CHAPTER

6

Abrupt Climate Change and Pre-Columbian Cultural Collapse

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Abstract

Holocene climate changes in the circum-Caribbean and Andean Altiplano are inferred by using paleolimnological methods. Paleoenvironmental data provide a climatic context in which the Maya (Yucatán Peninsula) and Tiwanaku (Bolivian–Peruvian Altiplano) cultures arose, persisted, and collapsed prior to European contact, ca. A.D. 1500. In the circum-Caribbean, the arid late Pleistocene period (>10,500 ^{14}C B.P., >10,470 B.C.) was followed by a relatively moist early to middle Holocene period (9000–4000 ^{14}C B.P., 8030–2490 B.C.), probably related to large differences between summer and winter insolation. The earliest Maya settlement dates to the Middle Preclassic period (1000–300 B.C.) and was associated with reduced seasonality and regional drying. In the northern part of the Yucatán Peninsula, the climate became even drier during the Classic period (A.D. 250–850). The driest episode of the middle to late Holocene occurred in the Maya lowlands at ca. A.D. 800–1000 and coincided with the Maya collapse, ca. A.D. 850.

In contrast to the circum-Caribbean area, the Andean Altiplano was relatively wet in the late Pleistocene period and experienced low seasonality and dry conditions in the early and middle Holocene. The southern basin of Lake Titicaca (Lago Wiñaymarka), currently >40 m deep, displayed a low stage between ca. 7700 and 3600 ^{14}C B.P. (6470–1930 B.C.). Chiripa culture developed in the Titicaca watershed ca. 1500 B.C. (3210 ^{14}C B.P.), and was associated with greater seasonality, increased moisture availability, and rising lake level. Tiwanaku culture emerged ca. 400 B.C. (2400 ^{14}C B.P.) and depended on raised-field agricultural technology from A.D. 600 to 1150. A prolonged dry period began in the Altiplano at ca. A.D. 1100, prompting abandonment of raised fields and cultural decline.

Climate changes in the Northern Hemisphere Maya lowlands and the Southern Hemisphere Andean Altiplano were out of phase on millennial timescales, when climate was apparently forced by shifts in seasonal insolation driven by the precession of the Earth's orbit (Milankovitch forcing). Shorter frequency climate changes in the Maya and Tiwanaku regions during the last ~3000 years may have been in phase and were driven by factors other than Milankovitch forcing. In both areas, population growth and cultural development occurred under favorable conditions for agriculture. Rapid cultural collapses in both regions were associated with protracted droughts. Paleoenvironmental data indicate that cultural development is limited by climatic thresholds and that abrupt, unpredictable climate changes can disrupt agricultural production and have devastating consequences for human populations.

Resumen

Cambios climáticos durante el Holoceno fueron inferidos usando métodos paleolimnológicos en las regiones Caribeñas y el Altiplano Andino. Datos paleoambientales proveen un esquema climático en la cual culturas Maya (Peninsula Yucatán) y Tiwanaku (Altiplano de Bolivia/Peru) surgieron, persistieron, y se desintegraron antes de contacto con los Europeos, ca. D.C. 1500. En la región del Caribe, el período del Pleistoceno tardío (>10,500 ^{14}C años A.P., >10,470 A.C.) fue caracterizado por un clima árido. El clima del Holoceno temprano y medio (9000–4000 ^{14}C años A.P., 8030–2490 A.C.) era mas húmedo, probablemente como consecuencia de amplias diferencias entre la insolación veranal e invernal. La colonización de la región por los Mayas empezó en el Preclásico Medio (1000–300 A.C.), asociado con condiciones mas secas y una reducción en la diferencia estacional. En el norte de la Peninsula Yucatán, el clima fue mas seco durante el Período Clásico (D.C. 250–850). En la región Maya, la época mas seca del Holoceno medio y tardío ocurrío cerca D.C. 800–1000 y coincidío con el colapso de la cultura Maya ~D.C. 850.

Los registros de la región Caribeña demuestran diferencias paleoclimáticas con las condiciones en el Altiplano Andino. El Altiplano fue relativamente húmedo durante el Pleistoceno tardío. Durante el Holoceno medio y tardío, el clima fue seco y diferencias estacionales no fueron muy pronunciadas. La cuenca sur del Lago Titicaca (Lago Wiñaymarka), que en el actual tiene una profundidad >40 m, tenía un nivel muy bajo entre cerca 7700 y 3900 ^{14}C años A.P. (6470–1930 A.C.). La cultura Chiripa se desarrolló en la cuenca de Titicaca ~1500 A.C. (3210 ^{14}C años A.P.), asociado con condiciones mas estacionales, un incremento en precipitación, y una subida en el nivel del lago. La cultura Tiwanaku emergío ~400 A.C. (2400 ^{14}C años A.P.) y dependío de una agricultura intensiva (campos elevados o camellones) entre D.C. 600 y 1150. Una sequía prolongada comenzó en el Altiplano alrededor de D.C. 1100, y incitó el abandonamiento de los camellones y el decaimiento de la cultura.

Cambios climáticos en regiones bajas de los Mayas (Hemisferio Norte) y en el Altiplano Andino (Hemisferio Sur) estaban fuera de fase sobre escalas de tiempo milenial, mientras que el clima era controlado por los cambios en insolación estacional, como consecuencia de precesión en la órbita del globo (ciclos Milankovitch). Durante los ultimos ~3000 años pasados, inferencias sobre cambios climáticos en las regiones Maya y Tiwanaku se aparecen mas o menos en fase y fueron causado por factores diferente a los ciclos Milankovitch. En ambas regiones, el crecimiento de poblaciones y el desarrollo cultural ocurrieron durante un tiempo cuando las condiciones climáticas eran favorables para la agricultura. El decaimiento en ambas culturas ocurrió rapidamente y fue asociado con sequías prolongadas. Datos paleoambientales indican que el desarrollo de una cultura es limitado por el clima. Cambios climáticos bruscos y no predecibles pueden alterar la producción agricola y tener graves consecuencias en las poblaciones humanas.

6.1. INTRODUCTION

6.1.1. Environmental Determinism in Pre-Columbian America

The concept of *environmental determinism* was debated in the anthropological and archaeological literature during the 1950s and 1960s. The fundamental tenet of the hypothesis posits that regional agricultural potential limits cultural development. Meggers (1954) concluded that the Maya lowlands of southern México and Petén, Guatemala (Fig. 1) have limited agricultural potential today. She suggested that the region would not have been conducive to the cultural florescence so evident in the archaeological record. A logical conclusion of the deterministic theory was that lowland Classic Maya culture (A.D. 250–850), with its monumental architecture, art, hieroglyphics, concept of zero, corbeled arches, calendrical system, and stela cult, did not develop *in situ*, but was imported from elsewhere and was destined to decline in its new geographic context. The simple theory apparently accounted for both the origin of the lowland Maya and their mysterious collapse in the ninth century A.D.

Meggers's (1954) claims were contested by those who argued that there was no archaeological basis for claiming that Classic Maya culture was imported from the highlands. Coe (1957) noted that several lowland Maya achievements, such as the corbeled arch, the Long Count calendar, and the stela cult, developed in the lowlands. Excavations in the lowlands yielded Early Preclassic (2000–1000 B.C.), *Formative phase* ceramics, supporting the claim that lowland Classic Maya cultural ontogeny had its roots in the low-elevation, tropical karst environment (Altschuler, 1958; Hammond et al., 1977, 1979). Today, it is generally acknowledged that the origins of lowland Maya civilization began in the Early Preclassic period, were complex, and involved interactions with peripheral coastal and high-

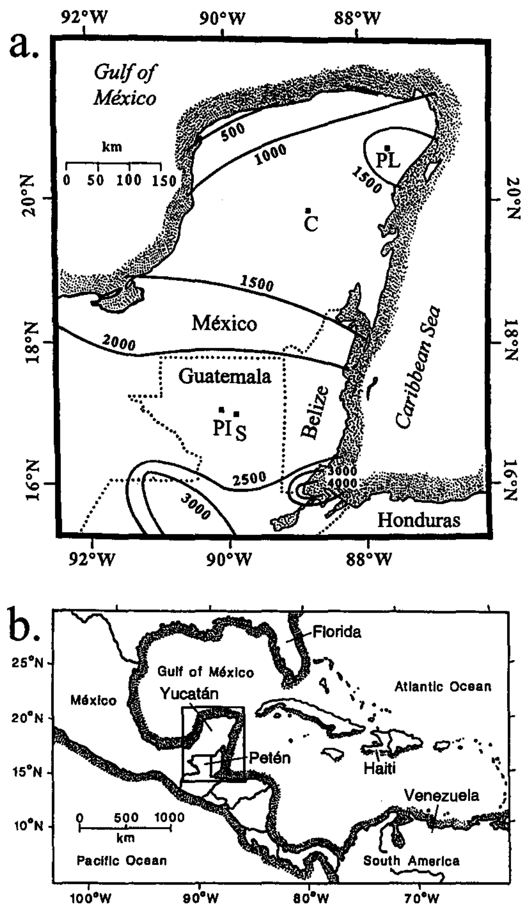

FIGURE 1 (a) Maya lowlands of the Yucatán Peninsula showing isohyets (mm) and locations of lake sediment records discussed in the text (PL = Punta Laguna, C = Chichancanab, PI = Petén-Itzá, S = Salpeten). Rainfall values are based on Wilson (1980). (b) The circum-Caribbean region showing the Yucatán Peninsula (box) and locations of paleoclimate studies referred to in the text, including Petén (Guatemala), Yucatán (México), Florida, Haiti, and Venezuela.

land regions (Sharer, 1994). Coe pointed out that there is no evidence for the *gradual cultural decline* referred to by Meggers. Instead, archaeological data point to abrupt, widespread collapse ca. A.D. 850, after 600 years of population growth and cultural expansion in the Early Classic (A.D. 250–550) and Late Classic (A.D. 550–850) periods.

