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Holocene paleohydrology and glacial history of the central Andes using multiproxy lake sediment studies

Mark B. Abbott [a,*], Brent B. Wolfe [b], Alexander P. Wolfe [c], Geoffrey O. Seltzer [d], Ramon Aravena [b], Brian G. Mark [d], Pratigya J. Polissar [e], Donald T. Rodbell [f], Harry D. Rowe [g], Mathias Vuille [e]

[a] Department of Geology and Planetary Science, University of Pittsburgh, 4107 O'Hara Street, Room 200 SRCC, Pittsburgh, PA 15260, USA

[b] Department of Earth Sciences, University of Waterloo, Waterloo, ON, Canada N2L 3G1

[c] Department of Earth and Atmospheric Sciences, University of Alberta, 1-13 ESB, Edmonton, AB, Canada T6G 2E3

[d] Department of Earth Science, Syracuse University, 204 Heroy Geology Laboratory, Syracuse, NY 13244-1070, USA

[e] Department of Geosciences, University of Massachusetts, Morrill Science Center, Amherst, MA 01003, USA

[f] Department of Geology, Union College, Butterfield Hall, Schenectady, NY 12308, USA

[g] Earth and Environmental Sciences, University of Illinois at Chicago, 845 West Taylor St., Chicago, IL 60607-7059, USA

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Abstract

Here we document at century to millennial scale the regional changes of precipitation–evaporation from the late Pleistocene to present with multiproxy methods on a north–south transect of lake sites across the eastern cordillera of the central Andes. The transect of study sites covers the area from $\sim$14°S to 20°S and includes core studies from seven lakes and modern calibration water samples from twenty-three watersheds analyzed to constrain the down-core interpretations of stable isotopes and diatoms. We selected lakes in different hydrologic settings spanning a range of sensitivity to changes in the moisture balance. These include: (1) lakes directly receiving glacial meltwater, (2) overflowing lakes in glaciated watersheds, (3) overflowing lakes in watersheds without active glaciers, and (4) lakes that become closed basins during the dry season. The results of our current work on multiple lakes in the Bolivian Andes show that while the overall pattern of Holocene environmental change is consistent within the region, conditions were not always stable over centennial to over millennial timescales and considerable decadal- to century-scale climate variability is evident [Abbott et al., Quat. Res. 47 (1997) 70–80, Quat. Res. 47 (1997) 169–180, Quat. Sci. Rev. 19 (2000) 1801–1820; Polissar, Master's thesis, University of Massachusetts (1999)]. Comparison of the paleoclimate record from one well-studied site, Lago Taypi Chaka Kkota (LTCK), with others within the region illustrates a consistent overall pattern of aridity from the late glacial through the middle Holocene. Previous work noted a difference between the timing of water-level rise in Lake Titicaca $\sim$5.0–3.5 ka B.P. [Abbott et al., Quat. Res. 47 (1997) 169–180; Cross et al., Holocene 10 (2000) 21–32; Rowe et al., Clim. Change 52 (2002) 175–199] and the onset of wetter conditions at 2.3 ka B.P. in LTCK, a lake that drains into the southern end of Lake Titicaca [Abbott et al., Quat. Res. 47 (1997) 70–80]. Sedimentary and oxygen isotope evidence from Paco Cocha (13°54′S) located in the northern reaches of the expansive 57 000 km^2 Titicaca watershed, which spans $\sim$14°S to 17°S, indicates that

* Corresponding author. Tel.: +1-412-624-1408; Fax: +1-412-624-3914.
E-mail address: mabbott1@pitt.edu (M.B. Abbott).

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glaciers returned to the watershed around 4.8 ka B.P. In addition, sedimentary and geochemical data from Llacho Kkota (15°07′S), located between LTCK (16°12′S) and Paco Cocha, indicate wetter conditions around 3.4 ka B.P. This suggests wetter conditions occurred first in the northern reaches of the Titicaca watershed and resulted in rising water levels in Lake Titicaca while the LTCK watershed remained unglaciated and seasonally desiccated. Comparison of the paleoclimate records from Paco Cocha, LTCK, and Potosi with other paleoclimate records from the region including Lake Titicaca and Nevado Sajama illustrates a consistent overall pattern of aridity from the late Pleistocene through the middle Holocene, but wetter conditions occurred in the northern areas first, and the aridity in the north was of shorter duration and less severe. The progressive increase in wet season (summer) insolation across the region during the Holocene likely resulted in an increasingly southward position and intensification of the zone of intense summer convection – known as the Intertropical Convergence Zone over the tropical Atlantic and Pacific. We suggest this is the cause of the overall pattern of increasingly later dates for the onset of wetter conditions moving north to south that we observe across the seven lakes discussed here that span four degrees of latitude. However, this does not explain the decadal- to century-scale variability that demonstrably exists at these same sites. Overall the last 2.3 ka has been the wettest period during the Holocene, but even during this relatively wet phase there are century-scale lowstands in lakes, including Titicaca, Llacho Kkota, Juntutuyo (17°33′S), and Potosi (19°38′S), and significant changes in the extent of glacial activity in the glaciated watersheds of Paco Cocha, LTCK, and Viscachani (16°11′S), indicating the continued sensitivity of the region to shifts in the moisture balance.

Keywords: paleolimnology; paleoclimatology; stable isotopes; South America; Bolivia; Peru

1. Introduction

The overall goal of our research in the central Andes is to document the regional pattern of climatic change from the late Pleistocene to present with multiproxy analyses of sediment cores collected from a network of seven alpine watersheds spanning ~14°S to 20°S (Fig. 1). The emphasis of this research is to reconstruct the precipitation–evaporation (P–E) balance of this semi-arid region. The study area is located at the southern edge of the annual migration of the zone of intense summer convection, which is connected with the Intertropical Convergence Zone (ITCZ) over the tropical Atlantic and Pacific (Fig. 2). This region of the Andes is particularly sensitive to P–E changes because of a single wet season, whereas closer to the equator there are two rainy periods during the spring and fall. The aim of this work is to understand the moisture-balance history of the tropics, in part because of the implications that changes in water resources have on both past and present human populations (agriculture, drinking water, fisheries, and hydropower). The glacial history of the central Andes is an important component of this work because today meltwater from small alpine glaciers is an important source of water that buffers lake and groundwater levels during the winter dry season. This water source is not available during periods when glaciers are absent, such as the early and middle Holocene, and possibly the near future.

The results of our current work using multiple lakes in the eastern cordillera of the central Andes show that while the overall pattern of Holocene environmental change is consistent across the region marked differences exist in the timing of events from north to south. The outstanding question raised by previously published results is how can the Lago Taypi Chaka Kkota (LTCK) watershed remained unglaciated and desiccated on a seasonal basis until 2.3 ka B.P., while Lake Titicaca rose ~100 m, flooding the small, shallow southern basin (Lago Wiñaymarka) by 3.5 ka B.P.? The answer to this question should yield information regarding how the regional climate system works.

