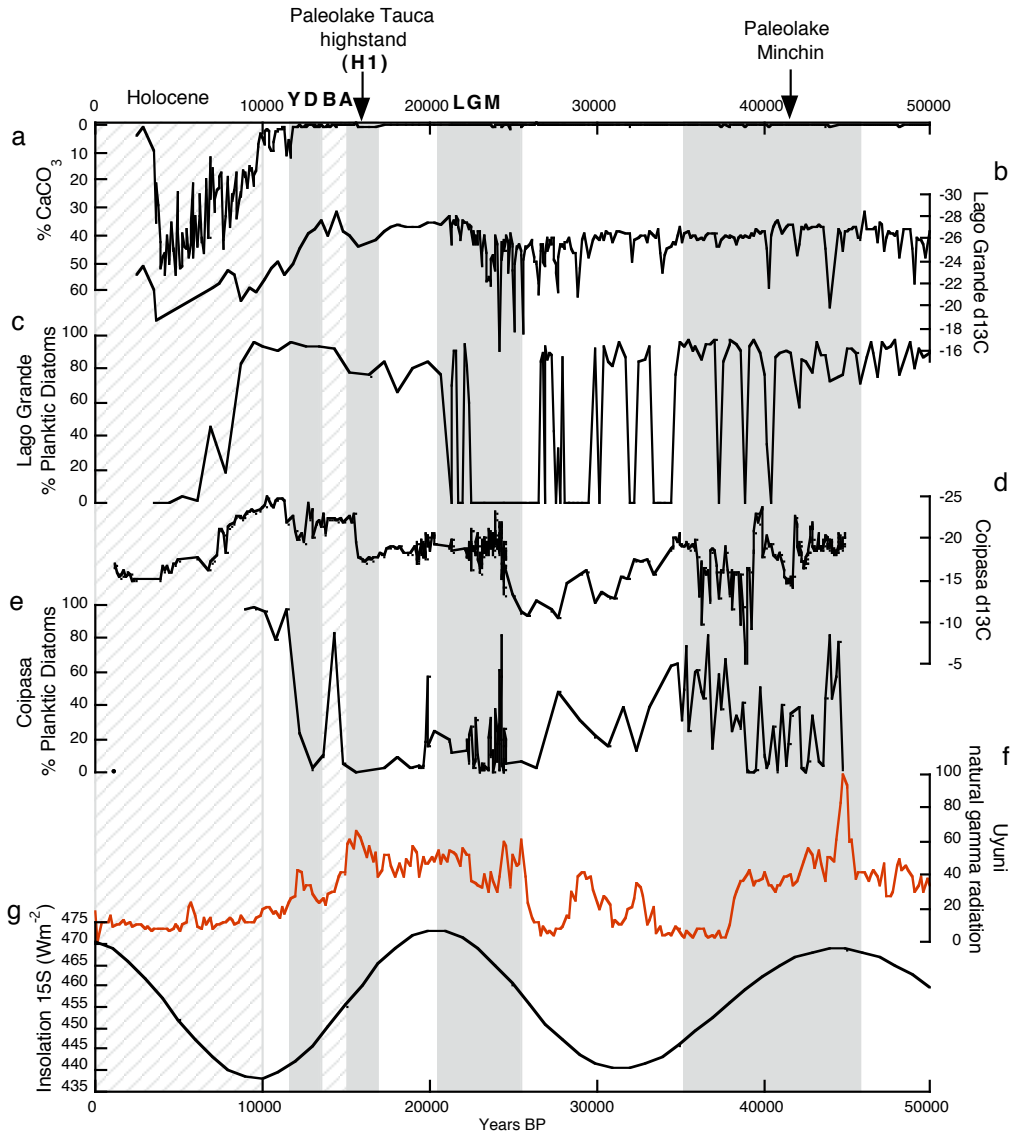

Figure 18: Lacustrine sediment records from major basins on the Altiplano for the last 50,000 cal yr BP plotted with major north Atlantic climate events (Younger Dryas [YD], Bølling-Allerød [BA] warming; LGM; Heinrich 1 [H1] cooling: a) Lake Titicaca % $CaCO_3$, b) $\delta^{13}C$, and c) % planktic diatom stratigraphy (Fritz et al., 2007); d) Salar de Coipasa $\delta^{13}C$ and e) %planktic diatom stratigraphy (Nunnery, 2012); f) Salar de Uyuni natural gamma radiation (Baker et al., 2001a; Fritz et al., 2004); and g) insolation at 15° south (Berger and Loutre, 1991).

water levels at elevations equal to or above ~3,680 masl. In the model, some climate conditions allow for connection but not equilibrium of the three basins, which produces a string of lakes connected by rivers rather than a single lake.

A comparison of sediment records from Lake Titicaca, the Salar de Coipasa, and the Salar de Uyuni for the interval ~36-26 ka (Figure 18) shows that the Salar de Uyuni was occupied by a perennial hypersaline lake, indicated by layers of bedded halite, and which coincided with a deep Lake Titicaca and a shallow lake at the Salar de Coipasa. These hydrologic conditions are simulated in THMB with a decrease in temperature of ~1.5 °C or an increase in precipitation of ~100 mm/yr relative to modern values. For these changes in climate Lago Poopó is ~16 m deep (barely overflowing), Coipasa is ~10 m deep, and Uyuni is ≤ 5 m deep. During this relatively dry interval the Rio Desaguadero contributes <10% of total input to the central and southern basins. For scenarios in which Lago Poopó is not overflowing (<3,700 masl, <16 meters depth) shallow lakes on the Salar de Coipasa and the Salar de Uyuni are maintained by local runoff and direct precipitation.

Based on paleoshoreline data (e.g. Servant and Fontes, 1978; Servant et al., 1995; Sylvestre et al., 1999; Placzek et al., 2006) and our records from Coipasa and Uyuni for the period between 26 to 21 cal kyr BP, lake levels in the terminal basin fluctuated ±43 m, with a minimum surface elevation of 3,657 masl (4 meters deep at Uyuni) and a

maximum of 3,700 masl (~47 meters deep at Uyuni) by the end of this interval. With an estimated temperature depression ~5 °C for the LGM (Hostetler and Mix, 1999), we calculate precipitation on the southern Altiplano was likely increased by ~250 mm yr basin wide compared to modern. During this interval flow from the northern Altiplano contributed ~60% of total input to the central and southern basin.

The paleolake Tauca phase (~16 Cal kyr BP) was the deepest of the paleolakes ever recorded on the southern Altiplano (e.g. Servant and Fontes, 1978; Clapperton 1993; Bills et al., 1994; Servant et al., 1995; Sylvestre et al., 1999; Placzek et al., 2006), reaching a maximum surface-water elevation of ~3,760 masl (~100m deep). In order to create a lake of this magnitude with precipitation change alone an increase of 600 mm/yr over modern annual average is necessary. In this situation, flux from the northern Altiplano to the southern Altiplano via the Rio Desaguadero accounts for ~40% of total input to the central and southern Altiplano.

Blodgett et al. (1997) found that a decrease of 10 °C was sufficient to create the paleolake Tauca highstand discussed above. Our results are similar, requiring a decrease in temperature of 11 °C to sustain a similar size lake. With a more reasonable decrease of temperature of just 5 °C (similar to LGM temperatures), we find that a basin wide precipitation increase of 350 mm/yr is required to sustain a lake at 3,760 masl.

Paleolake Tauca coincides with marine records of the Heinrich 1 cold event in the north Atlantic (Baker et al., 2001b). The dramatic rise of lake level on the southern Altiplano during the Heinrich 1 event is coincident with an increase of precipitation of 350-500 mm/yr relative to modern (assuming a 2-5 °C depression) implying a maximum precipitation over the Uyuni basin of 500 mm/yr, a maximum precipitation over Lake Titicaca of 1,300 mm/yr, and a contribution of northern Altiplano input to the central and southern total input of Altiplano of ~55%.

Around 14 cal kyr BP, a shift towards higher $\delta^{13}C$ values in the Coipasa record coincides with Antarctic meltwater pulse 1A (Fairbanks 1989; Stanford et al., 2006) and the Bølling-Allerød interstadial, a period of relatively warm North Atlantic temperature (+5 °C)(Stanford et al., 2006). The southern Altiplano experienced a shift to a negative water balance, with precipitation increased only slightly from modern values (<50 mm/yr) and temperature decreased between 1-2°C compared to modern. This resulted in the retreat of lakes on the southern Altiplano with water depths at Coipasa falling to as low as 3 m (3,659 masl) and Uyuni becoming nearly desiccated.

At approximately 12 cal kyr BP, a lake again formed in the Uyuni basin coincident with the Younger Dryas (Baker et al., 2001b). The lake increased in size rapidly to a maximum depth of ~50 meters (~3,700 masl), the result of cooling

temperatures (~2 °C compared to modern) and 250-325 mm/yr precipitation increase basin wide.

For much of the Holocene (<10 ka), sediment cores from the Salar de Uyuni and the Salar de Coipasa are composed of coarse halite deposits, which indicate that the southern Altiplano was dry at this time. Sediment cores from Lake Titicaca show lake level decline during the mid-Holocene (8.5-3 ka)(Figure 18). Carbonate percentage and the abundance of benthic and saline diatoms increase substantially, suggesting a shallower, more saline lake, which is confirmed by seismic studies describing a mid Holocene lake level drop of 85-100 m. The Coipasa record shows a shift to more positive $\delta^{13}C$ coupled with a decrease in planktic diatom percentage in the early Holocene (5-10 cal kyr BP) terminating in a salt cap as a result of late Holocene desiccation. At Uyuni a late Holocene (0-5 cal kyr BP) phase of desiccation is indicated by a very thick salt cap. The results from running THMB with increased temperature and decreased precipitation show that the region may have experienced a temperature increase relative to modern of 2-4 °C basin wide (maximum of 12 °C at LT; 11 °C at Uyuni) and precipitation decrease of 200 mm/yr basin wide over modern conditions during the mid-Holocene, creating desiccated basins in Poopó, Coipasa, and Uyuni and a closed basin lake at Titicaca.

2.5.3 Model limitations

The THMB model is run as a sensitivity test of Altiplano hydrology using mean annual values of precipitation and temperature. A more realistic approach would include seasonal precipitation and temperature variation, which would better represent how water balance is affected by inter-annual climate variability. Evaporation is calculated using a temperature-based method, which takes into account precipitation amount for calculation of cloud cover. This was done, because temperature and precipitation are two variables that are most reasonably related to observed measurements, such as lake level. However, more precise calculations of evaporation and other important but poorly known quantities, such as relative humidity and wind speed, should be considered. Also, the model resolution of 9 x 9 km, while high in comparison to earlier models (i.e. 15 x 15 km in Blard et al., 2009), is still quite coarse, leading to some inaccuracies for hydrological features, such as the dimensions of the Rio Desaguadero valley, the Strait of Tiquina, and the Rio Laka Jahuira, and subtle variations in the bathymetry of Lake Titicaca and Lago Poopó. Despite these drawbacks, the model output reproduces modern hydrology reasonably well and serves as a useful tool for determining how changes in climate affect long-term Altiplano hydrology.

2.6 Conclusions

Modest changes in precipitation and temperature are sufficient to connect the basins of Poopó, Coipasa and Uyuni in the southern Altiplano, creating a single lake at an elevation of 3,700 masl. During the Younger Dryas and LGM, lakes of this size likely formed as a result of a basin wide precipitation increase of 150-350 mm/yr over modern annual average values (dependent on a temperature depression between 2-5 °C through the glacial interval).

Output from hydrologic modeling suggests that the largest lake stage on the southern Altiplano, designated paleolake Tauca (16,000 cal yr BP), is consistent with a combined precipitation increase over modern annual average values of 350 mm/yr basin wide (assuming a 5 °C lower temperature relative to modern values). This produced a maximum precipitation of ~500 mm/yr over the Salar de Uyuni and ~1,300 mm/yr over Lake Titicaca. A smaller temperature decrease (~2 °C) during the event would have required a precipitation increase of ~500 mm/yr basin wide.

Runoff from the northern Altiplano (including Lake Titicaca overflow and runoff from the Rio Mauri) has a significant influence on the hydrology of the southern Altiplano, representing ~40-60% of total input to the terminal basin via the Rio Desaguadero. In order for the basins of Poopó, Coipasa and Uyuni to become connected as a single lake (3,700 masl) without contribution from the north, precipitation on the

southern Altiplano must increase 400-500 mm/yr basin wide (550-650 mm/yr over the Salar de Uyuni). To create a lake comparable in size to that of the paleolake Tauca without Lake Titicaca overflow, precipitation must increase 650-800 mm/yr basin wide (dependent on 2-5 °C temperature decrease) over the Salar de Uyuni from modern annual average values.

Assuming that temperatures on the Altiplano are likely to rise 1-2 °C over the 21st century (Vuille et al., 2003), Lake Titicaca will experience a significant decrease in lake level. A temperature increase of 2 °C, with no change in precipitation will cause Lake Titicaca to become a closed-basin lake and is sufficient to create a lake drop of >30 meters, which would have the effect of drying up Lago Huinaimarca and making Lago Grande a closed basin. In this scenario Rio Desaguadero flow would be significantly reduced, effectively creating a hydrological disconnect between the northern and central basins of the Altiplano. A temperature increase of 2 °C combined with a precipitation decrease of ~150 mm/yr is sufficient to create a lake level drop comparable to that during the mid-Holocene (85-100 meters). A decrease of lake level of this magnitude has serious implications for the availability of water for agriculture and consumptive use for inhabitants of the Altiplano.

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3. Evidence for a state change from a "playa lakes" phase to a "great lakes" phase ~60,000 yr BP on the southern Altiplano, Bolivia

3.1 Introduction

The Altiplano of Peru and Bolivia, located in the tropical Andes, is a vast region of dramatic landscapes shaped by intense tectonic and volcanic activity, creating a complex topography that yields an equally complex hydrologic system. Hundreds of thousands of years of basin evolution have led to the existence of some truly magnificent features, such as the large salt flats (salars) in the terminal basin in the southern Altiplano (the Salar de Uyuni and the Salar de Coipasa) and the expansive and voluminous Lake Titicaca in the northern Altiplano.

Understanding the processes that led to the modern configuration of lakes, rivers, and salars has been the subject of many research expeditions and studies, dating back to the 1800s (e.g. Forbes et al. 1861, Agassiz 1875; Musters 1877; Minchin 1882). It has long been known that the salars in the south have been the site of repeated expansion and contraction of large lakes and that the contemporary thick salt crusts are the result of desiccation of large saline lakes. In the north, the water level of Lake Titicaca has fluctuated significantly during both the past century (Servicio Nacional de

Meteorología e Hidrología [SENAMHI]) and on longer timescales (e.g. Seltzer et al., 1999; Cross et al., 2001; D'Agostino et al., 2002; Baker et al., 2001a; Fritz et al., 2007, 2010 and 2012), sometimes in apparent synchrony with the southern basin (Baker et al., 2001b; Fritz et al., 2004; Fritz et al., 2007; Nunnery, 2012). Hydrologic variability on the Altiplano is a product of both changes in precipitation amount and fluctuations in temperature, and both are linked to external forcings particularly long-term insolation variability and abrupt warming and cooling in the north Atlantic (e.g. Baker et al., 2001a, 2009) and tropical Pacific (Garreaud et al., 2009).