Ferdon (1959) reassessed Petén's agricultural potential and challenged Meggers's application of *environmental determinism* to the lowland Classic Maya. Using modern temperature, rainfall, soil, and landform criteria, he reclassified the lowlands as a region favorable for agriculture. The analysis did not refute deterministic theory as applied to the Maya, but rather, freed the civilization from expectations dictated by an environment with limited agricultural potential. Ferdon, however, argued that there is no correlation between natural agricultural potential and cultural development and attributed the cultural decline to invasion of agricultural plots by grasses that interfered with Classic Maya crop cultivation.

Adams (1973) summarized theories that purport to explain the ninth century A.D. Maya collapse. Among the proposed causes that continue to be explored are overpopulation combined with soil erosion and exhaustion (Rice, 1978; Deevey et al., 1979; Paine and Freter, 1996), demographic models that show skewed sex ratios in favor of males (Cowgill and Hutchinson, 1963), health problems and disease (Saul, 1973), social upheaval, warfare, religion (i.e., the collapse was preordained), and insect infestation (Sabloff, 1973). Natural catastrophes such as earthquakes and volcanic eruptions (Ford and Rose, 1995) have also been invoked as factors contributing to the collapse. Others have suggested a correlation between climate and Maya cultural development and decline (Gore, 1992; Gill, 2000; Dahlin, 1983; Gunn and Adams, 1981; Folan et al., 1983; Messenger, 1990; Hodell et al., 1995; Curtis et al., 1996). Several theories for the collapse have been tested by archaeological and paleoenvironmental study, but some of the hypotheses are supported by few, if any, data.

Outside the Maya lowlands, other pre-Columbian cultures also arose, persisted for centuries, and ultimately collapsed in seemingly harsh environments. The Chiripa (1500–200 B.C.) and Tiwanaku (400 B.C. to A.D. 1100) cultures developed in the Lake Titicaca watershed of the Andean Altiplano (Fig. 2), where nitrogen-poor soils on steep slopes are prone to erosion and moisture deficit. Nighttime freezes often kill crops prior to harvest (Kolata and Ortloff, 1989). Flatlands (pampas) near Lake Titicaca are inundated periodically and subject to soil salinization. The prolonged rise and abrupt fall of a great civilization in this region of seemingly limited resources appears to defy environmental determinism.

Human ingenuity, combined with state-level agricultural organization, can overcome natural environmental limitations to food production. Intensive agricultural practices, such as raised-field construction and terracing, were widely used by pre-Columbian civilizations. Remnant raised fields have been found along waterways and in wetlands of the Maya lowlands (Adams, 1980; Adams et al., 1981; Siemens and Puleston, 1972; Matheny, 1976) and in the river drainages and pampas of the Titicaca basin (Kolata, 1991). The ancient agricultural strategy is documented archaeologically throughout the tropics (Denevan, 1970; Denevan and Turner, 1974). In the Maya lowlands, constructed fields in wetlands served to drain land and raise crop

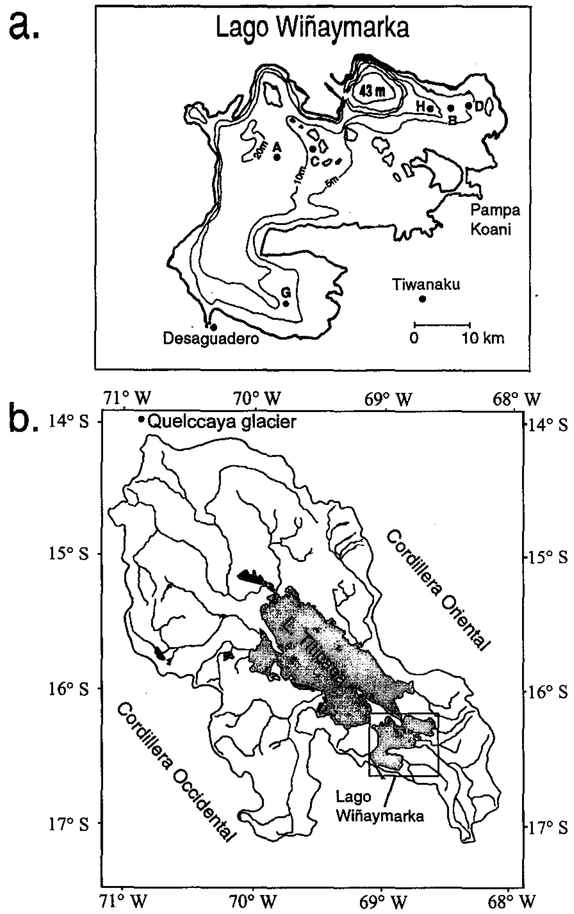

FIGURE 2 (a) Bathymetric map of the southern basin of Lake Titicaca (Lago Wiñaymarka) showing locations of the six sediment cores used to date late Holocene lake level changes. Cores A, B, C, D, G, and H were used to infer lake level shifts over the past ~3500 years. Also indicated are the Río Desaguadero outflow, the archaeological site of Tiwanaku, and the location of the Pampa Koani (see Fig. 8). (b) The Titicaca drainage basin showing the lake and major input rivers (redrawn from Boulangé and Aquize Jaen 1981). High mountain ranges of the Cordillera Oriental and Cordillera Occidental bind the watershed to the east and west, respectively. The Quelccaya ice cap (●) lies at the northwest margin of the drainage.

roots above the inundated soil zone (Pohl, 1990). Organic matter that accumulated in canals between raised fields was used to fertilize intensively farmed, nutrient-poor soils. Canals also provided an environment for cultivation of aquatic resources, including edible macrophytes, mollusks, fish, and turtles.

The multiple advantages of raised fields and their contribution to sustainable agricultual production have been demonstrated in the Andean Altiplano by both archaeological excavation (Kolata, 1991) and experimental study of reconstructed raised beds (Erickson, 1988; Kolata et al., 1996). On the pampas, raised fields elevate crop roots above the phreatic zone. Canals between fields absorb solar radiation during the day and store heat, protecting crops against nighttime freezes (Kolata and Ortloff, 1989). Cyanobacteria in canals fix atmospheric nitrogen (Biesboer et al., 1999), and the canals retain nutrients and provide fertilizer for N-limited soils (Binford et al., 1996; Carney et al., 1993). Canals between raised fields receive fresh stream water or groundwater from the base of surrounding hills, and low ion concentrations in these waters prevent soil salinization (Sanchez de Lozada, 1996).

Our purpose in revisiting the concept of environmental determinism is not to debate its validity as applied to the Maya and Tiwanaku cultures. Instead, we note that in the Maya and Tiwanaku cases an implicit assumption used to assess natural agricultural potential was indefensible, rendering the discussion moot. Although it is never explicitly stated, evaluations of natural agricultural potential ignored technological innovations and assumed that climate conditions were constant and similar to those of the present. Recent archaeological investigations indicate that many pre-Columbian societies utilized intensive agricultural practices (Denevan, 1970; Pohl, 1990). Paleoclimate studies in the circum-Caribbean (Hodell et al., 1991, 2000; Curtis and Hodell, 1993) and South American Altiplano (Abbott et al., 1997a; Thompson et al., 1985, 1995) provide evidence for late Holocene climate fluctuations. We contend that modern agricultural potential alone cannot establish whether past environments stimulated or constrained cultural development. Instead, we argue that although agricultural potential is, in part, dependent on landscape characteristics such as landforms, soils, and vegetation, crop production varies over time as a function of human social organization, land use practices, and climate change. About 3000 years ago, climate changes in the Maya lowlands and Andean Altiplano generated environmental circumstances that were conducive to human agricultural development. Two millennia later, climatic change leading to droughts exceeded environmental thresholds for Maya and Tiwanaku agricultural sustainability and led to the collapse of both civilizations.

In this chapter, we review paleoclimatological and archaeological results and discuss the correlation between Holocene climate and both Maya and Tiwanaku cultural development and collapse. The two cultural areas were chosen for several reasons. First, they have been subject to intensive archaeological study, so their demographic and agricultural prehistories are well known. Second, both civilizations arose, prospered, and collapsed in low-latitude areas adjacent to lakes that contain paleoclimate archives in their sediments. Third, although the Maya lowlands lie at low elevations (0–300 m above sea level [asl]) and the Tiwanaku

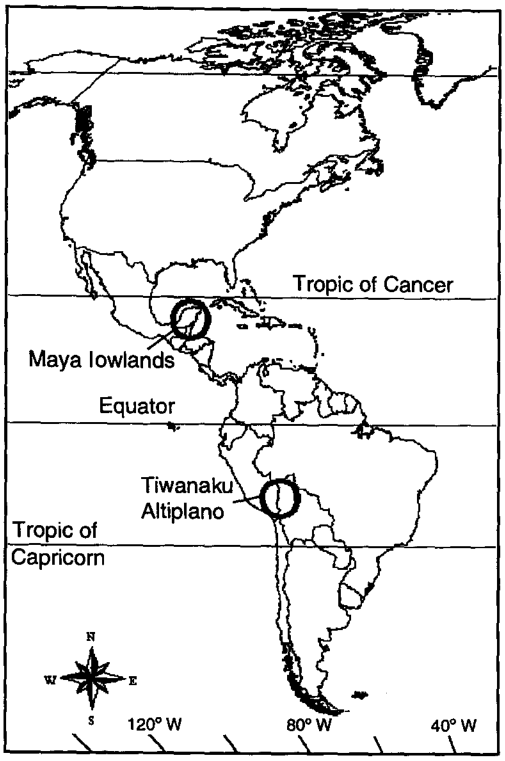

FIGURE 3 Map of the Americas indicating the Maya lowlands and Andean Altiplano where paleoclimate studies were done.

region is located at high altitudes (~3600 to >4000 m asl), both regions are in climatically marginal areas for agriculture, where small changes in available moisture can have profound impacts on crop yields. Finally, the two study areas lie along the Pole-Equator-Pole: Americas (PEP 1) transect (Fig. 3). Combined investigation of the Maya and Tiwanaku cultural areas allows temporal correlation of tropical climate changes north and south of the equator, establishment of interhemispheric climate linkages, and differentiation of forcing factors driving long-term and short-term climate change in the two regions.

