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Evidence for Younger Dryas global climate oscillation and human response in the American Southwest

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ABSTRACT

Whether or not abrupt Younger Dryas climate change affected regional paleoenvironments and late Pleistocene hunter-gatherer populations is an important topic in the archaeology of the American Southwest. This paper reviews multiple, age-resolved proxy evidence to gauge the magnitude and direction of Younger Dryas Chronozone (YDC) environmental changes in different settings and systems. There is no record of YDC pluvial lake highstands in Arizona or New Mexico, but there are impressive records of vegetation, faunal, stable isotope, and geomorphological change coincident with the YDC. These correlate with important adaptive changes in human hunting and land use, as revealed in the analysis of the spatiotemporal distribution of late Pleistocene hunting technologies. Clovis and Folsom projectile point distributions do not support extant models of paleoenvironmental conditions in these interpretations. Significant cultural changes that coincide with the YDC include the Clovis-to-Folsom transition, the demise of mammoth hunting and the development of a highly successful emphasis on bison, increased regionalization, and the abandonment of the northwestern Chihuahuan and the Sonoran deserts by mobile, big-game hunters.

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1. Introduction

The topic of Paleoindian adaptive changes related to the Younger Dryas (YD) relies on an assessment of the character, direction, and magnitude of environmental changes that took place in the American Southwest coincident with abrupt temperature anomalies recorded in the Greenland ice cores for a period of roughly 1300 years between approximately 11–10 ka¹⁴C BP (12.9 and 11.6 ka BP (Alley, 2000; Rasmussen et al., 2006; Steffensen et al., 2008)). An increasingly robust picture of paleoenvironmental change is available from a wide variety of regional records. A second challenge relates to the archaeological record immediately before, during, and after the YDC, a period characterized by scant archaeological evidence for what were clearly dynamic social and economic adaptations associated with the late Pleistocene human colonization and regionalization of the North American continent (Kelly and Todd, 1988; Barton et al., 2004; Meltzer, 2004; Beck and Jones, 2010; Erlandson et al., 2011).

Conflicting Clovis-to-Folsom paleoenvironmental reconstructions for the Southwest and the Southern High Plains, whether warmer/drier-to-cooler/wetter (Haynes, 1991), or the opposite (Holliday, 2000), are seemingly at odds with the distribution and abundance of Bølling-Allerød (BA)/YDC (Clovis) and fully YDC (Folsom/Midland) archaeological sites in both regions, based on expectations.

The concept that the last deglaciation terminated with an abrupt shift to cooler conditions throughout the entire northern hemisphere stems from paleobotanical and lithostratigraphic studies carried out at Swedish and Danish bog and lake sites more than a century ago, but Björck (2007) points out that it was not until YD oscillations were identified in highly resolved, proxy-based time-series from North America (Petee, 1987) that the importance of studying the nature and exact timing of these changes was widely recognized. The apparent global scale of YD climate change (Broecker, 1994; Hughen et al., 2000; Wang et al., 2001) – resulting from interhemispheric teleconnections related to the meridional overturning circulation – has since made this climate phenomenon an important consideration in interpreting the archaeological record of cultural changes that occurred during the late

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Pleistocene–early Holocene transition (Munro 2003). Comparisons between temporally well-resolved archives of paleoclimate from high- and mid-latitudes, with particular emphasis on those from continental interiors, are of paramount importance for improving the understanding of global climate change and its cultural impacts.

It is within this context that this paper explores multi-proxy evidence for paleoenvironmental changes in the North American Southwest during the YDC. There is a long history of paleoenvironmental research in the Southwest, focused in part on the latest Pleistocene and early Holocene (e.g., Antevs, 1955; Martin, 1963; Long, 1966; Haynes, 1968a, b; Irwin-Williams and Haynes, 1970; Van Devender and Spaulding, 1979); nonetheless, continuous records spanning the YDC are rare. This comparative study is made possible by several types of paleoenvironmental archives, including speleothem, packrat midden, paleontological, palynological, and stratigraphic records documented in the region.

This review considers regional paleoenvironmental evidence that brackets the BA/YDC transition before outlining the contemporaneous archaeological record in an effort to correlate significant changes in human subsistence, technology, demography, and land use patterns with the onset and end of the YDC. Continuous, site-specific records of environmental and cultural change are invaluable, but a limiting aspect of rare paleoenvironmental reconstructions based on rare archaeological sites is their probable bias in favor of unique places and resources. This is an important consideration in terms of ferreting-out the choices and decisions made by foragers. The authors acknowledge this bias and attempt to minimize it by using a landscape-scale sample of time-sensitive projectile point types as an anthropogenic proxy for favorable hunting conditions, which carries paleoenvironmental implications in terms of primary productivity and the abundance of plants and animals.

Regional and continental-scale analyses based on Paleoindian projectile point distributions have originated from the Southeastern United States and the Southern Plains (Anderson, 1990; Meltzer and Bever, 1995; Anderson and Faught, 1998; Anderson and Gillam, 2000; Prasciunas, 2011). Similar efforts in the Southwest have concentrated on smaller geographical areas (Judge, 1973)

or were limited in theoretical scope (Agenbroad, 1967; Huckell, 1982), but are becoming increasingly systematic and problem-oriented (Amick, 1996; Huckell, 2004).