Recent studies of a sediment core from the Salar de Uyuni (Baker et al., 2001; Fritz et al., 2004), which reconstructed the late-Quaternary history of lake levels on the southern Altiplano, suggested that prior to ~60,000 years before present (yr BP) the southern basin was more frequently dry than in the interval spanning 60,000-10,000 yr BP. The primary evidence supporting this hypothesis is the absence of both thick halite units and thick mud units in core sections > 60,000 yr BP. Instead thin interbedded halite, gypsum, and mud beds are interpreted as the product of shallow playa lake deposition (Fritz et al., 2004; Lowenstein et al., unpublished manuscript). After ~60,000 yr BP thick mud beds are interbedded with thick halite beds that were deposited respectively in long-lived relatively fresh lakes and in saline/hypersaline lakes. Cores recovered from the Salar de Atacama in neighboring Chile (Bobst et al., 2001;

Lowenstein et al., 2003) have efflorescent halite crusts and cements, indicative of hyper-arid conditions, for the interval 325,000 - 53,000 yr BP, and bedded salts, characteristic of shallow saline lakes, core intervals < 53,000 yr BP (Bobst et al., 2001; Lowenstein et al., 2003), suggesting that whatever caused the hydrologic change on the Altiplano ~60,000 yr BP was influenced by larger atmospheric processes that affected other regions of the tropical Andes. Within the Altiplano to the north, lacustrine sediment cores from Lake Titicaca (Baker et al., 2001b, Cross et al., 2001; Fritz et al., 2007, 2010, and 2012) show low effective moisture during Marine Isotope (MIS) Stage 5 (~125 ka -75 ka) and relatively high effective moisture during most of MIS 4-2 (~75 ka -11 ka). Yet, the Lake Titicaca drill core does not show a secular drying trend in intervals prior to MIS 5, suggesting that the hydrology of the southern and northern Altiplano may not have been as tightly coupled in the past as it is today.

In this study we examine evidence for hydrologic state change in the southern Altiplano. We also evaluate whether the shift from the "playa lakes" phase to a "great lakes" phase at ~60 ka that is observed in Salar de Uyuni (and Salar de Atacama) may have resulted from climate change or from topographic change in the hydrologic connection between the northern and southern basins. Strontium isotopic compositions of a 220-m core from the Salar de Uyuni are compared to published strontium isotopic data from modern source waters, sediment cores, and paleoshoreline deposits from

locations throughout the Altiplano to determine changes in source water to the Salar de Uyuni through time.

3.2 Site description

3.2.1 Geologic setting

The Altiplano (14° 30′ W to 21° 50′ S) is a large (187,000 km^2) and high-altitude (average elevation ~3,800 meters above sea level [masl]) internally draining basin located between the Cordillera Oriental and Cordillera Occidental in the tropical Andes of Peru and Bolivia (Figure 19). The Cordillera Oriental is primarily composed of folded and faulted Paleozoic sediments, intrusive rocks, and young ignimbrites. The Cordillera Occidental is primarily composed of Cenozoic volcanic rocks. The Altiplano has a sedimentary fill composed mostly of Cretaceous and Tertiary continental sediments, which overlie a basement of Paleozoic sediments (Marsh et al., 1995).

The Salar de Uyuni is the world's largest salt flat, with an area of ~10,500 km^2. The surface of the Salar consists of a hard halite crust with a maximum thickness of 11 m near the center of the Salar and a minimum thickness of a few mm (overlying lacustrine mud) along the marshy marginal zones (Eriksen et al., 1978; Rettig et al., 1980; Risacher and Fritz, 1991). Stratigraphy underlying the crust consists of alternating units of mud and coarse halite reflecting a history of lake level rise and fall.

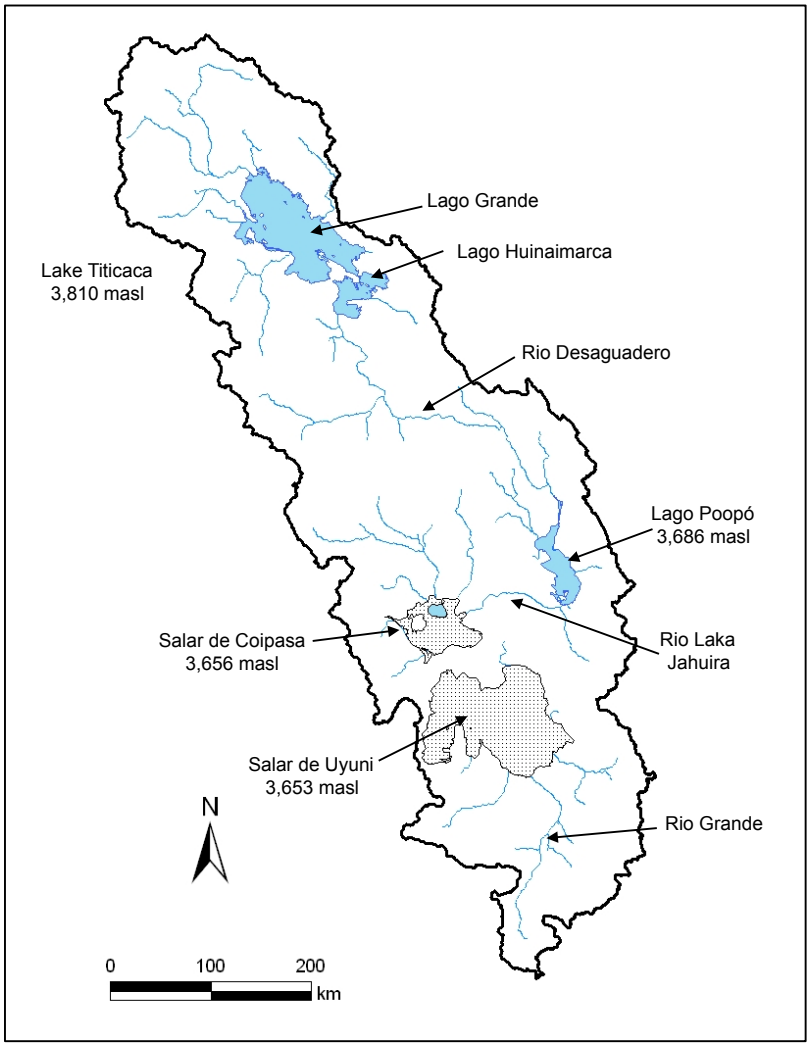

Figure 19: Map of the Peruvian/Bolivian Altiplano showing locations of major hydrologic features.

3.2.2 Hydrology and climate

The Altiplano is characterized by arid to semi-arid climate conditions, with the majority of precipitation falling during the austral summer from December to March, and a long dry season lasting from April to November (SENAMHI). Precipitation ranges from 800-1,000 mm/yr in the northern Altiplano, near Lake Titicaca, to 300-400 mm/yr in the central basin, near Lago Poopó, and to <100 mm/yr in the southern Altiplano, in the vicinity of the Salar de Uyuni. Mean annual temperatures range from 10 °C in the north, 6 °C in the central region, and less than 4 °C in the south, with a mean annual variation of ~5 °C. The temperature maximums fall during October-November instead of during the austral summer, due to increased cloud cover during summer months. Daily temperature fluctuates 15-20 °C (Boulange and Aquiz Jaen, 1981). An evaporation gradient from low in the north to high in the south (Mariaca, 1985) creates conditions for a negative water balance (precipitation < evaporation over open water) in the central and southern Altiplano, while maintaining a positive water balance over Lake Titicaca in the north (Montes de Oca, 1997). The major river flowing into the Salar de Uyuni is the Rio Grande, which has a mean annual discharge (estimated for the period 1945-1959) of 0.4×10^9 m^3 (Montes de Oca, 1997). The Salar de Uyuni floods during the rainy season, although not as extensively or for as long as the Salar de Coipasa to the north (Eriksen et al., 1978; Risacher and Fritz, 1991).

During prolonged phases of increased effective moisture, Lake Titicaca rises above its outlet (~3,804 masl)(Roche et al., 1992) and overflows via the Rio Desaguadero, eventually entering Lago Poopó (Figure 20).

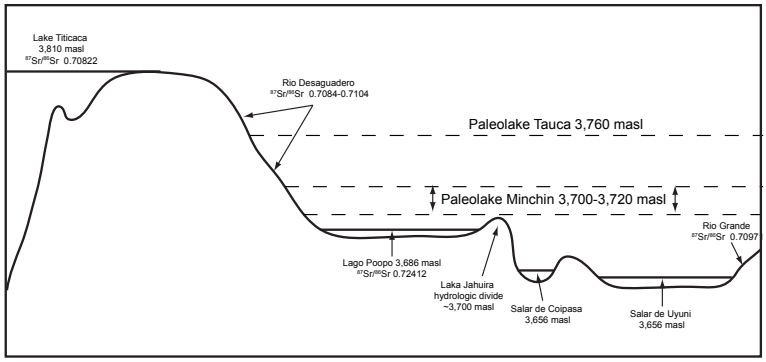

Figure 20: Longitudinal profile of the Bolivian Altiplano showing locations and elevations of Lake Titicaca, Lago Poopó, the Salars de Coipasa and Uyuni, as well as the Laka Jahuira hydrologic divide between Lago Poopó and the Salar de Coipasa. Paleolakes Tauca (3,760 masl) and Minchin (3,700-3,720 masl) are indicated by dashed lines. Strontium isotopic compositions for various locations reported by Grove et al., 2003. Vertical exaggeration is ~560x.

Likewise, during prolonged wet phases, higher local runoff combined with enhanced input from Lake Titicaca can cause Lago Poopó to fill to a surface water elevation of approximately 3,700 masl, at which point Lago Poopó overflows, via the Rio Laka Jahuira, westward into the Salar de Coipasa. The surface of Salar de Coipasa (3,657 masl) is only ~4 m above that of the Salar de Uyuni (SRTM data). At a surface water elevation of ~3,658 masl Salar de Uyuni and Salar de Coipasa would unite into a single lake. Once surface water in the united Coipasa/Uyuni basin rose to the level of Lago Poopó overflow (~3,700 masl), the three basins would combine to form a single large

lake connected by a series of straits (Figure 21). Thick lacustrine deposits in the upper 55 m (60 ka – present) of a 220 m core from the Salar de Uyuni (Baker et al., 2001b; Fritz et al., 2004) indicate a "great lakes" phase during which time the Poopó, Coipasa, and Uyuni basins were apparently connected, as described above. Mud and salt units below 55 m (> 60 ka) indicate a "playa lakes" phase.

3.2.3 Strontium isotopes

$^{87}Sr/^{86}Sr$ variability in surface water of a given hydrologic system, such as the Altiplano, is the product of mixing between multiple strontium sources, through riverine input, from geological regimes of differing Sr isotopic composition. Therefore,

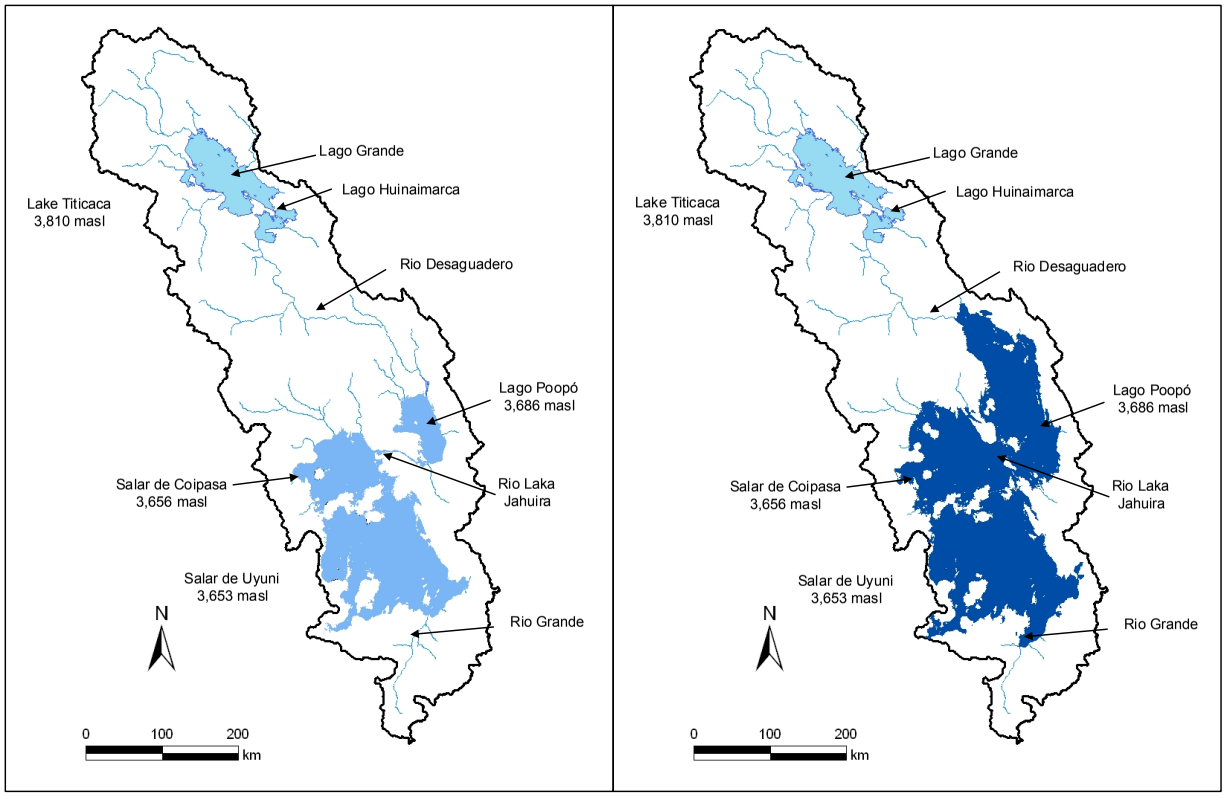

Figure 21: During intervals of increased effective moisture on the Altiplano the basins of Poopó, Coipasa and Uyuni combine to form a single large-area lake (assuming modern topography). The map on the left shows lake levels at an elevation at which all three basins form a single large lake (3,700 masl, indicated by shaded areas in Poopó, Coipasa and Uyuni basins). The map on the right shows the possible extent of paleolake Tauca (3,760 masl) as inferred from paleoshoreline deposits around the Poopó, Coipasa, and Uyuni basins (e.g. Bills et al., 1994).

values of $^{87}Sr/^{86}Sr$ in lake waters are indicators of Sr sources to the lake and are useful in determining patterns of flow within a hydrologic system (Bain and Bacon, 1994).

A previous study by Grove et al. (2002) utilized strontium concentrations and strontium isotopic ratios in natural waters and carbonate sediments from locations throughout the Altiplano to develop better constraints on the modern hydrologic mass balance and evaluate changes in the late Quaternary hydrologic mass balance. Results from this study suggested that during the paleolake Tauca highstand interval (17,000-15,000 cal yr BP), which was the deepest of the paleolakes in the basin, 70-83% of total riverine input to the central and southern Altiplano was derived from Lake Titicaca overflow (representing a 7-fold increase over modern discharge into Lago Poopó). This suggests that the formation of paleolake Tauca was not formed by decrease in evaporation alone, but was highly dependent on overflow from the northern Altiplano.

