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Paleowetlands and regional climate change in the central Atacama Desert, northern Chile

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Abstract

Widespread, organic-rich diatomaceous deposits are evidence for formerly wetter times along the margins of the central Atacama Desert, one of the driest places on Earth today. We mapped and dated these paleowetland deposits at three presently waterless locations near Salar de Punta Negra (24.5°S) on the western slope of the Andes. Elevated groundwater levels supported phreatic discharge into wetlands during two periods: 15,900 to ~13,800 and 12,700 to ~9700 cal yr BP. Dense concentrations of lithic artifacts testify to the presence of paleoindians around the wetlands late in the second wet phase (11,000?–9700 cal yr BP). Water tables dropped below the surface before 15,900 and since 8100 cal yr BP, and briefly between ~13,800 and 12,700 cal yr BP. This temporal pattern is repeated, with some slight differences, in rodent middens from the study area, in both paleowetland and rodent midden deposits north and south of the study area, and in lake level fluctuations on the adjacent Bolivian Altiplano. The regional synchroneity of these changes points to a strengthening of the South American Monsoon — which we term the "Central Andean Pluvial Event" — in two distinct intervals (15,900–13,800 and 12,700–9700 cal yr BP), probably induced by steepened SST gradients across the tropical Pacific (i.e., La Niña-like conditions).

Keywords: Atacama; Paleowetlands; Carbon-14; ENSO; La Niña; El Niño

Introduction

Paleoclimate researchers in the central Andes are engaged in a vigorous and important debate on the timing and causes of climate change at low latitudes during the late Quaternary. Some researchers studying ice records (Thompson et al., 1998, 2000) see strong links between global temperature shifts and central Andean climate change. Perhaps a majority of geologists, in this region as well as throughout the tropics and sub-tropics, view summer insolation over land as the chief control on climate variability at low latitudes (Martin et al., 1997; Baker et al., 2001a,b; Wang et al., 2004; Cruz et al., 2005). By contrast, the focus among the South American climate community in recent years has been on the potential role of millennial-scale variability in tropical Pacific sea-surface temperature (SST) gradients driving climate change not just within the Andes, but even globally (Garreaud et al., 2003; Cane, 2005). Although climate modeling suggests a role for seasonal insolation variations directly over the tropical Pacific Ocean (Clement et al., 1999, 2004), seawater temperature and salinity variations in the western tropical Pacific warm pool appear to correlate with Dansgaard/Oeschger cycles over Greenland, with El Niño-like conditions prevailing during stadials and La Niña-like ones enduring during interstadials (Stott et al., 2002). Among other things, this raises questions about the potential role of the North

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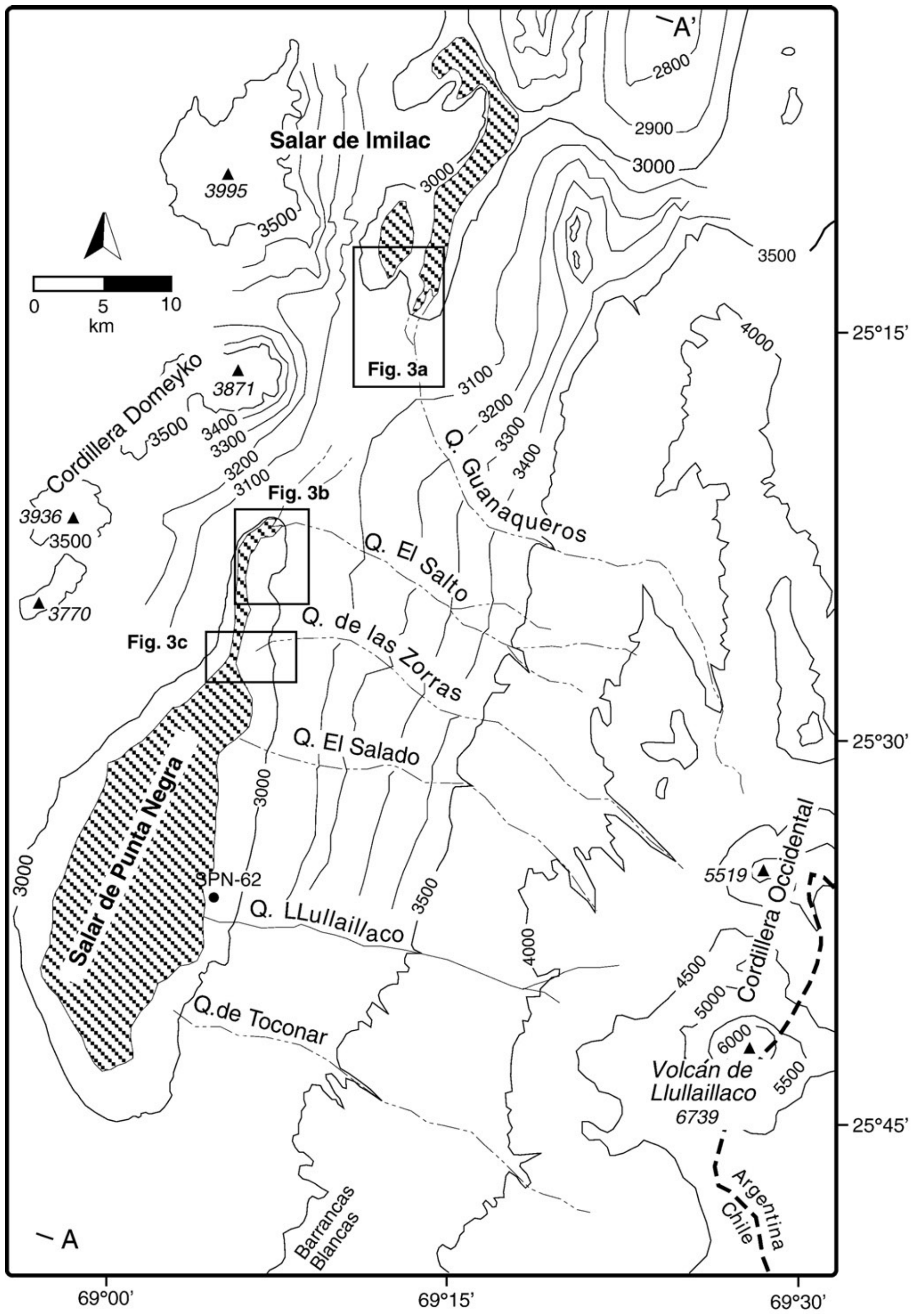

Figure 1. Study area, with locations of detailed maps shown in Figures 3a–c. See Figure 8 for regional location.

Atlantic thermohaline circulation in modulating tropical Pacific variability, and by extension Central Andes precipitation, at various time scales (Dong et al., 2006; Dong and Sutton, 2007; Timmermann et al., 2007).

Resolution of this far-reaching debate will require careful dating and replication of unambiguous paleohydrologic records in the central Andes, and careful comparison with marine and ice core records globally. We are fortunate that, in the Atacama Desert (21–26°S), there is conspicuous evidence of past climate variability in the datable stratigraphy of paleospring/paleowetland deposits that crop out along the western base of the Andes (Rech et al., 2002, 2003; Grosjean et al., 2005; Nester et al., 2007), and in plant records from fossil rodent middens preserved on dry, rocky hillslopes in the interfluves (Betancourt et al., 2000; Latorre et al., 2002, 2005, 2006; Maldonado et al., 2005). Rain today at this latitude is dominantly monsoonal and is carried by the tropical easterlies over and down the Pacific side of the Andes to about 3000 masl. Lying at the waning eastern margin of this system at ~3000 m elevation, our study area today is virtually plantless and waterless. Periodically in the past this area was wetter, as evident from abundant fossil rodent middens and by the scattered diatomaceous badlands, which our group previously studied in the Calama and Salar de Atacama Basins (22–24°S).

In 2001, we commenced work on some exceptionally well-exposed and extensive diatomaceous deposits in other salt basins south of Salar de Atacama, near Salares de Imilac and Punta Negra (Figs. 1, 2). Grosjean et al. (2005) recently published on one cluster of outcrops at the north end of Salar de Punta Negra and found evidence for wet conditions at this location between about 10.2 and 12.6 ka, correlative with a region-wide wet phase termed the "Central Atacama Pluvial Event" — or CAPE — by Latorre et al. (2006). In this paper we propose to (1) expand on the studies of Geyh et al. (1999) and Grosjean et al. (2005) by encompassing more and better exposed paleowetland deposits at three locations in the valley; (2) establish whether the organic-rich diatomites were deposited in shallow wetlands or in deep lakes, as suggested by Lynch (1986); (3) relate our findings to patterns of environmental change across the region, with a view of testing possible links to variations in seasonal insolation over land, in the tropical Pacific SST gradient, and in North Atlantic variability; and briefly (4) examine the relationship between the paleowetland deposits and the dense concentrations of lithic artifacts in the area. In this report, we also include preliminary analyses of a series of middens collected and dated from 3142 to 3400 m in dry canyons, Quebrada de las Zorras and Q. Toconar, which dissect the west slope of Volcán de Llullaillaco (Figs. 1, 2). These middens will be reported in detail elsewhere, but here we compare wet indicators such as grass abundances in middens with the paleowetland stratigraphy.

Setting of deposits

The Atacama Desert is a narrow coastal desert extending from about 16 to 27°S. It is sandwiched between the high (>6 km) Andean volcanoes to the east and the cold Pacific Ocean to the west. Large alluvial fans wash down the western slope of the Andes into a series of north–south oriented structural basins now floored by a string of salt pans (Figs. 1, 2). Notable among these are three salt pans: Salar de Atacama (~2300 masl; 2900 km^2 area), Salar de Imilac (~2970 masl, 10 km^2) and Salar de Punta Negra (~2950 masl, 250 km^2). Despite the linear alignment, there is no evidence now or in the recent past for a surface hydrological connection between the three basins. Our previous studies were focused on the Andean piedmont directly east of Salar de Atacama (Fig. 2; Rech et al., 2002, 2003). The present study was carried out in paleowetlands near Salares de Imilac and Punta Negra.

