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Lake-Level Chronology on the Southern Bolivian Altiplano (18°–23°S) during Late-Glacial Time and the Early Holocene

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Stratigraphic analyses of outcrops, shorelines, and diatoms from the southern Bolivian Altiplano (Uyuni-Coipasa basin) reveal two major lacustrine phases during the late-glacial period and the early Holocene, based on a chronology established by radiocarbon and U/Th control. A comparison of ^{14}C and $^{230}Th/^{234}U$ ages shows that during times of high lake level, radiocarbon ages are valid. However, during low-water periods, ^{14}C ages must be corrected for a reservoir effect. The lacustrine Tauca phase started a little before 16,000 ^{14}C yr B.P., and the lake level reached its maximum between 13,000 and 12,000 ^{14}C yr B.P. A dry event (Ticaña) occurred after ca. 12,000 and before 9500 ^{14}C yr B.P. A moderate lacustrine oscillation (Coipasa event) occurred between ca. 9500 and 8500 ^{14}C yr B.P., using a reservoir-corrected conventional ^{14}C chronology. Comparisons between the lake-level chronology in the Uyuni-Coipasa basin and data from other southern tropical areas of South America suggest that the lacustrine evolution may reflect large-scale climatic changes. © 1999 University of Washington.

INTRODUCTION

Closed lakes respond to climate changes with fluctuations in water level, salinity, and chemistry. Such lakes represent a powerful tool for paleoclimatic reconstructions if a good chronology can be obtained. Significant changes in lake levels occurred on the southern Bolivian Altiplano (Uyuni-Coipasa basin) during the late-glacial period. A large lake (Lake Tauca) filled the closed basin where salars (salt pans) now exist (Servant and Fontes, 1978). The timing of lake-level changes has been based on radiocarbon dating of gastropod shells, calcareous crusts, algal bioherms, and stromatolites (Servant *et al.,* 1995; Sylvestre *et al.,* 1996). However, in closed lake systems the radiocarbon inventory is not always in ^{14}C equilibrium with atmospheric CO_2. Contamination by old carbon ("reservoir effect") can contribute to inaccurate radiocarbon age estimates (Fontes and Gasse, 1991; Fontes *et al.,* 1992; Benson, 1993; Fontes *et al.,* 1996). In the northern Altiplano, the lake reservoir effects are minimal in the peat bogs and glacial lakes (Abott *et al.,* 1997a). In Lake Titicaca, the modern ^{14}C reservoir effect is 250 yr (Abott *et al.,* 1997b). In several lakes of the Atacama Altiplano (northern Chile), the ^{14}C reservoir effect is on the order of thousands years for modern lake water and live aquatic organic plants (Grosjean *et al.,* 1995). Although these authors suggest that a modern reservoir correction should be applied to revise the late-glacial/early Holo-

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cene lake-level chronology of the Chilean Altiplano, paleohydrologic conditions changed during the past and it is unlikely that a modern reservoir correction is applicable throughout (M.A. Geyh, M. Grosjean, and U. Schotterer, unpublished data). A direct comparison of apparent radiocarbon ages with other independent chronometers is needed. In this paper, we propose a revision of the radiocarbon chronology in the southern Bolivian Altiplano (Uyuni-Coipasa basin) based on a comparison between radiocarbon and $^{230}Th/^{234}U$ disequilibrium methods performed on the same samples.

THE BOLIVIAN ALTIPLANO

The Bolivian Altiplano is flanked by an ice- and snow-capped cordillera that rises above 6000 m altitude (Fig. 1a). The plateau that constitutes the Altiplano is the sedimentary infill of an intermontane trench (or series of trenches) of tectonic origin. The Altiplano is approximately 2000 km long and 200 km wide and extends between 15° and 23°S (Fig. 1a). It lies generally above 3800 m altitude and the drainage is endorheic.

Lake Titicaca lies 3810 m above sea level in the northern Altiplano. It receives an average rainfall of 895 mm/yr (Roche *et al.*, 1992). Its outlet, the Río Desaguadero (3804 m altitude), drains southward over a sill that separates the Titicaca basin from the larger southern basin of the Altiplano. Río Desaguadero provides the main inflow of freshwater to the basin occupied by Lake Poopó at 3686 m altitude. This lake experiences drier climatic conditions, with a mean annual rainfall of 390 mm/yr. Depending on the season and the year, the lake depth varies between 1.6 and 2.2 m. Lake Poopó feeds into a larger salt flat, the Salar of Coipasa (2500 km^2, 3657 m altitude), which lies adjacent to the more extensive Salar of Uyuni (10,000 km^2, 3653 m altitude). These two basins receive an average rainfall of 100 mm/yr. During the rainy season, the salars may be covered by brine to a depth of about 25 cm, but during the dry season, the brine usually evaporates to expose the salt surface (Risacher, 1992). Occasionally, there is a hydrologic connection between the Salar of Coipasa and the Salar of Uyuni, the threshold being located at 3655 m altitude. However, today, the Salar of Uyuni is fed mainly by the Río Grande of the Lipez, which comes from the southern part of the Bolivian Altiplano. The southern part of the Bolivian Altiplano (south Lipez) is a volcanic landscape, characterized by several small closed lacustrine basins located at about 4800 m altitude covered by shallow brines and/or salt crusts.

The region is characterized by a tropical semiarid to arid climate, with a short rainy season (December–March) and a long dry season (April–November). The seasonality of precipitation is controlled by the movement of the Intertropical Convergence Zone (ITCZ). During the austral summer, the ITCZ moves southward over South America. Water vapor coming from the Atlantic Ocean and the Amazon Basin brings precipitation to the Bolivian Altiplano in the rainy season; during the austral winter, the ITCZ moves northward, reducing rainfall in the area.

REGIONAL STRATIGRAPHY

The regional stratigraphy based on field studies has been established by Servant *et al.* (1995), Rouchy *et al.* (1996), and Sylvestre *et al.* (1996) (Fig. 1b). A synthetic stratigraphic diagram (Sylvestre *et al.*, 1996) across the region comprises from the base to the top (Fig. 2) lacustrine clay deposits (**M**) cropping out on the margin of the Salar of Coipasa. These are attributed to the upper part of the Minchin lacustrine phase. Fluviatile sands and gravels (**F1**) observed on the margins of the Salar of Coipasa and Uyuni overlie the lacustrine **M** clays. They are related to a strong erosional phase in the catchment area. Silts, clays, and diatomites (**T**) crop out between 3653 and 3735 m altitude around the two basins. These are attributed to the lacustrine Tauca formation. Algal bioherms (**B**) associated with this phase reach 3760 m altitude. Fluvial sands containing interbedded clayey–silty lens-shaped deposits (**F2**) at 3657 m altitude overlie the Tauca deposits. A calcareous crust (**C**), which supports small algal bioherms, represents the most recent lacustrine deposits (Coipasa formation). It is widely developed around the margin of the two basins at 3660 m altitude. A halite crust (**H**) covers the bottom of the salars. Numerous cores show that this crust is about 10 m thick in the salar of Uyuni (Risacher, 1992).