Placzek et al. (2010) used a new Sr isotopic data set from carbonates associated with paleolake phases on the central and southern Altiplano, combined with prior data from Grove et al. (2003), to determine the causes of lake-level fluctuations during the late Quaternary. Sr ratios in tufa deposits reflect variable contributions from the eastern and western Cordilleras, with highest ratios during the so-called paleolake Ouki interval (120–95 ka), lowest for paleolake Coipasa (12.8–11.4 ka), and intermediate for paleolake Tauca (18.1–14.1 ka). However, Ouki phase Sr samples were collected only in the Lago

Poopó basin, a region of more radiogenic runoff (Grove et al., 2003; Placzek et al., 2010), which explains the higher $^{87}Sr/^{86}Sr$ for this phase. In contrast to the findings of Grove et al. (2003,) Placzek et al. suggest that overflow from Lake Titicaca was a modest contributor to large paleolakes in the central and southern Altiplano throughout the late Quaternary and that the paleolakes of the southern Altiplano were sustained mainly by rainfall and local runoff.

Both of the strontium isotopic studies discussed here lead to a better understanding of hydrologic balance in the modern basin, as well as that of the Minchin, Tauca, and Coipasa paleolake stages. However, use of paleoshoreline $^{87}Sr/^{86}Sr$ measurements alone does not provide a continuous record for the determination of changes in Sr source throughout the late Quaternary. Grove et al. (2003) reported $^{87}Sr/^{86}Sr$ of carbonates in lake core sediments from Lago Grande and Lago Huinaimarca for the Holocene, indicating significant variability in lake level and in connection between the two basins. The records provide support for earlier findings of lake level fluctuation throughout the Holocene (Wirrmann et al., 1988; Mourguiart et al., 1995; Abbott et al., 1997; Seltzer et al., 1998; Cross et al., 2000; Baker et al., 2001a; Rowe et al., 2002; D'Agostino et al., 2002), but give little indication as to the state of hydrologic balance on the southern Altiplano.

3.3 Methods

3.3.1 Salar de Uyuni sample collection

In June 1999, a 220.6 m sediment core was recovered from the central region of the Salar de Uyuni (20° 14.97′ S, 67° 30.03′ W). The borehole was logged at 10-cm intervals measuring natural gamma radiation, which measures radiation produced by decay of U, Th, and K, and thus distinguishes well between halite and gypsum (low values) and volcanic ash and lacustrine mud (high values) (Baker et al., 2001a; Fritz et al., 2004). Calcium carbonate and salt samples were taken from the sediment core. Each of the major alternating salt and mud beds in the upper part of the sediment column were sampled.

3.3.2 Sample preparation

Calcium carbonate shells were separated from the bulk organic sediment, washed using distilled water, and dissolved in acetic acid for Sr analysis. Halite samples from the Salar de Uyuni and the Salar de Coipasa were coarsely ground with a mortar and pestle and dried at 60°C for at least 24 hours. Between 50-200 mg of powdered sample was weighed into centrifuge tubes and diluted to 50 mL using DI water. All diluted samples were loaded onto a shake tray for 24 hours and then placed in a centrifuge for 30 minutes to draw out remaining solids.

Samples for Sr isotope analysis were prepared by total dry-down of a sample aliquot containing approximately 3 μg of Sr. The dried sample was then digested in 3 N ultra-pure HNO_3 and extracted using Teflon micro columns containing Eichrom Sr resin. The extracted Sr was then dried again before digestion with TaCl solution and loaded onto Re filaments for a final desiccation. Strontium isotopes were analyzed by thermal ionization mass spectrometry. Samples were calibrated with multiple runs of an internal standard (NBS SRM-987). The average $^{87}Sr/^{86}Sr$ of the standard measured at Duke during this study was 0.710269 +/-0.000005 (SD). Strontium isotopic composition is expressed as a ratio ($^{87}Sr/^{86}Sr$). Strontium concentrations for samples from the Salar de Uyuni core were measured using inductively coupled plasma mass spectrometry (ICP-MS).

3.4 Results

3.4.1 Transition from "playa lakes phase" to "great lakes phase"

The core from the Salar de Uyuni can be divided into two major phases: the "great lakes" phase (upper 55 m, ~60,000 yr BP - present), characterized by thick mud and halite units, and the "playa lakes" phase (55-220 m, > 60,000 yr BP), characterized by thick sequences of massive mud, halite and gypsum typical of mud flats and saltpans (Figure 22).

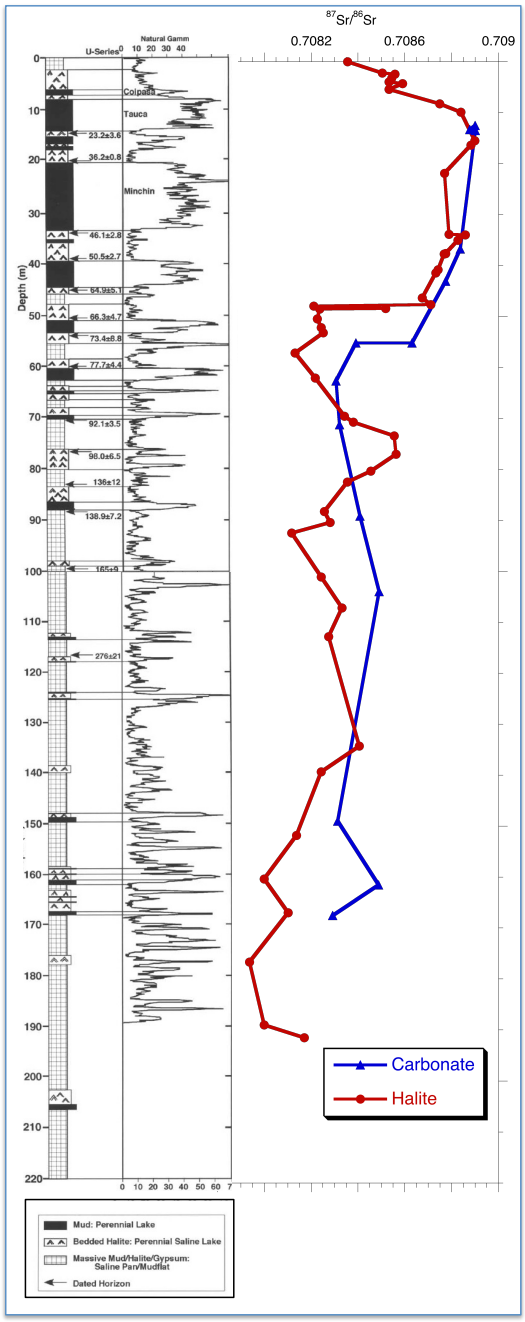

Figure 22: Stratigraphic column of the Salar de Uyuni core, with U-series ages (Fritz et al., 2004). Natural gamma radiation (Baker et al., 2001a) shown in the center. $^{87}Sr/^{86}Sr$ of carbonates and halites is shown on the right side.

Thick mud beds indicate long-lived perennial lakes of variable, but relatively low, salinity and variable water depth. With an accumulation rate of ~0.5-1 mm/yr a mud unit of 1 m represents 1,000-2,000 years of deposition. Thick carbonate rich lacustrine mud units are abundant during the great lakes stage, a period that includes the Minchin (46,000-36,000 yr BP), Tauca (26,000-15,000 yr BP), and Coipasa (~12,500 yr BP) paleolake stages. Some intervals within these units contain abundant planktonic diatoms and depleted oxygen isotopic values of the lacustrine carbonates (Fritz et al., 2004), both suggesting moderate lake depths. The few perennial lake sequences that are present from 140,000-60,000 yr BP are thin, suggesting a short lifespan, with compositions of brine shrimp fecal pellets, gypsum, and halite indicating shallow and saline conditions (Lowenstein, unpublished manuscript).

Thick halite beds indicate perennial saline to hypersaline lakes (Bobst et al., 2001; Lowenstein et al., 2003). Semi-permanent stratified hypersaline lakes produce salt beds devoid of the dissolution and diagenetic textures found in modern mudflat and saltpan deposits. Instead, these units are able to preserve pristine chevron and cumulate halite textures created during deposition, because they are not exposed to undersaturated waters (Lowenstein and Hardie, 1985; Smoot and Lowenstein, 1991). For sediments deposited at the Salar de Uyuni during the great lakes phase, the semi-permanent shallow saline lakes, brought on by relatively wetter conditions basin wide, produced

the bedded halite structures observed in the upper ~55 m of the core from the Salar de Uyuni.

Salts in the sub-surface of mudflats and saltpans are characterized by mineral dissolution and diagenetic growth, which is a product of periodic infiltration of undersaturated waters to the basin (Lowenstein and Hardy, 1985). In the case of the Altiplano, annual flooding of the Salar de Uyuni during austral summer induces the formation of these features. Sediments deposited during the playa lakes stage show an abundance of the dissolution and diagenetic features associated with mudflats and saltpans, indicating much drier conditions in the Salar de Uyuni basin (Lowenstein, unpublished manuscript).

3.4.2 Sr isotopic composition of the Salar de Uyuni core

$^{87}Sr/^{86}Sr$ of halites in the Salar de Uyuni core range from 0.7079-0.7089; $^{87}Sr/^{86}Sr$ in the carbonates range from 0.7083-0.7089 (Table 1, Figure 22). In the upper salt bed (Holocene age based on one ^{14}C analysis) $^{87}Sr/^{86}Sr$ compositions of 4 halite samples range between 0.7085-0.7086. These values are consistent with $^{87}Sr/^{86}Sr$ measurements of salt from the Salar de Uyuni crust ($^{87}Sr/^{86}Sr$ 0.70885, Placzek et al., 2010), but are lower than modern surface waters at the Salar de Uyuni that range from $^{87}Sr/^{86}Sr$ 0.70878-0.70894 (Grove et al., 2003; Placzek et al., 2010).

Strontium isotopic compositions for the period 60,000-10,000 cal yr BP range from $^{87}Sr/^{86}Sr$ 0.70821-0.70890 in halites and from $^{87}Sr/^{86}Sr$ 0.70877-0.70890 in carbonates. This interval includes several paleolake stages: the Tauca high-stand phase (15,000-17,000 Cal yr BP, $^{87}Sr/^{86}Sr$ 0.708749), the LGM (20,000-26,000 Cal yr BP, average $^{87}Sr/^{86}Sr$ 0.708892), and the Minchin phase (36,000-46,000 Cal yr BP, $^{87}Sr/^{86}Sr$ 0.7087-0.708858). The $^{87}Sr/^{86}Sr$ values in halites and carbonates for each of these lake phases are reasonably consistent with $^{87}Sr/^{86}Sr$ from tufa deposits associated with the Tauca phase in the Uyuni basin, measured by Placzek et al. (2010) as 0.70875-0.70888 and Grove et al (2003) as 0.70881, and similar to $^{87}Sr/^{86}Sr$ of Rio Desaguadero north of Lago Poopó (0.70864-0.70939 [Grove et al., 2003]). These values indicate that the $^{87}Sr/^{86}Sr$ of carbonates and halites deposited at the Salar de Uyuni during the great lakes phase reflect mixing of Sr sources originating in the relatively more radiogenic watersheds of the eastern cordillera (~0.709-0.724 [Grove et al., 2003; Placzek et al., 2010]), including significant riverine input from the northern Altiplano (> 50% [Grove et al., 2003; Nunnery, 2012]), and relatively less radiogenic watersheds of the western cordillera (~0.706-0.708 [Grove et al., 2003; Placzek et al., 2010]).

During the playa lakes phase, prior to ~60,000 Cal yr BP, $^{87}Sr/^{86}Sr$ of samples from the Salar de Uyuni core are less radiogenic compared to the upper great lakes phase, with values ranging from $^{87}Sr/^{86}Sr$ 0.70793-0.70856 in halites and $^{87}Sr/^{86}Sr$ 0.70829-0.70863

in carbonates (Table 3; Figure 22). $^{87}Sr/^{86}Sr$ of the playa lakes phase most closely resemble $^{87}Sr/^{86}Sr$ of streams and rivers to the west (i.e. Rio Kolcha, 0.70678 [Placzek et al., 2010]) and south (i.e. Rio Javilcha, 0.707373; Rio Salado, 0.70756; Rio Viscachillas, 0.708296 [Grove et al., 2003]) of the Salar de Uyuni. The less radiogenic $^{87}Sr/^{86}Sr$ values of halites and carbonates deposited at the Salar during the playa lakes phase indicate a stronger influence of Sr sources originating in the western cordillera and a weaker influence of more radiogenic Sr sources originating in the eastern cordillera. This suggests that prior to ~60,000 yr BP input to the salar from the northern Altiplano, via Lago Poopó to the east, was significantly diminished if not completely absent.

Grove et al. (2003) and Placzek et al. (2010) have measured Sr isotopic compositions for several important riverine and surface water sources throughout the Altiplano, producing differing results for mass balance of Sr and the resulting $^{87}Sr/^{86}Sr$ of the lake intervals that occupied the southern Altiplano during the late Quaternary. While Groves et al. and Placzek et al. disagree on the importance of some source waters to the terminal basin (specifically the importance of Lake Titicaca overflow input via Lago Poopó and the Rio Laka Jahuira), they agree that during wet periods the surface waters at Uyuni reflect with more radiogenic Sr sources, such as the Lago Poopó basin, while during arid periods the surface waters at Uyuni are representative of more local Sr sources, such as rivers and streams draining the less radiogenic Cordillera Occidental.

Table 3: Strontium isotopic data for halite and carbonate samples.

Sr isotopic data

depth (m)	age (yr BP)	description	87Sr/86Sr	depth (m)	age (yr BP)	description	87Sr/86Sr
2.23	5,135	halite	0.708504	48.44	61,394	halite	0.708518
2.50	5,731	halite	0.708558	48.55	61,496	halite	0.708236
3.30	7,466	halite	0.708544	50.50	63,340	halite	0.708226
4.00	8,947	halite	0.708532	52.15	64,966	halite	0.708242
4.35	9,675	halite	0.708591	53.21	66,046	halite	0.708251
5.47	11,945	halite	0.708532	55.20	68,160	carbonate	0.708631
8.30	17,312	halite	0.708749	55.22	68,182	carbonate	0.708391
9.95	20,209	halite	0.708839	57.20	70,409	halite	0.708131
12.55	24,454	carbonate	0.708900	62.65	77,305	carbonate	0.708306
13.35	25,685	carbonate	0.708878	69.60	88,135	halite	0.708341
13.55	25,988	carbonate	0.708901	70.72	90,134	halite	0.708380
15.50	28,834	halite	0.708899	71.25	91,107	carbonate	0.708321
16.40	30,086	halite	0.708882	77.05	102,985	halite	0.708563
21.90	36,984	halite	0.708769	80.35	110,844	halite	0.708455
33.90	48,829	halite	0.708789	82.50	116,437	halite	0.708354
34.00	48,916	halite	0.708858	89.20	136,499	carbonate	0.708409
35.05	49,827	halite	0.708827	103.95	197,188	carbonate	0.708490
36.73	51,265	carbonate	0.708837	112.80	246,609	halite	0.708274
37.62	52,020	halite	0.708774	139.35	469,038	halite	0.708242
37.70	52,088	halite	0.708774	149.05	583,309	carbonate	0.708314
37.75	52,130	halite	0.708768	160.35	742,448	halite	0.707999
40.80	54,705	halite	0.708742	161.55	761,102	carbonate	0.708488
41.44	55,247	halite	0.708733	167.00	850,283	halite	0.708102
43.10	56,661	carbonate	0.708774	167.55	859,696	carbonate	0.708292
46.35	59,497	halite	0.708674	176.65	1,026,894	halite	0.707936
47.70	60,714	halite	0.708711				
48.00	60,989	halite	0.708211				

Sr isotopic data for carbonate and halite samples in the core from the Salar de Uyuni (Baker et al., 2001b).