In addition to the seven lakes where long cores were studied we collected modern samples and recorded limnological measurements from twenty-two other lakes along the same transect from 14°S to 20°S to help constrain down-core interpretations. Lakes were selected that are representative of a range of hydrologic settings that

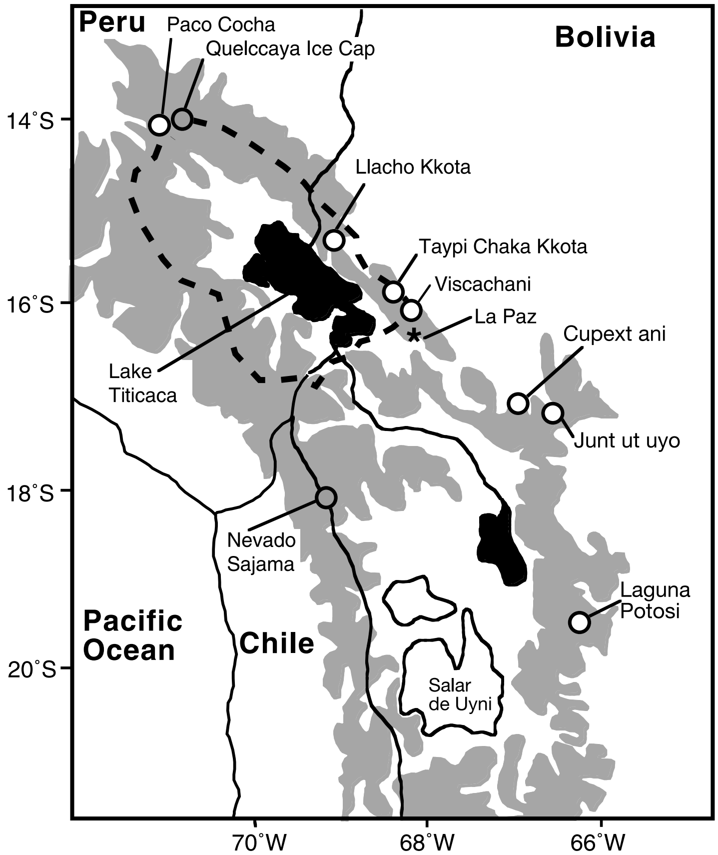

Fig. 1. Map showing the location of lakes on the north–south transect across the eastern cordillera of the central Andes spanning 14°S to 20°S. Cores were collected from Lagunas Paco Cocha, Llacho Kkota, LTCK, Viscachani, Cupextani, Juntutuyo, and Potosi. The dashed line surrounding Lake Titicaca outlines the 57 000 km^2 watershed.

have different sensitivities to changes in the moisture balance. These lake systems include: (1) lakes directly receiving glacial meltwater, (2) overflowing lakes in glaciated watersheds, (3) overflowing lakes in catchments without active glaciers, and (4) lakes that become closed basins during the dry season. Core transects from shallow to deep water were used to identify transgression and regression sequences and collect sediments to document and date these features. Cores with continuous sedimentary records from the central basin were analyzed with an integrated methodology that combined (1) sedimentology and geochronology, (2) geochemistry and stable isotopes, (3) diatoms, (4) sampling of surficial sediments, lake water, and limnological measurements to constrain down-core interpretations, and (5) development of hydrologic and isotopic mass-balance models to quantify past changes in P–E (Wolfe et al., 2001a). Cores were dated by accelerator mass spectrometry (AMS) ^{14}C measurements on macrofossils. Recent sediment accumulation rates were constrained by ^{137}Cs and ^{210}Pb measurements.

All radiocarbon ages used in this paper have been calibrated by using INTCL98 (Stuiver et al., 1998) and are presented in calendar years before present (ka B.P.).

2. Regional climate

The Altiplano, with an average elevation of 3700–4400 m (650–600 hPa), effectively separates

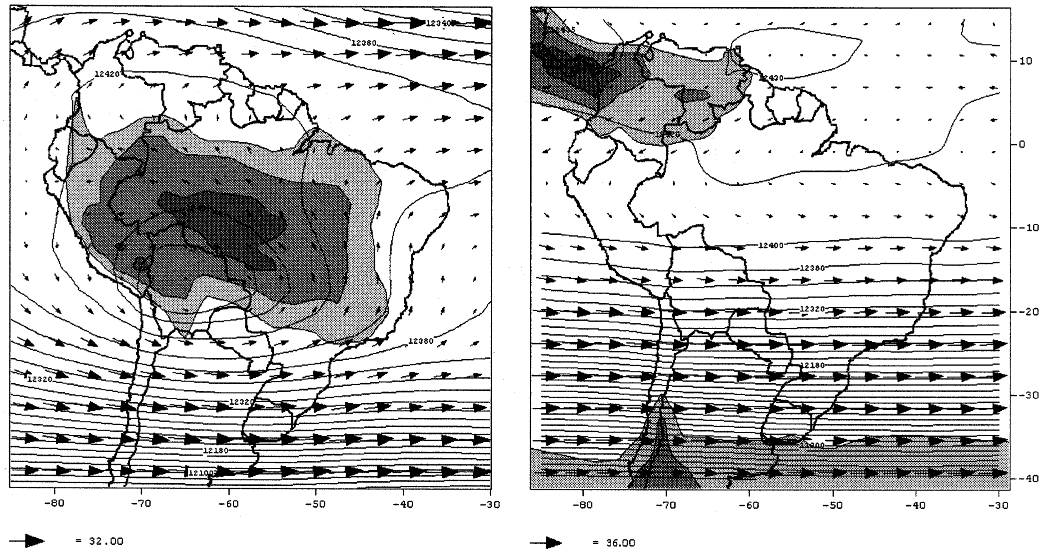

Fig. 2. Long-term mean 200 hPa wind and geopotential height (1968–96) and outgoing long-wave radiation (OLR, 1974–1998) based on NCEP-NCAR reanalysis and NOAA-OLR data for (left panel) austral summer (December, January, February) and (right panel) austral winter (June, July, August). Contour interval is 20 gpm, 10 gpm above 12 420 and 5 gpm above 12 430. Light, medium, and dark shades indicate OLR values less than 225, 210, and 195 W m^{-2}.

the low-level flow to the east and the west. While the Southeast Pacific Anticyclone in combination with cold sea surface temperatures produces dry and stable conditions to the west, with moist air trapped below the subsidence inversion at about 900 hPa, to the east a thermal heat low develops during summer months, and the lower troposphere there is characterized by warm and humid conditions. The Altiplano separating these two opposite climatic conditions is accordingly characterized by a very strong climatic gradient, with average annual precipitation below 200 mm yr^{-1} in the southwestern part and more than 800 mm yr^{-1} in the northeast. 50 to 80% of the annual rainfall amount falls during the austral summer months (December, January, February), associated with the development of convective cloudiness over the central Andes and the southwestern part of the Amazon (Fig. 2, left panel). During the austral winter (June, July, August), dry conditions prevailing over the Altiplano are associated with a strong zonal westerly flow, while the center of convective activity is displaced towards the northwest of the continent (Fig. 2, right panel). During the summer months, however, intense solar heating of the Altiplano surface leads to a destabilization of the boundary layer, inducing deep convection and moist air advection from the eastern interior of the continent (Vuille et al., 1998; Garreaud, 1999). In the upper troposphere an anticyclonic vortex, the Bolivian High, develops during the summer months (Fig. 2, left panel). The positioning and strength of this high is intrinsically linked to precipitation anomalies over the Altiplano, featuring an intensification and southward displacement during wet episodes, while a weakening and northward displacement can be observed during dry periods (Aceituno and Montecinos, 1993; Vuille et al., 1998; Lenters and Cook, 1999; Vuille, 1999; Garreaud et al., this issue). Over millennial timescales changes in insolation should influence the location and strength of the Bolivian High and strongly affect the precipitation regime of the region.