1.1. Study area

The American Southwest includes geologically and environmentally distinct physiographic regions, including the southern Basin and Range physiographic section, the Rio Grande Rift, and the southern Colorado Plateau. Reed (1964) offers a useful shorthand definition of the Southwest as extending from Durango, Colorado, to Durango, Mexico, and from Las Vegas, New Mexico, to Las Vegas, Nevada (Fig. 1). The landforms and landscapes of the Southwest are characterized by dozens of block-fault mountain ranges and desert basins. Some of the basins remain closed, but drainages such as the Gila River and the Rio Grande integrated most of them during the late Pliocene to early Pleistocene.

Influenced by the North American Monsoon (NAM), a shift in atmospheric circulation resulting from westward expansion of the Bermuda High and locally expressed by intense summer rains (Adams and Comrie, 1997), the Southwest experiences a semi-arid climate with two pronounced wet seasons interrupted by essentially warm/dry seasons. Elevations ranging from 50 m in the Gila River Valley of Arizona, to above 4000 m in the Southern Rocky Mountains of New Mexico, experience pronounced orographic controls in mean annual temperature (MAT) and precipitation (MAP). MAP generally increases from west-to-east across the study area, or with increasing elevation, with summer inputs increasing with decreasing latitude. In the southern Southwest in the vicinity of Tombstone, Arizona (1405 m), historic measurements show that winter precipitation is concentrated between December and February, when ~20% of MAP occurs. Intense summer rains (July–September) deliver most (~60%) of MAP, but are short in duration, producing surface runoff in the local arroyos and brief continuous flow in high-order streams. Maximum summer temperatures average ~34 °C between June and August, with minimum temperatures averaging ~1 °C between December and February. The Southern Colorado Plateau and Southern Rocky



Fig. 1. Approximate location of physiographic features, paleoenvironmental records, and archaeological sites discussed in the text.

Mountains of Arizona and New Mexico experience average maximum and minimum temperatures between 10° and 0.6 °C, but summer monsoon rains account for slightly less than 50% of MAP. Summer rains are less than 40% of MAP in the northern and westernmost portions of Arizona (Western Regional Climate Center, www.wrcc.dri.edu).

1.2. Methods

The paleoenvironmental summary is based on a literature review supplemented by new information provided by Holliday, Ballenger, Kowler, and Reitze. The authors acknowledge from the outset that special attention is paid to the Basin and Range of southeastern Arizona and New Mexico where these research programs are being conducted, and where the Paleoindian archaeological record is best documented. High elevation records from the Southern Colorado Plateau and Rocky Mountains are under-represented due to the sampling bias. The western boundary is Lake Mojave, which is located west of Las Vegas, Nevada, only insofar as for the paleolake review and to provide some contrast to the Chihuahuan records. The eastern boundary is the Blackwater Draw site (Fig. 1), which, although located in the Southern High Plains, shows that important changes occurred coincident with the YDC across the Southwest – Southern High Plains boundary.

Paleoindian projectile point type distributions are based on the PIDBA dataset (Anderson et al., 2010; online at <http://pidba.utk.edu/main.htm>), with the Clovis data supplemented by Haynes (2011) and Prasciunas (2011), and the Arizona Paleoindian and Paleoarchaic Projectile Point Survey (<http://azpaleosurvey.pidba.org/>), The Folsom data is supplemented by Amick (2006), and Amick and Lukowski (2006). Hill and Holliday (2011) also provide Clovis and Folsom projectile point counts based in large part on the collection of the late Robert H. Weber. These sources resulted in a sample of 1960 projectile points, including 109 Clovis points and 33 Folsom/Midland points from Arizona, and 292 Clovis points and 1526 Folsom/Midland points from New Mexico. Although both Folsom and Midland projectile point types are included in the dataset, their distributions are not compared, and they are referred to collectively as Folsom throughout the paper. The number of documented projectile points per 1000 km² for each county was calculated. Whereas these sources provide a robust sample of fluted Paleoindian projectile points from Arizona and New Mexico, this survey was not exhaustive.

Scientific bias clearly plays a fundamental role in the creation of these data, and this is evident in the Paleoindian record in the Southwest (Huckell, 2004). Additionally, surface geology, ground cover, cultivation practices, and other variables can affect projectile point exposure and visibility. Projectile point distributions and especially densities should therefore be interpreted cautiously (Prasciunas, 2011). This study assumes that the distribution of projectile points provides a reasonable proxy for the distribution of Paleoindian hunters and, to an uncertain extent, human demography.

Ice core, uranium-series, optically stimulated luminescence (OSL), and radiocarbon ages are cited using the 10³ notation “ka.” Ice core, uranium-series, and OSL dates are presented using the “BP” notation (before present = 1950). Uncalibrated radiocarbon dates are expressed using the notation “¹⁴C BP,” whereas radiocarbon ages calibrated using OxCal 3.10 (Bronk Ramsey, 2001) based on the Intcal09 dataset (Reimer et al., 2009) are expressed as “cal BP.” The radiocarbon calibration curve is notoriously problematic during the YDC (Muscheler et al., 2008), and is continually refined, so both calibrated and uncalibrated midpoints are cited where appropriate.

2. YDC environmental records in the Southwest

The YD was originally recognized based on paleobotanical and stratigraphic evidence, but the YDC is most apparent in subdecadal-scale Greenland ice core records (Alley, 2000). High-resolution $\delta^{18}\text{O}$ time-series derived from U-Th dated speleothem calcite offer a unique opportunity to observe decadal and possibly subdecadal changes in temperature and precipitation during the latest Pleistocene in the arid Southwest (Polyak and Asmerom, 2001; Betancourt et al., 2002, 2003; Baker and Genty, 2003), but speleothem records covering the LP/EH transition are rare and have only recently become the focus of concentrated study. Before this, regional paleoclimate reconstructions were informed by well-dated macrobotanical assemblages in packrat middens, and a fewer number of dated pluvial/paleolake shoreline, pollen, paleontological, and stratigraphic records, which are reviewed and summarized separately.