Moisture carried by the easterlies is intercepted by the Andes, including the towering Volcán de Llullaillaco, at 6739 m purportedly the sixth highest peak in the Andes and fourth highest volcano in the world, which drains into Salar de Punta Negra (Fig. 2). Today, upland recharge nourishes a number of springs along the base of the Andean slope (Fritz et al., 1981, Margaritz et al., 1989). In our study area, these include: (1) fault-controlled discharge such as Imilac Springs (Fig. 3a) and two small springs at Barrancas Blancas near the south end Salar de Punta Negra (Fig. 1); and (2) phreatic seepage along the eastern margin of Salar de Punta Negra (Fig. 2) and the lower reaches of Salar de Imilac, both well removed from the major active faults of the basin.

At the end of the Pleistocene, springs also discharged along the lower slope of the Andes, but discharge appears to have been

Figure 2. Landsat NLT true color oblique view of the study area looking toward the northeast (Image credit: NASA/JPL). For a sense of middle foreground scale, SPN-56 is about 40 km from Volcán de Llullaillaco.

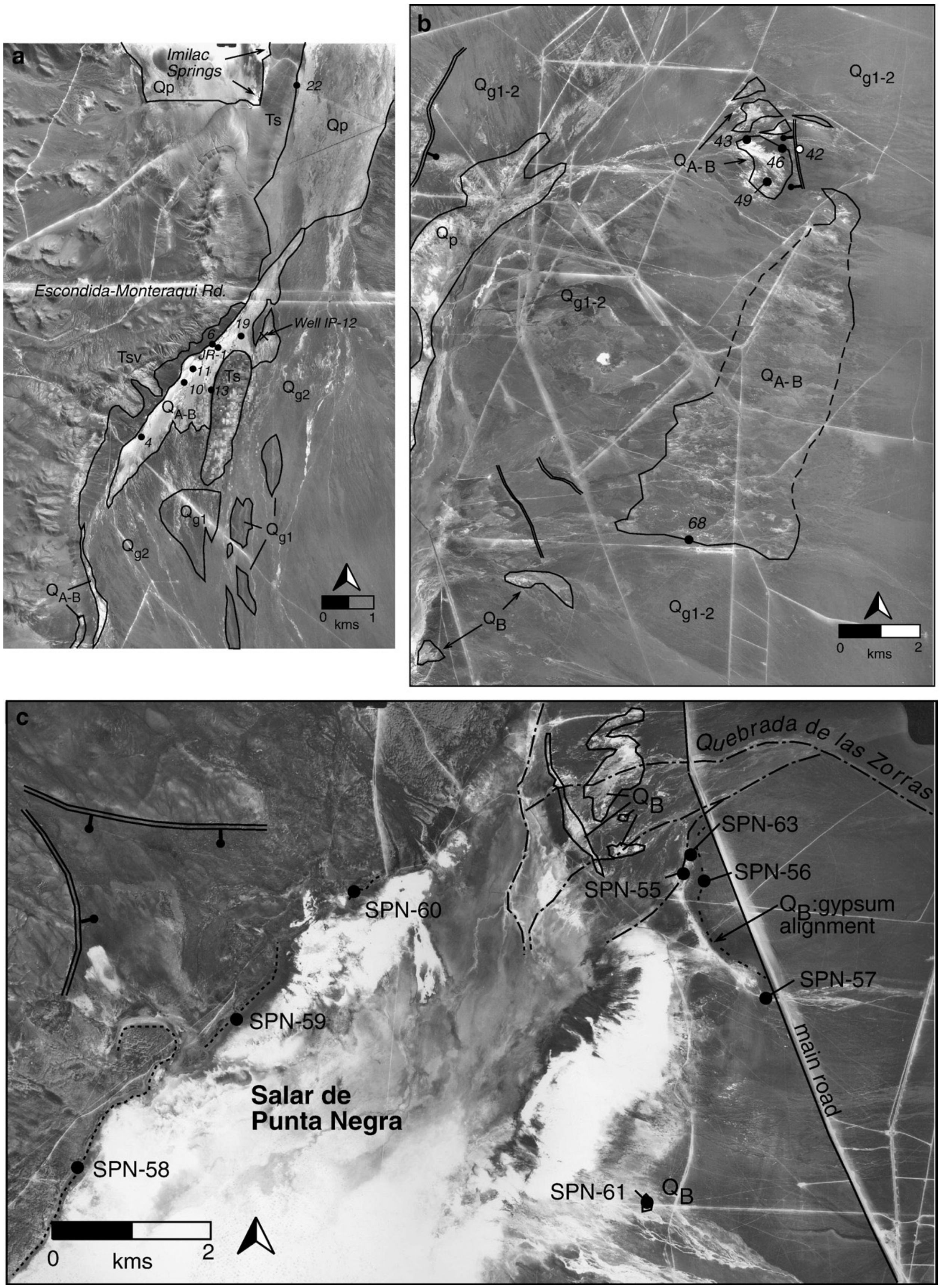
a
Imilac
Springs
Qp
22
Ts
Qp
Escondida-Monteraqui Rd.
19
Well IP-12
6
JR-1
Tsv
11
Ts
10
13
Qg2
QA-B
4
Qg1
Qg1
Qg2
QA-B
0
kms
1
b
Qg1-2
Qg1-2
43
42
QA-B
46
49
Qp
Qg1-2
QA-B
68
QB
Qg1-2
0
kms
2
c
Quebrada de las Zorras
QB
SPN-63
SPN-56
SPN-55
SPN-60
QB:gypsum
alignment
SPN-57
SPN-59
main road
Salar de
Punta Negra
SPN-58
SPN-61
QB
0
kms
2

Table 1
Radiocarbon dates from the study

AA #	Area	Sample #	Fraction [a]	$\delta^{13}C$	fmc [b]	s.d.	^{14}C yr	s.d.	cal yr	2 σ range	Unit
65865	Guanaqueros	Imilac 4-1	A	−29.7	0.1887	0.0016	13398	68	15934	401	B_1
65866	Guanaqueros	Imilac 4-3	A	−24.1	0.2641	0.0019	10694	57	12743	102	B_3
65867	Guanaqueros	Imilac 6-3	B	−23.6	0.2946	0.0025	9817	68	11153	250	B_3
60424	Guanaqueros	Imilac 10-1	A	−35.5	0.2105	0.0031	12520	120	14614	456	B_1
60425	Guanaqueros	Imilac 10-1	B	−28.5	0.2030	0.0016	12809	63	15158	263	B_1
60426	Guanaqueros	Imilac 10 (140–142)	A	−24.2	0.2860	0.0019	10056	53	11637	318	B_3
60427	Guanaqueros	Imilac 10 (140–142)	B	−23.0	0.2872	0.0019	10023	53	11522	237	B_3
60428	Guanaqueros	Imilac 10-2 (59–63)	A	−29.0	0.2084	0.0017	12599	64	14785	357	B_1
60429	Guanaqueros	Imilac 11 (50–53)	A	−29.8	0.2056	0.0016	12706	64	14976	270	B_1
60430	Guanaqueros	Imilac 11 (140–142)	A	−14.4	0.2808	0.0019	10203	54	11876	229	B_3
60431	Guanaqueros	Imilac 11 (140–142)	B	−21.3	0.2858	0.0019	10062	53	11677	279	B_3
60432	Guanaqueros	Imilac 11 (204–209)	A	−24.8	0.3058	0.0019	9518	51	10848	241	B_3
60433	Guanaqueros	Imilac 19-1	A	−30.8	0.1999	0.0016	12931	66	15293	294	B_1
60434	Guanaqueros	Imilac 19-1	B	−28.3	0.2000	0.0017	12930	67	15282	305	B_1
60435	Guanaqueros	Imilac 10 (150–155)	A	−28.9	0.2819	0.0019	10172	54	11841	224	B_3
60436	Guanaqueros	Imilac 13-4	A	−24.6	0.2943	0.0019	9827	52	11258	87	B_3
60448	El Salto	SPN-42-1	A	−25.3	0.2868	0.0019	10033	53	11523	245	hearth
60438	El Salto	SPN-43-1	A	−30.7	0.1985	0.0016	12989	65	15369	312	B_1
65868	El Salto	SPN-43-3	B	−27.2	0.2756	0.0019	10353	55	12277	288	B_3
65869	El Salto	SPN-43-5	B	−26.0	0.2799	0.0020	10229	57	12010	212	B_3
60445	El Salto	SPN-43-8	A	−24.1	0.3059	0.0019	9514	51	10847	240	B_3
60444	El Salto	SPN-46-4	A	−25.8	0.2760	0.0023	10342	68	12213	375	B_3
65870	El Salto	SPN-46-6	B	−26.2	0.2768	0.0019	10318	56	12111	280	B_3
60441	El Salto	SPN-46-9	A	−26.0	0.3383	0.0021	8702	49	9716	172	B_3
60450	El Salto	SPN-49-1	A	−29.8	0.1914	0.0020	13282	83	15766	411	B_1
65871	El Salto	SPN-49-2	B	−24.6	0.2561	0.0019	10943	59	12910	86	B_3?
60437	El Salto	SPN-49-3	A	−24.3	0.2979	0.0019	9729	52	11058	187	B_3
60447	SPN-north	SPN-55-3	A	−25.6	0.3114	0.0020	9371	51	10582	148	B_3
60451	SPN-north	SPN-56A-1	A	−25.5	0.2819	0.0019	10172	53	11840	221	B_3
60442	SPN-north	SPN-56A-2	A	−25.0	0.2990	0.0019	9699	51	11014	217	B_3
60443	SPN-north	SPN-56B-3	A	−24.2	0.3091	0.0020	9432	52	10787	273	B_3
60452	SPN-north	SPN-57-1	A	−23.6	0.3248	0.0032	9032	79	10153	243	B_3
60439	SPN-south	SPN-62-1	B	−25.5	0.2256	0.0017	11963	61	13841	136	B_1
60446	SPN-south	SPN-62-2	A	−23.6	0.3640	0.0079	8520	180	9582	551	B_3
60449	SPN-north	SPN-63-2	A	−24.6	0.3168	0.0020	9235	51	10402	149	B_3
60440	SPN-north	SPN-63-4	A	−25.2	0.3147	0.0020	9287	50	10464	182	B_3
65872	El Salto	SPN-68-1	B	−24.1	0.2844	0.0019	10101	55	11687	288	B_3
65873	El Salto	SPN-68-3	A	−23.4	0.2790	0.0019	10254	55	12053	294	B_3
38746	Guanaqueros	AT-628	B	−22.9	0.2780	0.0028	10,283	80	12123	407	B_3
38747	Guanaqueros	AT-629	B	−23.0	0.2793	0.0020	10,246	58	12035	312	B_3
38748	Guanaqueros	AT-630	A	−22.7	0.3989	0.0024	7,384	49	8100	53	B_3
38749	Guanaqueros	AT-631	A	−22.6	0.3358	0.0023	8,765	56	9836	283	B_3
38750	Guanaqueros	AT-632	B	−23.8	0.2795	0.0018	10,240	53	11990	237	B_3

SPN = Salar de Punta Negra.
[a] A = residue from acid–base treatment; B = humate fraction (base soluble fraction).
[b] fmc — fraction modern carbon.

much more extensive, supporting standing water that produced diatomaceous green muds mixed with organic-rich layers. We studied four areas of isolated diatomaceous badlands in two basins: Quebrada de Guanaqueros (abbreviated QG), El Salto, and Salar de Punta Negra north (SPN-north) and south (SPN-south). All four areas occur on slopes that face westward into the Salar de Punta Negra and Salar de Imilac structural basins (Figs. 1, 2), along drainages that head in Volcán de Llullaillaco and its immediate flanks. Today these areas are virtually plantless and completely waterless in the immediate vicinity of the badlands.