The playa lake stage is less radiogenic, consistent with less northern input, and the great lakes phase is more radiogenic, consistent with increased connection of the Poopó-Coipasa-Uyuni basins, as well as greater connections with Lake Titicaca. Sr isotopic compositions in the core from the Salar de Uyuni do not point to a specific dominant source for Sr during periods of lake expansion, rather $^{87}Sr/^{86}Sr$ during the great lakes phase more closely resembles $^{87}Sr/^{86}Sr$ of the Poopó basin and the Rio Desaguadero north of Lago Poopó, while $^{87}Sr/^{86}Sr$ during the playa lakes phase more closely resemble $^{87}Sr/^{86}Sr$ of less radiogenic sources east of the Salar de Uyuni. Recent modeling of Altiplano hydrology (Nunnery et al., 2012) indicates that overflow from Lake Titicaca constitutes a significant contribution of water (40-60% of riverine input) to the southern Altiplano during the highest lake stages. Based on these results, the $^{87}Sr/^{86}Sr$ of the Uyuni core during the great lakes stage is likely significantly influenced by overflow from Lake Titicaca.

3.5 Discussion

Although the literature has since become confused with different names assigned to different paleolake intervals, at least three periods of significant lake expansion in the northern Altiplano have been inferred from observations of high shoreline deposits above the modern surface of Lake Titicaca - designated as paleolakes Mataro (3,950

masl), Cabana (3,900 masl), and Ballivian (3,860 masl)(Agassiz, 1875; Bowman, 1914 and 1916; Ogilvie, 1922; Moon, 1939; Newell, 1949; Ahlfeld and Brasnia, 1960; Servant and Fontes, 1978). The ages of these paleolakes are not firmly established, though they are considered to be of early to mid-Pleistocene age, based on fossil analysis of each unit (Pompecki 1905; Hoffstetter et al., 1971) and K-Ar dating of volcanic ash beneath the oldest of these deposits (Lavenu 1984, 1986). The timing of the formation of these paleolakes has been suggested to be coincident with glacial stages of the mid-Pleistocene (Servant and Fontes, 1978).

There is no evidence in the southern basin for paleolakes coincident with the early paleolakes in the northern basin. Possible reasons for this are: (1) the sediment units attributed to lakes Mataro, Cabana, and Ballivian are not truly lacustrine and therefore do not represent an interval of lake expansion; (2) the northern and southern sub-basins were not hydrologically connected during these intervals; or (3) evidence for these lake phases in the southern basin has been lost through erosion.

On the southern Altiplano, old lacustrine deposits and algal bioherms located in the Lago Poopó basin (central Altiplano) and the basins of the Salar de Uyuni and the Salar de Coipasa (southern Altiplano)(Musters, 1877; Minchin; 1882; Troll, 1927; Moon, 1939; Ahlfeld and Brasnia, 1960; Servant and Fontes 1976; Servant, 1977; Servant and Fontes, 1978) indicate two major phases of lake expansion that are younger than the

Ballivian stage in the north, designated paleolakes Minchin and Tauca. The ages of these paleolakes have been the source of considerable controversy.

Servant and Fontes (1978) estimated the age of paleolake Tauca at 14,100-11,400 cal yr BP and the age of paleolake Minchin at 30,000-32,000 cal yr BP based on radiocarbon dates of shells found in outcropping sediments in the Lago Poopó basin. Later, Bills et al. (1994) determined the age of the highest tufa around the Coipasa and Uyuni basins, at a height of 3,760 masl and ~107 m above the present-day surface of the Salar de Uyuni, to be ~16,000 cal yr BP. Bills et al. called this lake phase paleolake Minchin, even though, based on timing instead of locality, it is the equivalent of the Servant and Fontes (1978) Tauca paleolake. This highest and largest lake stage was short-lived and desiccated completely during theBølling-Allerød time period (~15,000-13,000 yr BP). Subsequent dating of algal bioherms in the Coipasa and Uyuni basins agree with the 17,000-15,000 yr BP timing of the highest phase of paleolake Tauca (Servant et al., 1995; Rouchy et al., 1996; Sylvestre et al., 1999; Grove et al., 2003; Placzek et al., 2006). Rondeau (1990) used U/Th dating of algal bioherms and bracketed the Minchin interval between 30,000-73,000 cal yr BP. A third relatively shallow (~3,660 masl) and short-lived paleolake, designated paleolake Coipasa, was identified in late Pleistocene paleoshoreline deposits around the Salar de Uyuni, and it's age has been reported as ~13,400-11,500 cal yr BP, based on radiocarbon dating techniques (Rondeau,

1990; Servant et al., 1995; Rouchy et al., 1996; Sylvestre et al., 1999; Fornari et al., 2001; Placzek et al., 2006)

A first deep drill core (121 m) in the Salar de Uyuni (Risacher and Fritz, 2000; Fornari et al., 2001) identified alternating salts and muds beneath the surface. Age determination of the mud units within this core, based on the U/Th isochron method, suggested a paleolake Tauca interval of 12,000-16,000 cal yr BP and a paleolake Minchin interval of > 31,000 cal yr BP, with the oldest age estimates of ~68,000-73,000 yr BP (Fornari et al., 2001). Radiocarbon dating of subsurface sediments from a subsequent drill core and continuous downhole logging from the same general location (Baker et al., 2001b) permitted identification and dating of the various great lake stages in the Salar de Uyuni in super-posed stratigraphic order (as opposed to previous and subsequent work on high-stand deposits that are sampled without continuous stratigraphic context). Baker et al. (2001b) assigned dates of 46,000-36,000 yr BP for paleolake Minchin and 25,000-16,000 for paleolake Tauca. There is no representation of paleolake Coipasa in the core due to low-resolution recovery of this interval during collection. However, the natural gamma radiation log of the borehole indicates deposition of lacustrine muds centered at ~12,500 cal yr BP (Baker et al., 2001b) consistent with earlier age determinations for the Coipasa lake phase (Servant et al., 1995; Rouchy et al., 1996; Sylvestre et al., 1999; Fornari et al., 2001; Placzek et al., 2006). The timing of paleolake

Coipasa is contemporaneous with the Younger Dryas cooling event in the north Atlantic, supporting the hypothesis that southern tropical Andean precipitation is connected to temperature variability in the northern hemisphere (Baker et al., 2001b).

Placzek introduced new terminology for several presumed lake stages based on tufa deposits in the central and southern Altiplano: Sajsi (24,000-20,500 cal yr BP, ^{14}C age determination) based on low tufas in the Uyuni basin, Inca Huasi (47,000-45,000 yr BP, U/Th isochron dating) based on low tufas in the Uyuni basin, Salinas (96,000-80,000 yr BP, U/Th isochron dating) based on low tufas in the Uyuni basin, and Ouki (125,000-96,000 yr BP, U/Th isochron dating) based on high tufa deposits (up to 3,728 masl) limited to the Poopó basin. The "Sajsi" stage can be equated with the early phase of the long-lived and variable depth paleolake Tauca phase (26,000-17,000 cal yr BP) described in Baker et al. (2001). The "Inca Huasi" stage can be equated with the Minchin phase of Rondeau (1990) and Rouchy et al. (1996) and the long-lived (and mostly shallow) phase indicated in core sediments from the Salar de Uyuni between 46,000-36,000 yr BP (Baker et al., 2001). The "Salinas" phase may relate to one or more of three discrete shallow and short-lived lake phases that have been identified and dated in the drill core and logging record from the Salar de Uyuni (Fritz et al, 2004).

The "Ouki" phase named by Placzek et al. (2006) is problematic, because there are no observed long-lived lakes of this age in the drill core from Salar de Uyuni (Fritz et al.,

2004). Instead, the drill core, between 125,000-96,000 yr BP, shows evidence for long-lived saltpan conditions alternating with shallow perennial saline and hypersaline lakes (Fritz et al., 2004). Placzek et al. (2006) claim that the dating of the core by Fritz et al. (2004) was incorrect (by ~60-80 thousand years) and that there was a very long-lived great lake throughout the Poopó-Coipasa-Uyuni basin during the interval between 125,000-96,000 yr BP. However, the Ouki phase samples of Placzek were collected exclusively from the Lago Poopó basin, and thus far there have been no Ouki aged outcrops located in either the Salar de Coipasa or the Salar de Uyuni basins. Placzek et al. (2010) acknowledge the missing outcrops for the Ouki stage in the southern Altiplano basins, suggesting that the absence of contemporaneous outcrops indicates that the Poopó basin was minimally hydrologically connected to the Salar de Uyuni, via the Laka Jahuira hydrologic divide, at this time.

3.5.1 Downcutting of the Laka Jahuira hydrologic divide

Presently, the Rio Desaguadero transports overflow from Lake Titicaca to Lago Poopó in the central Altiplano. During extended periods of increased precipitation and decreased evaporation (Blodgett et al., 1997; Nunnery, 2012) increased flow of the Rio Desaguadero causes Lago Poopó to fill to it's outlet (~3,700 masl), overflowing westward to the Salar de Coipasa via the Laka Jahuira hydrologic divide. This divide generally

lies above 3,720 masl, with the exception of the Rio Laka Jahuira valley, which lies at or below 3,700 masl (Figure 23).

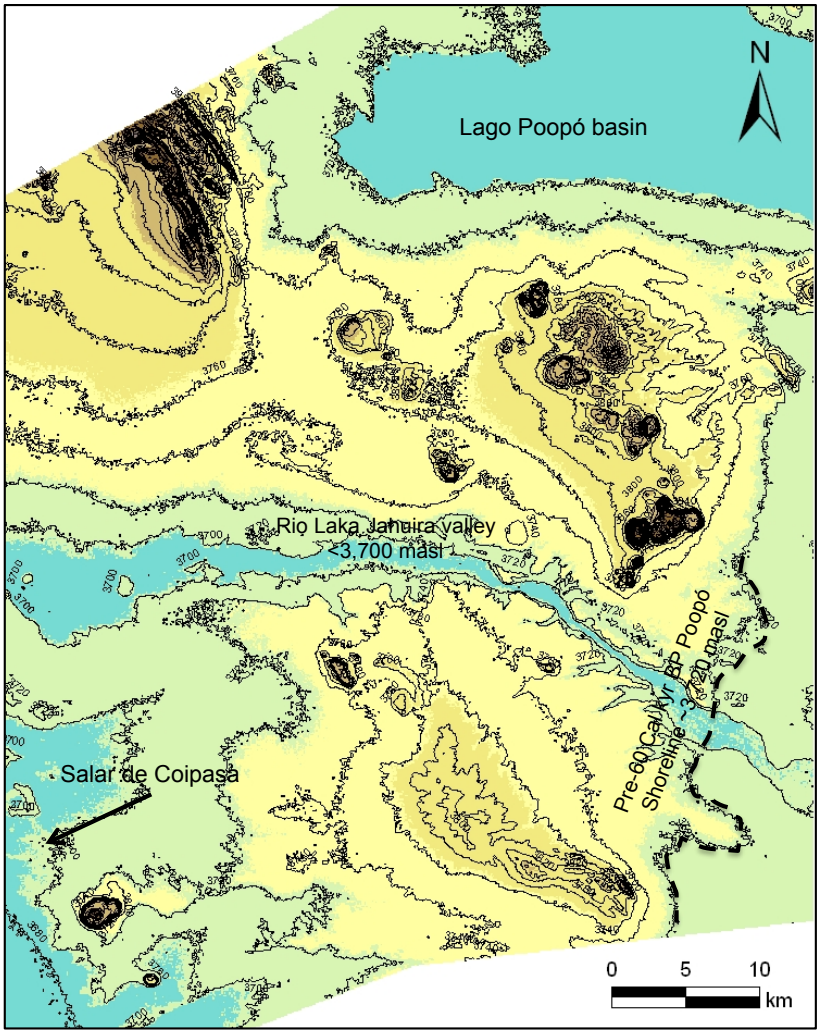

Figure 23: Topographic map of the region between Lago Poopó (east) and the Salar de Coipasa (west) known as the Laka Jahuira hydrologic divide. The Rio Laka Jahuira lies at < 3,700 masl and, during periods of increased lake level, flows westward towards the Salar de Coipasa. The dotted line along the eastern edge of the map indicates the 3,720-masl terrace, the location of a possible breach in the boundary of Lago Poopó at ~60,000 years ago

To the east and west of this rise, the Rio Laka Jahuira resembles a wide river flood plain, but in between it is an incised channel (as narrow as 1.2 km across in some locations). A conspicuous topographic feature at ~3,720 masl, which appears to be nearly continuous around the Lago Poopó basin, indicates an elevated shoreline consistent with the highest elevations of the Ouki paleolake stage. At the southeastern edge of the Poopó basin, where outflow of Lago Poopó enters the Rio Laka Jahuira, the river appears to have cut through the 3,720-masl terrace. Prior to the incision of this feature, lakes in the Poopó basin could have been raised to the level of the 3,720-masl terrace without inducing overflow westward to the Salar de Coipasa. In this basin configuration, lakes in the Coipasa and Uyuni basins would only be the product of increased local runoff and direct precipitation. We propose that the downcutting of the Laka Jahuira hydrologic divide occurred ~60,000 yr BP, such that afterward the basins of Poopó, Coipasa, and Uyuni became hydrologically connected. Sr isotopic composition of carbonates and halites from the Salar de Uyuni core, presented in this study, support the idea that a downcutting of the Laka Jahuira hydrologic divide at ~60,000 yr BP caused a state change from the playa lakes phase to the great lakes phase. This timing is coincident with diatom stratigraphy of Lake Titicaca sediments (Fritz et al., 2012) suggesting that the hydrogeomorphic character of the Lake Titicaca drainage basin changed at ~60,000 yr BP, as well as evidence from the Lake Titicaca drill core (Fritz et al.

2007) of major lake-level rise and of expanded catchment glaciation at ~60,000 yr BP. This wet period overlies an interval from ~120,000 to 60,000 yr BP of reduced or no sediment accumulation.

3.5.2 Sr isotopic composition evidence for downcutting

Prior to the 60 ka state change, average $^{87}Sr/^{86}Sr$ in carbonate and halite was 0.70829, consistent with local less radiogenic Sr sources, such as the streams draining the Cordillera Occidental and hydrothermal waters within the Salar de Uyuni basin (Groves et al., 2003; Placzek et al., 2006). Subsequent to the 60 ka state change, average $^{87}Sr/^{86}Sr$ in carbonate and halite was 0.70871, consistent with relatively more radiogenic Sr sources originating from Cordillera Oriental, similar to the Lago Poopó basin and the Rio Desaguadero north of Lago Poopó (Grove et al., 2003; Placzek et al., 2006).