6.1.2. The Maya Lowlands

Here, we consider the Maya area of the Yucatán Peninsula encompassed by what is today Belize; the northern department of Petén, Guatemala; and the Mexican states of Campeche, Yucatán, and Quintana Roo (Fig. 1). The region is characterized principally by karst topography that varies in elevation from sea level to ~300 m asl. Annual rainfall across the peninsula is heterogeneous, ranging from <500 mm / year in the extreme northwest to ~2000 mm / year (Fig. 1) in central Petén (Wilson, 1980). The precipitation gradient is reflected in soil development and vegetation (Flores and Carvajal, 1994). Soils in the flat, northern part of the peninsula are thin, with limestone bedrock exposed at the ground surface. Farther south, in the karst hills of Petén, soils are deeper and more productive. Dry-adapted vegetation in the semiarid northwest is of low stature and diversity. The Petén forest, sometimes referred to as a "quasi-rain forest" (Lundell, 1937), is taller and more diverse.

Rainfall on the Yucatán Peninsula is highly seasonal, with dry conditions from January through April (Deevey et al., 1980). Rainfall is high in May, June, September, and October. A drier interval, referred to as the *canicula,* usually occurs in July or August (Magaña et al., 1999). Interannual variation in the amount and timing of rainfall can be pronounced (Wilson, 1980). Under the influence of the Northeast Trades, summer rains on the peninsula coincide with the northward migration of the Intertropical Convergence Zone (ITCZ) (Hastenrath, 1976, 1984). The Azores–Bermuda high-pressure system also moves northward in summer. Sea surface temperatures (SSTs) are warm in the tropical / subtropical North Atlantic and Caribbean during the Northern Hemisphere summer, providing ample moisture for precipitation. Summertime rainfall is frequently delivered during violent, convective thunderstorms, and total rainfall in any given year can be influenced by tropical storms and hurricanes that contribute substantial precipitation within a short time period.

Dry conditions prevail on the Yucatán Peninsula in the Northern Hemisphere winter, when the ITCZ and Azores–Bermuda high-pressure system move southward (Hastenrath, 1976, 1984). Low winter precipitation is a consequence of relatively low SSTs in the tropical North Atlantic, a steep pressure gradient on the equatorward side of the Azores–Bermuda High, and a strong temperature inversion associated with enhanced trade winds. Differences in precipitation, rather than temperature, define interannual climate variability on the Yucatán Peninsula.

Both pre-Columbian and modern agriculturists on the Yucatán Peninsula have had to contend with strong seasonality and unpredictability of annual precipitation. Although the intraannual pattern of rainfall is known, delayed onset of summer rains or other disruptions of seasonal precipitation can have disastrous consequences for farmers, many of whom still depend on slash-and-burn (swidden) techniques. In rural Yucatán, the contemporary Maya still perform the ancient Cha Cha'ac ceremony to invoke rainfall. Ubiquitous portrayal of the rain god, Chac, on monumental architecture in the northern Maya lowlands (Fig. 4 [see color insert]) demonstrates that the quantity and timing of rainfall were also of great importance to the ancient Maya.

High-resolution Holocene paleoclimate inferences for the Yucatán Peninsula are based on stable isotopic ($\delta^{18}O$) study of multiple ostracod or gastropod shells in samples taken at 1-cm intervals from lake sediment cores. Here, we summarize core results from Lakes Chichancanab (Hodell et al., 1995) and Punta Laguna (Curtis et al., 1996) on the northern part of the peninsula and from Lake Petén-Itzá (Curtis et al., 1998) in Petén, Guatemala (Fig. 1). We are currently studying cores from Lake Salpeten, in the Petén lake district (Fig. 1), to examine the details of climate change in the southern Maya lowlands. The $\delta^{18}O$ of shell carbonate from these lake sediment cores was governed principally by three factors: (1) the $\delta^{18}O$ of the lake water from which the shells were precipitated; (2) taxon-specific fractionation, i.e., *vital effects;* and (3) water temperature. In these stratigraphic studies, we used shells of monospecific, adult snails and ostracods in each core to minimize vital effects. Temperature changes during the Holocene were probably not responsible for the dramatic shifts (>3‰) in shell $\delta^{18}O$, because a 1‰ shift in $\delta^{18}O$ requires a ~4°C change in temperature (Craig, 1965). It is unlikely that long-term temperature fluctuations were solely or largely responsible for the stratigraphic shifts in shell $\delta^{18}O$, as this would require a change in mean water temperature of >12°C. Likewise, intraannual temperature fluctuations probably did not contribute significantly to stratigraphic changes in shell $\delta^{18}O$. Seasonal differences in diurnal water temperatures from these tropical lakes are typically on the order of only 3°–4°C (Covich and Stuiver, 1974; Deevey et al., 1980), and shells collected from 1-cm sample intervals in the cores integrate, on average, material deposited over ~5- to 50-year periods. Although temperature changes may contribute minimally to stratigraphic $\delta^{18}O$ variation, the primary determinant of large shifts in the $\delta^{18}O$ of shell carbonate during the Holocene has been the $\delta^{18}O$ of the lake water from which the carbonate was derived.

In tropical lakes that lack overland outflows, the two major factors that control the $\delta^{18}O$ of lake water over time are the isotopic signature of input waters (rain, runoff, and groundwater) and the relation between evaporation and precipitation (E/P) (Fontes and Gonfiantini, 1967). Rozanski et al. (1993) reported weighted mean isotopic values for several low-elevation, circum-Caribbean sites that were part of the International Atomic Energy Agency (IAEA)/World Meteorological Organization (WMO) precipitation monitoring network. Samples were collected between 1961 and 1987, and in each case, weighted means integrated data obtained over a period of at least 14 years. The isotopic signatures of rainfall from stations at Veracruz, México (−4.13‰, n = 169); Howard Air Force Base, Panama (−5.65‰, n = 165); Barranquilla, Colombia (−5.09‰, n = 85); and Maracay, Venezuela (−4.01‰, n = 64) are similar.

Our mean values for $\delta^{18}O$ in precipitation samples from Petén, Guatemala (−2.86‰) (Rosenmeier et al., 1998) and near Punta Laguna, México (−3.91‰) (Curtis et al., 1996) reflect only a few rainfall events, but are only slightly higher than weighted mean values for the circum-Caribbean reported by Rozanski et al. (1993). Isotopic values for groundwaters in Petén (−3.38‰) and near Punta Laguna (−3.92‰) are similar to their respective rainfall values, suggesting that input waters to Yucatán lakes, whether they are delivered via direct rainfall, runoff, or subterranean infiltration, possess the same isotopic signature.

Lake waters on the Yucatán Peninsula yield isotopic values that are more positive relative to input waters, indicating that ^{18}O is evaporatively concentrated in the water bodies (Covich and Stuiver, 1974; Hodell et al., 1995; Curtis et al., 1996; Rosenmeier et al., 1998). Differences between $\delta^{18}O$ values for precipitation and lake water show enrichment ranging from ~3.6‰ at Lake Petén-Itzá (Curtis et al., 1998) to >7‰ at Lakes Chichancanab (Hodell et al., 1995) and Salpeten (Rosenmeier et al., 1998). Although we cannot demonstrate definitively that the source water $\delta^{18}O$ signature has not changed over the course of the Holocene, we believe that the major process affecting lake water $\delta^{18}O$ on the Yucatán Peninsula for the last ~10,000 years has been shifting E/P. Thus, stratigraphic study of $\delta^{18}O$ in mollusk and crustacean shells in Yucatán sediment cores can be used to reconstruct past changes in moisture availability. Ostracods and snails often occupy different habitats within lakes and display different growth characteristics. Whereas ostracods molt as they pass through instar stages, mollusks simply add shell as they grow. The two taxonomic groups are ecologically and developmentally different, but yield similar stratigraphic $\delta^{18}O$ signals, indicating that the sedimented shell remains faithfully reflect past changes in the oxygen isotope ratio of the lake water.

Shallow lakes (<30 m) and shallow areas of deep lakes on the Yucatán Peninsula were first covered by water and began depositing lacustrine sediments between ca. 8030 and 5840 B.C. (9000–7000 ^{14}C B.P.) (Leyden et al., 1994, 1998). Lakes and sinkholes (cenotes) filled in the early Holocene, after a long, late Pleistocene arid period (Leyden et al., 1994). On the northern part of the peninsula, early Holocene basin filling was a consequence of increasing moisture availability as well as of sea level rise, which raised the local water table. Farther south in the Petén, where the local aquifer lies deep below the land surface (Gill, 2000), increased rainfall at the Pleistocene–Holocene transition filled lakes and supplied moisture for tropical forest synthesis (Leyden, 1984).