3. Methods

3.1. Sedimentology

Whole-core magnetic susceptibility was measured at 1-cm increments, after which cores were opened, photographed, and sampled following our routine laboratory protocols. Once cores were split, half was archived in ODP D-Tubes and stored in a dark refrigerated (1–3°C) location. The work half was sampled for bulk density, carbon content, biogenic silica, elemental and stable-isotope geochemistry, and diatoms. Samples were sieved to locate macrofossils for AMS ^{14}C measurements. Core lithology was determined from smear-slide mineralogy and detailed inspection of sediments for Munsell color, texture, sedimentary structures, and biogenic features. Total and organic carbon were measured by loss on ignition (LOI; 500°C) or with a Leco CS-300 Carbon–Sulfur Analyzer. Biogenic silica analyses was performed by time-series dissolution experiments (DeMaster, 1981). Magnetic susceptibility was measured with a Bartington Susceptibility Bridge, corrected for mass differences with bulk density measurements.

3.2. Geochronology

We used an integrated approach that included ^{137}Cs and ^{210}Pb assays to determine an age model for the last 150 years and AMS ^{14}C measurements on identifiable macrofossils thereafter. Sampling protocols for radiocarbon measurements by AMS are continuously being refined (Abbott and Stafford, 1996; Eglinton et al., 1997; Turney et al., 2000; Rowe et al., in press). We targeted terrestrial material that can be identified for AMS ^{14}C measurements and did not have a problem finding sufficient macrofossils. In fact, there is not a single age reversal in these cores that are currently dated at 500–1000 year resolution. Hard-water reservoir effects were assessed at several sites by: (1) dating living plants that should be $\sim$112% Modern, (2) measuring the ^{14}C activity of aquatic-plant matter from the A.D. 1850 stratum as identified by ^{210}Pb assay, and (3) measuring aquatic and terrestrial macrofossils from the same stratigraphic level to determine if there was a reservoir age in the past. No evidence for a reservoir effect exists from the transect of alpine lakes. Radiocarbon plateaus are an impediment to the development of high-resolution chronologies, although these effects are minimal during the last 4 ka and radiocarbon ages were calibrated using INTCAL98 (Stuiver et al., 1998). Sediment–water interface cores were collected and extruded vertically in the field for ^{137}Cs and ^{210}Pb dating according to our standard field protocols.

3.3. Elemental and stable-isotope geochemistry

Sediment cores were sub-sampled for organic elemental and stable-isotope analysis following Wolfe et al. (2001b). The pretreatments included removal of carbonate with 1*N* HCl at 60°C, followed by rinsing with de-ionized water, freeze-drying, and sieving to remove coarse debris. Carbon and nitrogen content and isotopic composition of this residue was determined by an elemental analyzer interfaced with a continuous-flow isotope-ratio mass spectrometry (CF-IRMS). Evaluation of elemental carbon/nitrogen ratios was used to identify the origin of the organic matter (Meyers and Lallier-Vergès, 1999). Samples characterized by dominantly aquatic organic matter were subjected to further treatment involving solvent extraction, bleaching, and alkaline hydrolysis to remove non-cellulose organic constituents, followed by hydroxylamine leaching of iron and manganese oxyhydroxides was measured by dual-inlet mass spectrometry on CO_2 derived by nickel-tube pyrolysis (Wolfe et al., 2001b). Cellulose carbon and oxygen isotope composition were measured by CF-IRMS. Sediment sample preparation and analysis were performed at the University of Waterloo (LTCK and Potosi) and at the University of Minnesota (Paco Cocha). Stable-isotope measurements on sediment cellulose discussed in this paper are reported as $\delta^{18}O_{cell}$. Lake water $\delta^{18}O$ ($\delta^{18}O_{lw}$) is inferred from $\delta^{18}O_{cell}$ values using a cellulose–water oxygen isotope fractionation factor of 1.028 ± 0.001 (Epstein et al., 1977; DeNiro and Epstein, 1981).

4. Results and discussion

4.1. Modern surface-water samples

Oxygen and hydrogen isotope ratios were measured for water samples from twenty-three lakes

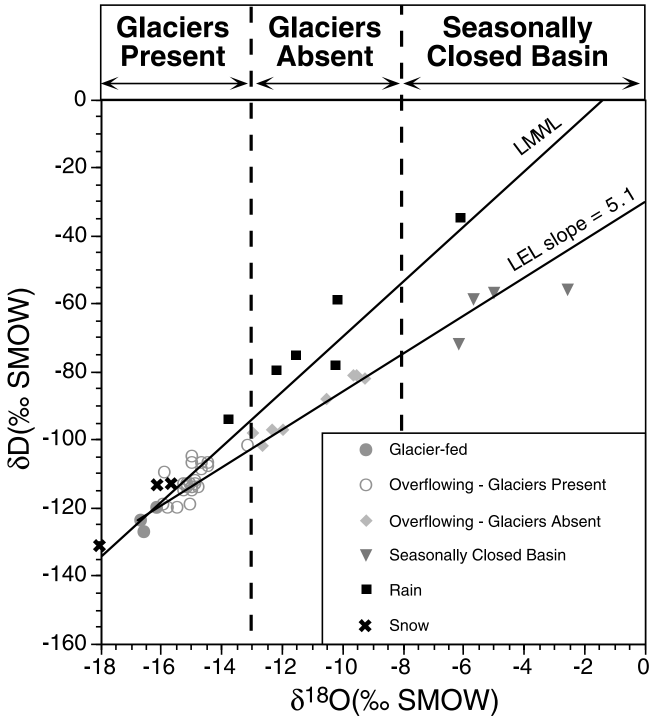

Fig. 3. Isotope composition of lake waters from the central Andes sampled in June 1997 (Abbott et al., 2000). Also shown are the Global Meteoric Water Line (LMWL: $\delta D = 8\delta^{18}O+10$) and the Regional Evaporation Line (LEL: $\delta D = 5.1\delta^{18}O-35.5$). Progressively increasing offset from the LMWL is found for lakes directly receiving glacial meltwater, lakes in glaciated watersheds, overflowing lakes in non-glaciated watersheds, and seasonally closed lakes. Oxygen isotope values spanning more than 10‰ are indicative of the strong variation in local hydrological settings in this region.

collected during the dry winter season along a north–south transect spanning the Cordillera Vilcanota (14°S), Cordillera Real (16°S), Cordillera Tunari (17°S), and Cordillera Chichas (19°S) (Fig. 1). These lakes are representative of different hydrologic settings, including: (1) lakes directly receiving glacial meltwater, (2) overflowing lakes in glaciated watersheds, (3) overflowing lakes in watersheds without active glaciers, and (4) lakes that are seasonally closed basins (Abbott et al., 2000). Results show a wide range of isotopic compositions with $\delta^{18}O$ ranging between −16.6‰ and −2.5‰ and δD between −127‰ and −56‰ (Fig. 3). The dashed vertical lines in Fig. 3 are used to divide the modern surface-water data into three categories to help interpret the down-core data shown in Figs. 4–6: (1) $\delta^{18}O$ values less than −13‰ occur today in watersheds where glacial ice is present, (2) $\delta^{18}O$ values between −8‰ and −13‰ occur today in watersheds where lakes are overflowing but with glacial ice absent, and (3) $\delta^{18}O$ values greater than −8‰ occur today in watersheds that are seasonally closed basins.