2.1. Speleothems

Speleothem $\delta^{18}\text{O}$ measurements in the Southwest are assumed to be near-direct proxies for variation in precipitation $\delta^{18}\text{O}$ (Asmerom et al., 2010), which records changes in the contribution of summer and winter precipitation, precipitation amount, and temperature (Dansgaard, 1964; Rozanski et al., 1993). The former is possible because the stable oxygen isotope composition of modern precipitation correlates with season and amount more so than it does with temperature in the Southwest; modern summer monsoon precipitation originating in the tropical Pacific tends to be more enriched in ^{18}O than winter frontal precipitation originating from the temperate Pacific (Wright et al., 2001; Wahi et al., 2008). Colder climates that persist at high latitudes push the westerly storm track (polar jet) further south, resulting in a greater contribution of winter precipitation to MAP in the Southwest (Van Devender and Spaulding, 1979; Kutzbach, 1987; Asmerom et al., 2010). Three records from the Southwest demonstrate strong directional and chronological correlations between high latitude temperature oscillations inferred from the Greenland ice core records, and the growth and/or $\delta^{18}\text{O}$ composition of regional speleothems. A fourth record does not indicate a terminal YDC rebound to antecedent conditions, but it does not completely bracket the period of interest.

The FS-2 stalagmite from Fort Stanton Cave, southeastern New Mexico, is based on 68 high-precision U-Th dates documenting the rate of stalagmite growth occurring between 55.9 and 11.4 ka BP (Asmerom et al., 2010). Assuming that $\delta^{18}\text{O}$ values in precipitation are controlled mostly by air temperature, seasonality, and rainfall amount, the FS-2 speleothem is interpreted to indicate abrupt and sustained warming and/or increased summer moisture beginning at approximately 15 ka BP. The YDC is marked by a sharp decrease in $\delta^{18}\text{O}$ values ($\sim 2\text{‰}$) shown to begin around 13 ka BP and recovering to the range of pre-YDC values by 11.7 ka BP. When calculated using the 1950 datum, the $\delta^{18}\text{O}$ decrease begins at ~ 12.9 ka BP based on the age model (Asmerom et al., 2010:SI). These values are interpreted to indicate increased winter precipitation and/or cooler temperatures throughout most of the YDC (Asmerom et al., 2010).

Uranium-series dating of six small speleothems from three separate caves in the Guadalupe Mountains in southeastern New Mexico records stalagmite formation between approximately 30 and 10.5 ka BP, based on 37 U-Th dates (Polyak et al., 2004). Because calcite growth is moisture limited, Polyak et al. (2004) argue that variable speleothem growth in the Guadalupe speleothems indicates drier conditions during the BA and earliest YDC, followed by wetter conditions by 12.5 ka BP and drying no later than 10.5 ka BP, coincident with maximum summer insolation (Berger and Loutre,

1991). However, Polyak et al. (2004) infer an increasingly wetter YDC from beginning to end. $\delta^{18}\text{O}$ values from Pink Panther Cave (PP1), also in the Guadalupe Mountains, do not bracket the YDC but indicate Holocene-range values by 12.1 ka BP (Asmerom et al., 2007).

The Cave-of-the-Bells (COB-01-02) stalagmite record from the Santa Rita Mountains in southeastern Arizona brackets the BA/YDC transition (Wagner et al., 2010). The LP/EH portion of the COB-01-02 speleothem chronology is controlled by 61 U-Th dates ranging between approximately 52.7 and 10.3 ka BP (Wagner, 2006). However, the ages are not in order (multiple reversals occur). Assuming that speleothems record mostly winter precipitation amount, a $>2\%$ increase in $\delta^{18}\text{O}$ values (ice-volume corrected) is interpreted to indicate drier conditions beginning at approximately 15.3 ka BP. A 0.7% decrease in $\delta^{18}\text{O}$ values is argued to indicate the onset of moister conditions at 13 ka BP (Wagner et al., 2010), but $\delta^{18}\text{O}$ values do not decline below Older Dryas minimums until ~ 12.9 ka BP based on the age model. The end of moister winter conditions is estimated to have occurred around 11.5 ka BP, but this date is limited by a hiatus in speleothem growth at this time.

In summary, three of the four best-resolved archives of climate change in the arid Southwest show abrupt changes associated with BA warming and/or drying by 15 ka BP, and increased winter precipitation and/or cooler conditions at ~ 12.9 (Asmerom et al., 2010; Wagner et al., 2010), or ~ 12.5 ka BP (Polyak et al., 2004). The PP1 stalagmite record (Asmerom et al., 2007) is anomalous.