The QG deposits are deeply and continuously exposed over 8–10 km^2 (Fig. 3a) at the toe of the Quebrada de Guanaqueros fan. There are no obvious structural features in the area, and paleodischarge appears to represent simple intersection of the regional water table with the fan toe. The badlands at El Salto

Figure 3. Detailed maps of the (a) Guanaqueros, (b) El Salto, and (c) Salar de Punta Negra north areas. Map locations in Figure 1. Numbers and solid circles refer to stratigraphic logs shown in Figures 5a–c. Map abbreviations: T = Tertiary, Q = Quaternary; s = sedimentary and v = volcanic rocks, p = playa deposits, A = Unit A, B = Unit B; g1, g2 = gravel units 1 and 2. Double lines = faults with dot and bar on the hanging wall.

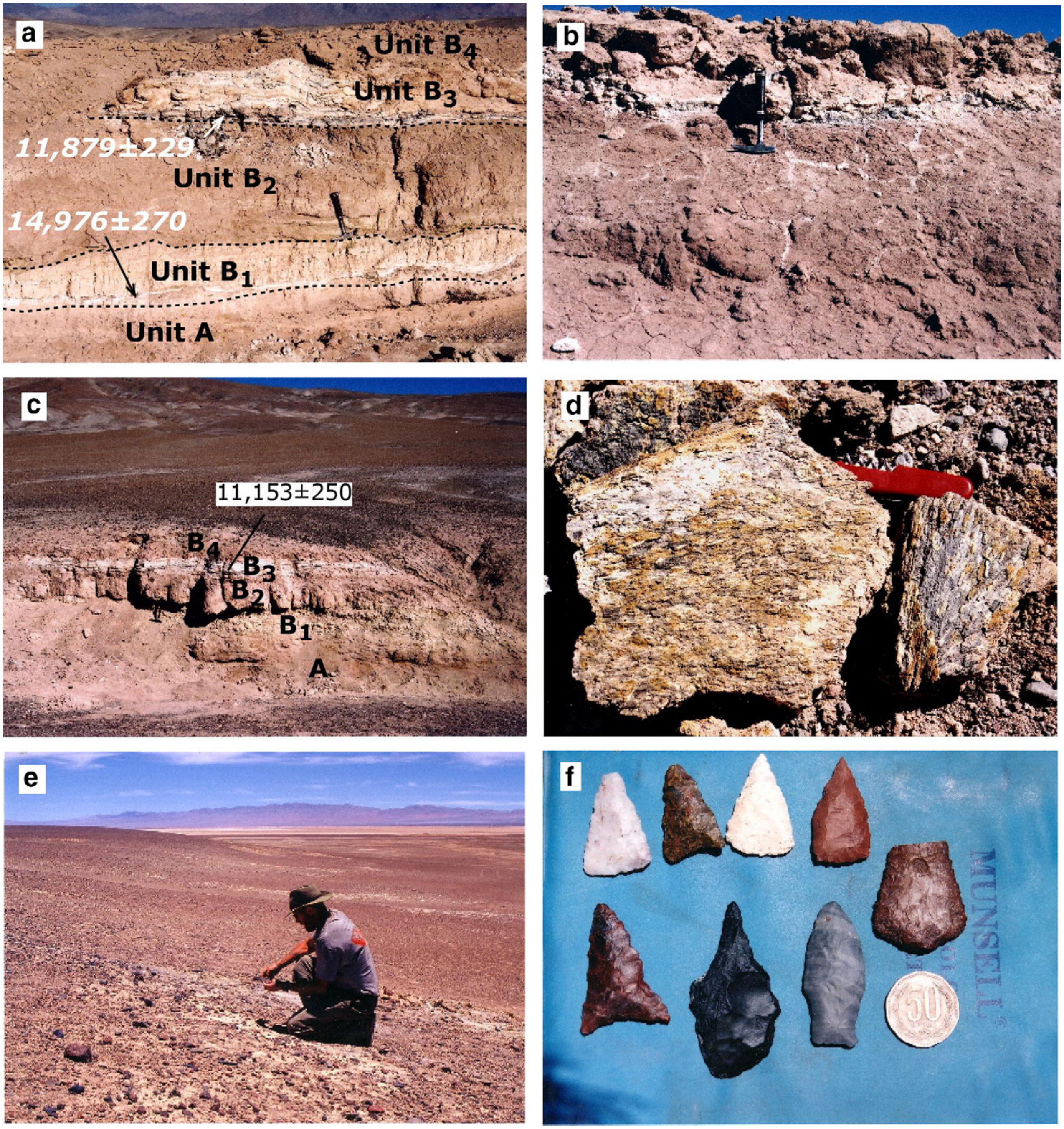

Figure 4. (a) Section Imilac 11 at Quebrada de Guanaqueros. See Figure 3a for location, and Figure 5a for stratigraphic section. Units B_1 and B_2 are composed of diatomite and greenish siltstone deposited in wetlands with locally abundant, dated (in calendar years) organic layers. Unit B_4 is composed of gypsum blocks cementing diatomite and brown silt, representing salt accumulation in a phreatic playa; (b) desiccation cracks extending down into Unit B_2, locally filled with white diatomaceous silt from overlying Unit B_3; (c) Section Imilac 6. Note the complete lack of vegetation in the area today. Near this location, water table is about 27 m below the surface; (d) matted, partially carbonized grasses in Unit B_3, El Salto area; (e) archeological site at Section SPN-56, looking southeast toward Salar de Punta Negra. Layered diatomite and organic matter underlie the surface (see Fig. 5c, sections SPN-56A, B). Note the absence of vegetation at the site; (f) photograph of surface find artifacts colleccted at SPN-56 (Fig. 4e). The triangular points on the left and above are Tuina points (Núñez, 1983). The gray projectile point immediately to left of the coin is a "fish-tail" point, the first one to be identified from the Atacama Desert of Chile. All artifacts are currently housed in the Museo Arqueológico "Padre Le Paige", San Pedro de Atacama, Chile. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

occur in a mid-fan rather than terminal fan position, and their distribution is likely related to a fault that bounds the uphill extent of the deposits (Fig. 3b). At SPN-north and SPN-south, badland outcrops are very dispersed and shallow, and concentrated along fan margins around Salar de Punta Negra (Fig. 3c). These appear to represent paleospring discharge similar to that occurring on the eastern edge of Salar de Punta Negra today, only positioned farther uphill by 8–16 m in elevation.

Laboratory methods

Samples were examined under 10–40× reflected and transmitted light microscopes, and wherever possible, fragments of macroscopic organic matter were hand-picked for ^{14}C dating. In other cases, fragments were too diffuse and small to isolate, and whole-sediment samples were processed. Each sample was subjected to the standard acid–base treatment. In three samples, both the acid and base fractions were submitted for dating. With other samples, either the acid–base residue fraction ("A" fraction, Table 1) or the base soluble fraction ("B" fraction, Table 1) was submitted, depending on whether sufficient carbon survived the base pretreatment. All pretreatments were performed at the Desert Laboratory of the University of Arizona, followed by conversion of samples to graphite. Dating was carried out by the University of Arizona Accelerator Mass Spectrometer Facility. For conversion of carbon-14 dates (expressed as ^{14}C yr BP) to calendar ages (expressed as cal yr BP), we used CALIB 5.01 (IntCal04) (Stuiver and Reimer, 1993; Stuiver et al., 2005) to generate a median and a 2-sigma probability range for each ^{14}C date in calendar years (Table 1).

Grass abundance was measured in middens as point occurrence on a 120-cell rectangular grid overlaid on a sorting tray. A sediment matrix splitter was used to randomly segregate 100 mL of plant debris from each washed and dried. Midden debris was then spread uniformly across a 120-cell rectangular grid (each cell is 6.45 cm^2). The abundance of grass taxa was calculated from the number of cells or "hits" in which grass blades, florets, and seeds were identified, divided by 120.

Results

Dating experiments and observations

Carbon in the fine-grained deposits tends to occur in thin (<1 cm), discrete layers of locally pure carbon (Figs. 4a, d). These layers break along laminae to reveal thin mats of broken to complete plant fragments, some readily identifiable as grasses (Fig. 4d). Such terrestrial plant material will not suffer from ^{14}C reservoir effects and thus is highly desirable for dating. We focused in the field and laboratory on sampling this delicate material. In some layers, however, organic carbon was very fine and it was not possible to isolate large organic fragments for dating. In such cases we cannot preclude the possibility that some non-plant material is included in our samples, material potentially subject to ^{14}C hard-water effects, since groundwater discharge must have fed these depositional systems. Fortunately, we were able to densely sample many outcrops, and at 2 σ no dating inversions were observed, supporting the idea that non-plant organic material, if present in our samples, was not far out of ^{14}C equilibrium with the atmosphere.