4.2. Lago Paco Cocha

Multiproxy analyses of Laguna Paco Cocha (13°54′S, 71°52′W) include studies of sedimentology, geochemistry, physical properties, magnetic susceptibility, and stable isotopes. The age model for the core was produced by linear interpolation between 11 calibrated AMS radiocarbon dates on individual macrofossils (Table 1) (Stuiver et al., 1998). Fig. 4 shows the results from the analysis of organic matter content (LOI at 500°C), bulk density, magnetic susceptibility, cellulose-inferred $\delta^{18}O_{lw}$, and mass accumulation rates of organic and mineral matter. The abrupt shift to glacial values at 4.8 ka B.P. for all parameters is high-

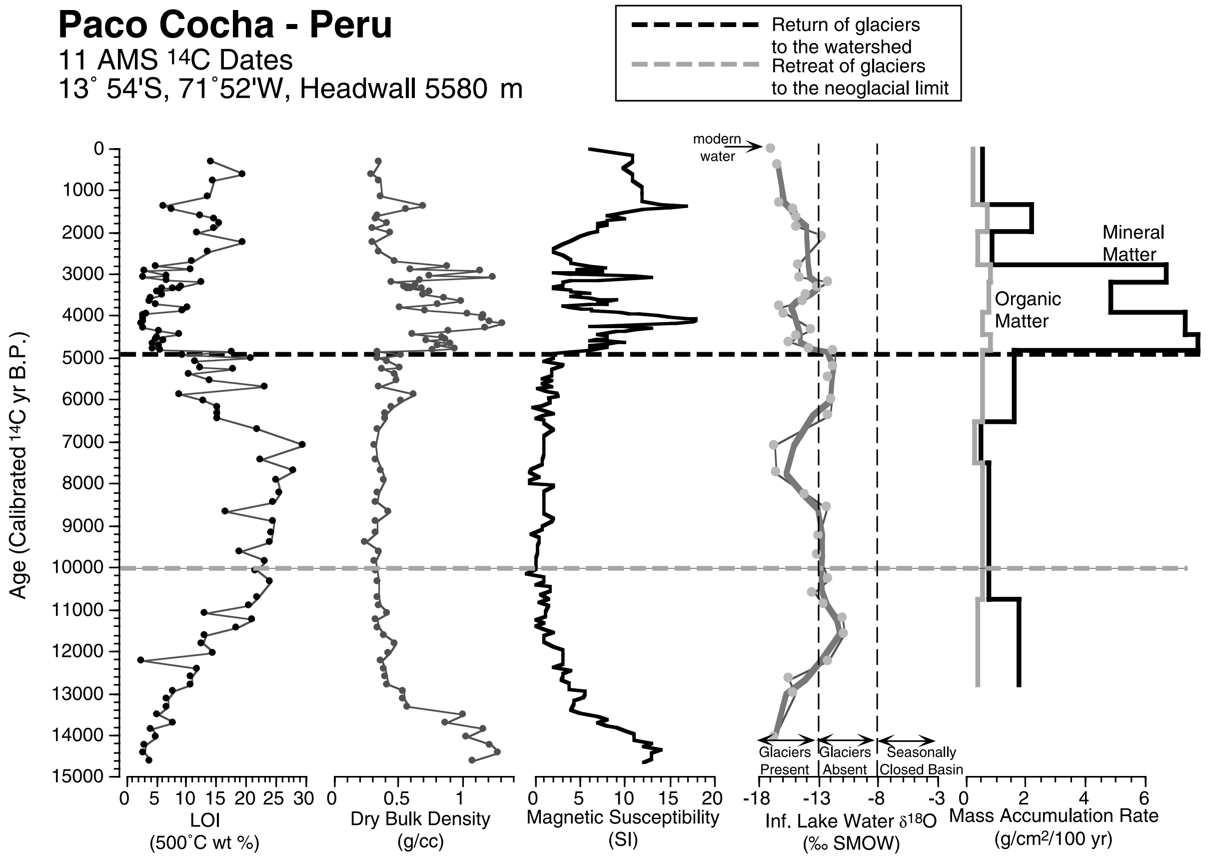

Fig. 4. Sediment-core analyses from Laguna Paco Cocha including organic matter, dry bulk density, magnetic susceptibility, cellulose-inferred $\delta^{18}O_{lw}$ (thick line is the three-point running average), and mass accumulation rates of organic and mineral matter. The dashed horizontal gray line indicates the retreat of glaciers to at least the neoglacial limit by 10.0 ka B.P., and the horizontal dashed black line shows the return of glaciers to the watershed at 4.8 ka B.P. The vertical dashed lines on the graph of inferred $\delta^{18}O_{lw}$ provide a framework to interpret the status of the watershed based on the modern calibration samples shown in Fig. 3. The cellulose-inferred $\delta^{18}O_{lw}$ values were corrected for a systematic offset arising from methodological differences between the University of Waterloo and University of Minnesota laboratories (Beuning et al., 2002).

lighted by a dashed black line. This includes decreased organic matter values from >15 to <5 wt%, higher dry bulk density from <0.5 g/cc to >0.9 g/cc, higher magnetic susceptibility from <5 to >10 SI, and an abrupt $\delta^{18}O_{lw}$ decrease of 3‰. We interpret the results of the analyses presented in Fig. 4 to show that glaciers in the Paco Cocha watershed retreated rapidly beginning prior to ~12.7 ka B.P. and were gone from the watershed by 10.0 ka B.P., as indicated by the dashed gray line (Mark et al., 1999). Although glaciers were probably absent from the watershed between 10.0 and 4.8 ka B.P., the lake remained at the overflowing stage during this period, as suggested by analyses of organic matter, sediment density, and magnetic susceptibility. If the lake had desiccated during this period we would expect oxidation of organic matter leading to low values which we do not see. Additionally, cellulose-inferred $\delta^{18}O_{lw}$ values remained <−10‰ suggesting that the lake did not become a closed basin during this period. Increased mineral and decreased organic matter accumulation rates after 4.8 ka B.P. also support the return of glacial ice to the watershed at this time and the lack of ice in the drainage basin during the early and middle Holocene. After glaciers returned to the watershed at 4.8 ka B.P. they have been present until