CHRONOLOGY

Radiocarbon Ages

There are 44 ^{14}C ages (Table 1) for deposits from the Uyuni-Coipasa basin (Servant *et al.*, 1995; Sylvestre, 1997). Because of the absence of terrestrial organic macrorests, radiocarbon dates were obtained on autochtonous inorganic carbon (12 samples of calcareous crust, 1 sample of sediment, 1 sample of aragonite) and on autochtonous inorganic carbon of biological remains (16 samples of mollusc shells, 10 samples of characeae, 3 samples of algal bioherms). Mineralogic and isotopic data were used (Sylvestre, 1997) to detect whether some radiocarbon ages could be anomalously young because of the introduction of younger carbon through recrystallization of carbonate material. Radiocarbon dating was performed by two techniques: conventional liquid scintillation at Laboratoire de Géochronologie (ORSTOM, Bondy, France) and accelerator mass spectrometry (AMS) at Beta Analytic (Miami, FL).

Uranium–Thorium Ages

Uranium–thorium dating is reliable under closed-system conditions. The method is more difficult to apply in continental environments than in marine ones because closed-system conditions are rarely encountered in continental areas. Noticeable exceptions are permafrost conditions and hyperarid environ-

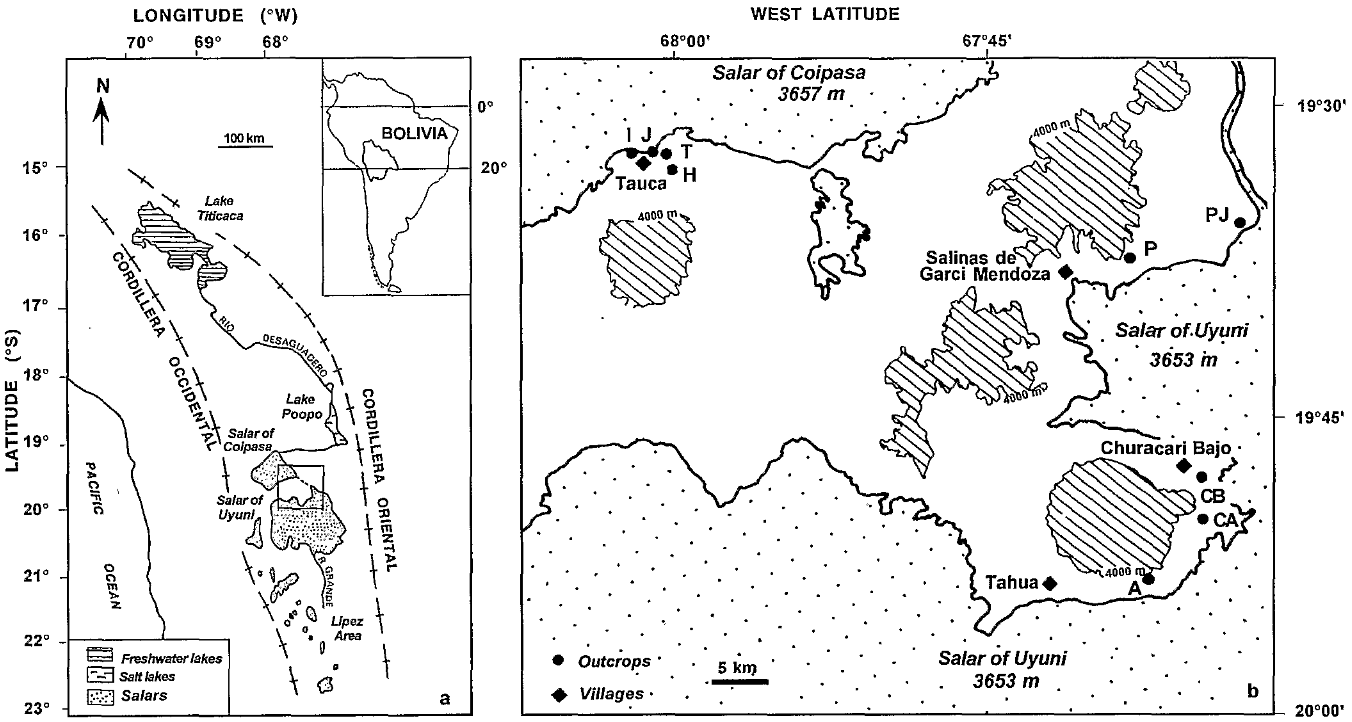

FIG. 1. (a) Map of the Bolivian Altiplano showing location of the study area in the Coipasa-Uyuni basin; (b) Map of the study area and location of the study sites in the margin of the Uyuni-Coipasa basin. I and J: Tauca sites (19°30'S, 67°98'W), H: Huacuyo (19°30'S, 68°00'W), T: Ticaña (19°30'S, 68°00'W), PJ: Pakollo Jahuira (19°32'S, 67°31'W), P: Pisalaque (19°32'S, 67°31'W), CB: Churacari Bajo (19°72'S, 67°31'W), CA: Churacari Alto (19°72'S, 67°31'W), A: Alianza (19°75'S, 67°40'W).

ments (e.g., Causse and Vincent, 1989; Causse *et al.,* 1989; Gaven *et al.,* 1981). An additional difficulty is the composition of continental carbonates, which are very often "dirty" (Schwarcz and Latham, 1989). Detrital ^{232}Th, linked with an excess of unsupported ^{232}Th, is frequent in continental settings, and calculated ages must be corrected for this detrital contamination. In the case where no gain or loss of uranium has occurred, an estimate of the actual age can be obtained by the isochron method (Schwarcz and Skoflek, 1982; Ku and Liang, 1984; Causse, 1992). Plotting ^{230}Th/^{232}Th against ^{234}U/^{232}Th allows the influence of detrital thorium to be determined, assuming that the detrital supply possesses a constant ^{230}Th/^{232}Th ratio. This activity ratio is commonly not different from 1, and especially in the case of low detrital contribution, this value can be used to correct for it (Geyh and Hennig, 1986; Causse and Vincent, 1989). This alternative correction was used here to calculate ages expressed as A_1 (Table 2).

We applied the uranium-thorium dating to six samples, for which ^{14}C was also analyzed (Table 2). The U/Th measurements were performed by thermal-ionization mass spectrometry (TIMS) at GEOTOP, Université du Québec (Montréal). The samples were selected after a petrologic study (Rouchy *et al.,* 1996) to represent specific paleohydrological environments (Sylvestre *et al.,* 1996): one sample (PK) is related to a shallow water level at the beginning of a lacustrine transgression, one sample (HUA44) corresponds to a high water level, and two samples (BO.23H, BO.23M) correspond to a moderate positive lacustrine oscillation superimposed on a long-term regression. Two analyses (CHU 93/24, CHU 91/6) were also performed on characeae in order to test the applicability of the ^{230}Th/^{234}U desequilibrium method to this material.

The comparison between ^{14}C and U/Th dates shows a good agreement for the PK and HUA44 samples (Table 2). For instance, sample PK gave a radiocarbon age of 15,430 ^{14}C yr B.P., calibrated to 18,400 cal yr B.P., and the U/Th age is 18,860 yr. There is a discrepancy between ^{14}C and U/Th ages for samples BO.23H and BO.23.M. The difference between ^{14}C and U/Th ages is ~1900 yr for BO.23H and ~2300 yr for BO.23M. There is also a discrepancy between ^{14}C and U/Th ages on characeae samples CHU93/24 and CHU91/6, which is difficult to assess because of a large detrital contribution to the U/Th system. However, both dating methods place these samples in the late-glacial period.