The shift in $^{87}Sr/^{86}Sr$ composition from the playa lakes phase to the great lakes phase implies an increase in basin wide precipitation, with increased runoff to the Salar de Uyuni from more radiogenic sources. However, there were wet and dry periods during both the great lakes phase and the playa lakes phase, as indicated by alternating mud and salt beds in the Uyuni core (Figure 22)(Fritz et al., 2004). Additionally, both phases include similar duration intervals of perennial saline lakes (indicated by bedded halite)(Fritz et al., 2004). If $^{87}Sr/^{86}Sr$ composition was purely dependent on changes in

precipitation amount, the $^{87}Sr/^{86}Sr$ of these saline lakes should be similar throughout the record. Instead, the record shows more radiogenic $^{87}Sr/^{86}Sr$ values for halites deposited during the great lakes phase (average $^{87}Sr/^{86}Sr$ 0.7087) and less radiogenic $^{87}Sr/^{86}Sr$ values for halites deposited during the playa lakes phase (average $^{87}Sr/^{86}Sr$ 0.7082). Therefore, the increase of $^{87}Sr/^{86}Sr$ composition at the 60 ka state change implies a permanent change in source to the Salar de Uyuni, consistent with a major morphological change, such as downcutting of the 3,720 masl terrace, which allowed greater flow from Sr sources originating in the more radiogenic watersheds surrounding the Lago Poopó basin.

3.5.3 Cause of the downcutting of the Laka Jahuira divide

Downcutting of the 3,720 masl terrace at ~60,000 year BP provides a reasonable explanation for some important problems of Altiplano hydrologic history: (1) the absence of Ouki stage, and older (i.e. Mataro, Cabana, and Ballivian), paleoshoreline deposits in the Coipasa and Uyuni basins, (2) the state change apparent in the lithology of Uyuni core due to increased input from Lake Titicaca via Lago Poopó, and (3) the shift in Sr isotopic composition in the Uyuni core indicating a change in Sr source. Although the cause of the postulated downcutting of the 3,720-masl terrace, is unknown, it would necessitate high lake level, approximately equivalent to the 3,720-masl terrace,

sustained by a prolonged period of increased precipitation, in order to breach this divide.

Based on hydrologic modeling of modern topography (Nunnery, 2012), an increase of precipitation alone of ~120 mm/yr above modern values is sufficient to raise Lago Poopó to 3,700 masl, which is the elevation of the modern Poopó overflow outlet. With an increase of precipitation of ~200 mm/yr and a temperature decrease of ~5 °C (compared to modern values), Lago Poopó is at equilibrium with the Coipasa and Uyuni basins at 3,720 masl. Similar climate conditions are estimated for the paleolake Minchin phase (46,000-36,000 yr BP), with an estimated decrease in temperature consistent with glacial stage estimates (~5°C [Hostetler and Mix, 1999]) and an estimated basin wide precipitation increase of 100-200 mm/yr above modern values (Nunnery, 2012) capable of sustaining lake levels between 3,700-3,720 masl.

A combination of increased precipitation and decreased temperature, comparable to the paleolake Minchin phase and consistent with the last glacial period, was likely sufficient to maintain Lago Poopó at an elevation that either enabled the breach of the 3,720-masl terrace, or caused lake waters to rise and flood the Laka Jahuira hydrologic divide above the 3,720-masl terrace, subsequently downcutting the Laka Jahuira divide. The underlying cause of this shift in precipitation amount can be attributed to larger global-scale climate forcing.

Insolation is one control on the hydrology, lithology, and Sr isotopic composition of the Salar de Uyuni core. Comparison of the Uyuni core stratigraphy to 15°S insolation (Figure 24) shows that during the playa lakes phase short-lived saline-to-hypersaline lakes correlate with insolation maxima, and mudflat and saltpan conditions correlate with insolation minima. During the great lakes phase, thick mud units corresponding to the Minchin and Tauca paleolake phases are coincident with insolation maxima, and shallow saline lakes (indicated by bedded halite) correspond to insolation minima. The amplitude of insolation variability is greater during the playa lakes phase than during the great lakes phase, suggesting that prolonged wet conditions on the Altiplano are somehow connected to lower amplitude insolation oscillations.

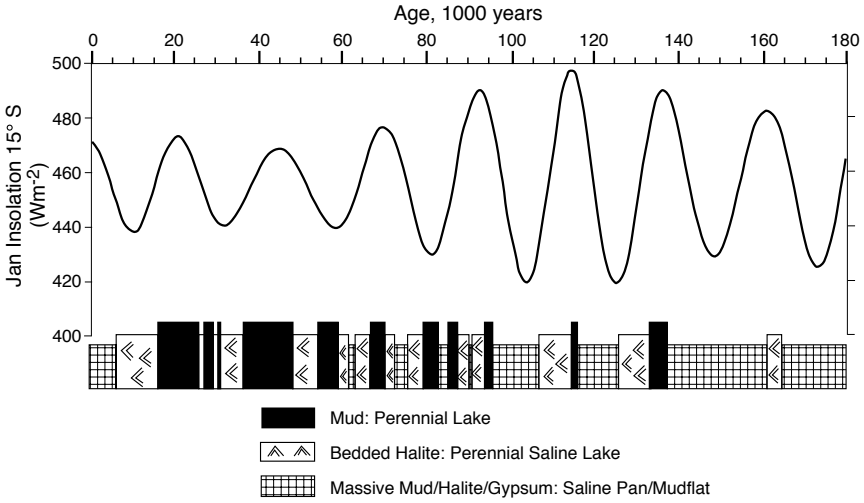

Figure 24: Comparison of the stratigraphic column of the upper 100 m of the Salar de Uyuni core and January insolation at 15°S (Berger and Loutre, 1991)

There is no evidence in the Salar de Uyuni core that prior periods of low amplitude insolation variability and global glacial periods (i.e. MIS-6) led to the creation of large lakes in the southern Altiplano (Figure 24). Therefore, even though insolation forced glaciation cycles may have affected the rise and fall of Lake Titicaca (Fritz et al. 2007), and to some degree Salar de Uyuni wetness (indicated by alternating from mudflat/saltpan conditions to shallow saline lakes) the 60 ka state change from playa lakes to great lakes was likely not the result of increased glacial stage precipitation alone. If this were the case we would expect to see similar state changes throughout the core corresponding to glacial/interglacial transitions. Instead, the 60 ka state change observed in the Salar de Uyuni core represents a unique event in the hydrological history of the basin, in which the hydrology of the central and southern Altiplano was permanently altered by downcutting of the 3,720-masl terrace at the Laka Jahuira hydrologic divide during the onset of MIS-4.

To the south of the Altiplano, but still within the tropical/sub-tropical Andes and apparently within the influence of the SASM, lies the Salar de Atacama. Bobst et al. (2001) and Lowenstein et al. (2003) report on stratigraphy of three cores from this salar showing a transition from a regime of mud flats and saltpans to one of perennial lakes approximately identical in timing to the southern Altiplano shift from the playa lakes phase to the great lakes phase. The timing of the Atacama shift is ~75,000-60,000 yr BP,

which is consistent with the transition from an interglacial interval (MIS 5) to a glacial interval (MIS 4). The correlation of transition from arid to wet conditions over such a large range of latitudes in the Andes supports the hypothesis that the processes governing the onset of the great lakes phase at the Salar de Uyuni were probably related to changes to continental-scale atmospheric processes, which were in turn likely influenced by global-scale glacial cycles (Figure 25).

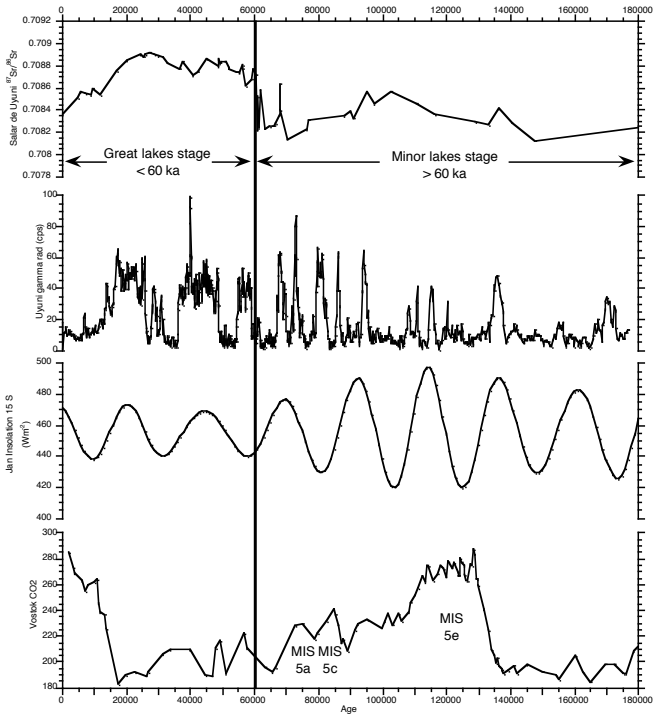

Figure 25: $^{87}Sr/^{86}Sr$ (this study) and natural gamma radiation (Baker et al., 2001) of the Salar de Uyuni core compared to January mid-month insolation at 15°S (Berger and Loutre, 1991) and Vostok CO_2 (Petit et al., 1999). The "great lakes" stage (< 60 ka) is consistent with cooler global climate conditions, while the "minor lakes" stage (> 60 ka) appears to be influenced by the MIS 5 interglacial.

This suggests that while the connection of the northern and southern Altiplano was likely caused by a relatively rapid change in morphology sometime between 30-60 ka, the overriding cause of the eventual connection was a global-scale shift from interglacial to glacial conditions, creating higher precipitation and lower evapotranspiration across the tropical and sub-tropical Andes.

3.6 Conclusions

A 220-meter core from the Salar de Uyuni shows a state change, at ~60 ka, from a regime of mud flats and saltpans to one of perennial fresh-to-hypersaline lakes. This change is indicated by changing textures of salt deposits and thickening up-sequence of lacustrine mud units. Sediments deposited prior to 60 ka are designated the "playa lakes" phase and sediments deposited subsequent to 60 ka are designated the "great lakes" phase. Strontium isotopic analysis of core sediments from the Salar de Uyuni support the interpretation for a significant hydrologic change at ~60 ka, with less radiogenic $^{87}Sr/^{86}Sr$ prior to the 60 ka state change suggesting a drier climate and more radiogenic $^{87}Sr/^{86}Sr$ following the 60 ka state change suggesting wetter conditions. The hydrologic shift observed in the record from the Salar de Uyuni is likely the result of a combination of global-scale climate change, as observed in Antarctic ice core records, and local morphological changes, which ultimately connected the northern and southern

Altiplano basins. As glacial-stage conditions increased precipitation amount in the tropical Andes, increased flow from the north to the central Altiplano created conditions allowing the Laka Jahuira hydrologic divide, located between Lago Poopó and the Salar de Uyuni to be downcut, providing a pathway for Lake Titicaca overflow to flow downstream, eventually contributing water to the advance of large lakes on the southern Altiplano at the Salar de Uyuni and the Salar de Coipasa.

Conclusions

Each study presented in this thesis gives insight into how the basins of the Altiplano have filled through time, how they interact with one another, and how they are likely to respond to future changes in climate. The most interesting result of this work is the verification that the basins in the central and southern Altiplano have different hydrologic histories. This is shown by the comparison of the new sediment core from the Salar de Coipasa to a sediment core and bore hole gamma-radiation log from the Salar de Uyuni, and Sr isotopic analysis from the Salar de Uyuni indicating separation between the central and southern Altiplano prior to ~60 ka. Results from hydrologic model simulations support these findings showing that, while there is connection between the central and southern Altiplano with large increases in precipitation (or decreases in temperature), with modest increases in precipitation (or decreases in temperature) the basins function as a series of lakes of varying surface elevation connected by straits.

Output from hydrologic modeling suggests that to create the conditions observed in paleolake records, climate on the Altiplano during the last 60 ka (described as the "great lakes" phase) fluctuated with basin wide changes in precipitation between 50-350 mm/yr compared to modern values based on maximum temperature depressions of 2-5°C from modern values. The largest lake stage on the southern Altiplano during

the last 60 ka, designated paleolake Tauca (16,000 cal yr BP), was likely caused by a precipitation increase over modern annual average values of 350 mm/yr basin wide assuming a 5 °C lower temperature relative to modern values (based on approximations for temperature on the Altiplano during the LGM). The model also shows that runoff from the northern Altiplano has a significant influence on the hydrology of the southern Altiplano, representing ~40-60% of total input to the central and southern basins via the Rio Desaguadero. The results from strontium isotopic analysis of sediments from the Salar de Uyuni during this "great lakes" phase show strong evidence of influence of source waters from the more radiogenic Lago Poopó basin of the central Altiplano, which agrees with the model suggesting a significant input to the central and southern large lakes from Lake Titicaca overflow.

Prior studies suggested that a significant change in hydrology occurred on the southern Altiplano sometime prior to 60,000 years ago (Baker et al., 2001; Bobst et al., 2001; D'Agostino et al., 2002; Lowenstein et al., 2003; Groves et al., 2004; Fritz et al., 2004; Lowenstein, unpublished), which had the effect of shifting the southern Altiplano from a phase of "playa lakes" to one of "great lakes". Strontium isotopic analysis of core sediments from the Salar de Uyuni support the interpretation for a significant hydrologic change at ~60 ka. The hydrologic shift observed in the record from the Salar de Uyuni can be attributed to a combination of global-scale climate change toward

cooler north Atlantic temperatures (having the effect of increased precipitation on the Altiplano) and the downcutting of the hydrologic divide between the central Altiplano (occupied by Lago Poopó) and the southern Altiplano. I propose the following interpretation of the Sr isotopic data from the Salar de Uyuni: As precipitation on the Altiplano increased basin wide, overflow from Lake Titicaca, combined with direct precipitation and local runoff, caused Lago Poopó to become deeper. As the lake deepened it began to overflow to the west, inundating the Laka Jahuira hydrologic divide that was generally at an elevation > 3,720 masl, and eventually cutting the Rio Laka Jahuira valley down to an elevation < 3,700 masl. Prior to this downcutting the southern basins of Coipasa and Uyuni experienced variation of wet and dry conditions, as indicated by the lithology of the 220-meter core from the Salar de Uyuni, with alternation between conditions of mud-flats/saltpans and hypersaline shallow lakes dependent on local runoff and direct precipitation and no input from Lake Titicaca overflow. After the downcutting of the Laka Jahuira hydrologic divide, Sr isotopic measurements indicate that the southern basins received a significant amount of input from the northern and central basins of the Altiplano.

The results from these three studies, combined with previously collected paleorecords throughout the region, allow for a more refined reconstruction of Altiplano basin history and a better understanding of basin dynamics, which enables more

accurate predictions of how the region will respond to expected tropical Andean temperature increases in the future. Very often, paleoclimate reconstructions for vast regions are based on very few records, mostly because of the difficulty of locating and obtaining these records. However, it is impossible to characterize the paleoclimate and paleohydrology of a large hydrologic basin, such as the Altiplano, based solely on paleoshoreline records, lacustrine sediment cores, or ice cores from adjacent glaciers. The three studies discussed above provide a multi-faceted approach at refining the paleohydrology of the Altiplano. The research presented in this thesis is a contribution to the many records collected over more than a century of expeditions to the region and the decades of hydrologic modeling that have worked towards a better understanding of how this basin has evolved during the late Quaternary.

Appendix A

Table 4: Complete carbon and nitrongen isotope data table.