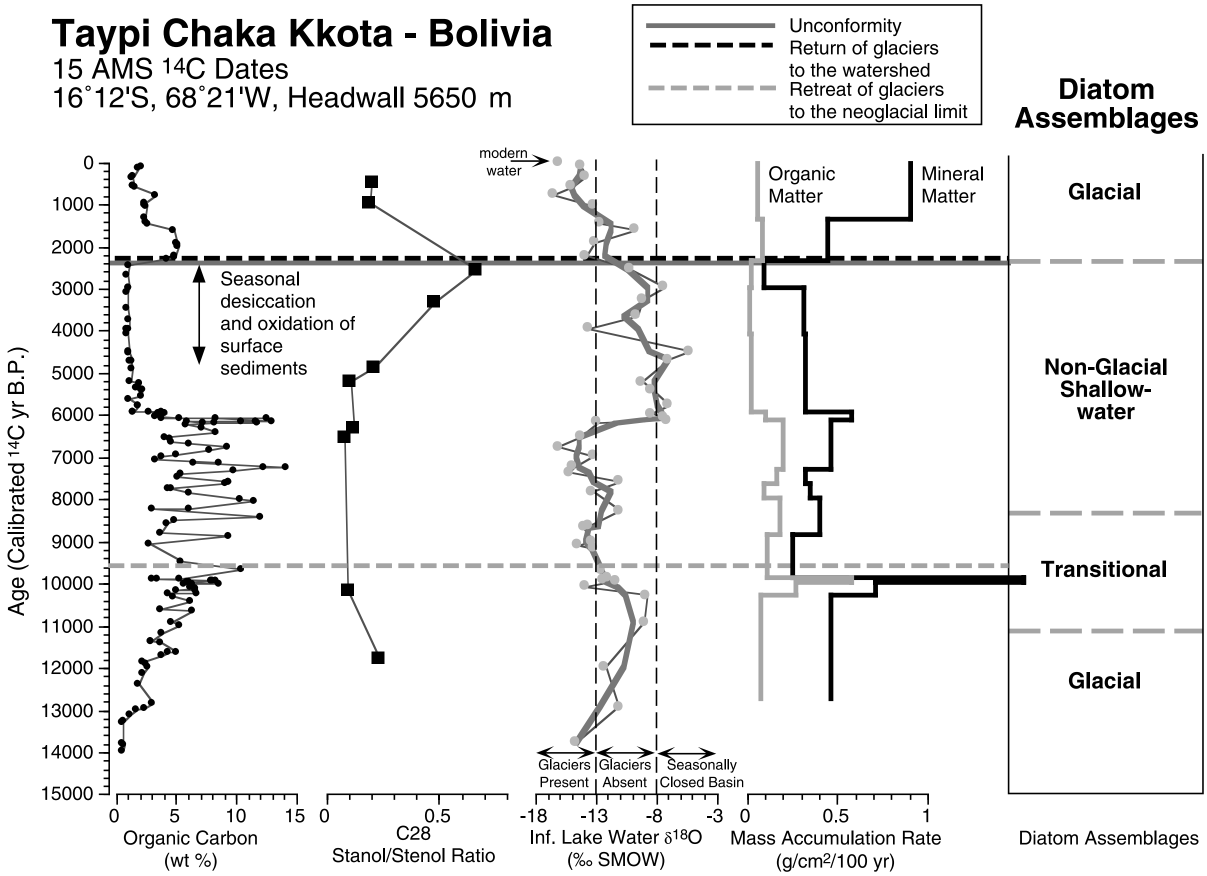

Fig. 5. Sediment-core analyses from LTCK including organic carbon, organic geochemistry, cellulose-inferred $\delta^{18}O_{lw}$ (thick line is the three-point running average), mass accumulation rates of organic and mineral matter, and interpretation of diatom assemblages. The dashed gray line indicates the retreat of glaciers to at least the neoglacial limit by 9.6 ka B.P., and the dashed black line shows the return of glaciers to the watershed at 2.3 ka B.P. The solid gray line in denotes an unconformity ending with the return of ice to the watershed at 2.3 ka B.P. The vertical dashed lines on the graph of inferred $\delta^{18}O_{lw}$ provide a framework to interpret the status of the watershed based on the modern calibration samples shown in Fig. 3.

today, but analyses of organic matter, sediment density, and magnetic susceptibility in addition to changing accumulation rates suggest significant fluctuations during this period. Cores from Laguna Llacho Kkota, which is located to the south (15°07′S, 69°08′W), show the onset of wetter conditions at 3.4 ka B.P.

4.3. *Lago Taypi Chaka Kkota*

Multiproxy analyses of LTCK (16°12′S, 68°21′W) include studies of sedimentology, geochemistry, magnetic susceptibility, stable isotopes, lipid biomarkers, and diatoms. The age model for the core was produced by linear interpolation between fifteen calibrated AMS radiocarbon dates on individual macrofossils (Abbott et al., 1997a). The results of these analyses shown in Fig. 5 indicate that glacial ice in the LTCK watershed retreated rapidly beginning prior to ∼10.0 ka B.P. and was gone from the watershed by 9.6 ka B.P., as indicated by the dashed gray line (Abbott et al., 1997a, 2000). The lake, which is overflowing and glacial-fed today, was shallow and seasonally closed for prolonged periods between ∼8.5 and 2.3 ka B.P. Supporting evidence for seasonal desiccation during this period comes from low organic carbon values due to oxidation and the stanol/

stenol ratio shown in Fig. 5, which indicates a marked increase between ~5 and 2.3 ka B.P. (Polissar, 1999). Additional evidence is provided by the presence of an unconformity formed just prior to 2.3 ka B.P., as indicated by the solid black line. Glaciers returned to the watershed after 2.3 ka B.P., as indicated by the dashed black line, and have been present until today. Supporting evidence for the return of glacial ice is supplied by higher organic carbon values, a lower stanol/stenol ratio, decreased cellulose-inferred $\delta^{18}O_{lw}$, and a sharp increase in the mineral matter accumulation rate. Additional evidence comes from large shifts in diatoms from non-glacial, shallow water to glacial assemblages across the boundary at 2.3 ka B.P. (Abbott et al., 2000). Cores from Laguna Viscachani (16°11′S, 68°07′W), which is located < 30 km to the southeast, show the same pattern and timing of changes as is seen in LTCK (Abbott et al., 1997a).

4.4. Laguna Potosi

Multiproxy analyses of Laguna Potosi (19°38′S, 65°41′W) include studies of sedimentology, geochemistry, magnetic susceptibility, and stable isotopes (Wolfe et al., 2001a,b). The age model for the core was produced by linear interpolation between nine calibrated AMS radiocarbon dates on individual macrofossils (Wolfe et al., 2001a,b) and the base of the ^{210}Pb chronology. Fig. 6 shows the results from the analysis of organic matter content (LOI), bulk density, magnetic susceptibility, $\delta^{18}O$ of fine-grained cellulose, and mass accumulation rates of organic and mineral matter. The results of these analyses shown in Fig. 6 indicate that glacial ice in the Potosi watershed retreated rapidly beginning prior to ~10.0 ka B.P. as indicated by the dashed gray line, and was gone from the watershed by 9.0 ka B.P. The lake, which is a seasonally closed basin today, has likely varied between overflowing and seasonally closed for prolonged periods between ~8.5 ka B.P. and present.

4.5. Lake Titicaca

Over the past decade, numerous studies on Lake Titicaca sought to constrain lake-level changes during the latest Quaternary (Wirrmann and Fernando De Oliveira Almeida, 1987; Wirrmann and Mourguiart, 1995; Abbott et al., 1997b; Binford et al., 1997; Mourguiart et al., 1998; Cross et al., 2000, 2001; Baker et al., 2001; Rowe et al., 2002). Studies using sedimentology, diatoms, and stable isotopes all suggest that the lake was ~100 m lower during the middle Holocene before rising sometime between ~5 and 3.5 ka B.P. Sediment cores were collected from the southern basin of Lake Titicaca on a transect from shallow to deep water to study lake-level changes during the late Holocene. Detailed sedimentary analyses of sub-facies and age control with 60+ radiocarbon measurements from four representative cores documented four periods of prolonged drought during the last 3.5 ka (Abbott et al., 1997b). After a prolonged middle-Holocene lowstand water levels rose between 3.5 and 3.3 ka B.P. Four pronounced century-scale droughts resulting in low lake stands occurred from 3.2 to 2.8 ka B.P., 2.4 to 2.2 ka B.P., 1.7 to 1.5 ka B.P., and 0.9 to 0.6 ka B.P. Several of the low lake levels coincided with cultural changes in the region, including the collapse of the Tiwanaku civilization (Binford et al., 1997).