2.2. Packrat middens

Packrat (*Neotoma*) middens from the Southwest generally do not indicate significant warming at ~ 15 ka BP and a return to cooler/wetter conditions during the YDC, but rather stable piñon-juniper-oak woodlands throughout the late Pleistocene under a winter-dominated precipitation regime (Van Devender and Spaulding, 1979). In the northern Chihuahuan Desert, late Pleistocene piñon-juniper-oak woodlands persisted until ~ 11 ka¹⁴C BP (~ 12.9 ka cal BP), followed by xeric early Holocene (EH) communities ranging from juniper woodlands to shrub oak-grasslands (Van Devender, 1990a). The upslope retreat of piñon (*P. monophylla*) from the Sonoran Desert also occurred ~ 11 ka cal BP, leaving juniper-oak woodlands, chaparral shrub, and modern desert plants including saguaro (Van Devender, 1990b; Van Devender et al., 1994). In the summer-dry central Mojave Desert, increased aridity is signaled by the departure of piñon by 11.5 ka¹⁴C BP (~ 13.3 ka cal BP) (Koehler et al., 2005). By 8 ka¹⁴C BP juniper is gone from most records and grasslands and desert scrub expand, implying warmer temperatures and a switch to summer-dominated precipitation (Van Devender and Spaulding, 1979). The lack of grass-dominated Pleistocene packrat middens has led Van Devender (1990a, 1995) to suggest that desert grasslands were extensive only during relatively brief interglacial periods, and that this explains the sporadic presence of bison in the Southwest after 11 ka¹⁴C BP.

More recently, Holmgren et al. (2006) have reported a decline in piñon (*P. edulis*) beginning around 13 ka¹⁴C BP (15.6 ka cal BP) in the Peloncillo Mountains of southeastern Arizona, coincident with BA warming, followed by its latest occurrence during the YDC at 10.3 ka¹⁴C BP (~ 12.1 ka cal BP). Of interest is the antiquity and diversity of summer annuals and grasses, because general circulation models for the late Pleistocene predict the near-elimination of warm season rains due to a southern displacement of the westerlies (COHMAP Members, 1988; Kutzbach et al., 1998). Packrat middens with YDC radiocarbon dates indicate diverse summer annuals and C₄ grasses at mid elevations as evidenced by two middens from the Playas Valley in southwestern New Mexico (Holmgren et al., 2003), and three from the Peloncillo Mountains

of southeastern Arizona (Holmgren et al., 2006). In fact, Holmgren et al. (2007) model late Pleistocene C₄ grasslands extending from central Texas to southern Arizona below 35° N. The packrat midden records of the southern Southwest are contrasted by those of the southern Colorado Plateau, where summer grasses are scarce in late Pleistocene middens (Betancourt, 1990), indicating that the northern reach of summer monsoon moisture was truncated at that time (Holmgren et al., 2007). In the Grand Canyon, however, packrat pellets (Cole and Arundel, 2005) as well as bat guano (Wurster et al., 2008) experience $\delta^{13}\text{C}$ minima during the YDC.

2.3. Pollen records

Well-dated pollen records bracketing the YDC are available from a limited number of upper- and mid-elevation bogs and small lakes, and still fewer intermediate elevation valley sites (Hall, 2005). Anderson (1993) briefly summarizes pollen records from three small lakes located on the Southern Colorado Plateau (2200–2800 m) in central Arizona, elevations presently dominated by extensive ponderosa pine (*P. ponderosa*) forests. The dominance of spruce and fir pollen types during the middle and late Wisconsin show generally cooler and moister conditions, but the chronology of these records relative to the YDC is not good. At Potato Lake, the BA/YDC transition is bracketed by two dates between about 14.4 and 9.9 ka¹⁴C BP (~ 17.5 – 11.5 ka cal BP), an interval characterized by closed spruce forests and maximum lake levels until ~ 12.3 ka cal BP when near-modern vegetation was established with the appearance of ponderosa pine.

A uniquely preserved paleoenvironmental record occurs at Chihuahueños Bog (2925 m), a small, spring-fed basin located in the Jemez Mountains of northern New Mexico. Anderson et al. (2008) obtained two dates, 11.85 and 9.1 ka¹⁴C BP (~ 13.7 – 10.3 ka cal BP), bracketing the BA/YDC transition. Pollens and macrofossils show the establishment of spruce parkland after 13.7 ka cal BP that persisted through the YD oscillation. Anderson et al. (2008) do not observe climatic fluctuations at the onset of the YDC, but their sampling was not continuous. However, using elemental and stable C and N isotopes, Cisneros-Dozal et al. (2010) have distinguished the early YDC in terms of a reduction in terrestrial productivity, followed by an increase in aquatic productivity during the late YDC. An increase in ponderosa pine, a reduction in *Artemisia*, and the transition from a small pond to a wetland occurred after 11.7 ka cal BP based on the age model, marking warming and possibly an expansion of the summer monsoon after the YDC (Anderson et al., 2008).

The only radiocarbon dated pollen record from a low elevation lake comes from Montezuma Well (1125 m), an artesian spring surrounded by Sonoran Desert scrub and grasslands in central Arizona. Davis and Shafer (1992) obtained radiocarbon dates of approximately 10.98 and 9.5 ka¹⁴C BP (~ 12.8 and 10.9 ka cal BP) from the lower portion of an 11 m sediment core, effectively spanning the YDC-EH transition. It is difficult to assess the reliability of the chronology due to a single age reversal within the YDC, as well as a possible hard water effect. Taking the chronology at face value, the YDC is not associated with a significant change in pollen composition, although the authors note the presence of EH piñon-juniper-oak woodlands (Davis and Shafer, 1992).

The late Pleistocene pollen evidence from New Mexico is summarized by Hall (2005), who describes environments ranging from alpine tundra to sagebrush grasslands, but the YDC is not securely bracketed in those sequences. Two YDC pollen records are available from the upper San Pedro River Valley in the Chihuahuan desert of southeastern Arizona. Pollens from Clovis and YDC (black mat) deposits at the Lehner site (1280 m) indicate a modern “desert

grassland” during Clovis times, with grass pollens accounting for 20–25% of the pollen sum (Mehringer and Haynes, 1965); *Pinus* did not exceed 5% in the surface samples and did not exceed 10% in Clovis-age sediments. Pollen samples from the overlying “black mat” are marked by a significant increase in *Ambrosia* and a decline in grass pollens that continues well beyond the YDC (Mehringer and Haynes, 1965).