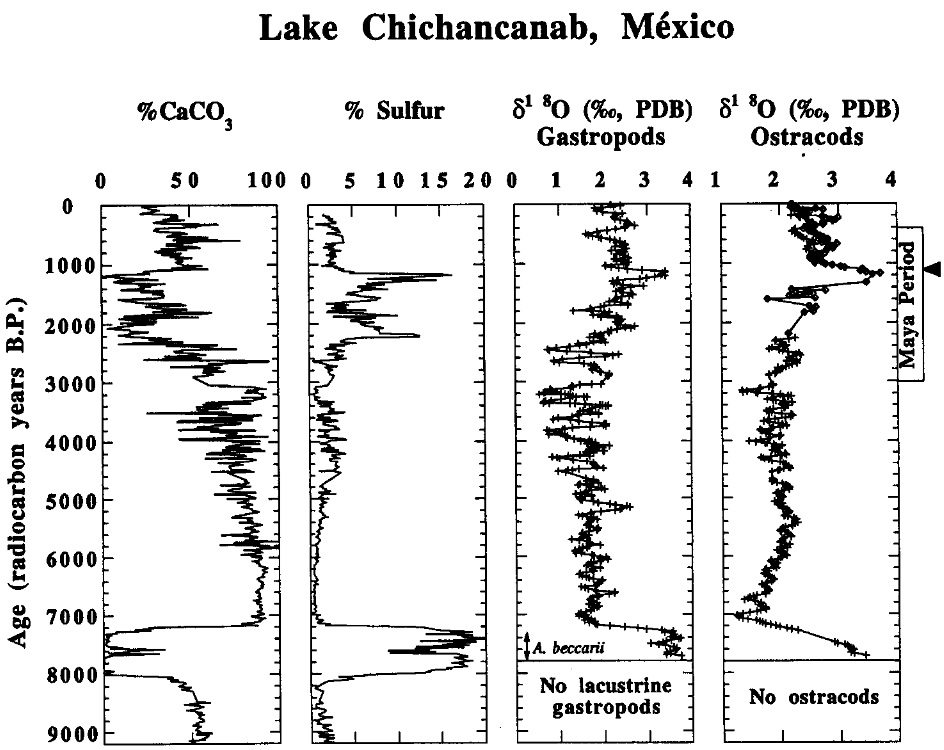

FIGURE 5 High-resolution paleoclimate record from Lake Chichancanab, in the northern Maya lowlands. Percent $CaCO_3$, percent sulfur (gypsum), $\delta^{18}O$ of gastropod shells (*Pyrgophorus coronatus*), and $\delta^{18}O$ in shells of ostracods *Physocypria* sp. (+) and *Cyprinotus* cf. *salinus* (◇) at 1-cm intervals, plotted against ^{14}C years B.P. Oxygen isotope results are 3-point running means and are expressed relative to the PDB (PeeDee Belemnite) standard. Arrows between about 7800 and 7300 ^{14}C years B.P. within the gastropod $\delta^{18}O$ plot indicate depths at which the foraminifer *Ammonia beccarii* was found. The arrow at the right margin indicates the driest episode of the late Holocene and is dated at 1140 ± 35 ^{14}C years B.P. (A.D. 922), coinciding with the Classic Maya collapse. (From Hodell, D. A. et al. (1995), with permission.)

The stable isotope ($\delta^{18}O$), geochemical, and microfossil stratigraphies (Fig. 5) of a core from Lake Chichancanab (19°50′ N, 88°45′ W) provide a high-resolution record of changes in E/P over the last ~8200 ^{14}C years (Hodell et al., 1995). Radiocarbon dating of gastropod and ostracod shells from Yucatán lakes is confounded by the effects of hard water lake error (Deevey and Stui-ver, 1964), which makes dates on lacustrine carbonates in Lake Chichancanab ca. 1200 years too old (Hodell et al., 1995). Therefore, the age/depth relation for the core (Fig. 5) was developed by regression using only accelerator mass spectrometer (AMS) ^{14}C dates on terrestrial carbon (Hodell et al., 1995).

Initial filling of Lake Chichancanab ca. 8200 ^{14}C B.P. (7250 B.C.) is marked by gypsum precipitation, relatively high $\delta^{18}O$ values for gastropod and ostracod shells, and the presence of the benthic foraminifer *Ammonia beccarii* (Fig. 5). *A. beccarii* can tolerate a wide range of temperatures (10°–35°C) and salinities (7–67 g/L), but is capable of reproducing only at salinities between 13 and 40 g/L (Bradshaw, 1957). Dry conditions in the earliest Holocene are inferred from the biological and geochemical indicators that point to low lake stage, saline waters, and high E/P. After ~7200 ^{14}C B.P. (6000 B.C.), the lake filled rapidly. Gypsum precipitation was replaced by carbonate deposition, stable isotope values were lower, and *A. beccarii* disappeared from the record, indicating wetter conditions. Relatively moist conditions persisted from ca. 7200 to 3000 ^{14}C B.P. (6000–1250 B.C.) (Fig. 5).

Palynological evidence suggests that swidden activity may have begun in northern Guatemala as early as ~5600 ^{14}C B.P. (4410 B.C.) (Islebe et al., 1996), but this gradual decline of forest in Petén may be attributable to the onset of a regional drying trend. The archaeological record indicates humans did not establish sedentary settlements in the Maya lowlands during the relatively moist early to middle Holocene. Archaeologists date the earliest sedentary populations in the region to the Middle Preclassic period, ca. 1000–300 B.C. (Rice and Rice, 1990; Turner, 1990). Initial settlement in the Maya lowlands corresponds generally to a period of increased regional drying at about 3000 ^{14}C B.P. (1250 B.C.). This pronounced drying trend is reflected in the

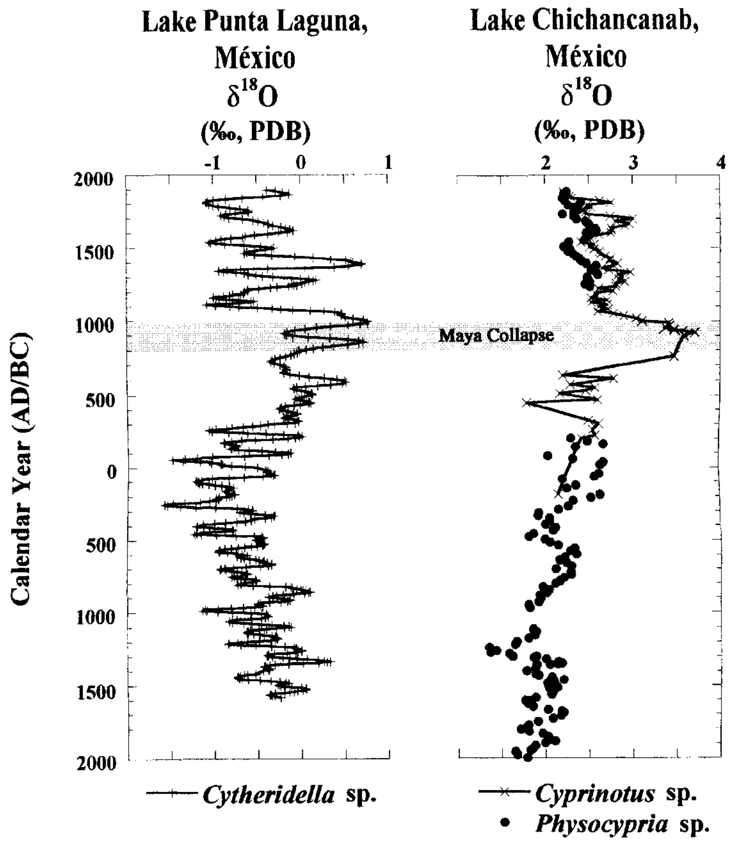

FIGURE 6 Oxygen isotope ($\delta^{18}O$) records based on ostracods from Lakes Punta Laguna and Chichancanab (México), spanning the last 3500 calendar years. Both $\delta^{18}O$ records display increasing or relatively high values, i.e., high E/P, within or just following the interval from A.D. 800–1000, corresponding to the period of the Maya collapse.

Chichancanab sediment core by increased sulfur (gypsum) concentrations and higher $\delta^{18}O$ values for gastropod and ostracod shells (Fig. 5). Concurrent late Holocene drying has also been documented in Lake Miragoane, Haiti (Hodell et al., 1991), and Lake Valencia in northern Venezuela (Bradbury et al., 1981; Leyden, 1985).

The paleoclimate record from Lake Chichancanab sediments indicates that early Maya agriculturists faced several challenges. Despite the drying trend that began ca. 1250 B.C., general conditions were apparently suitable for shifting agriculture. Nevertheless, in addition to interannual variations in the timing and amount of rainfall, relative E/P on the Yucatán Peninsula between ca. 1250 B.C. and ca. A.D. 1000 varied substantially over decadal to centennial scales. This variability is seen most clearly in the high-resolution, 3500 ^{14}C-year paleoclimate record from Punta Laguna (Fig. 6), which lies about 150 km northeast of Chichancanab at 20°38′ N, 87°37′ W (Fig. 1). Hard water lake dating error at Punta Laguna is also on the order of 1200–1300 years, and the age/depth relation for the core was developed by linear interpolation between five AMS ^{14}C-dated terrestrial wood samples, assuming linear sedimentation between dated horizons (Curtis et al., 1996). Dates were calibrated by using CALIB 3.0 with a 100-year moving average of the tree-ring data set (Stuiver and Becker, 1993; Stuiver and Reimer, 1993).