RECONSTRUCTIONS OF LAKE-LEVEL CHANGES IN THE UYUNI–COIPASA BASIN

We propose a revised lake-level chronology in the Uyuni-Coipasa basin during the late-glacial period and the early Holocene based on the regional stratigraphy (Fig. 2), on radiocarbon and U/Th ages (Tables 1 and 2), and on diatom studies (Fig. 3; Sylvestre *et al.,* 1996). Lake levels are estimated by the

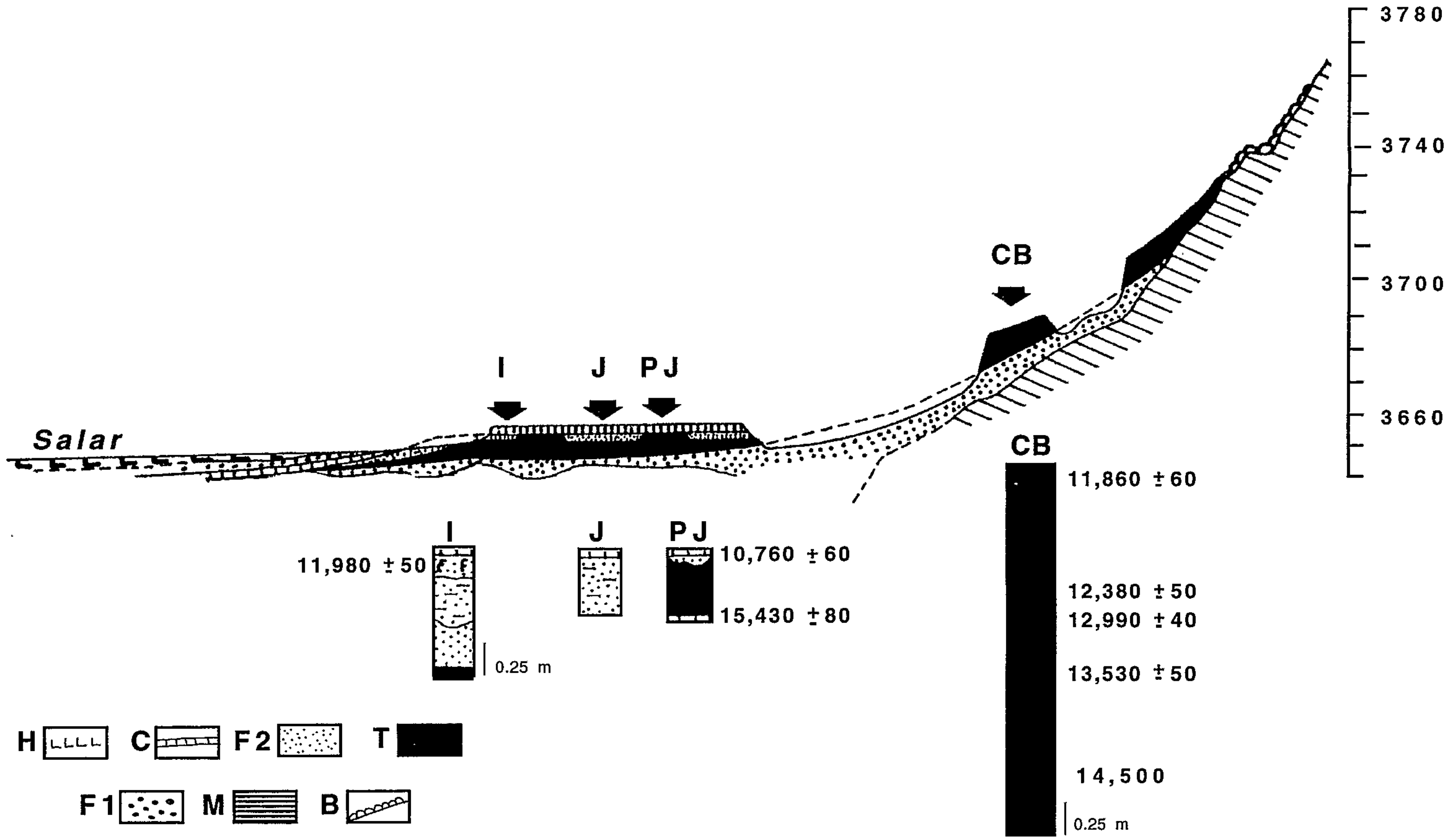

FIG. 2. Synthetic stratigraphic diagram of the Late Quaternary deposits in the Uyuni-Coipasa basin. (Ages in conventional ^{14}C yr B.P.) (**H**) halite crust, (**C**) calcareous crust with small algal bioherms, (**F2**) fluvial sands with clayey-silty lens shaped deposits, (**T**) lacustrine formation Tauca, (**F1**) fluvial sands, (**M**) Minchin lacustrine formation, (**B**) algal bioherms.

TABLE 1
List of Radiocarbon Ages in the Uyuni-Coipasa Basin

Regional/stratigraphy Uyuni-Coipasa basin		Sample number	Laboratory number	Site (Fig. 1)	Approximate altitude (m)	Material	^{18}O/PDB (‰)	^{13}C/PDB (‰)	Age (^{14}C yr B.P.)	Age (cal. ^{14}C yr B.P.) (Bard *et al.*, 1993)
Lacustrine *Coipasa* event (calcareous crust **C**)	V	42	Orsay	T	3660	Calcareous crust	—	1.18	10,450 ± 160	12,552
		92/11	OBDY 1055	PJ	3657	Calcareous crust	−9.38	2.19	10,580 ± 50	12,588
		91/5	OBDY 927	PJ	3657	Calcareous crust	−9.57	2.47	10,760 ± 50	12,761
		92/3	OBDY 1046	T	3660	Calcareous crust	−10.2	2.31	10,810 ± 50	12,808
		126	Orsay	PJ	3660	Calcareous crust	—	2.98	10,830 ± 180	12,934
		92/10	OBDY 1045	PJ	3660	Calcareous crust	−9.65	1.39	10,960 ± 50	12,947
		92/3	OBDY 1057	T	3660	Calcareous crust	−10.39	2.4	11,020 ± 60	13,012
		93/20	OBDY 1362	T	3660	Calcareous crust	—	2.65	11,260 ± 50	13,245
		91/1	OBDY 925	PJ	3660	Calcareous crust	−9.46	2.41	11,390 ± 50	13,389
Ticaña event (fluviatile sands **F2**)	IV	93/17	OBDY 1290	T	3657	Mollusc shells	−4.58	1.19	11,980 ± 50	14,100
		93/18/A	Beta73080	T	3660	Mollusc shells	—	1.4	12,960 ± 60	15,524
		93/18/B	Beta73081	T	3660	Mollusc shells	—	0.5	13,130 ± 60	15,777
Lacustrine *Tauca* event (lacustrine deposits **T**)	III	230	Orsay	H	3695	Mollusc shells	—	1.04	11,730 ± 350	14,107
		92/8	OBDY 1035	CB	3690	Characea	−12.16	4.32	11,860 ± 60	13,958
		92/8	OBDY 1025	CB	3690	Mollusc shells	−11.53	1.65	12,880 ± 70	15,413
		91/13	OBDY 911	H	3700	Mollusc shells	−11.82	1.9	11,920 ± 40	14,020
		91/13	OBDY 914	H	3700	Characea	−12.57	3.63	12,090 ± 40	14,237
		876	OBDY 364	A	3745	Mollusc shells	−10.65	4.7	12,100 ± 370	14,628
		86/113	OBDY 257	H	3720	Mollusc shells	−10.85	1.73	12,210 ± 270	14,650
		167	Orsay	RS[a]	3720	Mollusc shells	—	2.46	11,090 ± 280	13,012
		167	Orsay	RS[a]	3720	Carbonates	—	3.31	12,260 ± 130	14,516
		876	OBDY 997	A	3657	Characea	−9.57	1.27	12,870 ± 50	15,379
		92/2	OBDY 1049	H	3740	Bioherms	−11.31	4.22	12,290 ± 50	14,516
		93/22	OBDY 1214	M	3745	Bioherms	—	3.8	12,270 ± 50	14,488
		91/7	OBDY 919	CB	3690	Mollusc shells	−10.96	1.94	12,380 ± 50	14,647
		91/7	OBDY 916	CB	3690	Characea	−11.91	3.91	12,990 ± 40	15,55
		91/14	OBDY 918	H	3690	Mollusc shells	−10.48	0.28	12,390 ± 50	14,662
		90/13	OBDY 700	CB	3690	Characea	−11.82	3.99	12,490 ± 80	14,830
		90/13	OBDY 681	CB	3690	Mollusc shells	−11.39	2.69	12,830 ± 80	15,348
		91/10	OBDY 929	CA	3735	Characea	−11.75	4	12,560 ± 130	14,989
		91/10	OBDY 923	CA	3735	Mollusc shells	−11.01	1.53	12,930 ± 50	15,469
		40	Orsay	C[b]	3720	Mollusc shells	—	1.24	12,790 ± 120	15,334
		91/9	OBDY 926	CA	3735	Mollusc shells	−10.63	1.6	13,030 ± 80	15,649