The discovery of YDC alluvium in lower Palominas Arroyo, located 5 km upstream from the Lehner site, provides a pollen record from the riparian corridor of the river (Fig. 2). *Pinus* percentages at Palominas (1296 m) exceed 40% between approximately 10.7 and 9.9 ka¹⁴C BP (~12.6–11.3 ka cal BP), whereas grass pollen are no more than 3% (Ballenger, 2010a). YDC deposits in Palominas Arroyo contain pollens from riparian vegetation including willow, caittail, and sedges, with an upland component indicating *Pinus-Juniperus*-oak woodlands. A pollen sample from an adjacent core, dated to approximately 11.1 ka¹⁴C BP (~12.9 ka cal BP), also indicates that upland vegetation was dominated by *Pinus* woodlands. Unlike the “desert grassland” evidenced at the Lehner site, grasses were a minor component in the floodplain until about 8.85 ka¹⁴C BP (~10 ka cal BP) based on pollens and stable carbon isotopes from soil organic matter (Ballenger, 2010a). At face value, the pollen records from Lehner and Palominas indicate either a transition from desert grasslands to piñon-juniper-oak woodlands at ~12.9–12.8 ka cal BP, or accumulation and taphonomic vagaries between sampling sites. Additional high-resolution sampling is necessary to better reconstruct paleovegetation changes spanning the Clovis-YDC interval.

2.4. Paleontology

The best-dated late Pleistocene paleontological records of the Southwest mark the disappearance of a suite of large animals, namely mammoth, horse, camel, peccary, tapir, Dire wolf, and the short-faced bear in the San Pedro Valley (Haynes and Huckell, 2007; Saunders and Baryshnikov, in preparation), and the gomphothere *Cuvieronius* in northern Mexico (Sanchez et al., 2009), contemporary with the appearance (and disappearance) of Clovis in the stratigraphic record. Pleistocene extinctions are dated to approximately 10.9 ka¹⁴C BP (~12.8 ka cal BP) at Lehner, Murray Springs, and Blackwater Draw (Morgan and Lucas, 2005; Haynes

and Huckell, 2007), and sometime after 11 ka¹⁴C BP (~13 ka cal BP) at El Fin del Mundo (Sanchez et al., 2009). Grasslands or grassland savannas are strongly indicated by the terminal Rancholabrean fauna of the Southwest, but with disharmonious faunal associations indicating more equable climates and “no-analog” vegetational associations (Lundelius, 1989; Morgan and Lucas, 2005). Hall (2005) summarizes faunal evidence of a late Pleistocene sagebrush steppe in southern New Mexico.

Paleovegetation reconstructions based on the paleontological record increasingly rely on stable isotope values from close-interval samples of tooth enamel, which record the proportion of C₃/C₄ plants in herbivore diets (Cerling et al., 1997). Modern C₄ feeders have average δ¹³C values between 1 and 3‰, whereas C₃ feeders have average values of –12 to –13‰. Stable carbon isotope values from mammoth, horse, camel, and bison tooth enamel from Murray Springs and Blackwater Draw point to the consumption of C₄ grasses by these taxa during Clovis times (Connin et al., 1998). Clovis-age bison at Murray Springs show δ¹³C values between 0.9 and –1.7‰; Clovis-age bison at Blackwater Draw exhibit values from 0.4 to –1.3‰, but the sample sizes are small in number. Metcalfe (2009) reports sinusoidal δ¹³C values among Clovis-age mammoths in the San Pedro Valley, indicating a strong summer monsoon. YDC fauna are rare, but bison tooth enamel fragments from the Boca Negra Wash Folsom site in the middle Rio Grande Valley also show a reliance on warm season grasses, with δ¹³C values rarely less than –7‰ at Boca Negra Wash (Mullen, 2008:57). δ¹³C values of bison bone from the Folsom kill site show a similarly large contribution of C₄ plants (Meltzer, 2006:198).

2.5. Pluvial lakes and playas

Additional indicators of YDC climatic change in the Southwest come from pluvial lake deposits that occupy undrained intermontane basins in the southern Basin and Range, most of which occur within the Mexican Highlands section between the Rio Grande Rift and the Peloncillo Range to the west, along the New Mexico-Arizona border. Prominent shoreline features are found in the Animas, Mimbres, Playas, Hatchita-Moscós, Tularosa, Estancia, San Agustín, Salt, and Willcox Basins (Meinzer and Kelton, 1912; Hawley, 1993; Allen, 2005; Connell et al., 2005). Several of these valleys contain firm geochronological evidence, from shoreline