In the Chichancanab core, the highest gypsum concentrations and $\delta^{18}O$ values of the middle to late Holocene period were found in samples at ~65 cm depth (Fig. 5). An AMS ^{14}C date on a terrestrial seed at 65 cm shows that the protracted dry episode culminated at 1140 ^{14}C B.P. (A.D. 922), coinciding closely with the Classic Maya collapse (Hodell et al., 1995). In the Punta Laguna section, mean $\delta^{18}O$ values for the period 1750–940 ^{14}C B.P. (A.D. 300–1100) are higher than mean values for the preceding and following periods (Fig. 6). The $\delta^{18}O$ peak at A.D. 862 (1210 ^{14}C B.P.) in the Punta Laguna section may correspond to the Late Classic dry event recorded in the Chichancanab core.

Paleoclimate records from Lakes Chichancanab and Punta Laguna, in combination with the archaeological record, demonstrate a temporal correlation between drought and cultural demise. These water bodies lie in the dry, northern Maya lowlands. This region, however, was apparently least affected by the "collapse" (Lowe, 1985). Demographic consequences of the ninth-century decline were felt most acutely in the wetter, southern lowlands. This pattern may be a consequence of the fact that groundwater is less accessible in the southern lowlands than it is in the north because the depth to the water table increases southward. Therefore, the southern lowlands were more reliant upon surficial water supplies (Gill, 2000). Archaeological survey and test excavation of house mounds in several Petén watersheds yielded estimates of Late Classic Maya population densities in excess of 200 persons per square kilometer (Rice and Rice, 1990). By the Terminal Classic period (A.D. ~900), population densities in nearly all the studied drainage basins had declined to <100 persons per square kilometer, and by the Late Postclassic period (A.D. 1500), many watersheds were abandoned. Drought is implicated in the Classic Maya collapse. Evidence for this climatic change should therefore be found in Petén lake cores.

Recent attention has turned to paleoclimate reconstruction in the Petén lake district (Fig. 1). Lake Petén-Itzá (16°55′ N, 89°52′ W), the largest Petén water body (area, A = 99.6 km^2), yielded a ~9000 ^{14}C-year paleoenvironmental record based on sediment geochemistry, magnetic susceptibility, pollen, and stable isotopes (Curtis et al., 1998). Chronology for the core was based on AMS ^{14}C dating of terrestrial wood and charcoal samples (Curtis et al., 1998). Ages between dated horizons were interpolated by assuming constant linear sedimentation. Three $\delta^{18}O$ records from the basin, two based on snail taxa and another on ostracods, display little variability over the past 5000 ^{14}C years (Fig. 7). Rather than indicating climatic constancy during the past five millennia, however, they may simply reflect the fact that lake water $\delta^{18}O$ in large-volume lakes is insensitive to short-term changes in E/P. In large, deep lakes with long residence times, even protracted droughts may not reduce lake volume sufficiently to alter the lake water isotopic signature.

Lake Salpeten (16°58′ N, 89°40′ W) lies near Lake Petén-Itzá (Fig. 1), but is smaller (A = 2.6 km^2) and more saline (TDS = 4.76 g/L). Lake Salpeten is being studied because its sediments possess abundant ostracod and gastropod shells, it is at saturation with respect to gypsum, and its lakewater $\delta^{18}O$ is demonstrably enriched by >7‰ relative to rainfall $\delta^{18}O$. The relatively greater ^{18}O enrichment in Lake Salpeten suggests it is more effectively "closed" than Lake Petén-Itzá. Seismic reflection studies in Lake Salpeten were completed in summer 1999. Imaging of the sediment stratigraphy indicates periods of low lake stage during the Holocene, and consequent sedimentation hiatuses at shallow-water locations. Efforts to reconstruct late Holocene climate conditions from Lake Salpeten sediments are therefore focusing on cores collected at deep-water sites in the basin.

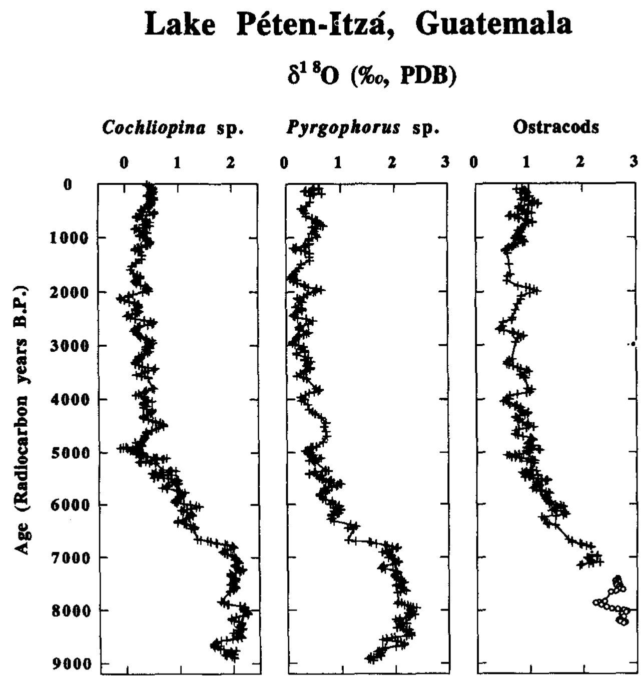

FIGURE 7 Smoothed, 5-point running mean, oxygen isotope ($\delta^{18}O$) records from the Lake Petén-Itzá core, Petén, Guatemala, are plotted against ^{14}C years B.P., and based on two gastropods (*Cochliopina* sp. and *Pyrgophorus* sp.) and two ostracods (*Cytheridella ilosvayi*, (solid line) and *Candona* sp., (open circles). (From Curtis, J. H., et al. (1998), with permission.)

Lake Salpeten is of particular interest because the history of human occupation in the watershed is well known (Rice and Rice, 1990). A large Maya population occupied the Salpeten drainage basin in the Late Classic Period, and human-mediated vegetation removal in the watershed has been documented palynologically (Leyden, 1987). Massive deforestation can alter watershed hydrology (Bosch and Hewlett, 1982; Stednick, 1996), which may complicate interpretation of isotope records and make it difficult to discern whether drought conditions prevailed in the southern Maya lowlands during the Late Classic.

There is some evidence that dry conditions at the end of the Late Classic period extended beyond the Maya area. Horn and Sanford (1992) reported increased fire frequency in Costa Rica at 1180–1110 ^{14}C B.P. (A.D. 880–970). Metcalfe et al. (1994) found evidence of a dry episode from 1200 to 1100 ^{14}C B.P. (A.D. 830–970) in sediments from Lake Pátzcuaro in the central Mexican highlands. A core from La Piscina de Yuriria, Guanajuato, México, also displayed evidence of drying during the period 1500–900 ^{14}C B.P. (A.D. 570–1070) (Metcalfe et al., 1994), as did the Zacapu basin, ca. 1000 ^{14}C B.P. (A.D. 1020) (Metcalfe, 1995).

6.1.3. The Andean Altiplano

We focus on the region of the Bolivian Altiplano (Fig. 3) that surrounds the southern basin of Lake Titicaca (Lago Wiñaymarka) (16°20′ S, 68°50′ W) (Figs. 2a and 2b) and lies between 3800 and 4200 m asl. Climate conditions in the Lake Titicaca watershed present special challenges for Andean agriculturists. At very high elevations (~5100 m) on peaks at the edge of the Titicaca basin, agriculture is precluded by mean annual temperatures <0°C (Roche et al., 1992). The large area (8560 km^2) and volume (~900×10^9 m^3) of Lake Titicaca (Wirrmann, 1992) and its relatively warm water temperatures (10°–14°C) have profound thermal effects on the drainage basin, helping maintain mean annual temperatures around the lake between 7° and 10°C (Roche et al., 1992). Nevertheless, the mean minimum monthly temperature near the lake, which occurs in July, is about 1.8°C. Crops are at risk of nighttime freezing during the austral winter.

Long-term, annual rainfall over the entire Titicaca basin averages about 758 mm/year, but is spatially quite variable and influenced by lake effects and orography (Roche et al., 1992). Values range from ~500 mm/year at sites far from the lake to >1000 mm/year over the water body. Precipitation in the Altiplano is derived largely from north-easterly winds that transport moist air from the Amazon basin over the Eastern Cordillera. Rainfall delivery is highly seasonal, with a wet period centered on January and extending from December to March. June marks the middle of the dry season, which runs from May to August. Seasonal rainfall in the Andean Altiplano is out of phase with the timing of precipitation in the Northern Hemisphere Maya lowlands. Intraannual precipitation variability in the Altiplano is best expressed by the percentage of annual rainfall delivered to the Titicaca basin during the 4-month wet season (70%), the 4-month dry season (5%), and the two intermediate 2-month periods (25%) (Roche et al., 1992). Mean monthly evapotranspiration exceeds average rainfall from March to December, leading to soil moisture deficit and salinization (Binford et al., 1997).