TABLE 1—*Continued*

Regional/stratigraphy Uyuni-Coipasa basin		Sample number	Laboratory number	Site (Fig. 1)	Approximate altitude (m)	Material	^{18}O/PDB (‰)	^{13}C/PDB (‰)	Age (^{14}C yr B.P.)	Age (cal. ^{14}C yr B.P.) (Bard *et al.*, 1993)
	II	93/24	OBDY 1291	CB	3685	Characea	—	3.93	12,850 ± 50	15,349
		93/23	OBDY 1289	CB	3685	Characea	—	3.61	13,190 ± 50	15,855
		92/9	OBDY 1007	P	3685	Characea	−11.56	4.42	13,620 ± 60	16,441
		91/6	OBDY 879	CB	3685	Characea	−11.26	3.96	13,530 ± 50	16,317
	I	91/4/B	Beta73088	PJ	3657	Aragonite	0.39	4.2	15,430 ± 80	18,458
Recrystallized samples		91/8	OBDY 969	CB	3690	Calcareous sediment	−5.41	−0.1	11,300 ± 60	
		91/12	OBDY 924	PJ	3658	Calcareous crust	−4.41	0.15	6950 ± 50	
		92/5	OBDY 1037	TJ	3700	Bioherms	−11.46	3.96	9970 ± 50	
		92/4	OBDY 1058	TJ	3700	Mollusc shells	−10.92	1.33	11,050 ± 70	
		91/4/A	OBDY 928	PJ	3658	Calcareous crust	−12.09	3.75	13,570 ± 50	
		91/2	OBDY 930	PJ	3660	Calcareous crust	−11.08	4.3	14,450 ± 50	

Note. Laboratories: Orsay: Laboratoire d'Hydrologie et de Géochimie isotopique, Université d'Orsay; Beta: Beta Analytic Inc. Miami; OBDY: Laboratoire de Géochronologie ORSTOM-Bondy.

[a] RS: Rio Salado, 19°30'S/67°00'W (Servant and Fontes, 1978).

[b] C: Coipasa, 19°10'S–69°15'W (Servant and Fontes, 1978).

altitude of the lacustrine deposits above the present bottom (the halite crust of the Salar of Uyuni at 3653 m altitude). The lake-level fluctuations are confirmed by analyses of fossil diatom flora in outcrops PJ (3657 m altitude), I, J (3660 m altitude), and CB (3685 m altitude), and of fossil diatoms in algal bioherms located on the highest shoreline at 3760 m altitude (Sylvestre *et al.,* 1996; Fig. 1b). Figure 3 shows for PJ and CB the percentage of periphytic, tychoplanktonic, and planktonic diatoms, which are indicators of shallow, intermediate, and deep water conditions, respectively.

Lacustrine Tauca Period (ca. 16,000–12,000 ^{14}C yr B.P.)

Lacustrine transgressive conditions began at least at ca. 15,500 ^{14}C yr B.P. The initial lacustrine transgression (Tauca Ia) is represented at the base of section PJ (3657 m) by a thin carbonate crust composed of aragonite dated at 15,430 ± 80 ^{14}C yr B.P. Radiocarbon and U/Th ages for this aragonite crust are in good agreement. The diatom assemblages are dominated by a benthic species *Denticula subtilis* (69%), indicating a shallow water body. The lake level was at ca. +4 m above the present salar. After 15,430 ± 80 ^{14}C yr B.P., the lake level had reached the altitude of the base of the CB outcrop at ca. +27 m above the present salar (Tauca Ib). At 3657 m (PJ), the dominance of the planktonic diatom *Cyclotella striata* implies deep water. At the base of the CB outcrop (~3682 m altitude) the sediments contain the tychoplanktonic diatoms *Fragilaria atomus* and *F. construens subsalina,* which live in intermediate water between littoral and deep conditions. These diatoms indicate water inputs to the basin.

The second phase of the Tauca period (Tauca II) is recorded at 3685 m altitude in outcrop CB by finely laminated sediments which contain in succession a dominance of epiphytic diatoms (*Achnanthes brevipes, Cocconeis placentula* var. *euglypta, Rhopalodia gibberula*) and tychoplanktonic diatoms (*Fragilaria pinnata*). These assemblages indicate shallow water on the margin of the basin with episodic fresh water inputs (Sylvestre, 1997). A sample from the middle of these deposits dates to 13,530 ± 50 ^{14}C yr B.P. An estimated age based on the sedimentation rate places the based of this phase at ca. 14,500 yr B.P. At lower altitude (3657 m) in PJ section, the Tauca II phase is recorded by planktonic diatoms *Cyclotella striata.* The water table is estimated to lie ca. 40 m above the present bottom.