Rowe et al. (2002) showed that the down-core $\delta^{13}C$ of bulk organic matter in Lake Titicaca sediments is an indicator of paleoshoreline proximity and used this dataset to reconstruct water levels during the late Pleistocene and Holocene when core tansects were impractical due to deep-water conditions. Heavier $\delta^{13}C$ values reflect deposition of near-shore aquatic macrophyte detritus, suggesting the core site was near to the shoreline during deposition. Lighter $\delta^{13}C$ values reflect algal-dominated deposition, suggesting a greater distance between core site and shoreline. Quantitative estimates of lake level were developed using transfer functions based on the $\delta^{13}C$ of modern lacustrine organic endmembers (algae, macrophytes) and mean lake-level values for the modern system (3810 m asl), the middle-Holocene lowstand (3725 m asl), and the late-Pleistocene highstand (3815 m asl). The results are consistent with other methods used to reconstruct water levels in Lake Titicaca but offer a nearly continuous se-

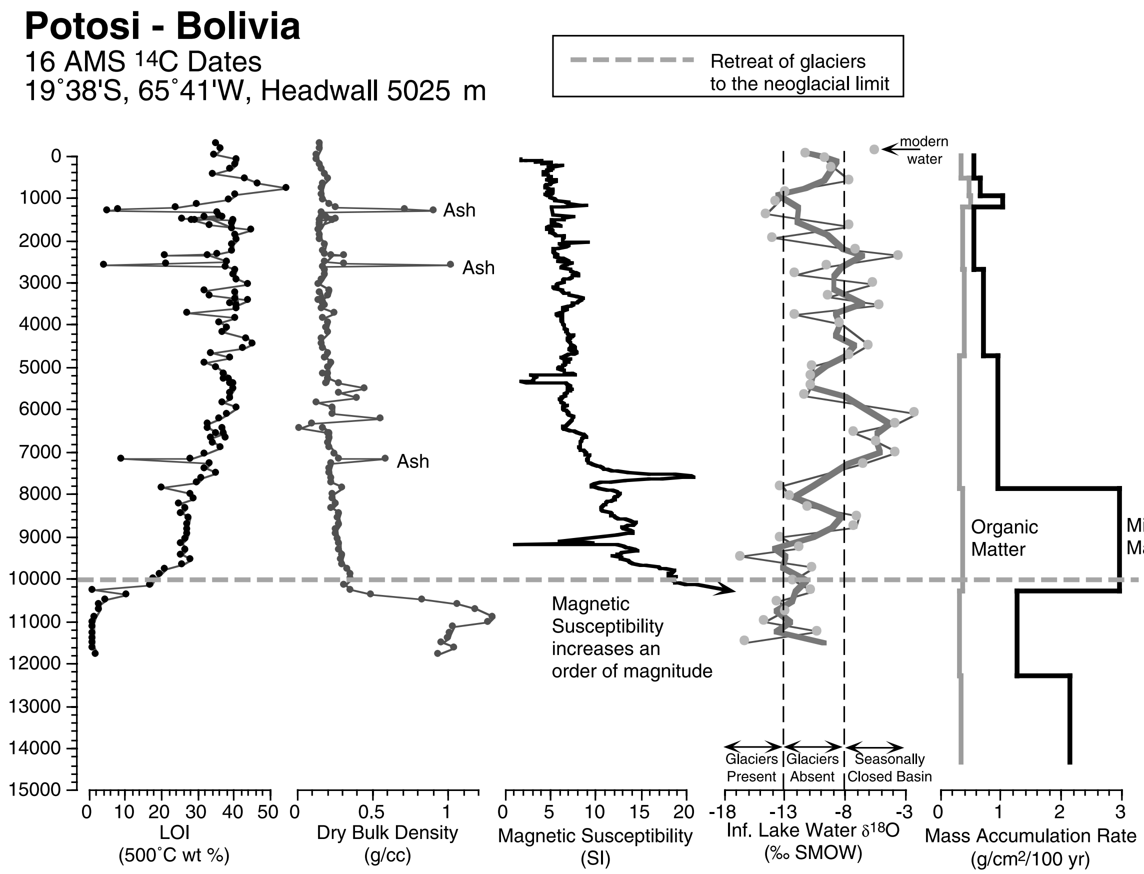

Fig. 6. Sediment-core analyses from Laguna Potosi, including organic matter, dry bulk density, magnetic susceptibility, cellulose-inferred $\delta^{18}O_{lw}$ (thick line is the three-point running average), and mass accumulation rates of organic and mineral matter. The dashed gray line indicates the loss of glaciers from the watershed at 10.0 ka B.P. The vertical dashed lines on the graph of inferred $\delta^{18}O_{lw}$ provide a framework to interpret the status of the watershed based on the modern calibration samples shown in Fig. 3.

quence unlike other methods (Wirrmann and Fernando De Oliveira Almeida, 1987; Wirrmann and Mourguiart, 1995; Abbott et al., 1997b; Mourguiart et al., 1998; Cross et al., 2000, 2001; Baker et al., 2001; Rowe et al., 2002).

4.6. Regional paleoclimate records

The results of our current work on multiple lakes in the Bolivian Andes show that while the overall pattern of Holocene environmental change is consistent within the region, conditions were not always stable over centennial to over millennial timescales (Abbott et al., 1997a,b, 2000; Polissar, 1999). There is a notable discrepancy between the timing of water level rise in Lake Titicaca around 3.5 ka B.P. and the onset of wetter conditions in the LTCK watershed at 2.3 ka B.P. This suggests wetter conditions occurred in the northern reaches of the Titicaca watershed first resulting in rising water levels in Lake Titicaca while the LTCK watershed remained unglaciated and desiccated on a seasonal basis until 2.3 ka B.P. First, it is important to recognize that Lake Titicaca, with its expansive 57 000 km^2 watershed as highlighted by the dotted line in Fig. 1, integrates climatic conditions over a much greater area than the small alpine catchments described above that are $\ll$100 km^2. Data from cores collected from the network of

Table 1
Radiocarbon ages from Paco Cocha, Peru

Lab number	Depth (cm)	Material	Measured ^{14}C age (^{14}C yr B.P.)	Median calibrated ^{14}C age (Cal ^{14}C yr B.P.)
CAMS-51355	17	macrofossil	1 450 ± 55	1 320
OS-18383	49.5	macrofossil	2 060 ± 80	1 995
AA-27028	66.5	macrofossil	2 690 ± 110	2 770
OS-18384	100.25	macrofossil	2 995 ± 45	3 190
CAMS-51356	149.5	macrofossil	3 610 ± 100	3 890
AA-27029	187.5	macrofossil	4 000 ± 40	4 435
CAMS-51357	224.5	macrofossil	4 235 ± 50	4 830
OS-18617	282.5	macrofossil	5 750 ± 70	6 530
AA-27030	295.75	macrofossil	6 675 ± 100	7 490
AA-27031	365.25	macrofossil	9 560 ± 105	10 755
AA-27032	365.25	macrofossil	10 870 ± 70	12 795

small alpine watersheds should reveal more localized climatic information that can be used to identify the spatial pattern of climatic change through time.