Interannual climate change in the Altiplano is expressed as yearly rainfall, which is controlled by the strength of summer monsoon circulation, the position of the ITCZ, and El Niño/Southern Oscillation (ENSO) events. Strong El Niño years are associated with dry conditions in the Altiplano (Roche et al., 1992). Lake Titicaca stage levels have been measured at Puno, Peru, since 1914. Although the relation between lake stage and rainfall is confounded somewhat by inputs of glacial meltwater, the lake level serves as a good proxy for rainfall and illustrates interannual variability in precipitation. The lake outflow sill to the Río Desaguadero is at 3804 m, and the mean lake stage for the period of measurement, 1914–1989, was ~3809 m. The lowest stage over the period of measurement was recorded in 1943 and was 6.37 m below the highest stage, measured in 1986. There was a lake level decline of nearly 5 m between the mid-1930s and mid-1940s (Roche et al., 1992). Such rapid shifts in rainfall and lake level create problems for farmers in the drainage basin. Crops receive insufficient water during very dry periods, but rot or are drowned during wet episodes. Furthermore, extensive, nearshore cultivable flatlands are inundated or exposed by lake level excursions of only a few meters (Fig. 8 [see color insert]). Ion-rich lake waters (conductivity = 1400 μS/cm) retreat during dry periods, leaving soils encrusted with salts.

Rocky, upland soils in the Altiplano are nitrogen deficient (Binford et al., 1996) and prone to erosion on steep slopes. Raised-field canal mucks and intermittently inundated soils have higher total N content (Binford et al., 1996) as a consequence of nitrogen fixation (Biesboer et al., 1999). The construction of raised agricultural fields by the Tiwanaku was a strategy that simultaneously addressed problems of water control, freezes, soil fertility, and salinization. Prehistoric raised fields in the combined Catari basin and Tiwanaku valley covered 130 km^2 (Kolata and Ortloff, 1996) and bore testimony to the efficacy of this intensive farming practice. Both the widespread use and long persistence (ca. A.D. 600–1100) of this agricultural technique are evidence of its importance.

In the late Pleistocene, 13,000–12,000 ^{14}C B.P. (13,500–12,050 B.C.), Lake Titicaca was deeper and fresher than it is now (P. Baker, personal communication). The extensive paleolake, or Tauca stage, was attributed to a 30–50% increase in precipitation over the Altiplano (Hastenrath and Kutzbach, 1985). At ca. 12,000 ^{14}C B.P., a drying trend began, and water levels in Lake Titicaca consequently fell as much as 85 m (Seltzer et al., 1998). Lago Wiñaymarka essentially desiccated, leaving only scattered, water-filled depressions. The low stage lasted until ca. 3600 ^{14}C B.P. (1930 B.C.) (Mourgiart et al., 1995; Wirrmann et al., 1987, 1992). An increasing E/P trend from the late Pleistocene to the middle Holocene is also inferred from studies of glacial retreat and higher snow lines (Seltzer, 1992), as well as lake sediment records from the Eastern Cordillera (Abbott et al., 1997b). As is predicted by insolation forcing, this long-term history of moisture availability from the Andean Altiplano is out of phase with conditions in the Northern Hemisphere Maya lowlands.

Late Holocene changes in Lake Titicaca's water level were inferred from paleolimnological study of lake sediment cores and serve as proxy estimates of moisture availability in the Altiplano over the last ~3500 ^{14}C years. Cores were taken at six sites throughout Lago Wiñaymarka (Fig. 2). Retrieved lacustrine sediment came from elevations between ~0.5 and 14.5 m below the lake outlet level at 3804 m. Lake stage reconstruction was based on lithostratigraphy of the cores and 60 AMS ^{14}C dates on gastropod shells and sedge (*Schoenoplectus tatora*) achenes found near erosion surfaces (Abbott et al., 1997a). Hard water lake dating error was shown to be about 250 years in Lake Titicaca (Abbott et al., 1997a), and this value was subtracted from ^{14}C ages before dates were calibrated with CALIB 3.0 (Stuiver and Reimer, 1993).

All cores collected from Lago Wiñaymarka displayed laminated lake sediments overlying gray clay soils (Binford et al., 1997). Underlying clay deposits were formed by gleization, implying that the soils developed under anoxic, waterlogged conditions. This conclusion is supported by the presence of littoral *S. tatora* achenes at the clay/organic sediment boundary (Abbott et al., 1997a). At all six coring sites, inception of lacustrine deposition following the protracted low-stage episode dates to between 3560 and 3160 ^{14}C B.P. (2030–1420 B.C.) (Binford et al., 1997). The timing of this lake level increase is consistent with the earlier findings of Wirrmann et al. (1987). Between the initial late Holocene lake level rise and the current high stand of Lake Titicaca, there have been four dry episodes (Fig. 9) when the lake level declined: 2900–2800, 2400–2200, 2000–1700, and 900–500 cal. B.P. (Abbott et al., 1997a).

In the early and middle Holocene, riparian agriculture in the Altiplano was precluded because of prevailing dry conditions. As the climate ameliorated ca. 1500 B.C. (3210 ^{14}C B.P.), Chiripa culture emerged in the basin. By ca. 400 B.C., Tiwanaku civilization began to develop, and raised-field cultivation was established by A.D. 600. The agricultural technology was widespread in the region by A.D. 800 and persisted until the Tiwanaku collapse, ca. A.D. 1150. Archaeological excavations in raised-field contexts indicate the technology fell into disuse after A.D. 1150, coincident with drought conditions on the Altiplano (Kolata and Ortloff, 1996) that are documented in sediment records from Lago Wiñaymarka (Abbott et al., 1997a; Binford et al., 1997) and in ice core records from the Quelccaya ice cap in Peru (Thompson et al., 1985).

During the late Tiwanaku IV and Tiwanaku V periods (A.D. 600–1100), dense population centers in the Catari basin and Tiwanaku valley were established near raised-field complexes. Following the state collapse, small populations were dispersed on the landscape and had no apparent relation with raised fields (Kolata, 1993). Populations declined as a consequence of the agricultural collapse (Ortloff and Kolata, 1993), and thereafter, Tiwanaku inhabitants, and later Inca populations, relied on terracing and flatland cultivation.

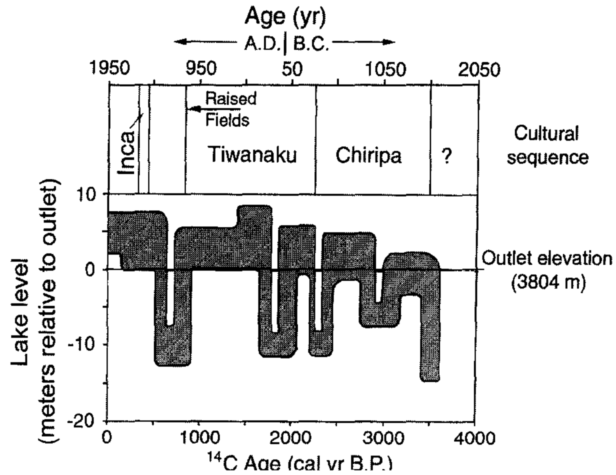

FIGURE 9 Water level record for Lago Wiñaymarka over the last ~3600 calendar years plotted relative to the outlet elevation at the Río Desaguadero (3804 m). The cultural sequence is plotted on a B.C./A.D. scale to show the correlation with the lake-level curve. Tiwanaku raised fields were widely utilized from about A.D. 600–1150. Fields fell into disuse and populations declined coincident with the onset of the last low-water-level event in the record. (Modified from Abbott et al., 1997.)

The protracted dry conditions in the Altiplano lasted more than a century and had negative impacts on raised-field agriculture. Low direct rainfall on crop planting surfaces reduced soil moisture beyond the permanent wilting point. Lack of groundwater recharge reduced or eliminated water flow to input springs and streams that fed canals between raised fields, exacerbating the dry conditions faced by crops. In addition, canal drying removed the benefits of freeze protection and nitrogen fixation. Without continuous flushing, fields were also prone to soil salinization.

6.1.4. Climate Forcing in the Circum-Caribbean

Modern studies of interannual rainfall variability in the tropical Atlantic demonstrate that wet years are associated with an enhancement of the annual cycle, driven by greater than normal summertime northward

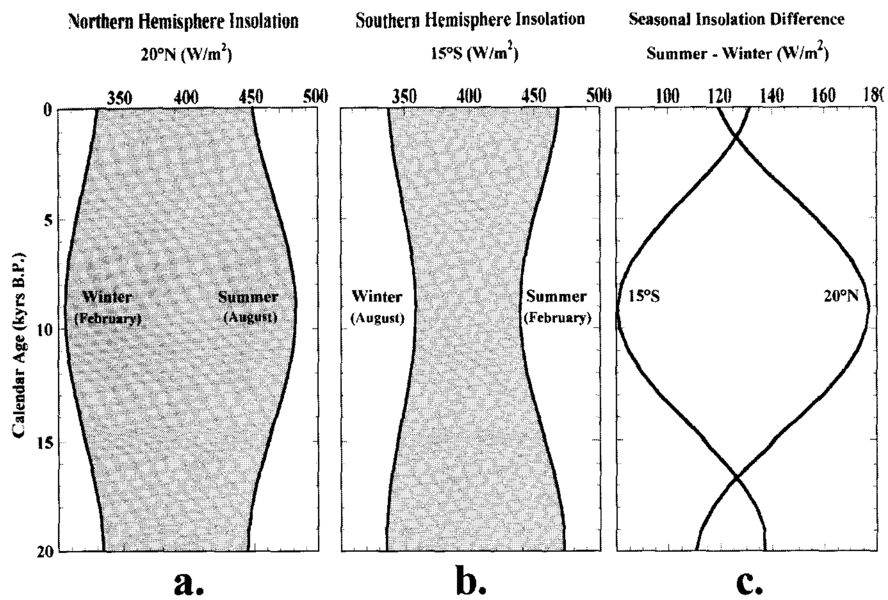

FIGURE 10 (a) Summer (August) and winter (February) insolation curves for the Maya lowlands (20°N) over the last 20,000 calendar years. (b) Summer (February) and winter (August) insolation curves for the Tiwanaku area (15°S) over the last 20,000 years. Values are from Berger and Loutre (1991). (c) Seasonal insolation difference (summer minus winter) at 20°N and 15°S, illustrating that long-term trends in seasonal insolation between the Northern Hemisphere and Southern Hemisphere tropics were out of phase. Since the end of the Pleistocene, the greatest seasonality at 20°N occurred in the early Holocene, whereas the greatest seasonality at 15°S occurred in recent millennia.