A larger concern was the fact that many of our samples dissolved completely during base (2% NaOH) pretreatment. This is not surprising, since partially or completely carbonized organic matter is typically base soluble (Quade et al., 1998). This is a potential problem because this prevents us from isolating primary humic material from secondary humic contaminants (see also Rech et al., 2003). Hyperarid conditions, producing the near complete lack of plant cover in the area since the early Holocene would, however, minimize the potential for contamination by secondary humic acids. To test this we analyzed four A–B pairs (Table 1: Imilac 10-1, Imilac 10 (140–142), and Imilac 11 (140–142); Imilac 19-1) and found that dates from two pairs overlapped at 1 σ, and all four pairs at 2 σ. If representative, this suggests that secondary contamination is not an issue, and that we can assume that the humate fraction of our samples is primary, not secondary.

Stratigraphy

The uniformity of the stratigraphy across our study area allowed us to define five distinct stratigraphic units: A, B_1, B_2, B_3, and B_4 (Fig. 5). Subsequent radiocarbon dating confirmed the stratigraphic correlations made between physically widely separated outcrops in the study area, and to the regional stratigraphic scheme for the late Quaternary of Rech et al. (2002). Unit B is a prominent stratigraphic feature of that scheme, representing widespread late Quaternary paleowetland and paleolake deposition in the Calama–Salar de Atacama basins. In this paper we further subdivide Unit B into sub-units B_1 to B_4. Incision is deepest at Quebrada de Guanaqueros and El Salto, and the complete stratigraphic succession is visible in many outcrops in both areas (Figs. 5a, b). In contrast, incision is much shallower at both SPN-north and SPN-south, and only Units B_3, B_4, and (at a few places) B_2 are exposed (Fig. 5c).

Unit A

Unit A (<70 cm) consists of soft, coarse sand and sub-angular gravel, and ranges in color from pale brown (7.5YR6/3-7/2d) at QG to mottled brown (Fig. 4a) and green (2.5Y7/3d) at El Salto. The base is not exposed, whereas the upper contact with Unit B is sharp and gently undulating. Sedimentary structures are rare and small-scale and fossils are entirely lacking. The upper half of Unit A can be extensively root bioturbated.

No organic matter was found in Unit A, and thus it remains undated. The oldest ages from overlying Unit B_1 show that Unit A must be >15.9 ka.

Unit B_1

Unit B_1 (<70 cm) is composed of massive, pale green (2.5 Y 7/2d) sandy silt (Figs. 4a, c) interbedded with multiple white stringers (<5 cm) of diatomite and diatomaceous silt, and thin partings of partially carbonized plant material. Many horizons also contain uncarbonized, well-preserved plant epidermis and vascular plant bundles. Aquatic bivalves were found in two locations at QG. The overlying contact with Unit B_2 can be sharp or diffuse.

Ten dates from Unit B_1 span from 15,934±401 to 13,841±136 cal yr BP (Table 1). Samples from near the lower contact are stratigraphically unambiguous and replicated in multiple locations with dates ranging from 15,934±401 to 15,158±263 cal yr BP. The upper boundary of Unit B_1 can be diffuse and is generally carbon-poor. Samples Imilac 10-1 and 10-2 yielded the youngest ages at 14,614±456 to 14,785±357 cal yr BP from

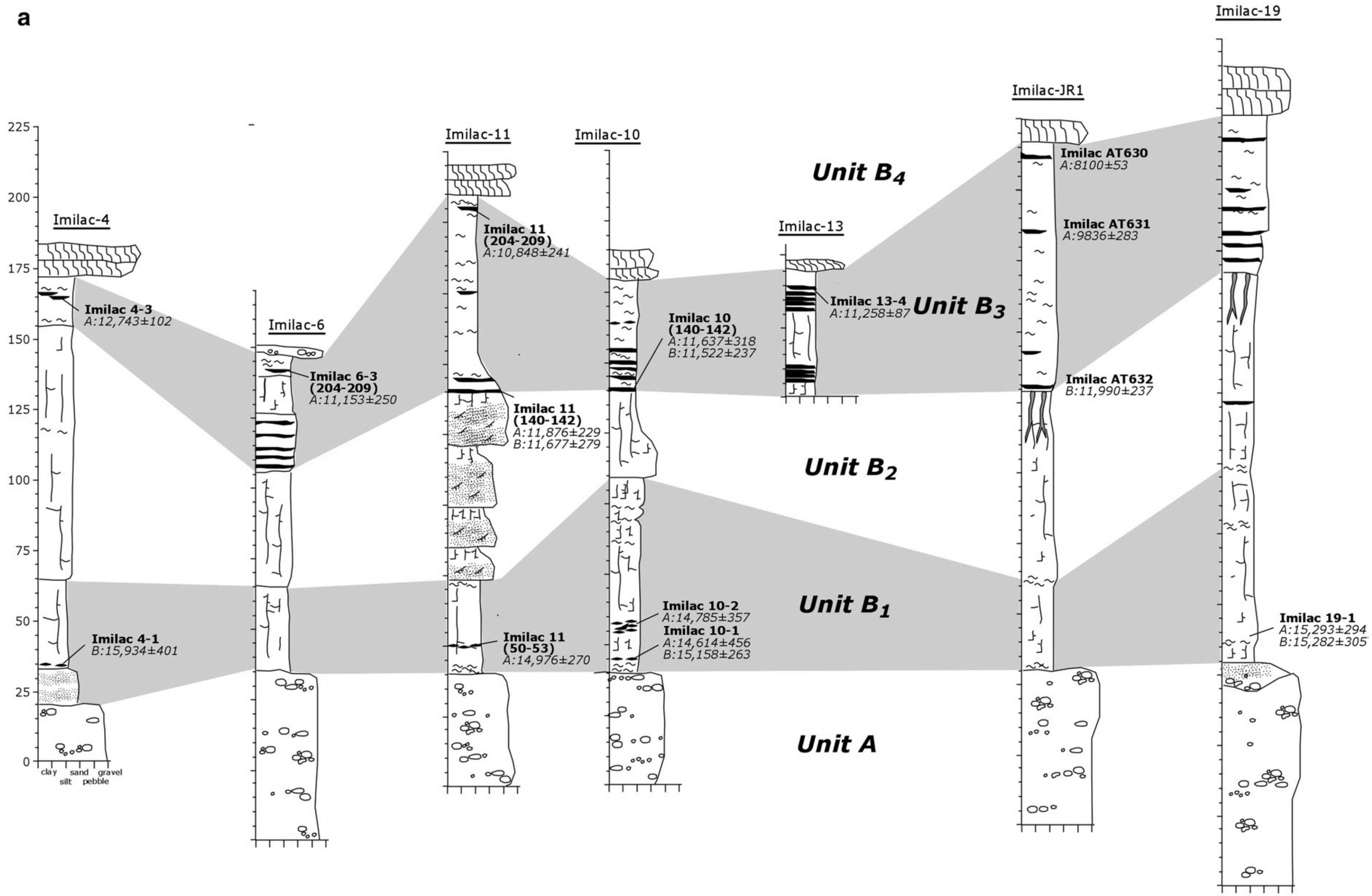

Figure 5. Stratigraphic sections (numbered, locations in Fig. 3) from the (a) Guanaqueros, (b) El Salto, and (c) Salar de Punta Negra north areas. Dates shown are in calendar years BP. Letters preceding dates are A = acid-base residue, B = humate fraction. Legend in panel c.

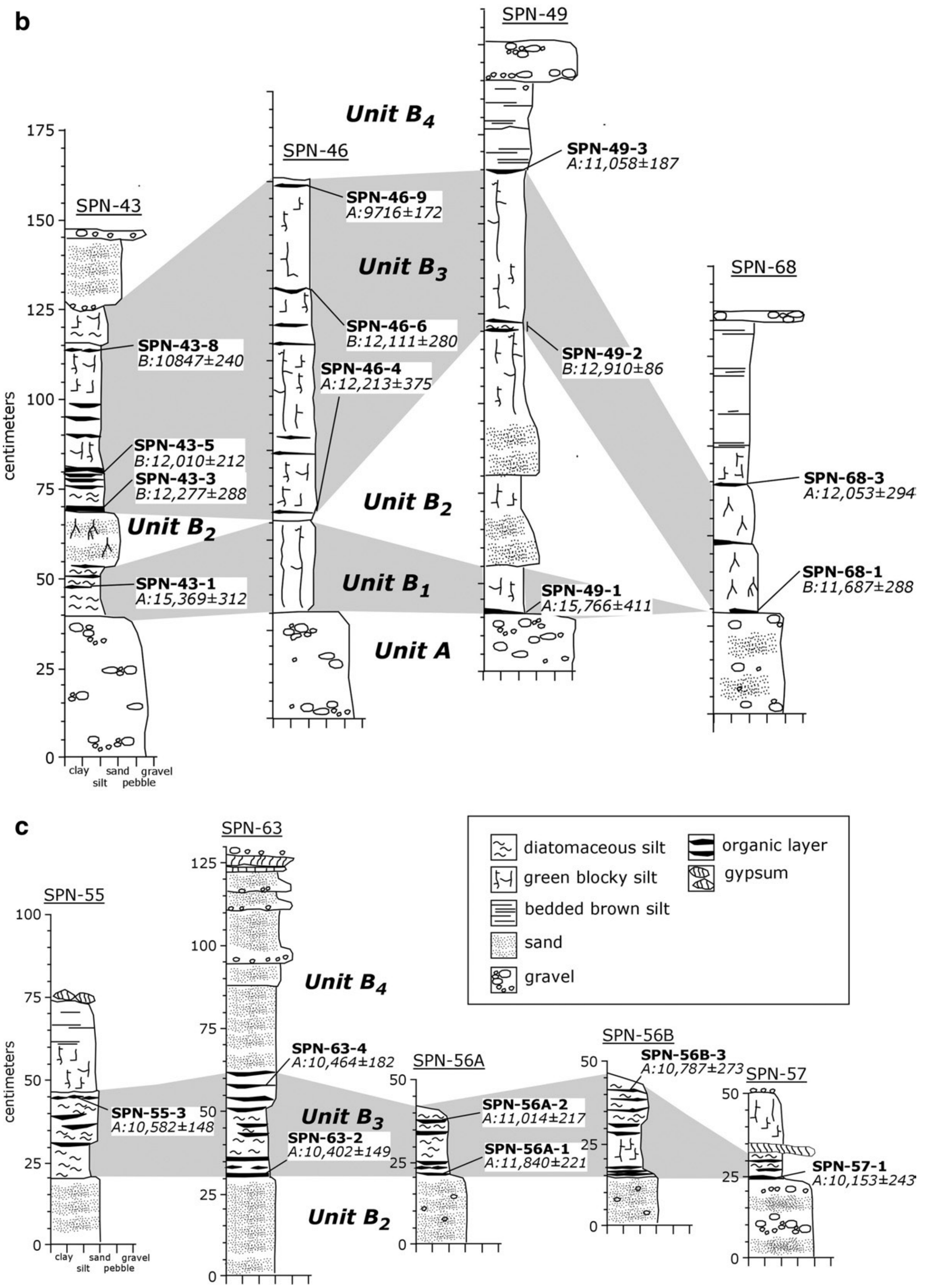

Fig. 5 (*continued*).

what is definitely Unit B_1, but they come from well below the top of the unit (Table 1; Fig. 5a). Sample SPN-62-1 from SPN-south appears to lie at/or near the top of Unit B_1, and it dates to 13,841 ± 136 cal yr BP.