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC1-0	0.3	1,147	43.7	0.01	0.08	-0.2	-16.7	9.33
SC1-0	0.3	1,147	51.8	0.05	0.41	7.8	-16.4	9.57
SC1-10	0.32	1,255	45	0.03	0.44	12.1	-15.3	17.04
SC1-12	0.34	1,364	54.7	0.04	0.49	9.6	-15.6	13.53
SC1-14	0.36	1,474	66	0.05	0.49	8	-15.7	12.09
SC1-16	0.38	1,585	49.4	0.05	0.51	8	-15.6	11.90
SC1-18	0.4	1,696	45.1	0.05	0.45	5.2	-14.8	10.45
SC1-2	0.42	1,808	56.8	0.01	0.14	17.6	-15.7	11.46
SC1-20	0.44	1,921	50.9	0.05	0.49	5.3	-14.9	10.63
SC1-22	0.46	2,035	51.2	0.02	0.22	2.8	-15.0	10.73
SC1-4	0.48	2,150	57.6	0.02	0.34	15.2	-15.2	23.47
SC1-6	0.5	2,265	51	0.02	0.39	13.7	-15.4	21.66
SC1-8	0.52	2,382	48.4	0.05	0.43	7.6	-16.0	10.03
SC2-0	0.77	3,904	43.1	0.02	0.14	-1.4	-16.5	8.17
SC2-2	0.79	4,031	53.1	0.02	0.16	7.8	-16.0	9.33
SC2-4	0.81	4,160	53.6	0.02	0.17	3.7	-16.5	9.05
SC2-6	0.83	4,288	60.2	0.02	0.18	4.6	-15.8	9.94
SC2-8	0.85	4,418	49.4	0.02	0.22	4.9	-15.8	10.62
SC2-10	0.87	4,549	50.3	0.01	0.12	6.9	-16.9	14.00
SC2-12	0.89	4,680	43	0.02	0.14	2.6	-16.7	8.17
SC2-14	0.91	4,812	48.7	0.02	0.18	3.9	-17.2	10.75
SC3-0	1.1	6,108	46.3	0.03	0.25	8.3	-17.5	9.72
SC4-0	1.2	6,819	50.2	0.02	0.21	6.1	-16.1	10.38
SC4-2	1.22	6,964	49.6	0.02	0.21	6.7	-17.0	11.42
SC4-4	1.24	7,110	47.6	0.03	0.29	7.1	-17.3	11.78
SC4-6	1.26	7,256	54.2	0.03	0.3	8.3	-17.9	11.67
SC4-8	1.28	7,403	54	0.06	0.72	9.7	-20.9	14.97
SC4-10	1.3	7,551	47.7	0.05	0.67	9.7	-20.8	15.33
SC4-12	1.32	7,700	47.6	0.04	0.54	9.6	-20.3	14.31
SC4-14	1.34	7,850	50	0.03	0.25	8	-18.2	9.72
SC4-16	1.36	8,000	45.4	0.06	0.74	10	-21.1	15.19

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC4-18	1.38	8,151	49	0.05	0.76	10.3	-21.4	16.38
SC4-20	1.4	8,303	47.2	0.06	0.85	10.3	-21.7	16.73
SC4-22	1.42	8,456	51.4	0.07	0.81	9.5	-22.4	13.50
SC4-24	1.44	8,610	47.2	0.05	0.82	9.9	-22.5	18.33
SC4-26	1.46	8,764	42.2	0.06	0.86	9	-22.1	16.80
SC4-28	1.48	8,919	42.1	0.06	0.81	9.2	-22.3	16.91
SC4-30	1.5	9,075	58.5	0.05	0.62	8.4	-23.0	14.47
SC4-32	1.52	9,232	47.3	0.05	0.78	8	-23.1	16.63
SC4-34	1.54	9,390	47.9	0.05	0.76	8.8	-23.2	17.36
SC4-36	1.56	9,549	46	0.06	0.83	8	-23.0	17.40
SC4-38	1.58	9,708	42.4	0.06	2.41	8.2		46.86
SC4-40	1.6	9,868	45.2	0.05	0.73	8	-23.2	16.60
SC4-42	1.62	10,029	54.6	0.05	0.69	7.7	-23.6	17.95
SC4-44	1.64	10,191	48.8	0.04	0.66	7.7	-23.7	17.52
SC4-46	1.66	10,353	51.8	0.06	0.72	7.9	-24.7	14.00
SC4-48	1.68	10,517	47.2	0.05	0.77	9	-23.8	17.45
SC4-50	1.7	10,681	48.1	0.04	0.87	16.7	-23.6	23.94
SC4-52	1.72	10,846	45.8	0.06	0.78	10.6	-24.3	15.17
SC4-54	1.74	11,012	48.9	0.07	0.81	9	-24.6	13.50
SC4-56	1.76	11,178	50.4	0.07	0.98	9.8	-24.6	16.33
SC4-58	1.78	11,346	46.8	0.10	1.47	11.2	-24.3	17.15
SC4-60	1.8	11,514	51.8	0.11	1.68	11	-21.6	17.82
SC4-62	1.82	11,683	52.4	0.10	1.5	10.2	-22.3	17.50
SC4-64	1.84	11,853	59.6	0.09	1.32	10.6	-22.5	17.11
SC4-66	1.86	12,024	49.1	0.11	1.45	11.1	-20.7	15.38
SC4-68	1.88	12,195	49.8	0.07	0.87	11.4	-19.7	14.50
SC4-70	1.9	12,368	45.8	0.08	1	10.1	-20.5	14.58
SC4-72	1.92	12,541	55.4	0.07	0.84	9.3	-19.3	14.00
SC4-74	1.94	12,715	54.1	0.14	1.91	8.2	-23.9	15.92
SC4-76	1.96	12,889	48.4	0.16	2.05	8.5	-22.2	14.95
SC4-77	1.97	12,977	49.6	0.21	2.77	5.2	-23.3	15.39
SC4-78	1.98	13,065	46.4	0.20	2.88	6.8	-20.0	16.80
SC4-80	2	13,241	51.3	0.10	1.16	1.8	-21.7	13.53
SC4-82	2.02	13,418	57.1	0.14	1.64	8	-22.0	13.67
SC4-84	2.04	13,596	57	0.15	1.72	10.6	-21.3	13.38
SC4-85	2.05	13,686	62.7	0.13	1.46	6.1	-22.8	13.10

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC4-86	2.06	13,775	47.7	0.15	1.88	7.2	-20.3	14.62
SC4-88	2.08	13,955	45.1	0.11	1.23	7.1	-22.2	13.05
SC4-90	2.1	14,135	50.9	0.11	1.19	6.7	-22.1	12.62
SC4-92	2.12	14,317	49.8	0.11	1.24	6.7	-22.1	13.15
SC4-94	2.14	14,499	52.5	0.10	1.18	6.1	-21.9	13.77
SC4-96	2.16	14,681	59.6	0.10	1.09	6.2	-22.1	12.72
SC4-98	2.18	14,865	49.3	0.10	1.11	6.5	-21.8	12.95
SC5-2	2.22	15,235	59.8	0.15	1.75	4.1	-22.4	13.61
SC5-4	2.24	15,421	39.9	0.11	1.14	5	-21.8	12.09
SC5-5	2.25	15,514	51	0.21	1.96	4.8	-22.8	10.89
SC5-6	2.26	15,608	48.1	0.12	1	6.8	-19.5	9.72
SC5-8	2.28	15,796	61.6	0.09	0.66	7	-17.7	8.56
SC5-10	2.3	15,984	47.7	0.07	0.48	6.8	-17.3	8.00
SC5-12	2.32	16,174	48.4	0.07	0.42	6.8	-17.0	7.00
SC5-14	2.34	16,364	55.8	0.07	0.44	6.8	-17.5	7.33
SC5-16	2.36	16,555	51.9	0.07	0.41	7.3	-17.3	6.83
SC5-18	2.38	16,747	57.8	0.06	0.35	6.6	-17.7	6.81
SC5-20	2.4	16,939	54.4	0.06	0.29	6.3	-17.6	5.64
SC5-22	2.42	17,133	48.3	0.05	0.29	6.7	-18.0	6.77
SC5-24	2.44	17,327	50.1	0.05	0.25	6.1	-17.8	5.83
SC5-26	2.46	17,522	51.2	0.05	0.24	5.7	-18.7	5.60
SC5-28	2.48	17,718	48.2	0.05	0.26	6.6	-18.8	6.07
SC5-30	2.5	17,914	46.1	0.05	0.23	5.8	-18.8	5.37
SC5-32	2.52	18,112	48.1	0.04	0.22	6.5	-18.7	6.42
SC5-34	2.54	18,310	54.7	0.05	0.21	6	-19.0	4.90
SC5-36	2.56	18,509	55.8	0.05	0.21	6.5	-18.2	4.90
SC5-38	2.58	18,709	52.3	0.04	0.19	6.7	-18.1	5.54
SC5-40	2.6	18,910	58.6	0.04	0.2	6.6	-18.1	5.83
SC5-42	2.62	19,111	47.3	0.05	0.2	6	-18.9	4.67
SC5-44	2.64	19,263	44.9	0.04	0.16	7	-17.8	4.67
SC5-46	2.66	19,295	50.2	0.05	0.22	6.9	-17.3	5.13
SC5-48	2.68	19,327	47.7	0.05	0.22	6	-16.9	5.13
SC5-50	2.7	19,359	47	0.05	0.2	6.1	-19.3	4.67
SC5-54	2.7	19,359	50	0.05	0.21	6.7	-17.5	4.90
SC5-52	2.72	19,391	51.8	0.05	0.2	6.4	-17.4	4.67
SC5-56	2.76	19,455	59.9	0.05	0.23	6.6	-18.1	5.37

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC5-58	2.78	19,487	49.2	0.05	0.2	5.5	-19.1	4.67
SC5-60	2.8	19,519	59.8	0.05	0.26	6.6	-19.7	6.07
SC5-62	2.82	19,550	51.6	0.05	0.21	6.3	-18.1	4.90
SC5-64	2.84	19,581	56	0.05	0.25	7	-18.7	5.83
SC5-66	2.86	19,613	43	0.04	0.17	5.8	-19.3	4.96
SC5-68	2.88	19,644	46.9	0.05	0.23	6.5	-18.1	5.37
SC5-70	2.9	19,675	42.1	0.05	0.21	6.1	-17.7	4.90
SC5-72	2.92	19,706	54.4	0.05	0.21	6.3	-17.5	4.90
SC5-74	2.94	19,737	43.3	0.05	0.21	5.5	-19.1	4.90
SC5-76	2.96	19,768	53.4	0.05	0.23	6.5	-18.1	5.37
SC5-78	2.98	19,799	56.9	0.05	0.21	6.1	-17.7	4.90
SC5-80	3	19,829	47	0.05	0.21	6.3	-17.5	4.90
SC5-82	3.02	19,860	54.6	0.04	0.17	5	-20.4	4.96
SC5-84	3.04	19,890	47.4	0.05	0.2	5.3	-19.2	4.67
SC5-88	3.08	19,951	52.2	0.04	0.18	5.5	-19.0	5.25
SC5-90	3.1	19,981	42.1	0.05	0.19	5	-20.1	4.43
SC5-92	3.12	20,011	60.4	0.06	0.25	5.6	-18.4	4.86
SC6-6	3.26	20,219	54.4	0.05	0.33	5.8	-20.2	7.70
SC6-8	3.28	20,248	48	0.06	0.46	6.9	-20.4	8.94
SC6-10	3.3	20,277	48.4	0.05	0.45	7.2	-20.3	10.50
SC6-12	3.32	20,306	53.8	0.05	0.38	6.3	-20.0	8.87
SC6-14	3.34	20,335	44.7	0.05	0.3	6	-19.2	7.00
SC6-16	3.36	20,364	60.9	0.05	0.42	6.7	-19.2	9.80
SC6-80	4	21,236	54.6	0.05	0.32	6.1	-19.0	7.47
SC6-82	4.02	21,262	51.8	0.05	0.3	6.2	-18.5	7.00
SC6-86	4.06	21,313	51.9	0.05	0.25	5.6	-19.6	5.83
SC6-88	4.08	21,338	47	0.05	0.28	6.5	-19.2	6.53
SC6-92	4.12	21,388	51.7	0.05	0.25	6.3	-19.0	5.83
SC6-94	4.14	21,413	9.4	0.25	1.25	5.1	-19.7	5.83
SC6-96	4.16	21,438	47.3	0.05	0.3	6.9	-18.3	7.00
SC7-62	4.82	22,206	52.4	0.04	0.3	6.7	-18.5	8.75
SC7-64	4.84	22,227	55.9	0.05	0.33	7.4	-18.2	7.70
SC7-66	4.86	22,249	46.3	0.04	0.29	6.6	-18.0	8.46
SC7-68	4.88	22,270	54.7	0.04	0.27	7.1	-17.7	7.88
SC7-70	4.9	22,292	51.2	0.04	0.31	7.3	-17.6	9.04
SC7-72	4.92	22,313	51.2	0.03	0.27	7.5	-17.6	10.50

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC7-74	4.94	22,334	58.4	0.04	0.29	7.1	-18.5	8.46
SC7-76	4.96	22,355	52.9	0.03	0.24	7.6	-17.4	9.33
SC7-78	4.98	22,376	46.6	0.04	0.31	7.4	-17.4	9.04
SC7-80	5	22,397	45.7	0.04	0.39	9	-19.2	11.38
SC7-82	5.02	22,417	47.6	0.05	0.28	6.4	-18.5	6.53
SC7-84	5.04	22,438	55.2	0.04	0.24	5.7	-19.3	6.72
SC7-86	5.06	22,458	54.1	0.04	0.23	6.2	-18.9	6.76
SC7-88	5.08	22,479	47.8	0.05	0.26	4.9	-19.0	5.89
SC7-90	5.1	22,499	45.2	0.05	0.22	5.7	-19.3	5.13
SC7-92	5.12	22,519	48.5	0.05	0.25	6	-19.1	6.14
SC7-94	5.14	22,539	55.1	0.02	0.22	7.4	-16.1	11.89
SC7-96	5.16	22,559	44.6	0.04	0.45	9.1	-20.8	14.09
SC7-98	5.18	22,579	51.3	0.07	0.66	7.1	-20.7	11.00
SC8-0	5.2	22,599	46.4	0.05	0.26	7.2	-18.4	6.18
SC8-4	5.24	22,638	49.4	0.05	0.28	7.7	-18.6	6.33
SC8-6	5.26	22,658	42.9	0.05	0.27	6.9	-18.4	6.30
SC8-8	5.28	22,677	50	0.05	0.29	7	-18.3	6.51
SC8-12	5.32	22,716	46	0.05	0.26	6.4	-18.8	5.98
SC8-14	5.34	22,735	55.4	0.05	0.29	6.3	-18.7	6.77
SC8-16	5.36	22,754	46.5	0.05	0.31	6.7	-18.2	6.52
SC8-20	5.4	22,792	47.7	0.06	0.28	6.6	-17.9	5.96
SC8-22	5.42	22,811	45.6	0.06	0.29	6.9	-19.0	5.64
SC8-24	5.44	22,829	48.8	0.06	0.31	6.9	-18.7	6.42
SC8-28	5.48	22,866	50.8	0.05	0.26	6.4	-19.5	5.67
SC8-30	5.5	22,885	55.2	0.05	0.25	6.2	-20.5	5.83
SC8-32	5.52	22,903	50.4	0.05	0.24	6.7	-19.1	6.10
SC8-36	5.56	22,939	56.6	0.05	0.27	6.6	-19.0	6.54
SC8-38	5.58	22,957	48.4	0.05	0.28	6.3	-20.0	6.53
SC8-40	5.6	22,975	50.3	0.06	0.38	7.3	-17.4	7.26
SC8-44	5.64	23,010	50.9	0.07	0.37	7.7	-16.5	6.60
SC8-46	5.66	23,028	54.7	0.08	0.48	7.4	-16.6	7.00
SC8-48	5.68	23,045	45.4	0.06	0.38	7.5	-15.8	6.87
SC8-52	5.72	23,080	48.6	0.07	0.4	7.7	-15.7	7.06
SC8-54	5.74	23,097	48.4	0.06	0.37	7.3	-18.1	7.19
SC8-56	5.76	23,114	45.5	0.06	0.39	7.7	-16.3	6.99
SC8-60	5.8	23,148	55.3	0.06	0.35	7.8	-16.8	6.81