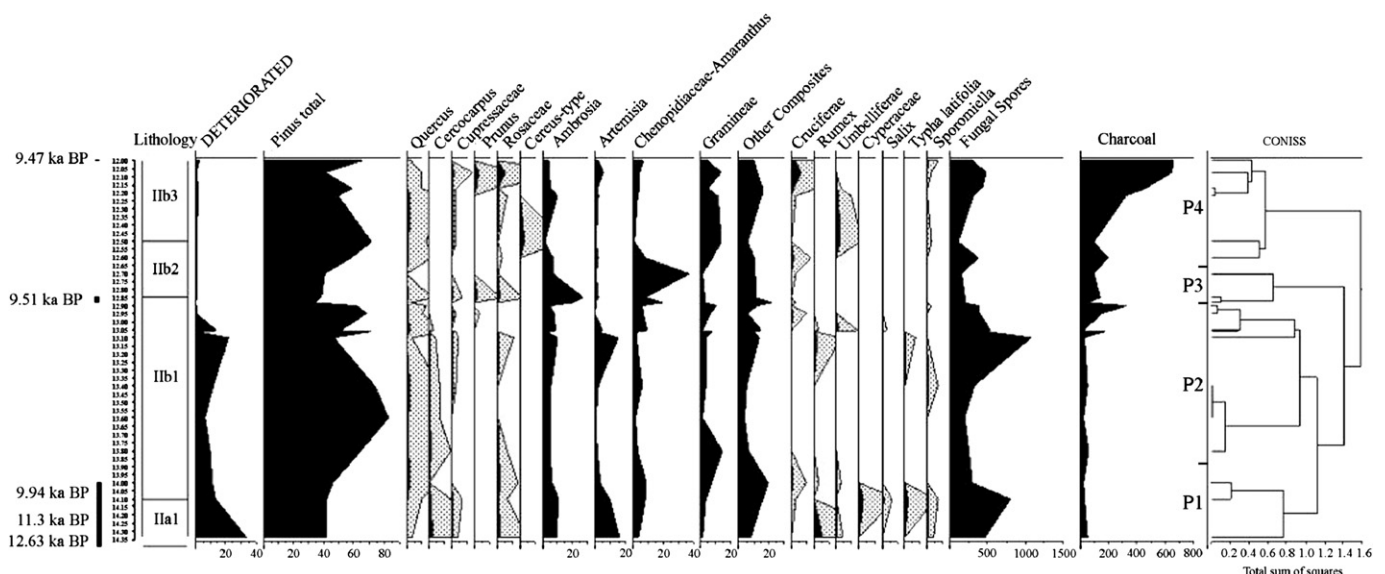


Fig. 2. Palominas Arroyo, pollen percentages for select taxa from Core 2 and CONISS cluster analysis of pollen types (Grimm, 1987). Rare taxa exaggerated (10×) (shaded).

deposits and their related deep water facies, for extensive pluvial episodes occurring both during the LGM and Late Glacial—including Paleolakes San Agustin, Playas, Cloverdale, and Cochise in the Mexican Highlands, and Paleolake Estancia in the adjacent Rio Grande Valley (Waters, 1989; Wilkins and Currey, 1997; Allen and Anderson, 2000; Anderson et al., 2002; Langford, 2003; Hill and Holliday, 2011). While deposits comprising the highest geomorphic shoreline features in most of these basins have yielded dates between 20 and 9 ka¹⁴C BP, none date to the YDC. Paleolake records can be valuable sources of paleoclimatic information, depending on how faithfully they record paleolake surface area—the proper gauge of a lake's response to a change in steady-state climatic conditions. While the presence of paleoshorelines enables absolute measurement of paleolake surface area, records based on these features are chronologically discontinuous. Nonetheless, well-dated shoreline records enable a comparison of steady-state effective moisture levels during successive humid intervals with enough intensity and/or duration to result in the creation of a resilient shoreline feature.

2.5.1. Paleolake Estancia

The ¹⁴C-based chronology of lake level fluctuations for Paleolake Estancia, located southeast of Albuquerque (Fig. 1), is the synthesis of two different data sets: 1) the sedimentology and ostracode biostratigraphy-salinity associations of basin center deposits (Allen and Anderson, 2000; Anderson et al., 2002), and 2) the geomorphology of eolian and shoreline landforms, and stratigraphy of lacustrine deposits buried beneath, and 3) the stratigraphy of draw fill along the northern margin of the lake basin. Taken together, these records indicate that Paleolake Estancia experienced maximum freshening and reached a minimum elevation of 1885 m on several occasions, from >19.0 to 15.0 ka¹⁴C BP (22.6–18.3 ka cal BP), at ~13.8 ka¹⁴C BP (16.9 ka cal BP), and by 12.5 ka¹⁴C BP (14.7 ka cal BP) until 12.0 ka¹⁴C BP (13.9 ka cal BP) or later, with strong evidence for an extended period of saline conditions from ~15.0–13.8 ka¹⁴C BP (18.3–16.9 ka cal BP). By 11.5 ka¹⁴C BP (13.4 ka cal BP), the lake had either desiccated or dropped to low stand levels, returning only to 1862 m by 9.7 ka¹⁴C BP (11.1 ka cal BP) and to 1875 m by 9.2 ka¹⁴C BP (10.4 ka cal BP). Subsequent deflation of the basin center section precludes direct knowledge about conditions during the YDC.

Combined with the above record, there is geomorphic evidence for complete desiccation during part or all of the period between 11.5 and 9.7 ka¹⁴C BP (13.4 and 11.1 ka cal BP), interrupted by a short-lived low stand reaching ~1862 m. Reworked from and burying the laminated lake bed sediments, a gypsum sand sheet provides direct evidence for the previously surmised interval of deflation, while overlying dune-like landforms indicate the subsequent reworking of the gypsum bed. Based on the peculiar shape of these eolian landforms, Anderson et al. (2002) postulated that they formed as dunes and were subsequently modified by subaqueous processes, concluding that a short-lived low stand occurred at some point during this interval.