Unit B_2

Unit B_2 (< 175 cm) is mainly pale brown (5YR7/2-7/3d) sandy silt (Fig. 4a) with some well-preserved bedding, alternating laterally with thin (< 1 m) sandy channel deposits, and in a few outcrops, very thin stringers of diatomite. Unlike underlying Unit B_1 or overlying Unit B_3, B_2 is not overprinted with an extensive network of root and rootlet channels. It is very conspicuously cross-cut by deep cracks that reach down from the contact with overlying Unit B_3 (Fig. 4b). Typically, the contact with the diatomite of overlying Unit B_3 is sharp. Unit B_2 is thickest at QG (up to 1.75 m), thinner and locally eroded at El Salto, and generally below the depth of natural exposure at SPN.

With one possible exception at SPN-south, there are no dates from Unit B_2, due to a lack of organic carbon. The top must be older than 11,876±229–12,910±86 cal yr BP (see below), the lower bound of Unit B_3, and the base of Unit B_2 must be younger than 13,841±136 cal yr BP. Thus, Unit B_2 spans much or all of a ~1000–2000 cal yr dating gap in the record.

Unit B_3

At QG, Unit B_3 consists of massive to thinly bedded white silt (Figs. 4a–c), alternating locally with thin to thick mats of plant debris, often grasses. The silts contain variable numbers of diatoms (in a few cases pure diatomite), sponge spicules, and algal testae. Bedding can be quite contorted, implying wet-sediment deformation. The entire lower half of the unit is filled with extensive root channels and plant castes. The plant castes tend to be fine (diameter <1 mm) at QG and El Salto, and coarse (diameter <1 cm) to fine at SPN-north. At El Salto, Unit B_3 is diatomaceous only at the base, and green to brown silt toward the top, locally interbedded with thin mats of plant fragments, diminishing upwards. A few horizons in upper B_3 at El Salto contain Succinids, a semi-aquatic snail.

Dates from the lower 10 cm of Unit B_3 at multiple locations range from 12,910±86 to 11,876±229 cal yr BP. The top of Unit B_3 is poorly defined in many locations, but extends up to 8100±53 at QG (Table 1), 9716±172 at El Salto, 10,153±243 at SPN-north, and 9582±551 cal yr BP at SPN-south. Only at QG does the top of Unit B_3 appear uneroded.

Unit B_4

Unit B_4 has two sharply different expressions in the area. Unit B_{4a} is dominated by coarse prismatic columns of gypsum and lesser carbonate (Fig. 4a); this occurs in QG and locally at SPN-north and SPN-south. Unit B_{4b} is composed of soft brown silt (Fig. 4c), locally well bedded, with minor brown sand, and gravel. The extensive root overprinting typical of Units B_1 and B_3 is conspicuously lacking. Organic matter was not found in Unit B_4 and hence only its lower age limit is defined, 8100±53 at QG (Fig. 5a), and 9716±172 cal yr BP at El Salto (Fig. 5b).

Discussion

Distribution of deposits: lake or shallow wetland?

Lynch (1986) suggested the possibility that the diatomaceous deposits of Salar de Punta Negra were once deposited in a continuous deep lake that occupied the basin in the late Quaternary. Lynch (1990) later revised this view, citing the lack of firm evidence for paleoshorelines. Geyh et al. (1999) and Grosjean et al. (2005) refer to the existence of paleowetlands rather than a paleolake associated with Unit B.

Our evidence firmly supports a shallow paleowetland origin, and not a deep paleolake, for Unit B deposition. The SPN and El Salto sites occupy the SPN hydrographic basin. The El Salto deposits stand at ~3030–3040 m, at least 70 m above the floor of the SPN playa, requiring a deep lake origin if they were lacustrine. The evidence against the existence of a lake of this depth or shallower includes:

(1) a lack of unambiguous shorelines at any elevation, either topographically or sedimentologically. What could be interpreted as possible shorelines from air photos turned out to be either sinuous faults (Fig. 3c) or berms possibly produced by subsidence of salts along playa margins (Fig. 3c; Stations 58–60). Bulldozer cuts through the berms show them to be underlain by angular alluvial fan gravels, with no evidence of shoreline features familiar to us from our recent paleolake research in Bolivia (Placzek et al., 2006), such as rounded, sorted gravels, cross-bedded sands, mollusks, diatoms, ostracodes, and/or tufas.

(2) the lack of diatomite or other fine-grained sediment in the center and west sides of the valley, as viewed in a number of deep pits associated with well digging operations north of the Salar de Punta Negra. Diatomaceous silts are only found along the eastern margin of the valley. This is to be expected of groundwater-fed systems (Quade et al., 1995), since the main recharge areas, the Andes, border the east side of the study area. This asymmetry is also visible with modern phreatic discharge, which is concentrated on the eastern, but not western, margin of Salar de Punta Negra (Fig. 2).

QG is in the Imilac hydrographic basin but several of the same observations apply, including a lack of shoreline features and of diatomites except at QG. Local closure or bottle-necking of fresh water is required to form the paleowetland deposits, but local water depths need not to have been greater than a meter, as suggested by modern analogs discussed below.

The lack of clear evidence of deep (>1 m) paleolakes in our study area is in accord with their absence in other large basins at similar (2900 m) or lower elevations such as Salar de Atacama (2300 m). Our own field and air photo observations suggest that late Quaternary paleolakes in this area of the central Andes were confined to elevations of ≥3500 m, where lower temperatures and greater moisture favored lake development. Nearby examples of basins with clear paleoshorelines include Laguna Miscanti (4120 m; Grosjean et al., 2001), Salar de Pajonales (3540 m) and Salar de Aguas Calientes (3650 m).

Depositional setting and history

We can divide the deposits in the study area into three basic depositional settings (Fig. 6): perennial wetlands, phreatic playas, and ephemeral streams and dry alluvial fans.

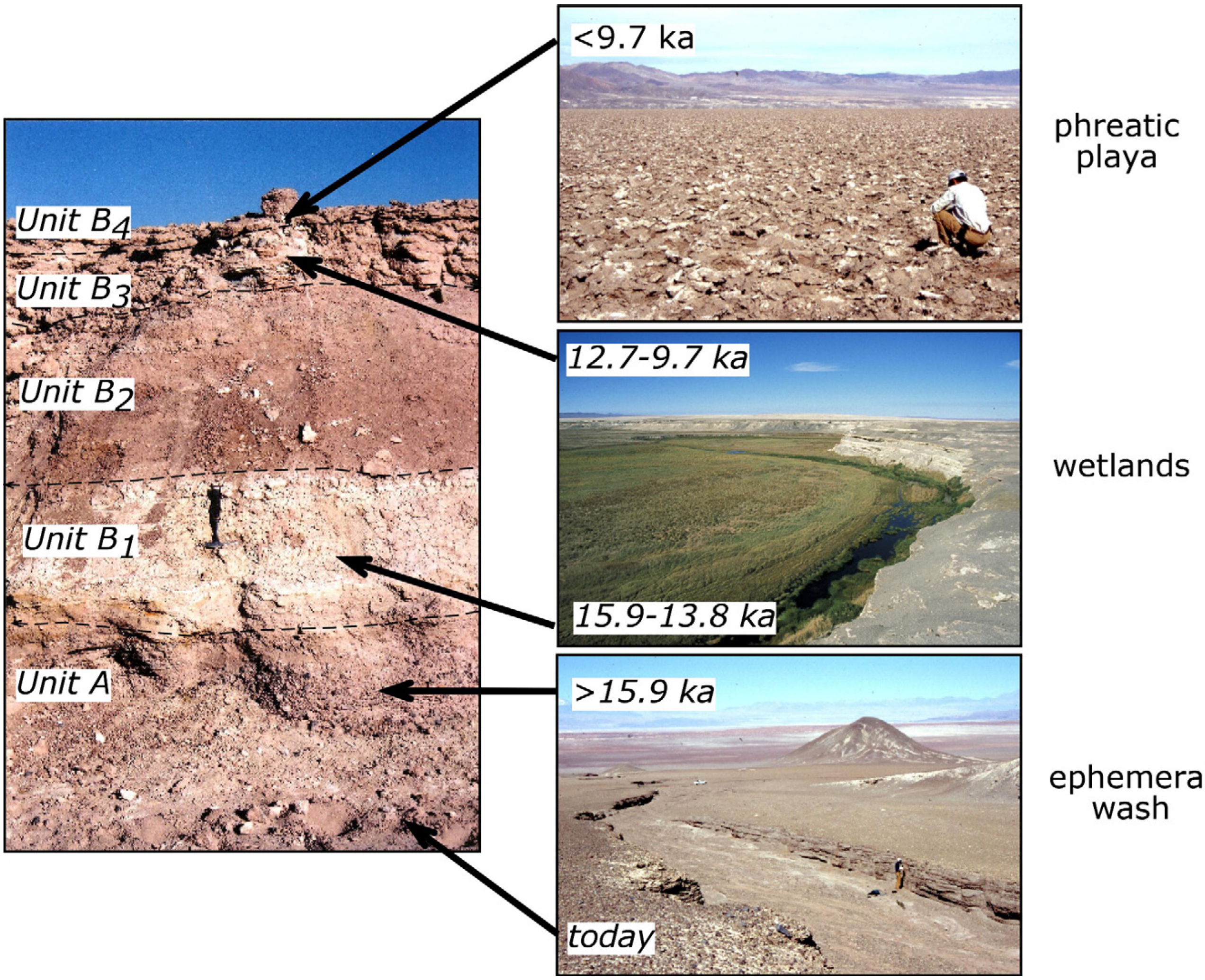

Figure 6. Depositional environments and modern analogs. The stratigraphic Section Imilac 19 (see Fig. 5a) is shown on left, and analog environments on right. The phreatic playa is Salar de Atacama, the wetlands are from the upper Rio Loa north of the study area, and the alluvial wash is on the northwestern edge of Salar de Atacama.