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC8-62	5.82	23,165	48.8	0.06	0.31	7.3	-18.2	6.03
SC8-64	5.84	23,181	53.5	0.06	0.34	7	-17.7	6.45
SC8-68	5.88	23,214	48.5	0.06	0.32	6.7	-17.8	6.33
SC8-70	5.9	23,231	45.2	0.05	0.26	6.2	-19.7	6.07
SC8-72	5.92	23,247	56.3	0.05	0.27	6.5	-18.7	6.48
SC8-76	5.96	23,279	53	0.04	0.23	6.9	-18.2	6.02
SC8-78	5.98	23,295	48.6	0.05	0.25	6.5	-19.9	5.83
SC8-80	6	23,311	58.5	0.05	0.26	6.6	-18.6	6.33
SC8-84	6.04	23,342	44.7	0.05	0.29	6.4	-18.1	6.24
SC8-86	6.06	23,358	50.4	0.05	0.26	6.2	-19.9	6.07
SC8-88	6.08	23,373	45.9	0.05	0.25	6	-18.0	5.65
SC8-94	6.14	23,419	55.3	0.05	0.26	5.7	-19.3	6.07
SC9-2	6.22	23,479	48.4	0.06	0.27	6	-19.3	5.25
SC9-4	6.24	23,494	51.3	0.05	0.28	5.8	-18.1	6.35
SC9-8	6.28	23,523	55	0.05	0.27	5.9	-19.2	5.88
SC9-10	6.3	23,537	46.1	0.05	0.24	5.8	-19.4	5.60
SC9-12	6.32	23,551	50.8	0.05	0.25	5.9	-18.8	6.08
SC9-16	6.36	23,580	43.7	0.05	0.25	6.4	-18.1	5.77
SC9-18	6.38	23,594	47.9	0.05	0.25	5.6	-19.1	5.83
SC9-20	6.4	23,608	52.8	0.04	0.22	5.8	-19.2	6.04
SC9-24	6.44	23,635	54.5	0.04	0.22	6	-18.8	5.88
SC9-26	6.46	23,649	46.2	0.04	0.17	5.7	-20.8	4.96
SC9-28	6.48	23,662	45.9	0.04	0.2	6	-19.5	5.64
SC9-32	6.52	23,689	55.8	0.05	0.21	5.7	-18.8	4.90
SC9-34	6.54	23,702	48.1	0.05	0.21	5.3	-19.9	4.90
SC9-36	6.56	23,715	52.1	0.05	0.21	5.5	-19.1	4.90
SC9-40	6.6	23,741	49.3	0.05	0.2	5.6	-18.9	4.67
SC9-42	6.62	23,754	45.4	0.05	0.19	5.1	-20.5	4.43
SC9-44	6.64	23,767	52.9	0.04	0.17	5.7	-19.2	4.96
SC9-48	6.68	23,792	47.6	0.05	0.21	6.1	-19.2	4.90
SC9-50	6.7	23,804	59	0.05	0.22	5.5	-20.4	5.13
SC9-52	6.72	23,817	51.7	0.05	0.24	6	-19.0	5.60
SC9-56	6.76	23,841	54.8	0.05	0.21	5.9	-18.7	4.90
SC9-58	6.78	23,853	52.9	0.05	0.24	5.3	-19.7	5.60
SC9-60	6.8	23,865	48	0.05	0.25	6.2	-18.9	5.83
SC9-64	6.84	23,889	50.8	0.05	0.28	6	-17.8	6.53

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC9-66	6.86	23,900	52.9	0.06	0.33	5	-19.4	6.42
SC9-68	6.88	23,912	63.5	0.03	0.17	5.7	-18.5	6.61
SC9-72	6.92	23,935	52.2	0.06	0.29	4.1	-19.4	5.64
SC9-74	6.94	23,946	49.3	0.06	0.28	6.3	-20.7	5.44
SC9-76	6.96	23,957	49.5	0.05	0.29	6.5	-19.8	6.77
SC9-80	7	23,979	49.6	0.05	0.3	7	-20.6	7.00
SC9-82	7.02	23,990	50.7	0.05	0.26	6	-22.2	6.07
SC9-84	7.04	24,001	64.8	0.05	0.29	6.4	-22.9	6.77
SC9-88	7.08	24,022	51.2	0.06	0.36	5.5	-19.8	7.00
SC9-90	7.1	24,032	51.4	0.05	0.28	5.8	-20.3	6.53
SC9-92	7.12	24,043	42.2	0.06	0.32	6.6	-19.2	6.22
SC10-2	7.22	24,093	46.5	0.06	0.34	5.8	-19.3	6.61
SC10-8	7.28	24,122	42.2	0.06	0.38	5.1	-18.6	7.04
SC10-10	7.3	24,131	54.6	0.06	0.41	5.4	-18.6	7.97
SC10-12	7.32	24,141	47.2	0.06	0.34	5.3	-19.4	6.37
SC10-14	7.34	24,150	51.6	0.06	0.38	5.6	-19.7	6.95
SC10-18	7.38	24,168	51.5	0.06	0.35	5.8	-18.0	6.81
SC10-20	7.4	24,177	51.2	0.06	0.4	6.3	-18.7	7.53
SC10-24	7.44	24,195	45.9	0.06	0.3	4.6	-19.3	6.35
SC10-26	7.46	24,204	48	0.05	0.2	5.2	-20.3	4.67
SC10-32	7.52	24,229	49.4	0.06	0.3	4.7	-19.2	6.26
SC10-34	7.54	24,237	50.1	0.06	0.34	5.3	-18.7	6.61
SC10-38	7.58	24,254	44.8	0.05	0.24	4.7	-20.5	5.28
SC10-42	7.62	24,270	54.1	0.05	0.24	5.2	-21.3	5.60
SC10-44	7.64	24,277	50.1	0.03	0.12	4	-21.7	5.19
SC10-48	7.68	24,293	54.9	0.04	0.19	4.9	-21.2	5.46
SC10-50	7.7	24,300	44.8	0.06	0.37	6.5	-20.6	7.19
SC10-52	7.72	24,308	53.4	0.06	0.4	5.6	-21.6	7.84
SC10-56	7.76	24,322	50.6	0.06	0.33	5.4	-21.4	6.63
SC10-58	7.78	24,329	53.2	0.06	0.31	6.9	-20.4	6.03
SC10-60	7.8	24,336	49.6	0.06	0.33	5.9	-19.9	6.40
SC10-70	7.9	24,370	52.8	0.08	0.44	7.1	-17.9	6.42
SC10-76	7.96	24,389	50.1	0.07	0.49	7	-16.8	7.75
SC10-78	7.98	24,395	56	0.08	0.51	8.6	-15.9	7.44
SC10-80	8	24,401	42.6	0.08	0.54	7.1	-15.8	7.77
SC10-82	8.02	24,407	41.2	0.08	0.57	7.8	-15.1	8.09

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC10-84	8.04	24,413	54.6	0.08	0.54	7.4	-15.3	8.32
SC10-86	8.06	24,418	44.1	0.08	0.53	8.1	-15.0	7.73
SC10-90	8.1	24,430	48.3	0.08	0.65	7.8	-15.4	9.86
SC10-94	8.14	24,440	53.1	0.07	0.54	8.8	-15.3	9.00
SC10-96	8.16	24,446	48.9	0.03	0.24	6.5	-16.0	8.02
SC11-2	8.22	24,461	55.5	0.06	0.42	7.8	-16.3	8.17
SC11-4	8.24	24,466	46.9	0.06	0.43	7.1	-16.7	8.07
SC11-8	8.28	24,475	55	0.07	0.53	7.5	-16.8	8.55
SC11-10	8.3	24,479	50.1	0.08	0.6	8.5	-16.0	8.75
SC11-12	8.32	24,484	41.6	0.08	0.64	7	-16.4	8.88
SC11-14	8.34	24,488	48	0.08	0.58	6.2	-17.6	8.81
SC11-16	8.36	24,493	43.9	0.10	0.84	6.4	-17.9	9.78
SC11-18	8.38	24,497	59.4	0.12	1.06	8.1	-17.5	10.31
SC11-20	8.4	24,501	44.6	0.12	0.96	7.2	-17.8	9.68
SC11-22	8.42	24,505	48.4	0.11	0.89	7.6	-18.0	9.60
SC11-24	8.44	24,509	47.2	0.10	0.9	8	-18.0	10.68
SC11-26	8.46	24,512	45.1	0.12	1.12	8.4	-17.7	10.89
SC11-28	8.48	24,516	51.6	0.08	0.61	6.1	-17.8	8.54
SC11-30	8.5	24,520	54.2	0.07	0.49	5	-18.5	7.78
SC11-32	8.52	24,523	44.8	0.07	0.55	4.6	-16.8	8.64
SC11-34	8.54	24,526	62.9	0.09	0.66	6.4	-17.5	8.56
SC11-36	8.56	24,530	51.6	0.08	0.51	4.3	-19.2	7.71
SC11-38	8.58	24,533	54.8	0.07	0.45	4.4	-19.5	7.70
SC11-40	8.6	24,536	55.7	0.07	0.54	4.4	-18.5	8.58
SC11-42	8.62	24,539	55.9	0.07	0.4	4.9	-18.8	6.67
SC11-44	8.64	24,542	51.8	0.06	0.42	4.4	-18.2	8.28
SC11-46	8.66	24,545	48.1	0.06	0.44	4.6	-18.6	8.01
SC11-48	8.68	24,547	49.9	0.06	0.37	5.1	-18.6	7.03
SC11-50	8.7	24,550	42	0.06	0.37	4.9	-18.1	7.19
SC11-52	8.72	24,552	54.7	0.06	0.36	4.6	-18.3	7.07
SC11-54	8.74	24,555	44.3	0.05	0.3	4.1	-17.4	6.53
SC11-56	8.76	24,557	57.4	0.08	0.74	4.3	-14.1	10.65
SC11-58	8.78	24,559	51.6	0.06	0.66	4.5	-13.2	12.83
SC11-60	8.8	24,561	57.4	0.08	0.66	3.9	-16.4	9.85
SC11-62	8.82	24,563	56.9	0.08	0.65	4.1	-16.7	9.89
SC11-66	8.86	25,025	52.5	0.04	0.41	3.4	-13.2	11.96

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC11-70	8.9	25,494	45.9	0.14	1.69	2.8	-11.1	13.97
SC11-74	8.94	25,959	45.5	0.16	2.21	3.2	-10.7	16.11
SC11-78	8.98	26,419	49.9	0.12	1.3	3.2	-12.5	12.89
SC11-82	9.02	26,874	43	0.10	5.91	1.9		68.95
SC11-86	9.06	27,324	46	0.16	2.17	2	-11.3	16.22
SC11-90	9.1	27,770	59.1	0.18	3.26	2.2	-10.4	21.13
SC11-94	9.14	28,211	52.7	0.08	0.75	4.6	-14.6	11.35
SC12-2	9.22	29,079	53.4	0.05	0.44	4.1	-15.6	10.20
SC12-6	9.26	29,506	45.3	0.07	0.74	4.9	-16.0	12.33
SC12-10	9.3	29,928	46.6	0.11	1.32	3	-12.2	14.16
SC12-14	9.34	30,345	50.1	0.08	0.8	4.6	-13.4	11.67
SC12-18	9.38	30,758	48.4	0.09	1.03	3.5	-12.9	13.09
SC12-22	9.42	31,165	50.3	0.08	1.13	4	-12.7	16.48
SC12-26	9.46	31,569	50.7	0.07	0.6	4.2	-15.3	9.95
SC12-30	9.5	31,967	46.1	0.07	0.58	5	-15.1	9.67
SC12-34	9.54	32,361	53.4	0.07	0.5	4.8	-17.2	8.19
SC12-38	9.58	32,750	54.8	0.06	0.5	5.4	-17.1	9.72
SC12-42	9.62	33,134	49.7	0.06	0.49	4.5	-17.2	9.80
SC12-46	9.66	33,514	56	0.06	0.59	5.1	-15.6	11.47
SC12-50	9.7	34,486	47.6	0.07	0.39	6.1	-18.6	6.79
SC12-54	9.74	34,587	50.8	0.06	0.31	7.3	-18.8	6.03
SC12-58	9.78	34,688	54.4	0.06	0.29	6.6	-19.9	5.46
SC12-62	9.82	34,789	42.7	0.07	0.29	7.2	-19.8	4.83
SC12-66	9.86	34,890	50.3	0.07	0.41	6.4	-19.5	7.01
SC12-70	9.9	34,991	44.4	0.07	0.38	6.3	-20.5	6.33
SC12-74	9.94	35,092	55.8	0.06	0.31	6.4	-19.2	5.66
SC12-78	9.98	35,193	46.3	0.06	0.27	6.9	-19.0	5.25
SC12-82	10.02	35,293	48.5	0.07	0.34	5.6	-18.9	5.62
SC12-86	10.06	35,394	49	0.06	0.32	6.3	-19.2	6.22
SC12-90	10.1	35,495	46.2	0.07	0.37	6.6	-19.3	5.79
SC12-94	10.14	35,596	47.4	0.06	0.34	6.6	-18.5	6.61
SC13-2	10.22	35,798	50	0.06	0.33	6.9	-19.1	6.42
SC13-6	10.26	35,899	44.7	0.06	0.35	6.3	-18.6	6.59
SC13-10	10.3	36,000	47.1	0.05	0.27	7	-19.7	6.30
SC13-14	10.34	36,101	56	0.07	0.44	6.8	-17.6	6.94
SC13-18	10.38	36,202	39	0.06	0.84	7.7	-13.0	16.33