Both the LTCK record (Abbott et al., 1997a) and the Lake Titicaca record (Abbott et al., 1997b) are generally consistent with other paleorecords in the region (Abbott et al., 2000), but are apparently at odds with one another between 3.5 and 2.3 ka B.P. (Fig. 8). One way to explain this apparent discrepancy is if wetter conditions occurred first in the northern reaches of the Lake Titicaca watershed and migrated southward over a multicentury timescale. The data from Laguna Paco Cocha presented in Fig. 4 supports this assertion by showing progressively lighter isotope values during this period. Further evidence is presented in Fig. 8 showing selected results from our north–south transect. Because of space limitations proxies are shown in Fig. 8 that best illustrate the records from Lagunas Llacho Kkota, Viscachani, Cupextani, and Juntutuyo. Farther to the north, results of $\delta^{18}O$ analyses of carbonates in Lake Junin (11°S) show conditions became progressively wetter during the Holocene (Seltzer et al., 2000). Interestingly, the $\delta^{18}O$ record in Junin is relatively smooth with no abrupt shifts, unlike what we observe farther to the south. In contrast, Paco Cocha, Llacho Kkota, LTCK, and Viscachani all contain abrupt shifts to wetter conditions with the onset of increased humidity occurring later at sites in the south. Junin may be located far enough to the north that it is not as affected by the progressive southward migration of the zone of summer convection as the other sites further to the south discussed in this paper.

In summary, Fig. 7 shows a series of seven lake-core records from this study illustrating the onset of wetter conditions during the middle to late Holocene at progressively later dates at sites located farther south in the transect. North to south: (1) glacial ice returns to the Paco Cocha watershed at 4.8 ka B.P., (2) wetter conditions persist at Llacho Kkota after 3.4 ka B.P., (3) glacier ice returns to the watersheds of LTCK and Viscachani after 2.3 ka B.P., (4) the watershed in Cupextani is stabilized and clastic input is decreased between 2.3 and 1.6 ka B.P., (5) Juntutuyo is desiccated for a prolonged period prior to 2.3 ka B.P., and (6) Potosi has wetter conditions after 2.1 ka B.P. Watersheds south of 17°S do not reestablish glaciers in the watershed during the late Holocene, although these catchments were glaciated during the late Pleistocene. The lower headwalls and decreased precipitation in the southern sites are the likely reason glaciers have not reestablished themselves in these watersheds during the Holocene.

The results from Lake Titicaca combined with results from the Sajama ice core (Thompson et al., 1998) and small alpine lakes in the eastern cordillera of the Titicaca Watershed (Abbott et al., 1997a, 2000; Wolfe et al., 2001a) all suggest a prolonged period of aridity on the Altiplano during the middle Holocene (Fig. 8). The onset of extreme middle-Holocene aridity between ∼7.0

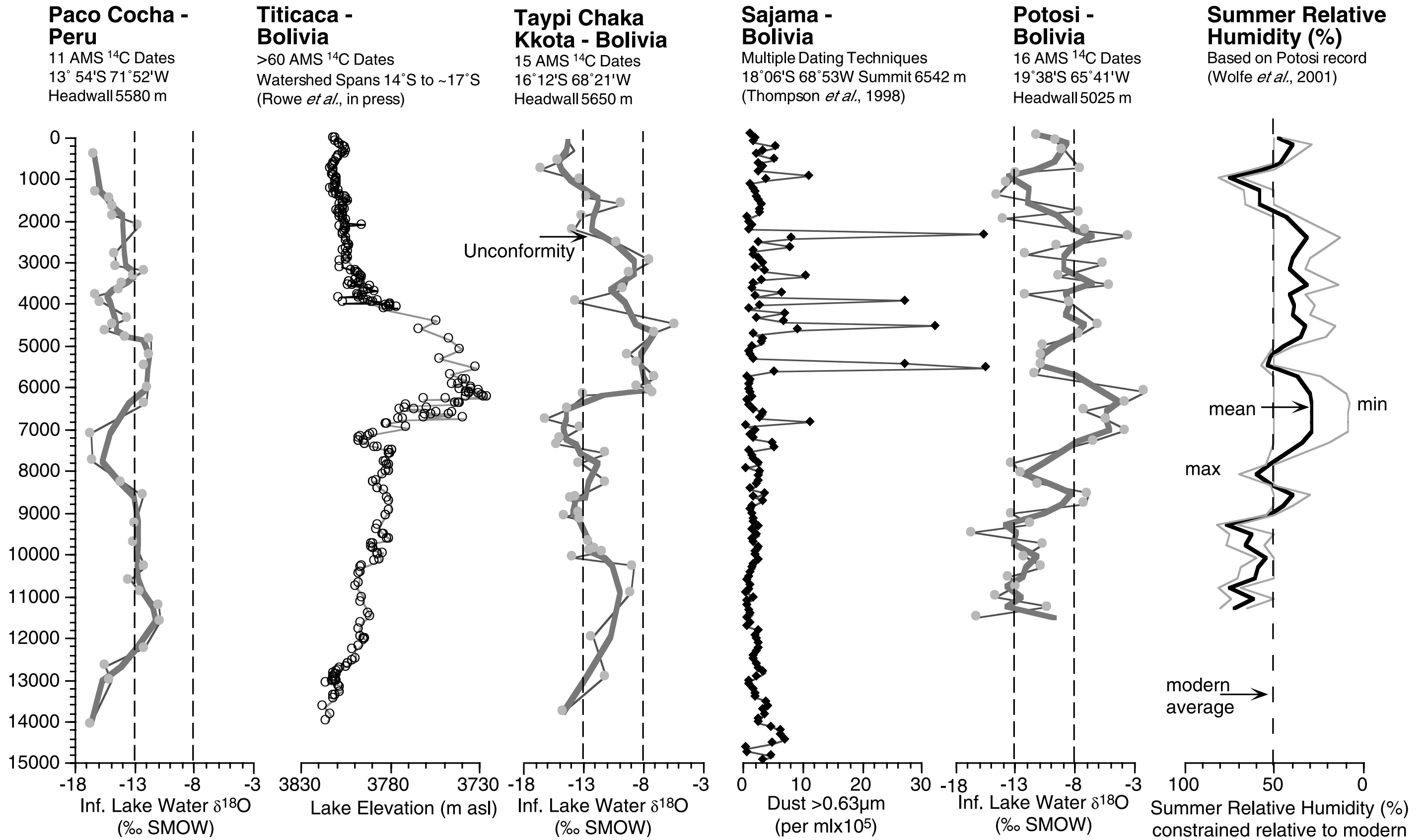

Fig. 7. Results from lake cores collected from sites on a north–south transect spanning the Titicaca watershed – Paco Cocha, Llacho Kkota, LTCK, Viscachani, Cupextani, Juntutuyo, and Potosi and also from Junin (Seltzer et al., 2000) which is situated well to the north at 11°S. There is a trend toward wetter conditions at northern sites, starting with Paco Cocha at 4.8 ka B.P., Llacho Kkota at 3.4 ka B.P., and both Taypi Chaka Kkota and nearby Viscachani at 2.3 ka B.P. The Junin record suggests progressively wetter conditions during the Holocene (Seltzer et al., 2000), but does not show abrupt hydrologic change, possibly because it is well north of the region affected by millennial-scale north–south migration of the tail end of austral summer convection (ITCZ). Results from lake cores collected from sites located south of the Lake Titicaca watershed (south of 17°S) indicate that these sites remained unglaciated throughout the Holocene. Cupextani shows lower magnetic susceptibility values during the latest Holocene associated with an increase in organic matter content. Juntutuyo was desiccated for a prolonged period prior to 2.3 ka B.P. Fluctuating hydrological conditions at Lago Potosi during the late Holocene correlate with lake-level changes in Lake Titicaca identified in the high-resolution record of Abbott et al. (1997b) and will be the subject of future high-resolution work. Due to limited space the analyses presented above were selected to highlight the properties of the sediments at each site.

and 6.0 ka B.P. is well illustrated in each of the records presented in Fig. 8. The duration of the arid phase is greatest, and the intensity highest, at Laguna Potosi, the southernmost location of the north–south transect. In contrast, the period of dry conditions is shorter and less severe at Laguna Paco Cocha, the northernmost site. The water-level rise in Lake Titicaca began about the time that glacial ice returned to the Paco Cocha watershed, suggesting that precipitation in the northern reaches of the catchment is an important control on water levels in Titicaca, at least during the middle to late Holocene. Finally, the timing of the beginning and end of the arid phase in the LTCK and Sajama records is strikingly similar, suggesting that the southern end of the north–south transect responds differently from the northern end. This is the topic on ongoing research on the sites at the southern end of the north–south transect.