Perennial wetlands

An ensemble of features of Units B_1 and B_3 suggests perennial standing water conditions. Diatoms and sponge spicules are common in all occurrences of Unit B_1 and most of B_3. In addition, organic matter, especially grasses, is abundant, as is a dense overprint of root channels, suggesting that wetlands were extensively vegetated. The presence of Succinids is also relevant. In deserts, these semi-aquatic gastropods live on moist ground marginal to permanent water (Wu, 1993; Pigati et al., 2004). The greenish color of silts points to sustained reduction in locally water-saturated, O_2-deficient conditions.

Many features of Units B_1 and especially B_3 can be seen in the modern and mid-Holocene-age wetlands found north of the study area, along the Rio Loa and Rio Salado (Rech et al., 2002) (Fig. 6). The close association of layered diatomite, extensive mats of plants, especially grasses, and wet-sediment deformation, is reproduced in this area. The wetlands are characterized by a nearly continuous carpet of sedges, grasses and small salt-tolerant shrubs growing on the margins such as *Tessaria* and *Atriplex*. Water depths over most of the Rio Loa wetlands are typically <1 m.

Phreatic playas

Unit B_4 at QG and SPN is characterized by coarse blocky cementation by gypsum and minor carbonate. Gypsum thicknesses exceed 50 cm in places (Fig. 4a) and display vertical or radiating crystal growth. Typically, the upper portion of cemented blocks is pure gypsum, whereas at depth the gypsum overprints white silt of the underlying Unit B_3. This style of blocky cementation by gypsum typifies the surfaces of phreatic playa settings in the region, such as the Salar de Atacama (Fig. 6). Such deposition denotes a very shallow water table (1–2 m deep) in which water and salt are wicked to the surface by evaporation.

Ephemeral stream and dry alluvial fan

Unit A and parts of Unit B_2 are characterized by reddish (5YR7/2-7/3d) to green (2.5Y7/3d) gravels and sands. Gravels are heterolithic and up to 6 cm in diameter, poorly sorted but clast supported. Gravels and sands in Unit A can be Fe-stained and bioturbated by rootlets below the contact with Unit B_1. These deposits are very similar in appearance to ephemeral wash gravels today (Fig. 6). Unit A therefore denotes very dry conditions and a deep water table. Unit B_2 displays a mix of features — a few thin

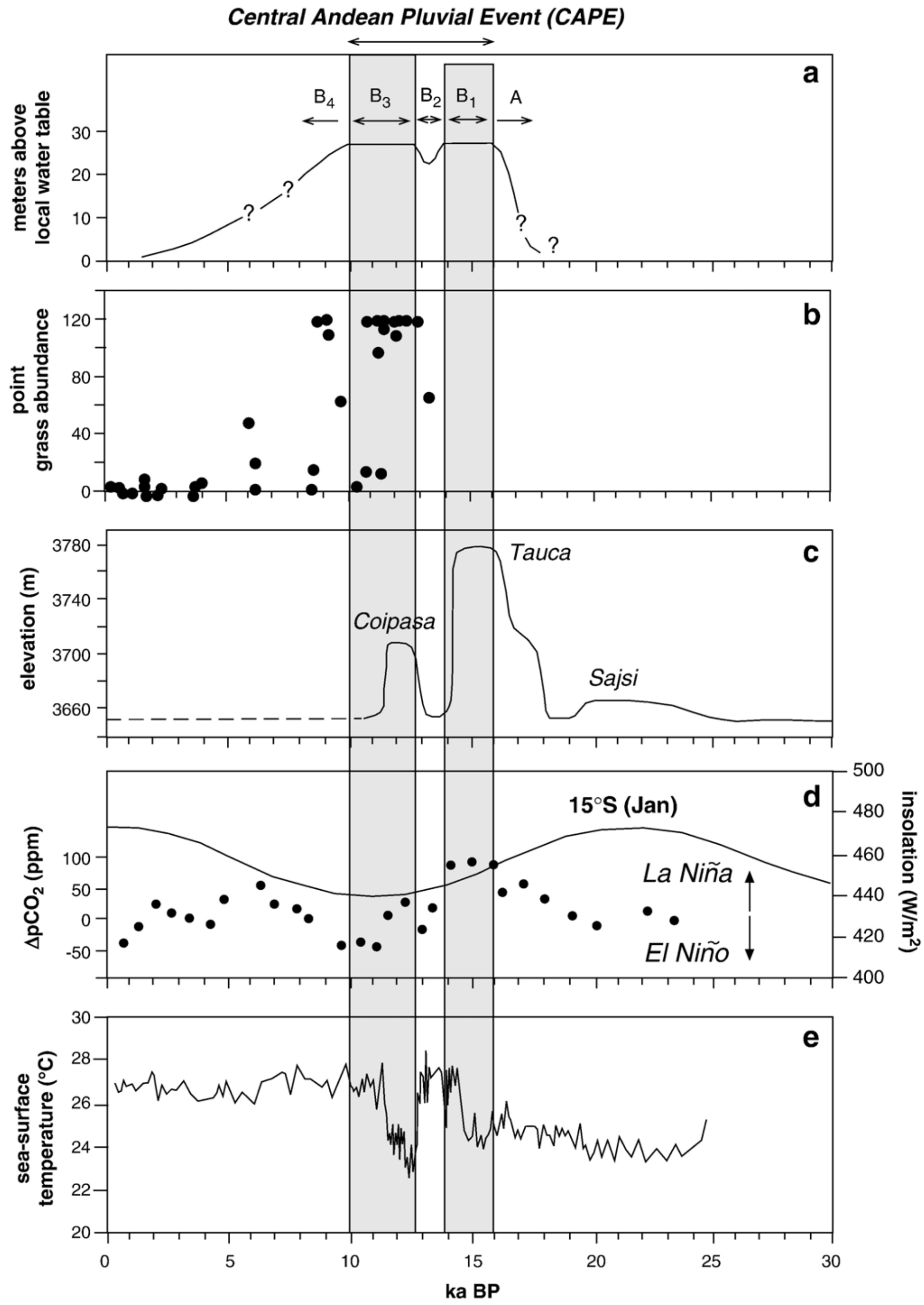

Figure 7. Comparison of paleohydrologic records from the central Andes south of 18°S from: (a) paleowetlands in this study, (b) rodent middens in point count abundance (see Laboratory methods) from the Quebrada de las Zorras and Q. de Toconar (Fig. 2), and (c) the Poopó/Coipasa/Uyuni lake system (Placzek et al., 2006), with lake phases shown. Panel (d) shows inferred ENSO variation from an Equatorial Pacific record (solid dots; Palmer and Pearson, 2003) and January insolation at 15°S (solid line; Berger and Loutre, 1991). Panel (e) shows reconstructed sea-surface temperature for the Cariaco Basin (modified from Lea et al. (2003)).

diatomite stringers interlayered with reddish silt and cross-bedded sand — that would point to very locally wet conditions in a more generally dry, oxidizing fluvial setting.

Wetland history

Formation of the diatomaceous deposits in our record requires a very substantial rise in the water table. Water levels in exploration well IP-12 (Fig. 3a) at QG and two observations at SPN constrain these water level changes. IP-12 lies in the northern part of the QG and depth to water below Unit B is 27 m. IP-12 and other boreholes at QG are exploratory and groundwater in this area has not been pumped (M. Parada, pers. comm., 2002). Thus, 27 m can be taken as the natural change in water table depth since Unit B deposition. In contrast, pumping from extensive well fields may have artificially lowered water levels around El Salto and

SPN, and thus we view any estimate of natural water level changes there with caution. Phreatic salt crusts all along the eastern margin of SPN, formed by shallow groundwater within ~1–2 m of the surface, provide at least one good fix on pre-pumping water levels. Since Unit B time, water levels have dropped by ≥16 m along the northern edge of SPN in the vicinity of SPN-56, and by ≥8 m in southwestern SPN.

We can reconstruct changes in water depth through time by combining these constraints with our observations in the previous section on the relation between water depth and depositional environment (Fig. 7a). During Unit A deposition (>15,900 cal yr BP), the water table was well (>5 m) below the surface, with no evidence of surface discharge or near-surface wicking of salts. During Unit B time, wetlands developed in response to two periods of high water tables, represented by Unit B_1 between 15,900 and ~13,800 cal yr BP and by Unit B_3 between 12,700 and ~9700 cal yr BP. Of the two deposits, B_3 is thicker and more widespread, as seen in the extensive exposures at QG. Unit B_2 divides these paleowetland units and provides evidence of a brief but clear dip in the water table between 13,800 and 12,700 cal yr BP. Minor stringers of diatomite in Unit B_2 suggest very local persistence of surface water at this time, suggesting that the water table drop was no more than a few meters, and very localized surface water remained fresh.

Within Unit B_3 itself, the lower portion is dominated by diatomaceous silt, suggesting standing water. Sometime shortly after 12,100 cal yr BP some of the wetland appears to have dried, but grass mats and root bioturbation persisted through 9700 cal yr BP. All indicators of surface or very near-surface wetness in Unit B_3 had ceased everywhere by 9700 cal yr BP, except at one low elevation site in northern QG, where surface moisture may have persisted to 8100 cal yr BP. This was followed in most areas by deposition of blocky gypsum of Unit B_4 in a phreatic playa setting. The duration of this period of shallow subsurface water tables is unknown, but possibly >1000 yr based on the extent of thick salt accumulation. Sometime in the early to mid-Holocene, water tables dropped below the surface and did not remerge in this area. The modest mid-Holocene rise in water tables reported from sites to the north (Betancourt et al., 2000; Rech et al., 2002, 2003; but see also Grosjean et al., 2003) is not evident in our study area.