migration of the ITCZ and Azores–Bermuda High (Hastenrath, 1984). As a result, wet years occur when sea level pressure (SLP) is low on the equatorward side of the Azores–Bermuda High, the trade winds are displaced poleward (northward), SSTs between 10° and 20°N are warmer, and convergence and cloudiness are enhanced (Hastenrath, 1984). Dry years occur when there is a reduction in the annual cycle, associated with conditions opposite to those described previously (Hastenrath, 1984). Hodell et al. (1991) suggested that long-term E/P changes recorded in the stable isotope record from Lake Miragoane, Haiti, were driven by orbitally forced (Milankovitch) variations in solar insolation, which in turn controlled the intensity of the annual cycle. Long-term changes in seasonal insolation (Fig. 10) may explain the general pattern of a dry late Pleistocene followed by a moister early Holocene in the circum-Caribbean. The dry late Pleistocene–moister early Holocene transition has been documented by paleolimnological studies at sites around the Gulf of México and Caribbean, including Haiti (Hodell et al., 1991); Yucatán, México (Hodell et al., 1995; Whitmore et al., 1996); Guatemala (Leyden et al., 1994); northern Venezuela (Bradbury et al., 1981; Leyden, 1985); and Florida (Watts and Hansen, 1994). Orbital forcing, however, does not fully account for the magnitude of aridity inferred for the late Pleistocene (Hodell et al., 1991), nor the abrupt onset of moister conditions in the early Holocene (e.g., Leyden et al., 1994).

Milankovitch forcing is also thought to be responsible for reduced intensity of the annual cycle and the consequent drying trend in the late Holocene (Fig. 10). Empirical evidence for late Holocene drying around the Caribbean is reported from Lakes Miragoane, Haiti (Hodell et al., 1991); Chichancanab, Yucatán Peninsula, México (Hodell et al., 1995); and Valencia, northern Venezuela (Bradbury et al., 1981). The gradual reduction in the intensity of the annual cycle, however, cannot explain the rapid onset of drier conditions at ca. 3400–3000 ^{14}C B.P., documented by the $\delta^{18}O$ records from Lakes Miragoane (Hodell et al., 1991) and Chichancanab (Hodell et al., 1995). Also, orbital forcing cannot explain the dramatic decadal to centennial E/P fluctuations that are so apparent in the late Holocene portion of the paleoclimate records from Lakes Miragoane, Chichancanab, and Punta Laguna (Curtis et al., 1996). Other, as yet unexplained, forcing factors are responsible for these shorter-term excursions in moisture availability that had such devastating consequences for Maya agriculturists in the ninth century A.D.

6.1.5. Climate Forcing in the Andean Altiplano

Since the late Pleistocene, long-term changes in Lake Titicaca's water level have been driven by insolation forcing that influenced both annual rainfall and glacial advance and retreat. Wet conditions on the northern Altiplano are associated with the Bolivian High, which develops as a consequence of convective precipitation during the Southern Hemisphere summer (Aceituno and Montecinos, 1993; Lenters and Cook, 1997). Perihelion occurred during the Southern Hemisphere winter (July) at 8450 ^{14}C B.P. (7500 B.C.), and seasonality was reduced relative to the present (Berger, 1988; Kutzbach and Guetter, 1986). Abbott et al. (1997b) argued that reduced summer insolation and increased winter insolation at 8450 ^{14}C B.P. would have caused cooler summers and warmer winters, decreased moisture transport over the continent, and a net increase in E / P, with the latter caused largely by a decrease in precipitation. Low summertime insolation over the Altiplano lasted from ca. 10,000–7600 ^{14}C B.P. (9160–6420 B.C.) (Fig. 10). Likewise, the highest June–August insolation at the same latitude occurred from ca. 11,000–6600 ^{14}C B.P. (11,000–5500 B.C.), probably causing increased ablation of glaciers in the winter dry season. The combined effects of reduced summertime precipitation and increased disappearance of glacial ice in winter contributed to deglaciation and to the mid-Holocene dry episode on the Altiplano (Abbott et al., 1997b).

For the past ~8000 calendar years, austral winter insolation on the Altiplano has decreased and summer insolation has increased (Fig. 10). Higher summertime insolation was accompanied by greater precipitation, which could account for the filling of Lago Wiñaymarka ca. 3600 ^{14}C B.P. Orbital forcing, however, does not explain the abrupt shifts in water balance that the lake has experienced over the last three millennia. The lake level has been shown to be correlated with rainfall, which tends to be lower during El Niño events (Roche et al., 1992; Binford and Kolata, 1996). Protracted periods of strong El Niño activity, which established its modern periodicity ca. 5000 cal. B.P. (Rodbell et al., 1999), might have been responsible for the documented low stands.

6.1.6. Interhemispheric Correlations

Since the late Pleistocene, long-term patterns of moisture availability in the circum-Caribbean and Andean Altiplano have been out of phase, i.e., negatively correlated. Low-elevation, circum-Caribbean sites were characterized by a dry late glacial period, a moist early and middle Holocene, and a general drying trend over the last ~3000 years. In contrast, the Andean Altiplano was relatively wet during the late Pleistocene period, Tauca stage. Lago Wiñaymarka desiccated when the Altiplano became drier in the early and mid-Holocene. Moist conditions may have returned as early as 4500 ^{14}C B.P. (Baucom and Rigsby, 1999). Insolation is arguably the driving force behind the general trends, assuming that periods of greater summer insolation and increased seasonality were accompanied by higher rainfall in each of the respective hemispheres. Nonetheless, short-term, secular climatic variations in the two localities north and south of the equator may have been influenced by geographic characteristics or processes such as forest cover, orography, and glacial melting.

Sedentary agriculture developed in the two cultural regions of interest ca. 3000 years ago, when the climate became drier in the circum-Caribbean and wetter in the Altiplano. Although agricultural sedentism may not have been a direct response to climate change, environmental conditions together with agricultural innovations enabled population growth and cultural development in both areas. In the centuries that followed, raised-field technology and terracing in the Altiplano increased crop yields and permitted sustainable farming of planting surfaces. In the Maya lowlands, intensive practices such as raised-field construction, terracing, and perhaps arboriculture (Puleston, 1978) supplemented yields from slash-and-burn activities. In both regions, agriculture was probably under state-level control because intensive techniques required infrastructure and organized labor. Innovative technologies greatly enhanced the natural agricultural potential of the regions. These technologies, in combination with storage and long-distance trade, probably protected human populations against minor disruptions in agricultural output.

6.1.7. Abrupt Climate Change and Cultural Response

Abrupt, but persistent droughts are implicated in the collapse of the Maya (Hodell et al., 1995; Gill, 2000) and Tiwanaku cultures (Binford et al., 1997). General temporal correlations between late Holocene dry episodes in the two cultural areas have been commented upon previously. Curtis et al. (1996) noted that several dry periods between A.D. ~600 and 1400, identified isotopically in the Punta Laguna sediment core (Fig. 6), correlate with periods of high, large (>0.63 μm) microparticle (dust) concentrations in the Quelccaya, Peru, ice core. High concentrations of microparticles in the ice core are attributed to aeolian transport of soils that were exposed during initial raised-field construction (Thompson et al., 1988) as well as long-distance transport of soil and desiccated lake sediment during dry events.

Chepstow-Lusty et al. (1996) noted that palynological and sedimentological shifts in a core from Lake Marcacocha in the central Peruvian Andes (13°3′ S, 72°12′ W) are correlated with Peruvian ice core records from Quelccaya and Huascarán (Thompson et al., 1988, 1995). This suggests that these climatic shifts were regionally significant. Influx of silt and charcoal to Lake Marcacocha between A.D. 600 and 700 may correlate with a major dust event recorded at A.D. 620 in the 1500-year Quelccaya ice core record. The Quelccaya ice core displays a second dust event at A.D. 920, which marks the inception of drier, warmer conditions that prevailed from A.D. 1000–1400, during the Medieval Warm Period (MWP) (Lamb, 1982). This dry episode is correlated with the decline of Tiwanaku agriculture (Ortloff and Kolata, 1993; Binford et al., 1997). At Marcacocha, this climatic episode is recorded by the establishment of *Alnus* after A.D. 1040. A dust event in the 15,000-year Huascarán record occurred 2000 years ago and is correlated with a silty, charcoal-rich layer in the Marcacocha profile that is attributed to the effects of flooding and riparian agriculture. At both Quelccaya and Huascarán, relatively depleted $\delta^{18}O$ values reflect the *Little Ice Age* (LIA) (A.D. 1490–1900). This cool period is marked at Marcacocha by the *Alnus* decline. The LIA is also reflected by relatively depleted $\delta^{18}O$ values between ca. A.D. 1400 and 1800 measured on ostracod (*Limnocythere*) shells in core D (Fig. 2a) from Lago Wiñaymarka (Binford et al., 1996).