Farther to the west, in the Atacama, there is contradictory evidence. Studies of alluviation and lake deposits suggest a prolonged dry phase during the middle Holocene (Valero-Garcés et al., 1996; Grosjean et al., 2001; Nunez et al., 2001). Betancourt et al. (2000), however, presented studies of rodent middens and spring deposits that suggest wetter conditions on the Atacama during the middle Holocene. Placzek et al. (2001) also suggest wetter conditions during the middle Holocene on the basis of a radiocarbon date on lake deposits in southernmost Peru, with the maximum lake level attained before 6.9 ka B.P. and ending around 2.8 ka B.P. Further evidence for wetter conditions in northern Atacama (16°S) is provided by rodent middens that do not reflect the arid conditions that are observed in the eastern cordillera of the central Andes between ~6.0 and 3.5 ka B.P. (Holmgren et al., 2001). As Holmgren et al. (2001) state this should not necessarily be surprising because wet-season rainfall is poorly corralated among weather stations in the central Andes and the Atacama.

5. Conclusions

Sound age control and multiproxy analyses on a network of sites across the region allows us to compare the timing, direction, and magnitude of climatic change as well as investigate the spatial pattern of these changes with the goal of identifying forcing mechanisms. Evidence presented here from lakes spanning 14°S to 20°S indicates that there were time-transgressive millennial-scale climatic changes across the central Andes during the late Pleistocene and Holocene. Results from the north–south network currently sampled at 50–100-yr resolution show there was a prolonged period of aridity throughout the region of the Altiplano and eastern cordillera during the middle Holocene (Figs. 4–8). The onset of wetter conditions in the late Holocene occurred ~2000 years earlier in the northern reaches of the Titicaca watershed than in the south. We suggest that this time-transgressive change in the P–E balance is related to orbital forcing, which results in a progressively southward shift in the location and a strengthening of wet-season convection as summer insolation increases during the late Holocene. Overprinting these millennial- to century-scale climatic shifts are higher-resolution changes occurring in each of the records presented that are not directly related to insolation forcing. We suggest that these fluctuations in the regional water balance are most likely related to changes in the recurrence interval, sign, and amplitude of tropical Pacific SSTA, and thus are related to more concentrated periods of El Niño Southern Oscillation activity (cf. Clement et al., 2000) as there is a strong correlation across the Altiplano between dry conditions and El Niño, and wet conditions and La Niña (Vuille, 1999; Vuille et al., 2000).

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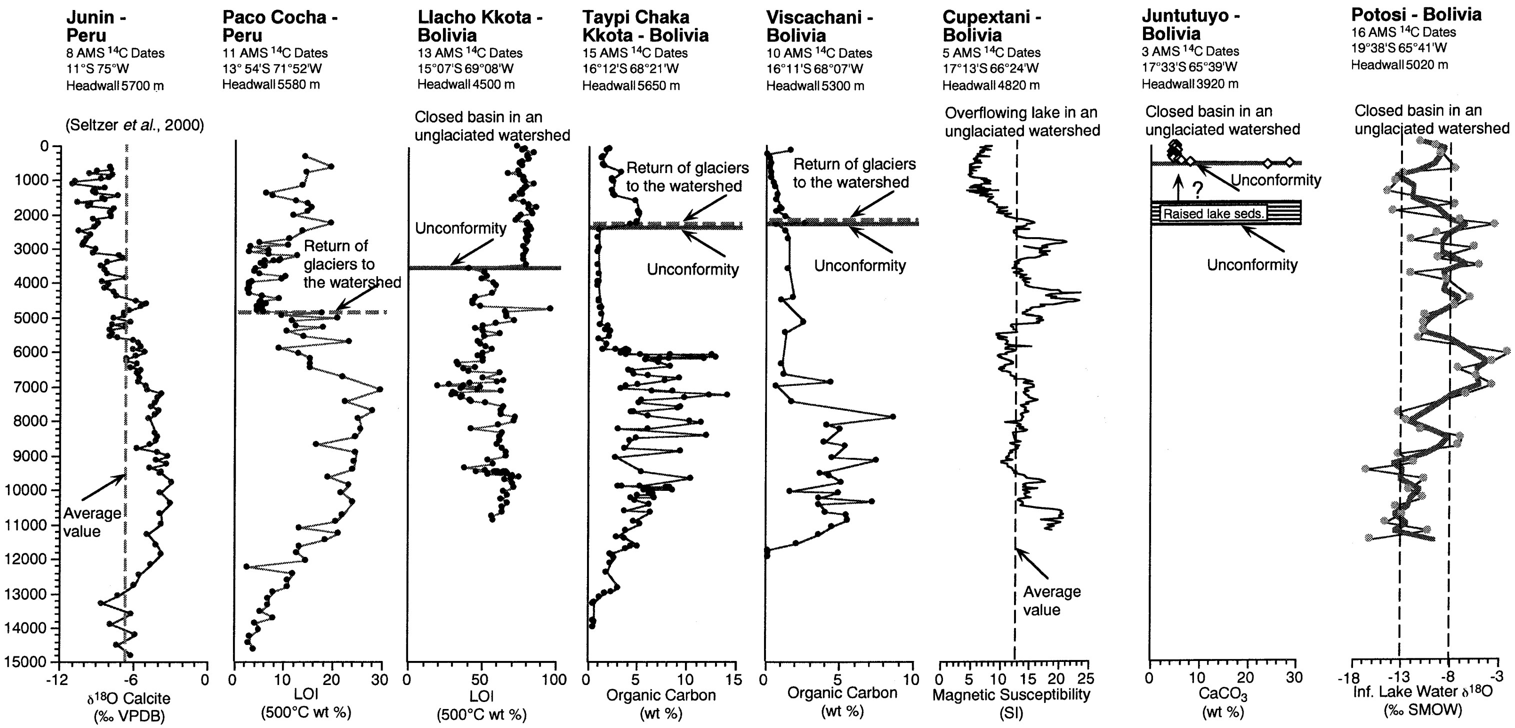

Fig. 8. Regional records of paleohydrology including: (1) the cellulose-inferred lake water $\delta^{18}O$ record from Paco Cocha, (2) water-level reconstruction of Lake Titicaca established using transfer functions based on the $\delta^{13}C$ of modern lacustrine organic matter endmembers (Rowe et al., 2002), (3) the cellulose-inferred lake water $\delta^{18}O$ record from Taypi Chaka Kkota (Abbott et al., 2000), (4) the Sajama ice core dust record (Thompson et al., 1998), (5) the cellulose-inferred lake water $\delta^{18}O$ record from Potosi (Wolfe et al., 2001a), and (6) summer relative humidity reconstruction derived from Laguna Potosi, which shows pronounced aridity between ~8.0 and 2.0 ka B.P. (Wolfe et al., 2001).

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