Paleoindian occupation

Our geologic survey (conducted by JQ and BQ in July, 2002) led to the discovery of numerous previously undocumented artifact concentrations on the alluvial flats immediately adjacent to all four paleowetland areas. Published archeological surveys of this region (Lynch, 1986, 1990) do not mention these occurrences. The locations, descriptions, and several artifacts, including a fish-tail point, from the many sites at SPN-north were turned over to Lautaro Núñez at the Museum of the Atacama, who went on to excavate SPN-1 in 2003, and report on it in Grosjean et al. (2005). Here we examine the relationship of the archeological sites to the paleowetland deposits in the area.

At QG, artifacts are found on patches of older alluvial surfaces within 10 m of the nearest outcrops of paleowetland deposits. No dates were obtained on these surface finds. At El Salto, many artifacts are scattered over a broad area on alluvial flats upslope from the paleowetland deposits (Fig. 3b). A ^{14}C date from charcoal collected at the surface in a hearth-shaped feature containing burnt camelid teeth returned an age of 11,523±245 cal yr BP (Table 1; Fig. 3b, site 42).

Paleowetlands at SPN are associated with impressively dense concentrations of artifacts (Fig. 4e). The artifacts, all surface finds, included many projectile points. Among them we recognized many triangular points of the Tuina tradition (Fig. 4f), a distinct morphology associated with the first human inhabitants of the Atacama (the early Archaic described by Núñez, 1983; Núñez et al., 2002). We also discovered a single complete "fish-tail" point of the Fell tradition (Fig. 4f), the first of its kind from northern Chile, and a distinctive morphology first found in association with extinct sloth and horse in southern Chile (Bird, 1938; Lynch, 1999). The main artifact concentrations are in the NE quadrant of the basin, along the string of gypsum mounds and underlying diatomite fringing the hillside.

Our dating of deposits in this area (SPN-north) matches very closely those of Geyh et al. (1999) and Grosjean et al. (2005), and confirms that only Units B_3 and B_4 are widely exposed in this area. We dug a series of shallow pits positioned to avoid archeological remains at SPN-56B-3 but expose the local diatomite stratigraphy. The youngest age from these pits nearest the artifacts is 10,787±273 cal yr BP (Fig. 5c; section SPN-56B-3; Table 1). As noted previously, all three former wetland areas — QG, El Salto, and SPN — are waterless and plantless today. The nearest freshwater at Salar de Punta Negra today discharges from a few springs located along the base of the Andes ~25 km away. Such large artifact concentrations only make sense as occupations if water was once locally present. Hence, we view the age of ~9600 cal yr BP from the top of Unit B_3 in the area — and therefore the last time that fresh water stood at the surface — as the minimum age for all of the artifacts. This agrees with the findings from excavations from SPN-56 (named SPN-1 in Grosjean et al., 2005) where dates of 10,320±80 and 12,375±255 cal yr BP bracket the stratigraphic levels containing the artifacts (Grosjean et al., 2005).

Grass abundances from local rodent middens

Forty-one fossil rodent middens were collected and dated from rock crevices and shelters in mostly Tertiary conglomerates within Quebrada de las Zorras and Quebrada de Toconar (Figs. 1, 2) and grass abundances were calculated. Living grasses are rare or nonexistent below ~3500 m elevation in both of these long canyons, and their abundance in middens is interpreted as increased precipitation compared to modern. Modern vegetation at these midden sites is sparse, and represented by scattered individuals of the perennial shrubs *Atriplex imbricata*, *Acantholippia deserticola*, and the annuals *Argylia tomentulosa*, *Cristaria andicola*, *Cistanthes* spp., *Phacelia* sp., and *Hoffmanseggia* sp. No grass grows today at any of the sites. The oldest midden dates to 13,400 cal yr BP, slightly younger than paleowetland Unit B_1 (Fig. 7b), and contains modest grass abundance. Abundances are highest between

12,900 and 11,400 cal yr BP early in the deposition of Unit B_3, and remain variably abundant through 8800 cal yr BP, followed by generally low abundances for the rest of the Holocene. Some of the early Holocene variability in grass abundance could arise from brevity and discontinuity in midden deposition, as well as site differences. Nevertheless, there is a surprisingly good match between variations in these grass abundances on canyon slopes at the foot of Volcán de Llullaillaco and groundwater levels tens of kilometers down the hydrologic gradient in the wetlands near Salares de Punta Negra and Imilac.

Comparison to other central Andean records

Evidence for the late Pleistocene Central Atacama Pluvial Event (CAPE) represented by Unit B comes from a number of other studies in the Atacama, among them Geyh et al. (1999), Rech et al. (2002), Grosjean et al. (2005), and Nester et al. (2007) studying paleowetlands; and Betancourt et al. (2000), Latorre et al. (2002, 2003, 2006) and Maldonado et al. (2005) drawing on fossil plant evidence from rodent middens. Evidence from these studies clearly shows that the Atacama Desert north of our study area (24.5°S) was consistently much wetter than today between about 16,200 and 10,000 cal yr BP. The period preceding CAPE (44–17 ka) was apparently cold and very dry in this area, judging from the lack of wetland deposits of that age, and from the nature of the fossil plants from the very few middens that actually date to that period (Latorre et al., 2002). However, there are some important exceptions to this pattern, such as along the perennial Rio Salado (22°S), a tributary of the Rio Loa, where Latorre et al. (2006) found wetter midden assemblages than today between 17,600 and 16,200 cal yr BP on the canyon slopes.

CAPE is also apparent in records as far south as Quebrada Chaco at 25.5°S (Maldonado et al., 2005). The wetland evidence there is similar to that presented in this paper, including diatomites and organic mats above current active spring elevations. The midden evidence includes low-elevation invasion by extra-local plants, especially by steppe grasses, which are found ~900 m below their current locations above ~3700 m. Lack of rainfall, not temperature, constrains the lower elevation limit of these plants today. And at Quebrada Chaco (25.5°S), midden evidence indicates wetter assemblages at >52,000 ^{14}C yr BP, 40,000–33,000 ^{14}C yr BP, 24,000–17,000, 17,000–14,000, 14,000–11,000 cal yr BP. The first three wet periods were interpreted to mean more winter precipitation, the period between 17,000 and 14,000 cal yr BP more precipitation in both seasons, and the period 14,000–11,000 cal yr BP more summer precipitation. Quebrada Chaco receives most of its presentation in winter, and Maldonado et al. (2005) suggest that, for the most part, winter and summer pluvials tend to be asynchronous in the central and southern Atacama. Winter rainfall may explain wetter events preceding CAPE at Quebrada Chaco, but the Rio Salado midden record is more perplexing and awaits regional replication on dry hillslopes far removed from a perennial stream in the central Atacama.

Two other notable features of the paleowetland deposits in our study area are (1) the clear break in wetland deposition represented by Unit B_2 between 13,800 and 12,700 cal yr BP, and (2) greater wetland extent during B_3 than B_1 deposition. Both these features are also visible in the midden records (e.g., Latorre et al., 2002). The recent midden study from Rio Salado (Latorre et al., 2006) indicates a clear break in CAPE at about 14,200 cal yr BP. This may correlate with the base of B_2, which is not as tightly constrained at 13,800 as is the top at 12,700 cal yr BP. Midden records from the margins of the Salar de Atacama basin display elevated grass-count percentages (Latorre et al., 2002) from 16,000 to 10,000 cal yr BP and five-fold increases in species richness when compared to modern floras along this hyperarid margin. Climate conditions obtained from these midden records indicate that it was clearly wettest between about 11,800 and 10,500 cal yr BP, overlapping considerably with Unit B_3 deposition. Comparison of the paleowetland and paleofloral records in the Atacama to paleolake records in the nearby Altiplano shows broad synchrony despite geochronologic uncertainties. A key advantage of the paleowetland and especially midden records is the use of terrestrial plants for accurate ^{14}C dating. Paleolake records from Salar de Atacama (26–15.9 ka; Bobst et al., 2001), Laguna Lejia (Geyh et al., 1999; 13,400–11,000 cal yr BP), Laguna Tuyajto (Geyh et al., 1999; 13,800 to 9100 cal yr BP), and other lakes all show evidence of expanded latest Pleistocene lakes. However, large uncertainties on U-Th isochron dates from salt at the Salar de Atacama (±3 to ±6 ka) and ^{14}C reservoir effects (1200–10,700 ^{14}C yr; Geyh et al., 1999) for Laguna Lejia, Miscanti, and most other lake records — all providing maximum ages on CAPE — make detailed comparisons to the wetland and midden records difficult. However, the overall duration of CAPE, starting between 15,400 and 13,800 and ending between 8400 and 8000 cal yr BP, estimated from lake studies summarized in Geyh et al. (1999) clearly overlaps the ages of Unit B (15,900 to 9700 cal yr BP). These studies also suggest that lakes were deepest late in this interval, between 12,900 and 10,400 cal yr BP, matching our view from the midden and wetland records that B_3 (12,700 to 9700) was a larger discharge event than B_1 (15,900 to 13,800 cal yr BP). Both the Laguna Miscanti (Grosjean et al., 2001) and Lejia (Grosjean, 1994) records indicate a short drop in lake levels during CAPE (equivalent to Unit B_2?), although the dating of this event is uncertain.