Two radiocarbon ages of 13,790 ± 70 ^{14}C yr B.P. were obtained for gastropod shells sampled in sandy sediments (Bills *et al.,* 1994). These placed the lake level at this time at ca. 98 m above the present bottom. However, our estimations place the paleolake at ca. 40 m. To explain this discrepancy, we suggest three hypotheses: (1) The two radiocarbon dates of Bills *et al.* (1994) were obtained from gastropod shells in sandy

TABLE 2
List of U/Th Data and Comparison between ^{14}C and U/Th Ages in the Uyuni-Coipasa Basin

Sample	Site (Fig. 1)	Material	^{238}U (ppb) $\pm 2\sigma$	^{232}Th (ppb) $\pm 2\sigma$	U/Th	$^{234}U/^{238}U$ $\pm 2\sigma$	$^{230}Th/^{234}U$ $\pm 2\sigma$	$^{234}U/^{232}Th$ $\pm 2\sigma$	$^{230}Th/^{232}Th$ $\pm 2\sigma$	A0[a] (yr)	A1[b] (yr)	Age (C^{14} yr B.P.)	Age (C^{14} cal yr B.P.)
PK	**PJ**	Aragonite	2645.276	352.719	7.49	1.542	0.185	35.341	6.533	21.98 ± 0.39	18,860	15,430	18,400
			10.717	3.14		0.01	0.003	0.38	0.116				
HUA 44	**H**	Bioherms	313.438	21.603	14.22	1.571	0.143	69.666	9.983	16.67 ± 0.4	15,070	12,930	15,330
			0.863	0.285		0.007	0.003	0.974	0.261				
CHU93/24	**CB**	Characeae	454.81	232.905	1.95	1.463	0.219	8.737	1.919	26.62 ± 1.8	13,600	12,800	15,300
			2.200	4.437		0.011	0.013	0.177	0.120				
CHU91/6	**CB**	Characeae	584.63	222.551	2.62	1.473	0.197	11.827	2.334	23.63 ± 0.4	14,100	13,500	15,800
			2.195	1.394		0.011	0.003	0.116	0.036				
BO.23H	**T**	Calcareous crust	307.033	132.416	2.3	1.657	0.177	11.741	2.075	20.92	11,340	11,370	13,200
			0.893	0.666		0.01	0.008	0.091	0.09	(+1.04/−1.03)			
BO.23M	**T**	Calcareous crust	235.594	245.085	0.96	1.565	0.293	4.596	1.347	36.9	10,900	11,340	13,200
			0.949	3.359		0.012	0.033	0.071	0.153	(+4.96/−4.76)			

[a] A0 = uncorrected age.
[b] A1 = corrected age for $^{230}Th_{excess}/^{232}Th$ AR values = 1.

sediments. Possibly these shells were of terrestrial origin, but available data do not allow evaluation of this hypothesis. (2) The periphytic/tychoplanktonic diatoms and the characeae in the CB outcrop were reworked from the littoral zone to the deepest depositional environment. (3) The laminated sediments were reworked and reflect slumping on the bottom of the lake. Hypothesis 2 is unlikely because in the CB outcrop the characeae are in life position and clearly were not reworked. This is also observed in outcrop P (Fig. 1b) where nonreworked characeae at 3690 m are dated at 13,620 ± 60 ^{14}C yr B.P. (Servant *et al.,* 1995). With respect to hypothesis 3, field observations indicate that the laminated sediments were not affected by slumping in the sampled outcrop. Thus, the lake elevation did not exceed ~3700 m during the Tauca II phase.

The third phase of the Tauca period (Tauca III) is represented at several sites by diatomite deposits that extend from 3660 to 3735 m altitude. Twenty-one radiocarbon dates place this phase between 13,000 and 12,000 ^{14}C yr B.P. One U/Th age (15,070 yr B.P., HUA 44) was obtained for a bioherm on the highest shoreline at 3760 m, which is in excellent agreement with the calibrated radiocarbon age (15,330 cal yr B.P.). The paleolake Tauca was deepest (ca. 110 m) between 13,000 and 12,000 ^{14}C yr B.P. This deduction agrees with the dominance of the planktonic diatom *Cyclotella striata* in all outcrops studied at 3657 m (PJ), at 3685 m (CB), and on bioherms on the highest shoreline at 3760 m. In all the studied outcrops, the deposits are cut by an erosion surface and we cannot exclude that the Tauca III phase ended after ca. 12,000 ^{14}C yr B.P.

Some layers included in the diatomite deposits contain a mixture of mollusc shells and characeae related to shallow-water environments and planktonic diatoms that imply deep-water conditions. Radiocarbon analyses of mollusc shells and characeae in the same samples from these deposits (92/8, 91/13, 876, 91/7, 90/13, 91/10, 167) show significant differences. For instance, mollusc shells and characeae yielded ages of 11,920 ± 40 and 12,090 ± 40 ^{14}C yr B.P., respectively for sample 91/13. We suggest that the materials characteristic of shallow-water environments on the margin of basins have been transported to deeper environments by wave action (e.g., Benson *et al.,* 1990). The age differences between the mollusc shells and the characeae could be explained by differential reworking during the transgression of paleolake Tauca.

Radiocarbon ages related to transgressive phases and highstands (Tauca I, II, III) are in good agreement with U/Th dates. This suggests that the radiocarbon ages are not affected by a reservoir effect. The most reliable hypothesis is that, at that time, the level of groundwater rose more rapidly than the lake level, the groundwater was recharged by lacustrine water, and the carbonates (e.g., PK, HUA44) were in atmospheric CO_2 equilibrium with lake water. Although the comparison between radiocarbon and U/Th is applied to paired samples, we assume that the previous radiocarbon chronology for the Tauca formation is reliable.

The Post-Tauca Period (Ticaña and Coipasa Events)

The **F2** alluvial sands that crop out at 3657 m altitude on the margins of the Coipasa and Uyuni salars (Fig. 2) were deposited under generally dry conditions. These sands contain mollusc shells. Two radiocarbon dates for the shells (outcrops I and J near Tauca village; Fig. 1) gave ages of 12,960 ± 60 and 13,130 ± 60 ^{14}C yr B.P., suggesting that they were reworked from the former Tauca deposits. Mollusc shells are more abundant in the silty lens-shaped deposits interbedded in the **F2**