2.5.2. Paleolake San Agustin

The Plains of San Agustin, located in east-central New Mexico (Fig. 1) yields Paleoindian archaeological data and limited paleo-environmental information for the late Pleistocene. The San Agustin Basin is ~90 km long and divided into three interconnected sub-basins connected by channels: the Horse Spring (2065 m), C-N (2102 m), and White Lake (2120 m). The highest lake stand covered most of the sub-basins and fluctuated between 2110 and 2120 m from ~20.9–19.0 ka¹⁴C BP (25.0–22.6 ka cal BP), coinciding with the first part of the LGM (Markgraf et al., 1983, 1984; Phillips et al., 1992). While palustrine conditions persisted in the adjacent and

higher C-N Basin from ~15.0 to 10.4 ka¹⁴C BP (~18.2–12.3 ka cal BP), the Horse Spring Basin—the lowest-lying and largest sub-basin—experienced several oscillations and an irregular decrease in lake level during this time. ¹⁴C dates from Bat Cave (W. Wills, pers. communication, 2005; Hill and Holliday, 2011) show that lake levels had dropped to 2105 m or lower by ~11.3 ka¹⁴C BP (13.2 ka cal BP), after which the remaining alkaline lake was confined to the Horse Spring Basin until 5 ka¹⁴C BP (5.7 ka cal BP) or later. Thus, during the YDC, the level of Paleolake San Agustin was near its Pleistocene minimum. However, a lack of knowledge about the chronology of shoreline landforms, the need for a more inclusive core record, and the geochronologic uncertainty of the Horse Spring Basin record, including a possible reservoir effect, renders knowledge about conditions associated with the YDC—including conditions immediately preceding and following the YDC—unclear.

2.5.3. Paleolake Playas

The poor preservation of shoreline features in the Lower Playas Valley, located in extreme southwestern New Mexico, has hampered lake level reconstructions, but a preliminary interpretation of limited available stratigraphic information suggests three distinct post-LGM lake stands. These include a low stand from >16.1 to 15.3 ka¹⁴C BP (19.3–18.6 ka cal BP), a high stand from >15.2 to 14.7 ka¹⁴C BP (18.5–17.9 ka cal BP) or later, and a low stand from >13.4 to 11.6 ka¹⁴C BP (16.6–13.4 ka cal BP) or later (Kowler, unpublished data). While dates constraining the more recent low stand are reliable, a reservoir correction of up to 1.3 ka may apply to earlier lake stands. With respect to the YDC, this suggests either that Paleolake Playas underwent complete desiccation, or that lake levels did not rebound as high as they had prior to 11.6 ka¹⁴C BP (13.4 ka cal BP).

2.5.4. Paleolake Cloverdale

The most prominent shoreline in the Upper Animas basin, located in “the boothill” of southwestern New Mexico, records a stand of Paleolake Cloverdale between 20 and 18 ka¹⁴C BP (23.9–21.5 ka cal BP) (Krider, 1998). Shoreline deposits from three subsequent lake stands, which rose to nearly the same elevation as the LGM high stand, all post-date 5.0 ka¹⁴C BP (5.7 ka cal BP). Other than this, there is no indication of lake level status during the latest Pleistocene and Holocene.

2.5.5. Paleolake Cochise

Geochronology of the prominent beach ridge along the western periphery of Willcox Playa, located in southeastern Arizona, has been the subject of several studies (Long, 1966; Haynes et al., 1987; Waters, 1989). Waters obtained ages for the lower portion of a gravel unit comprising the 1274 m shoreline, deposited between >13.8 and 13.4 ka¹⁴C BP (>17.0 and 16.4 ka cal BP), and Kowler (unpublished data) has since dated the upper portion of this gravel unit from >12.7 to 12.0 ka¹⁴C BP (15.0–13.9 ka cal BP) or later. Paleolake Cochise did not return to this level until 8.9 ka¹⁴C BP (>10.1 ka cal BP) or slightly earlier (Waters, 1989). The status of lake levels during the YDC is unknown at this time, but they were likely low relative to dated shorelines.

2.5.6. Paleolake Mojave

The paleohydrology of pluvial Lake Mojave located at the present terminus of the Mojave River in southeastern California provides pre-LGM evidence for climatic change in the Mojave Desert (Wells et al., 2003). A high-resolution stratigraphic sequence coupled with surface and subsurface shoreline features in the Silver Lake basin, one of pluvial Lake Mohave's depositional basins, indicates that two major high stands occurred from 18.4 to 16.1 ka¹⁴C BP (22.0–19.3 ka cal BP) (Lake Mojave I) and 13.7–11.4 ka¹⁴C

BP (16.8–13.3 ka cal BP) (Lake Mojave II). The Lake Mojave II high stand suggests increased precipitation during this time; however, surface water contributions to pluvial Lake Mojave are likely associated with increased Mojave River discharge and higher magnitude, higher frequency, floods (Wells et al., 2003). It is therefore uncertain how the paleohydrology of pluvial Lake Mojave relates to regional climatic conditions. Historically, however, increased precipitation in the Transverse Ranges corresponds well with an increase in annual precipitation across the Mojave Desert (McDonald et al., 2003). A lake stand occurred from 11.1 to 10.8 ka¹⁴C BP (13.0–12.7 ka cal BP) during the first part of the YDC. This occurred during the Intermittent Lake III phase (11.4–8.7 ka¹⁴C BP; ~13.3–9.6 ka cal BP), when the significant reduction of basin volume due to increased sediment storage led to an increase in the lake surface area to basin area ratio, and thus an increase in the sensitivity of lake surface area to changes in effective moisture. Further, diminished recharge from the Mojave River minimized its interference with the climate signal. An abundance of desiccation cracks infilled with eolian and fluvial sands suggest playa conditions persisted during the Intermittent Lake III phase, and complete drying of Lake Mojave did not occur until shortly after 9.6 ka¹⁴C BP (10.9 ka cal BP) (Wells et al., 2003).