Coupled ^{14}C and U-Th dates from shoreline carbonates from the Poopó/Coipasa/Uyuni paleolake system provide a final key comparison to lake level fluctuations south of 18°S (Geyh et al., 1999; Sylvestre et al., 1999; Placzek et al., 2006). This well-dated central Andean lake record shows a virtually identical temporal pattern to our central Atacama wetland records (Fig. 7c). Hence, CAPE should actually refer to the Central Andes rather than just Central Atacama Pluvial Event. Deep lakes formed in this basin between 16.4 and 14.1 ka (Tauca Phase; 3780 m shoreline), then again between 13 and 11 ka (Coipasa Phase; 3720 m shoreline), divided by a period of much lower lake 14.1 to 13 ka (Placzek et al., 2006). Interestingly, the relative magnitude of these events is reversed compared to the central (south of ~23°S) Atacama wetland, midden, and lake records. Unit B_1, which correlates with the deeper Tauca lake cycle, is a smaller event than B_3, which correlates with the shallower Coipasa phase lake. Similarly, midden records from

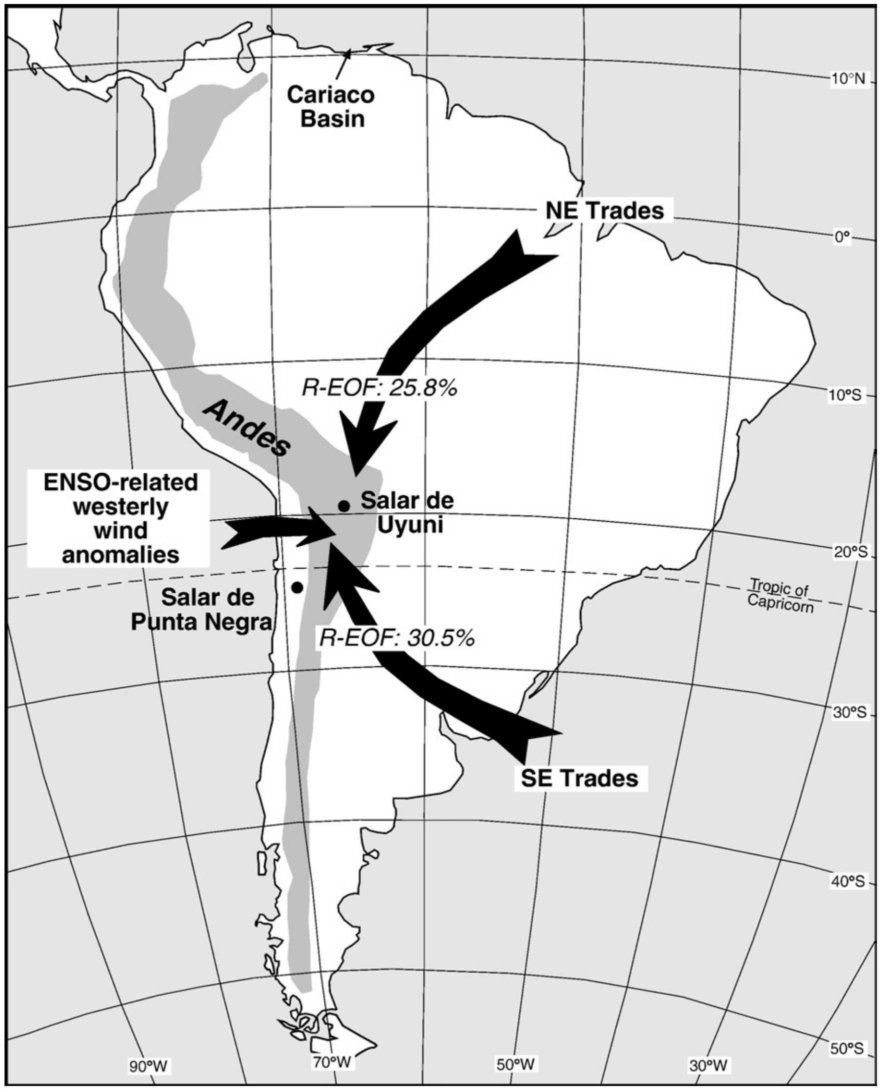

Figure 8. Modern weather systems controlling central Andean rainfall. REOF refers to the rotated empirical orthogonal function, which decomposes (shown in %) the regional precipitation components (from Vuille and Keimig, 2004). Monsoonal rainfall in the central Andes has two modes, as shown, one from the N–NW and the other from the SE. The N–NW mode, but not the SE mode, is modulated by ENSO. Strong westerly winds associated with El Niño disrupt the normal flow of moist N–NW tropical air, producing dry conditions, especially on the northern Altiplano. During La Niña years, N–NW air flow is strengthened and the Altiplano is wetter. The SE mode that dominates monsoonal rainfall in the southern Altiplano and Atacama is not correlated with by ENSO variation but rather with humidity in lowland southeastern South America.

the nearby northern Atacama at Rio Salado (22°S) also seem to show that B_1 is a wetter event than B_3 (Latorre et al., 2006). This latitudinal contrast in the relative intensity of B_1 and B_3 may have important implications for causality, which we examine in the final section.

Patterns and causes of regional climate change

Reductions in temperature and/or increases on rainfall must have produced the region-wide CAPE that we observe. Of the two causes, plant redistributions (Latorre et al., 2006) and lake-budget (Grosjean, 1994) and glacial ice modeling (Kull et al., 2002) show that a rainfall increase of up to 100%, not temperature decrease, was the main cause of CAPE in the region. Temperature, however, likely played some secondary role in increasing effective surface moisture, as global temperature between 16 and 10 ka was lower than today.

Patterns of inter- and intra-annual variability of modern rainfall in the region may provide key clues as to the causes of CAPE. The rains in the region are dominantly convective in nature and come mainly in the summertime (>80% DJFM), the fraction of summer rain gradually diminishing southward. This rainfall represents the distal end of the South American Summer Monsoon system, or SASM, a system traditionally viewed as sourced in the Amazon lowlands to the northeast of the central Andes (Lenters and Cook, 1997). Vuille and Keimig (2004), using cloud-coverage proxies for rainfall amount, identify not one, but two distinct modes of summer rainfall in the region (Fig. 8). One mode is sourced from the north–northwest and tends to dominate rainfall in that direction within the central

Andes, toward Salar de Uyuni. Rainfall from the N–NW falls off sharply to the southwest, especially along the western slopes of the Andes. The second mode is derived from the southeast and dominates rainfall in the direction of Salar de Punta Negra. The two modes are strongly anti-phased in most years, producing much less spatial coherence in summer rainfall variability on the Altiplano and Atacama than suggested by previous studies. The N–NW mode accounts for 25.8% of the rainfall variance in the Andes/Atacama extending from 15 °S to 30 °S, and SE mode 30.5% (Fig. 8).

What controls the strength of the two modes — N–NW and SE — of the summer monsoon system today? For both modes, the seasonal cycle of wet summers and dry winters are controlled by seasonal expansion of the equatorial easterlies in the upper troposphere; however, controls on the two modes (that is, on upper air zonal wind strengths) appear to differ. Much has been written recently about the importance of westerly winds associated with the El Niño (warm) phase of ENSO in producing dry years on the Altiplano, by disruption of the normal flow of the moist tropical easterlies (Vuille, 1999; Garreaud, 1999; Vuille et al., 2003). Vuille and Keimig (2004) confirm the importance of ENSO variation in modulating the N–NW mode of rainfall. They also found that humidity levels in source areas to the north and east was not important in determining the intensity of N–NW mode rainfall. By contrast, the main determinant of precipitation intensity of the SE mode was not ENSO variation but humidity levels in the source area of SE South America, a region where humidity varies much more than in Amazon source areas (to the N–NW mode) to the north. Hence, assuming the controls on modern interannual rainfall variability applied to the past, we might attribute the wet phase represented by the Tauca high lake stand in Bolivia and Unit B_1 in our study area to ENSO variations in the Pacific. This connection has already been made by Placzek et al. (2006), who point out that the Tauca highstand correlates with sustained La Niña-like conditions in the Pacific, as documented by Palmer and Pearson (2003) (Fig. 7d). La Niñas (El Niños) in the northern Altiplano today are correlated with wet (dry) summers.

Following this reasoning, the Coipasa phase of the Bolivian lake history, as well as the very wet conditions associated with Unit B_3, should not coincide with sustained La Niña-like conditions in the Pacific, and in fact there is little evidence for this (Palmer and Pearson, 2003). The Coipasa lake phase and Unit B_4 should be connected to changes in elevated humidity levels in southeastern South America. This idea is testable by study of deposits in that region for evidence of enhanced moisture coincident in age with Unit B_3.

The preceding interpretation is a strongly non-traditional view among geologists of what controls rainfall in tropical and sub-tropical South America. The much more widely held view is that variations in insolation over land, not ENSO or lowland humidity, control the large climate swings in the region (Martin et al., 1997; Baker et al., 2001a,b; Wang et al., 2004; Cruz et al., 2005). In this view, mechanical and sensible heating of the Andes in the summertime, which is increased during periods of high insolation, strengthens monsoonal rainfall in the central Andes.

The two views need not be mutually exclusive, however, since insolation variation may play some role extra-regionally in modulating ENSO (Clement et al., 1999, 2004). And on longer timescales and across many pluvial events, some well-dated South American climate records do show a precessional periodicity of 23 ka (e.g. Lamy et al., 1998; Wang et al., 2004). But insolation variation simply fails to predict the wet phases that we see in our late Pleistocene record. CAPE (15,900–9700 cal yr BP) falls between the insolation maximum (January; 15°S) at 21 ka and overlaps the minimum at 10 ka (Fig. 7d). During the insolation maximum at 21 ka, the paleolake in the Uyuni basins (Sajsi Lake Phase; Fig. 7d) was very shallow (Sylvestre et al., 1999; Placzek et al., 2006) and even dropped after 19 ka prior to the sharp rise in lake level thereafter during the Tauca lake phase. Today, insolation is near its maximum value, but conditions remain very dry (Fig. 7d). Hence, causes other than summer heating of the Andes must be sought to explain CAPE.

For an alternative view, Denton et al. (2006) refer to the period 17.5–14.6 ka represented by CAPE as the "Mystery Interval", in which, among other features, lake levels were high in both western North and now western South America. They suggest possible links between high lakestands and cold sea-surface temperature in the North Atlantic. This is supported by Broccoli et al. (2006) who modeled a connection between cooling in the north Atlantic and southward shift of the ITCZ from its current average summer position closer the equator. This shift could enhance rainfall in the central Andes but dry northernmost South America. Records from the Cariaco Basin at the northern edge of South America appear to support this connection for the late glacial period, showing dry conditions during the Younger Dryas (= Coipasa Lake phase) and pre-Bölling-Allerod (= Tauca Lake Phase) (Fig. 7e; Lea et al., 2003; Hughen et al., 2004).

This can only be a partial explanation for the patterns we observe in the Andes, since during the glacial maximum conditions were also cold in the North Atlantic, yet lakes were very low on the central Altiplano. Also, under current boundary conditions, North Atlantic cooling and the corresponding southward shift of the ITCZ tend to favor both increasing frequency of warm ENSO events, while weakening the annual phasing in the tropical Pacific, in freshwater hosing experiments with climate models (see Dong et al., 2006; Dong and Sutton, 2007; Timmermann et al., 2007 for explanation). This would actually tend to decrease, not increase, precipitation in the central Andes. Although the chronological limits of CAPE are now well defined from multiple archives, the causes of the Mystery Interval in South America and elsewhere remain to be firmly established.

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