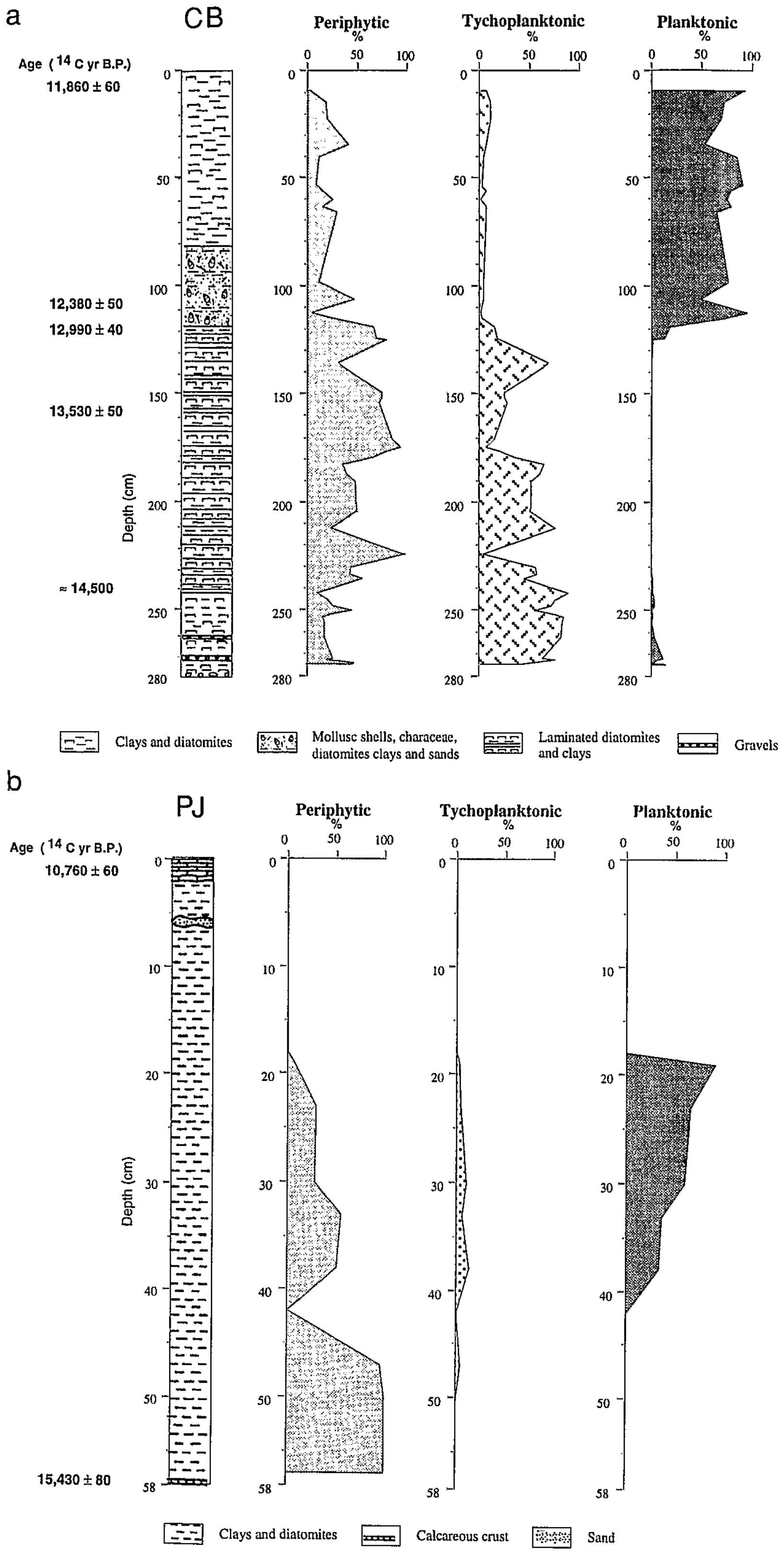

FIG. 3. Periphytic, tychoplanctonic, and planktonic diatoms expressed as a percentage for each sample in (a) the CB outcrop (3685 m) and (b) the PJ outcrop (3657 m).

alluvial sands, which contain the periphytic diatom *Denticula subtilis* (90%). Shallow residual ponds were scattered on the borders of the salars and were probably fed by springs. The age obtained for the shells (11,980 ± 50 ^{14}C yr B.P.) must be interpreted with caution because it is not possible to exclude a reservoir effect related to the feeding of these ponds by groundwater previously recharged during the former Tauca phase.

The lacustrine Coipasa event is represented by a calcareous crust (C) which is widely developed around the margin of the Uyuni-Coipasa basin at 3660 m altitude. This crust supports small algal bioherms, which contain a dominance of *Denticula subtilis,* indicating a shallow water body. We estimate that the lake level was at least at ca. 7 m above the present salar. We have not found deposits above this calcareous crust. Nine radiocarbon dates obtained from several sites around the two basins cluster between 11,390 ± 50 and 10,450 ± 160 ^{14}C yr B.P. Comparison of paired ^{14}C and U/Th analyses indicates that ^{14}C ages are ~2000 yr B.P. too old (Table 2). This is probably due to incorporation of old inorganic carbon coming from a discharge of groundwater recharged during the former Tauca transgression. We therefore assume that the lacustrine Coipasa event, previously dated between ca. 11,400 and 10,400 ^{14}C yr B.P. (Servant *et al.,* 1995; Sylvestre *et al.,* 1996), occurred in the early Holocene between ca. 9500 and 8500 ^{14}C yr B.P.

COMPARISON WITH OTHER SOUTHERN TROPICAL ANDEAN AREAS

Our new reconstruction can be compared with those for several other areas in the Bolivian and Chilean Altiplano, including (1) lake-level reconstructions from Lake Titicaca, from the southern part of the Bolivian Altiplano (Lipez area), from the Atacama in northern Chile, and (2) stratigraphic studies of changes in fluvial regimes in several nonglaciated Andean valleys from the Bolivian Altiplano and northern Chile.

At Lake Titicaca, lake-level changes were reconstructed from spores, pollen, algal taxa (Ybert, 1992), diatoms (S. Servant-Vildary, unpublished data) and sedimentological (Wirrmann *et al.,* 1988) data from cores TD and TD1, raised from a depth of ~19 m (Wirrmann, 1987). We have selected *Botryococcus* and *Pediastrum* taxa to illustrate this lake-level reconstruction (Fig. 4). The interpretation of fossil assemblages is based on a detailed study of modern assemblages in surficial sediments along two transects in Lake Titicaca (Ybert, 1992). *Botryococcus* and *Pediastrum* are rare (<20%) between 0 and 2 m depth, but increase from 20 to 70% between 2 and 4 m. Between 4 and 200 m, they compose 90% of the modern assemblages. In cores TD1 and TD, high percentages of *Botryococcus* and *Pediastrum* were observed, respectively, between 200 and 160 cm and between 380 and 320 cm (Fig. 4). These taxa are associated with the planktonic diatom *Cyclotella andina,* which is abundant in the oligosaline Lake Titicaca (Servant-Vildary, 1992). Thus, we conclude that salinity and depth were similar to the present conditions. In this situation we are able to apply the modern ^{14}C reservoir correction (−250 yr) for the ^{14}C age obtained on total organic carbon (13,180 ± 130 ^{14}C yr B.P.) near the maximum of *Botryococcus* and *Pediastrum,* at 182 cm in TD1. Thus, the Lake Titicaca highstand, indicated by the high percentage of *Botryococcus* and *Pediastrum,* can be correlated with the Tauca III phase.

An accurate altitude of the lake level cannot be estimated. An undated lacustrine terrace, located 5 m above the present level of Lake Titicaca and previously correlated to the Tauca phase by Servant and Fontes (1978) suggests that Lake Titicaca could have overflowed during this period into the southern basin (Poopó-Coipasa-Uyuni). A drastic drop of the lake level is registered by the disappearance of *Botryococcus* and *Pediastrum,* associated with the high percentages of littoral taxa, at 145 cm in the core TD1 before 7000 ± 230 ^{14}C yr B.P. and at 300 cm in the core TD just before 9620 ± 90 ^{14}C yr B.P. For that time, we cannot apply the modern ^{14}C reservoir correction to a paleolake of lower level than today. We tentatively correlate this event with the dry Ticaña event. The reappearance of *Botryococcus* and *Pediastrum* (40%) just after 9620 ± 90 ^{14}C yr B.P. (280–250 cm) in TD suggests that Lake Titicaca may have been somewhat deeper at that time (Fig. 4). This agrees with an early Holocene increase in lake level in the Uyuni-Coipasa basin (Coipasa event), but such an increase is not recorded in core TD1 and must be confirmed by further studies at Lake Titicaca.