Chepstow-Lusty et al. (1996) also point out that changes in the Marcacocha core correlate with events recorded in the profile from Lake Chichancanab, México. During the Maya Late Preclassic period (300 B.C.–A.D. 250), a dry episode is reflected in the Chichancanab core at about A.D. 1 by high gypsum concentrations, relatively enriched $\delta^{18}O$ values, and a change in ostracod taxa (Fig. 5). This dry episode in Yucatán is contemporaneous with the deposition of a silty, charcoal-rich layer in Marcacocha, above which cold-adapted *Plantago* spp. become established. The ninth-century dry episode on the Yucatán Peninsula may be linked to warming in Peru that is reflected by the somewhat delayed expansion of *Alnus*, ca. A.D. 1040. If these climatic events at distant sites on both sides of the equator prove to be synchronous, it would suggest global-scale disruption of atmospheric and oceanic fields.

Temporal correlation between climatic drying and cultural collapses at tropical sites north and south of the equator raises fascinating questions for paleoclimatologists and anthropologists alike. The data indicate that climatic thresholds for agricultural production can be exceeded, with disastrous consequences for human populations. Recent paleoenvironmental studies indicate that climate *surprises* during the Holocene, i.e., significant decadal to centennial climate variability, were more common than was previously believed (Overpeck, 1996). Paleoclimatologists must date late Holocene climate changes accurately, evaluate their causes, assess the geographic extent of areas affected, determine the quantitative reduction in moisture availability that occurred during drought events, and investigate the role of human activity (e.g., deforestation) in compounding the effects of climate change. Social scientists must study factors that make cultures resistant or vulnerable to environmental perturbations such as climate changes. Sheets (2000) notes that simple societies tend to recover from environmental stresses such as explosive volcanism more readily than complex societies. Complex societies, with their state-run trade, agriculture, economies, and other facilities, are unable to cope with sudden, unanticipated environmental perturbations. The Maya and Tiwanaku civilizations instituted intensive agricultural methods that overcame natural limits to food production. These technologies increased yields, promoted sustainability, permitted human populations to reach high densities, and created a vast state-level infrastructure. Thus, intensive agricultural methods may have ultimately contributed to the cultural collapses that occurred as a consequence of sudden, unpredictable climate changes.

Paleoclimatic and archaeological information from the Maya lowlands and Bolivian/Peruvian Altiplano sheds new light on the concept of environmental determinism. Moisture availability in the climatically marginal Maya and Tiwanaku regions fluctuated appreciably over time, illustrating that past regional agricultural potential cannot be assessed by using modern climate variables, but rather should be based on historic climate conditions inferred from paleoclimate proxies. The decadal to centennial E/P variability revealed by the paleoclimate record must have presented serious challenges for pre-Columbian agriculturists. Cultural continuity in light of documented climatic variability provides evidence for both the resilience of the Maya and Tiwanaku food production systems and the ingenuity that went into their development. The multidisciplinary approach to investigating human–environment interactions permitted us to demonstrate a strong temporal correlation between protracted drought and cultural collapse in the high-altitude Tiwanaku region and the Maya lowlands. Drought was a major stressor in both regions, which contributed to agricultural and subsequent cultural declines. The findings suggest that cultural development and survival are profoundly influenced by environmental conditions.

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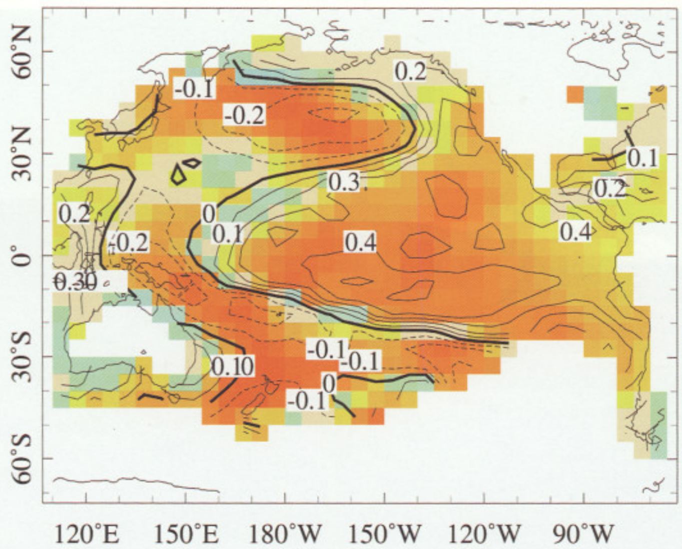
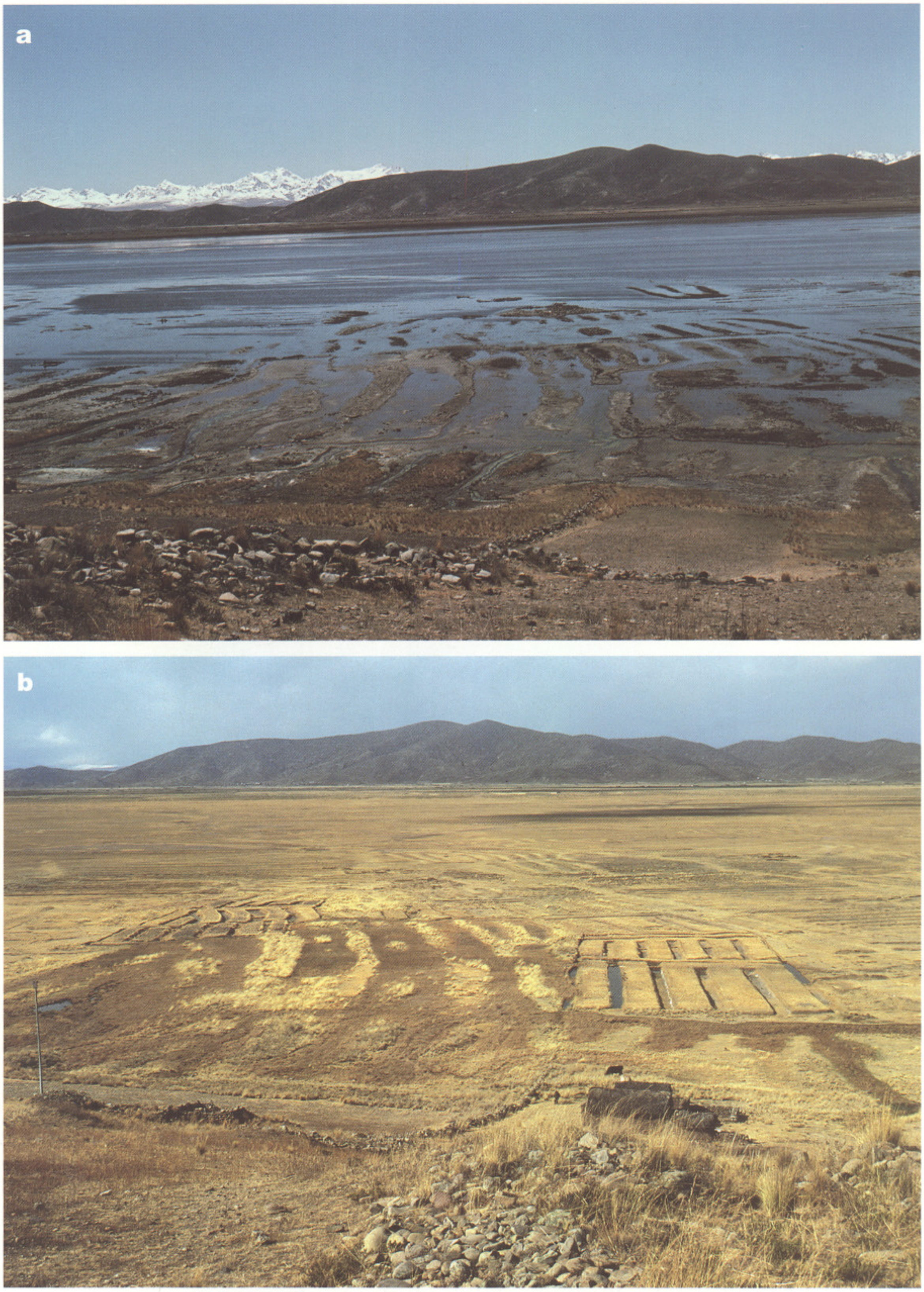


CHAPTER 4, FIGURE 7 Color plot of the verification correlation (Figure 6a) overlain by a contour plot of the correlation with the leading PC of the instrumental data from the tree-ring location (Figure 4b).

CHAPTER 6, FIGURE 4 A Chac (rain god) mask adorns a building at the archaeological site of Labná in the Puuc Region of the Maya lowlands, Yucatán, México. (Photo courtesy of M. Brenner.)



CHAPTER 6, FIGURE 8 Photos of the Pampa Koani, Bolivia, taken in 1986 (a), when Lake Titicaca was at a 20th-century high stand, and in 1996 (b) when the lake level had declined by several meters. Rehabilitated raised agricultural fields are seen in the foreground of (a). The "corduroy" appearance of the landscape is attributed to ancient Tiwanaku raised fields. Small changes in the relation between evaporation and precipitation (E/P) leave the flatlands either inundated or desiccated. (Photos courtesy of M. Brenner.)