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC13-22	10.42	36,303	44.8	0.05	0.37	5.8	-17.0	7.82
SC13-26	10.46	36,404	47.5	0.06	2.73	8.2	-9.7	53.08
SC13-30	10.5	36,505	51.8	0.07	0.62	7.9	-16.3	9.68
SC13-34	10.54	36,606	43.9	0.08	0.65	9.5	-15.6	9.48
SC13-38	10.58	36,707	55.2	0.09	0.7	8.1	-15.9	9.46
SC13-42	10.62	36,808	39.3	0.08	1.32	11.1	-12.3	19.25
SC13-46	10.66	36,909	46.8	0.08	0.68	8.7	-18.3	9.49
SC13-50	10.7	37,010	40.3	0.07	0.52	8.2	-18.6	8.67
SC13-54	10.74	37,111	48	0.09	0.59	8.2	-18.3	8.07
SC13-58	10.78	37,212	40.7	0.08	0.55	8	-16.6	8.02
SC13-62	10.82	37,313	51.7	0.09	0.69	8.2	-15.8	9.43
SC13-66	10.86	37,414	44.9	0.09	1.14	9	-12.8	14.78
SC13-70	10.9	37,515	49.5	0.06	0.57	7.8	-15.9	10.18
SC13-74	10.94	37,616	38.6	0.08	1.02	9.3	-15.9	14.88
SC13-78	10.98	37,717	52.1	0.08	0.61	8.3	-18.8	9.29
SC13-82	11.02	37,818	44.2	0.08	0.62	8	-18.3	9.04
SC13-86	11.06	37,919	45.5	0.09	0.74	8.7	-16.0	9.95
SC13-90	11.1	38,020	38.9	0.08	1.16	9.8	-13.1	16.92
SC13-94	11.14	38,121	45.9	0.07	0.69	9.2	-16.1	11.90
SC13-98	11.18	38,222	42.6	0.08	1.75	10.2	-11.5	25.52
SC14-2	11.22	38,323	43	0.08	0.86	8.9	-15.7	12.44
SC14-6	11.26	38,424	35.4	0.09	0.95	9.5	-15.7	12.31
SC14-10	11.3	38,525	48.5	0.07	0.79	8.4	-15.8	12.34
SC14-14	11.34	38,626	39.5	0.08	2.22	9.2	-9.7	32.38
SC14-18	11.38	38,727	47.9	0.09	1.06	8.4	-15.6	13.24
SC14-22	11.42	38,828	47.4	0.09	1.3	9.3	-11.8	16.85
SC14-26	11.46	38,928	53.5	0.07	1.41	8.3	-6.9	23.94
SC14-30	11.5	39,029	49.2	0.08	4.13	9.4	-3.3	60.23
SC14-34	11.54	39,130	46.9	0.10	1.04	7.6	-16.3	12.09
SC14-38	11.58	39,231	38.9	0.11	1.17	8.7	-15.9	12.41
SC14-42	11.62	39,332	49.3	0.06	1.09	7.5	-9.1	20.86
SC14-46	11.66	39,433	39.2	0.09	0.91	6.9	-18.3	11.80
SC14-48	11.68	39,484	43.3	0.10	1.01	7	-20.6	11.32
SC14-54	11.74	39,635	49.7	0.05	0.45	6.1	-22.8	10.50
SC14-58	11.78	39,736	56.3	0.01	0.06	7.3	-21.6	6.89
SC14-62	11.82	39,837	39.2	0.23	2.7	5.9	-22.8	13.70

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC14-66	11.86	39,938	45.8	0.17	1.91	8.7	-23.4	13.39
SC14-70	11.9	40,039	45.7	0.08	0.67	7.5	-17.6	9.77
SC14-74	11.94	40,140	50.5	0.08	0.69	6.1	-18.7	10.13
SC14-78	11.98	40,241	43.1	0.07	0.64	7.2	-20.1	10.67
SC14-82	12.02	40,342	45.4	0.07	0.52	6.6	-19.2	8.22
SC14-86	12.06	40,443	40.7	0.06	0.51	7.3	-18.9	9.92
SC14-90	12.1	40,544	50.4	0.04	0.33	7	-19.2	9.47
SC14-94	12.14	40,645	56.5	0.07	0.6	8.3	-17.9	10.00
SC15-2	12.22	40,847	41.9	0.07	0.61	8.2	-17.4	10.17
SC15-4	12.24	40,897	47.8	0.08	0.71	8.3	-17.5	10.35
SC15-8	12.28	40,998	52.9	0.07	0.52	8	-18.2	8.94
SC15-10	12.3	41,049	47.6	0.06	0.46	9	-17.9	8.94
SC15-16	12.36	41,200	46.6	0.08	0.57	9	-17.1	8.66
SC15-18	12.38	41,251	45.7	0.08	0.53	9.5	-16.1	7.73
SC15-20	12.4	41,301	46.8	0.09	0.65	9.8	-15.2	8.39
SC15-24	12.44	41,402	50.3	0.10	0.82	10.1	-14.9	9.75
SC15-26	12.46	41,453	39.4	0.09	0.78	10.5	-14.6	10.11
SC15-28	12.48	41,503	48.6	0.10	0.8	9.9	-14.8	9.78
SC15-30	12.5	41,554	49.8	0.10	0.95	11.2	-15.1	10.94
SC15-34	12.54	41,655	46.6	0.09	0.72	11.3	-14.4	9.33
SC15-36	12.56	41,705	47.2	0.11	0.99	10.6	-15.3	10.85
SC15-40	12.6	41,806	46.1	0.08	0.8	11.8	-14.0	11.42
SC15-42	12.62	41,857	43.8	0.10	0.82	9.8	-15.3	9.57
SC15-46	12.66	41,958	45.5	0.07	0.65	7.4	-22.4	10.56
SC15-50	12.7	42,059	47.3	0.07	0.55	8.1	-21.4	9.17
SC15-52	12.72	42,109	49.1	0.08	0.57	7.4	-18.8	8.47
SC15-53	12.73	42,134	51.9	0.07	0.51	7.8	-18.8	8.03
SC15-54	12.74	42,160	57.5	0.08	0.53	7.8	-17.9	7.80
SC15-55	12.75	42,185	49.9	0.08	0.52	7.6	-18.4	8.03
SC15-56	12.76	42,210	58.7	0.07	0.54	7.1	-19.5	8.47
SC15-57	12.77	42,235	57.7	0.07	0.5	7.8	-19.1	8.34
SC15-58	12.78	42,261	40.5	0.07	0.49	8	-18.9	8.17
SC15-59	12.79	42,286	62.3	0.08	0.57	8	-18.1	8.52
SC15-60	12.8	42,311	57	0.08	0.56	7.7	-17.3	8.55
SC15-61	12.81	42,336	53.5	0.08	0.62	7.8	-17.1	8.80
SC15-62	12.82	42,362	50.6	0.08	0.62	7.8	-17.0	8.55

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC15-63	12.83	42,387	52.5	0.08	0.63	7.7	-17.0	8.69
SC15-64	12.84	42,412	51.1	0.09	0.64	7.6	-17.1	8.74
SC15-65	12.85	42,437	51.1	0.08	0.59	8	-16.4	8.37
SC15-66	12.86	42,462	43.6	0.09	0.61	8.1	-16.4	7.91
SC15-67	12.87	42,488	54.3	0.09	0.66	8	-16.5	8.66
SC15-68	12.88	42,513	45.8	0.09	0.62	7.9	-16.3	8.46
SC15-69	12.89	42,538	48.5	0.08	0.63	7.2	-17.5	8.66
SC15-70	12.9	42,563	65.7	0.09	0.65	7.6	-17.1	8.89
SC15-71	12.91	42,589	56.9	0.08	0.62	7.6	-16.9	8.90
SC15-72	12.92	42,614	48.8	0.08	0.61	7.3	-17.0	8.86
SC15-73	12.93	42,639	46.4	0.08	0.61	7.3	-17.0	8.83
SC15-74	12.94	42,664	45.6	0.07	0.49	7.5	-17.5	8.17
SC15-75	12.95	42,690	52.3	0.08	0.65	7	-18.0	9.26
SC15-76	12.96	42,715	46.7	0.08	0.62	6.9	-18.1	9.28
SC15-77	12.97	42,740	51	0.08	0.62	7	-18.2	9.46
SC15-78	12.98	42,765	45.9	0.08	0.66	6.9	-18.3	9.43
SC15-79	12.99	42,791	48.9	0.07	0.6	6.9	-18.3	9.33
SC15-80	13	42,816	48.8	0.09	0.75	7.3	-20.0	10.16
SC15-81	13.01	42,841	48.3	0.07	0.61	6.9	-19.4	9.62
SC15-82	13.02	42,866	41.5	0.07	0.59	6.3	-21.2	9.83
SC15-83	13.03	42,892	65.3	0.08	0.74	6.3	-21.0	10.86
SC15-84	13.04	42,917	65.3	0.07	0.69	6.5	-21.1	10.95
SC15-85	13.05	42,942	63.4	0.08	0.72	6	-21.6	11.03
SC15-86	13.06	42,967	52.7	0.07	0.7	5.9	-21.0	10.89
SC15-87	13.07	42,993	59.4	0.07	0.64	5.9	-19.7	10.30
SC15-88	13.08	43,018	59.6	0.07	0.59	6.1	-18.9	9.24
SC15-89	13.09	43,043	51.9	0.07	0.56	6.5	-18.9	8.80
SC15-90	13.1	43,068		0.08	0.53	6.8	-18.4	
SC15-91	13.11	43,094	54.2	0.08	0.56	7	-18.4	8.42
SC15-93	13.13	43,144	51.9	0.08	0.55	7	-18.5	8.21
SC15-94	13.14	43,169	57.2	0.07	0.53	7.2	-18.5	8.31
SC15-95	13.15	43,195	47.5	0.08	0.56	7	-18.4	8.32
SC16-0	13.2	43,321	45.6	0.08	0.83	6.8	-20.4	11.41
SC16-1	13.21	43,346	45.3	0.08	0.67	6.2	-19.4	9.48
SC16-2	13.22	43,371	49.1	0.10	0.68	3.8	-19.1	7.88
SC16-3	13.23	43,396	59.8	0.08	0.61	7.3	-19.0	8.90

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC16-4	13.24	43,422	56.7	0.08	0.62	7.4	-18.6	8.69
SC16-5	13.25	43,447	53.5	0.10	0.65	5.3	-18.7	7.61
SC16-6	13.26	43,472	35.9	0.08	0.58	7.8	-18.9	8.46
SC16-7	13.27	43,497	52.6	0.08	0.65	7.5	-18.7	8.99
SC16-8	13.28	43,523	46	0.08	0.63	7.3	-18.6	8.63
SC16-9	13.29	43,548	44.9	0.09	0.64	7.1	-18.7	8.61
SC16-10	13.3	43,573	46.9	0.08	0.59	7.8	-18.5	8.38
SC16-11	13.31	43,598	49.5	0.08	0.58	6.8	-19.0	8.31
SC16-12	13.32	43,624	51.6	0.08	0.6	6.8	-19.2	8.88
SC16-13	13.33	43,649	47.2	0.09	0.75	6.2	-19.7	10.24
SC16-14	13.34	43,674	45.3	0.08	0.61	7.2	-19.1	8.90
SC16-15	13.35	43,699	47.3	0.08	0.73	6.6	-19.7	10.52
SC16-16	13.36	43,725	54.1	0.07	0.6	7	-19.8	10.60
SC16-17	13.37	43,750	44.9	0.08	0.77	7.1	-19.8	10.80
SC16-18	13.38	43,775	53.2	0.08	0.7	6.9	-19.9	10.68
SC16-19	13.39	43,800	45.6	0.08	0.71	6.7	-20.1	10.61
SC16-20	13.4	43,826	61.1	0.07	0.66	6.8	-20.1	10.93
SC16-21	13.41	43,851	53.1	0.06	0.58	6.6	-19.9	10.45
SC16-22	13.42	43,876	50.1	0.05	0.45	6.9	-19.7	10.50
SC16-23	13.43	43,901	50	0.07	0.56	5.6	-18.6	9.66
SC16-24	13.44	43,927	52.5	0.07	0.64	5.1	-18.8	10.93
SC16-25	13.45	43,952	59.9	0.07	0.78	5.1	-19.1	12.72
SC16-26	13.46	43,977	49.1	0.08	0.99	5	-19.0	13.69
SC16-27	13.47	44,002	45.9	0.08	0.89	4.7	-19.2	13.20
SC16-28	13.48	44,028	46.8	0.07	0.73	4.6	-19.6	12.87
SC16-29	13.49	44,053	51.5	0.08	0.93	4.9	-19.7	13.32
SC16-30	13.5	44,078	43.6	0.08	0.85	5.8	-19.6	12.40
SC16-31	13.51	44,103	46.9	0.07	0.85	5.4	-19.4	13.56
SC16-32	13.52	44,129	61.4	0.07	0.85	5.8	-19.1	14.04
SC16-33	13.53	44,154		0.07	0.85	5.6	-19.0	
SC16-34	13.54	44,179	48	0.07	0.81	5.5	-18.5	13.97
SC16-35	13.55	44,204	59.4	0.09	0.85	4.5	-18.2	11.52
SC16-36	13.56	44,230	46.6	0.08	0.63	4.1	-18.4	9.61
SC16-38	13.58	44,280	42.4	0.07	0.66	6.1	-18.6	11.00
SC16-39	13.59	44,305	49.1	0.07	0.69	6.2	-18.1	11.53
SC16-40	13.6	44,330	47.1	0.08	0.78	5.7	-17.9	11.90

Sample ID	Depth (m)	Age (yrs)	Sample wt. (mg)	Total N (%)	Total C (%)	$\delta^{15}N$	$\delta^{13}C$	C/N
SC16-41	13.61	44,356	51.6	0.07	0.76	6	-18.2	11.82
SC16-42	13.62	44,381	51.7	0.07	0.7	5.9	-18.3	11.91
SC16-43	13.63	44,406	44.8	0.06	0.67	6.3	-18.3	12.21
SC16-44	13.64	44,431	49	0.06	0.56	6.1	-18.2	11.40
SC16-45	13.65	44,457	56.4	0.06	0.63	6.1	-18.2	11.65
SC16-46	13.66	44,482	41.1	0.08	0.77	9.5	-18.8	11.23
SC16-47	13.67	44,507	46.8	0.08	0.79	5.9	-18.5	11.61
SC16-48	13.68	44,532	47.2	0.07	0.89	7.4	-18.7	14.51
SC16-49	13.69	44,558	50.5	0.06	0.61	7.3	-19.3	12.36
SC16-50	13.7	44,583	46.9	0.05	0.43	6.8	-20.0	10.43
SC16-51	13.71	44,608	50.4	0.06	0.59	8.3	-20.3	11.08
SC16-52	13.72	44,633	51.4	0.05	0.53	6.9	-19.6	12.08
SC16-53	13.73	44,659	45.9	0.04	0.4	6.9	-19.3	10.76
SC16-54	13.74	44,684	37.2	0.05	0.44	8.2	-18.6	10.27
SC16-55	13.75	44,709	48.6	0.05	0.74	6.9	-18.8	16.05
SC16-56	13.76	44,734	49.7	0.06	0.71	6.9	-19.1	15.00
SC16-57	13.77	44,760	46	0.06	0.93	6.4	-18.9	17.39
SC16-58	13.78	44,785	44.5	0.05	0.66	6.4	-19.2	14.40
SC16-59	13.79	44,810	49.7	0.05	0.59	6.4	-19.2	14.12
SC16-60	13.8	44,835	47.6	0.05	0.6	6.1	-19.1	13.62
SC16-61	13.81	44,861	58.6	0.03	0.34	5.8	-19.9	11.87
SC16-62	13.82	44,886	55.7	0.04	0.38	6.4	-19.9	11.08
SC16-63	13.83	44,911	50.6	0.04	0.41	5.5	-20.4	10.59
SC16-64	13.84	44,936	54.4	0.04	0.38	5.6	-20.5	10.58

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Biography

James Andrew Nunnery was born on October 12, 1978 in Murfreesboro Tennessee. In the fall of 1997 he entered Guilford College in Greensboro, North Carolina where he majored in geology. He graduated with a B.S. in geology in 2001. In 2002 he moved to San Francisco and worked as a hydrogeologist for MWH, an engineering and environmental consulting firm, and was involved with groundwater monitoring and remediation on several Air Force bases located in southern California. In 2005 he returned to Greensboro and worked as an environmental geologist for the environmental consulting group Trigon Engineering Consultants. In the fall of 2007 he enrolled in the Division of Earth and Ocean Sciences at Duke University where he earned an M.S. degree in 2009 studying mid-ocean ridge morphology. From 2009 to 2012 he remained at Duke shifting his focus to paleoclimate of the tropical Andes.