Lacustrine records are also available in northern Chile on the Atacama altiplano (Messerli *et al.,* 1993; Grosjean, 1994; Grosjean *et al.,* 1995) and in the nearby areas of southern Lipez in Bolivia (Servant-Vildary and Mello e Sousa, 1993; Sylvestre, 1997). In both regions, the lacustrine basins are small and their inflow is closely controlled by groundwaters inducing a strong reservoir effect. In northern Chile, the chronology revised by M.A. Geyh *et al.* (unpublished data) is essentially based on ^{14}C ages of terrestrial organic macrorests, "because cross-dating ^{14}C on shoreline carbonates and stromatolites with U/Th failed due to high detrital contamination or uranium leaching." In Laguna Lejía (23°30'S, 67°42'W), a lacustrine phase, called "Tauca" by analogy with the Bolivian terminology, began shortly after 11,480 ± 70 ^{14}C yr B.P. (the age of terrestrial organic macrorests in a basal gravel). A date of 10,305 ± 725 ^{14}C yr B.P. for bird bone is stratigraphically consistent with the former date. This lacustrine phase lasted until 8430 ± 75 ^{14}C yr B.P. (total organic fraction) in the salar Aguas Calientes. These data suggest that at least the upper part of this lacustrine phase can be correlated with the Coipasa event of Bolivia.

In the Bolivian Lipez area, which has the same geologic and geomorphologic setting as the Atacama Altiplano, diatom studies (Servant-Vildary and Mello e Souza, 1993) were performed on deposits attributed to the Minchin phase, and on lacustrine terraces tentatively correlated to the Tauca phase (Servant and

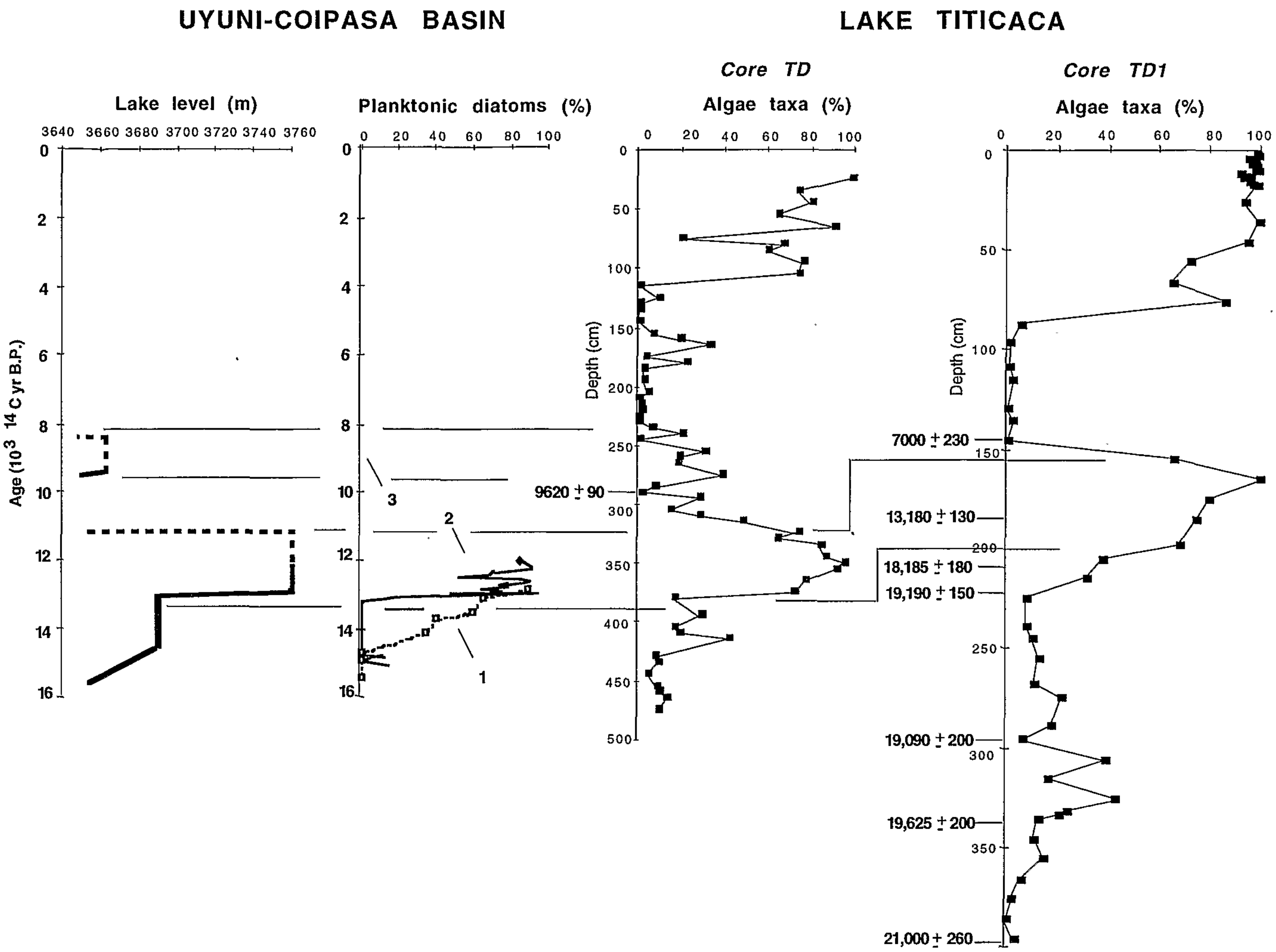

FIG. 4. Comparison between lake-level fluctuations on the southern and the northern Bolivian Altiplano. Lake-level fluctuations in the Uyuni-Coipasa basin are based on altitudes of lacustrine deposits and relative variations of planktonic diatoms studied in outcrop PJ (3657 m), outcrop CB (3685 m), and algal bioherms (3660 m). Lake-level fluctuations in Lake Titicaca are based on algal taxa (*Botryococcus* and *Pediastrum*) in the core TD and TD1 (Ybert, 1992). All ages are expressed in ^{14}C yr B.P.

Fontes, 1978). A detailed diatom study disclosed two lake-level highstands in these terraces that crop out on the margin of the small Chiar-Kkota and Honda basins (Sylvestre, 1997). The radiocarbon chronology cannot be used because the ages are inconsistent. However, three U/Th ages on calcareous crust and bioherms suggest that the late high lake level occurred between ca. 9640 and 8110 ^{14}C yr B.P. (Sylvestre, 1997).

The fluvial environments do not show a significant reservoir effect. Fluvial deposits were studied in several nonglaciated Andean valleys of Bolivia. Alluvial cones interbedded between two peats that dated ca. 16,000 and 12,350 ^{14}C yr B.P. indicated an intense channeling on slopes during early late-glacial time. This gullying phase can be correlated with the first steps of the lacustrine Tauca transgression (Tauca I and/or II). It is thought to have been caused by rainstorms similar to those that now occur during the summer on the Bolivian Altiplano (Servant *et al.*, 1981). Between ca. 13,000 and 10,000 ^{14}C yr B.P., peats and organic silty sediments are observed in the bottom of the nonglaciated valleys (Servant and Fontes, 1984). They were found in few sites, suggesting that they were strongly eroded just before 10,000 years ago. From 10,000 to 8000 ^{14}C yr B.P., peats are very well preserved in all the sites (Servant and Fontes, 1984). They indicate continued humid conditions in the valley bottoms where they do not now occur. This humid phase is synchronous with the lacustrine Coipasa phase, now dated as early Holocene in our revised chronology. Fluvial deposits older than 7300 ^{14}C yr B.P. are also present in northern Chile (Veit, 1996).

DISCUSSION

In spite of the unknown status of lake-level changes during early and middle late-glacial time (ca. 15,000–12,000 ^{14}C yr B.P.) in the Atacama Altiplano and the poor chronological control of the hydrologic changes in Lake Titicaca, we conclude that the lacustrine phases Tauca III (ca. 13,000–12,000 ^{14}C yr B.P.) and Coipasa (ca. 9500–8500 ^{14}C yr B.P.) were significant hydrologic events in the southern tropical Andes. Former interpretations tend to explain the lake-level highstands by meltwater input during glacier recession (Servant and Fontes, 1978) and/or precipitation higher than today (Hastenrath and Kutzbach, 1985). The first hypothesis is now considered unlikely because Clayton and Clapperton (1997) showed that a glacier advance coincided with a highstand of paleolake Tauca after ca. 13,300 ^{14}C yr B.P. They suggested that this glacier advance was related to increased summer moisture. However, a detailed chronology of glacier fluctuations and estimates of Lake Titicaca outflows into the southern Altiplano basins are still required for a thorough evolutionary reconstruction of the hydrologic system. Comparisons between lake-level changes in the southern tropical Andes and paleoenvironmental changes in the southern tropical lowlands of America offer a way of determining whether the lake-level changes are controlled either by regional forcing (e.g., meltwater input) or by large-scale climatic variations (e.g., shift of the ITCZ).

We must take into account the Amazonian lowlands because today they are the main source of water vapor feeding the precipitation over the tropical southern Andes. Palynological studies suggest that the areas which today are very humid did not experience drastic vegetation changes during the late-glacial period and the early Holocene (Colinvaux *et al.,* 1996). In now less-humid regions, strong vegetation changes occurred, as illustrated by a record in the Serra dos Carajás located in the southeastern part of Brazilian Amazonia (6°20'S, 50°25'W). The Serra dos Carajás is a narrow plateau 700–800 M high covered by an open or dense treelet-scrub savanna, surrounded by semidecideous forest vegetation. Sedimentological (Sifeddine *et al.,* 1994) and palynological (Absy *et al.,* 1991) studies were made of cores from small, shallow, freshwater lakes on the plateau. The chronologic control is based of radiocarbon dates on aquatic plant remains and total organic carbon. It is unlikely that these dates have been affected by a reservoir effect.

The modern pollen assemblages on the plateau are dominated by pollen of the surrounding vegetation, and the same situation is assumed to have existed for the fossil assemblages (Absy *et al.,* 1991). A hiatus from ca. 20,000 to 13,000 ^{14}C yr B.P. is interpreted as a drying of the lake. The appearance of lacustrine deposits at ca. 13,000 ^{14}C yr B.P. indicates wetter conditions. This change is correlated with the base of lacustrine Tauca III phase of the Bolivian Altiplano. However, pollen assemblages indicate herbaceous vegetation, suggesting long dry seasons or a dry climate (Salgado-Labouriau, 1997). From ca. 13,000 to 10,000 ^{14}C yr B.P., a progressive increase in arboreal taxa implies increased moisture, but pollen assemblages are still dominated by savanna taxa. This apparent contradiction between increasing lake level suggesting wetter conditions and palynological data suggesting dry conditions is also observed in other tropical areas during late-glacial time but it remains unexplained (e.g., Maley, 1991). Between ca. 9500 and 8000 ^{14}C yr B.P., the pollen assemblages are strongly dominated by forest taxa, indicating humid conditions in Carajás. This early Holocene humid phase is also recorded at other sites in Amazonia and can be correlated with the lacustrine Coipasa phase of the Uyuni-Coipasa basin.

In the tropical lowlands of central Brazil, at the same latitudes as the Bolivian Altiplano, the climate is controlled by the seasonal shift of the ITCZ and by the influence of polar air mass advections in the lower part of the atmosphere. In this region, now occupied by savanna, late-glacial and early Holocene records were obtained at several swampy sites. We assume that in such environments radiocarbon dates of organic remains are not affected by a significant reservoir effect. For late-glacial and Holocene time, the Salitre site (19°S, 46°46'W) yielded a well-dated, continuous record (Ledru, 1993). At the other sites, the chronology is poorly documented. In Aguas Emendadas (15°34'S, 47°35'W), late-glacial and early Holocene time are only represented by an undated thin silty-sand layer interbedded between organic layers, respectively dated at 21,450 ± 100 and 7220 ± 50 ^{14}C yr B.P. (Salgado-Labouriau *et al.,* 1997). At Cromínia (17°34'S, 49°25'W), clayey peats with macroscopic organic fragments were assigned to late-glacial/early Holocene time based on a single radiocarbon age (13,150 ± 50 ^{14}C yr B.P.). Pollen assemblages suggest that the clayey peats were deposited in a marsh surrounded by a poor grassland, probably in a semiarid climate or one with a very long dry season (Salgado-Labouriau *et al.,* 1997). In Salitre, between ca. 13,000–11,000 and 10,000–7500 ^{14}C yr B.P., moister conditions are indicated by the development of forest (Ledru, 1993). These two periods are, respectively, close to the lacustrine Tauca III and Coipasa phases of Bolivia. A very low percentage of arboreal taxa at ca. 10,500 ^{14}C yr B.P. indicates short and intense changes of the vegetation and the climate (long dry season, low temperature).

These comparisons are consistent with the hypothesis that the lacustrine variations of the Uyuni-Coipasa basin are related not only to regional forcing but also to large-scale climatic changes in the southern tropical zone of the South America. The lake highstand of Tauca III and the lacustrine Coipasa event coincided, respectively, with the appearance of a water body after a dry phase and with expansion of forest in Carajás, both suggesting wetter conditions. The low level during the Ticaña phase in Bolivia likely correlates with the dry event in Salitre at 10,500 ^{14}C yr B.P.

Atmospheric general-circulation models predict drier conditions than today in the southern tropical zones at 12,000 and 9000 ^{14}C yr B.P. due to reduced seasonality in solar forcing (COHMAP, 1988). The high lake levels observed in the southern tropical Andes contradict these results. The curve of lake-level fluctuations of Uyuni-Coipasa basin does not match that of insolation. The climatic variations of the Bolivian Andes cannot be fully explained by astronomical forcings. Thus, other mechanisms must be investigated to explain the positive hydrological changes during late-glacial and early Holocene time.

CONCLUSION

The chronological control of paired ^{14}C and U/Th methods validates the previous attribution of the strong lacustrine Tauca phase in the Uyuni-Coipasa basin to late-glacial (ca. 15,500–13,000 ^{14}C yr B.P.). In contrast, radiocarbon ages of the moderate lacustrine Coipasa phase must be revised. The U/Th dates suggest that this phase occurred in the early Holocene (ca. 9500–8500 ^{14}C yr B.P.). We assume that the inaccurate radiocarbon dates for the Coipasa phase can be explained by a delayed response of the groundwater table during the dry event (Ticaña) between the two lacustrine Tauca and Coipasa phases. Fossil water would have contributed to the feeding of the shallow Coipasa lake.

The lacustrine Tauca III phase seen in Lake Titicaca and the lacustrine Coipasa phase seen in the Atacama Altiplano (where it has been wrongly attributed to the Tauca phase), in the Lipez area, and perhaps in Lake Titicaca, were both significant hydrologic events in the southern tropical Andes, and appear to coincide with wetter conditions in the southern Amazonian lowlands.

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