2.5.7. Laguna Babicora

Only one dated lake record is known from northwestern Mexico, in Northwest Chihuahua, for which Ortega-Ramírez et al. (1998) reconstruct paleoenvironmental conditions in Laguna Babicora throughout the late Pleistocene and early Holocene. This record is based primarily on ostracode biostratigraphy-salinity associations and on sedimentological, geochemical, and stratigraphic data derived from lake-bottom sediment cores. Unfortunately, geochronologic uncertainties resulting from a dearth of ¹⁴C ages in conjunction with possible ¹⁴C reservoir effects limit the present age model. It may therefore be fortuitous that two ¹⁴C ages bracket the YDC, for which the authors infer fluctuating but increased levels of effective moisture.

2.5.8. Upland Playas

Small playas occur on the “West Mesa” of Albuquerque in the central Rio Grande valley. This upland includes older, higher terraces of the Rio Grande and basalt flows that mantle some of the higher surfaces (Crumpler, 1999; Smith et al., 1999; Connell et al., 2007). Eolian sands cover both the basalts and alluvial surfaces. A regional sand sheet on top of the basalt is dated ~23 ka BP (OSL) (Holliday et al., 2006a; Hall et al., 2008). Where preserved, it has a well-developed soil (Bt-Btk horization) sometimes buried by late Holocene sands. This soil demarcates the late Pleistocene landscape on which Paleoindians lived (Holliday et al., 2006a). There are numerous small depressions on the uplands where sand sheets cover hollows in the basalt flows, and these depressions acted as small lake and playa basins.

The stable carbon isotope values and C/N ratios (the latter indicative of terrestrial vs aquatic vegetation inputs) from the Boca Negra Wash Folsom site on West Mesa indicate generally unidirectional warming and drying trends occurring from the latest Pleistocene through the middle Holocene (Holliday et al., 2006a) (Fig. 3). $\delta^{13}\text{C}$ values from organic matter are more negative (more C₃ vegetation) in the earliest part of the record (~14 ka¹⁴C BP), ranging from –20‰ to –16‰. C/N values range between 5.0 and 8.0, but are consistently the most negative between 12 and 10.2 ka¹⁴C BP (~13.8–11.9 ka cal BP). The isotopic and elemental data suggest that more organic matter was derived from aquatic sources under cooler conditions between ~12 and 10 ka¹⁴C BP. During this time, the Playa basin likely held water more often than before, perhaps seasonally, and it was probably characterized by a dense

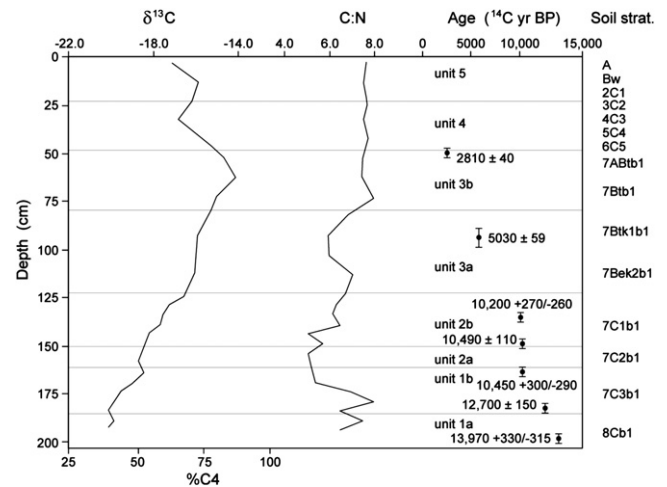


Fig. 3. Uncalibrated radiocarbon dates, stable carbon isotopes, organic carbon data, C/N ratios, and sand fractions from Trench 9, Boca Negra Wash Folsom site (from Holliday et al., 2006a).

cover of vegetation during the latest Pleistocene and earliest Holocene, including the YDC.

2.6. Alluvial stratigraphy

Alluvial stratigraphic records are a primary source of information in the study of Paleoindian environments because they contain some of the best preserved archaeological records. The Rio Grande Rift is a principal location for studying the alluvial chronology of Paleoindian sites in the Southwest. The upper Rio Grande flows through a rift that runs from south-central Colorado down into north-central Mexico and far western Texas (Fig. 1). The rift includes two hydrologic systems. The primary system is the Rio Grande itself, which flows through a series of interconnected structural basins. The Rio Grande Valley is defined as that part of the Rift with alluvial landforms and sediments related to the mainstream Rio Grande. The other principal hydrologic system is a series of closed depressions (reviewed above) in structural basins along the flanks of the river and the rift. Although dry now, these basins contained lakes during the late Pleistocene.

The Upper Rio Grande and its tributaries comprise one of the largest alluvial systems in the western United States. A variety of specific landscape and stratigraphic contexts encompasses the YDC. Thick deposits of alluvium are present throughout the valley in the form of well-preserved terraces, pediments, alluvial fans, and bajadas (e.g., Gile et al., 1981; Pazzaglia and Wells, 1999; Connell and Love, 2001; Connell et al., 2005). Large areas of eolian cover, including sand sheets and dune fields, occur locally on these surfaces.

Mainstream alluvium in the Albuquerque area is the most well documented late Quaternary stratigraphy along the upper Rio Grande (e.g., Connell et al., 2007). Sediments comprising axial-fluvial and alluvial fan deposits were deposited under high energy conditions, with the balance between deposition and erosion maintained throughout the system. At some point near or during the YDC, the floodplain was incised, creating the lowest terrace of the central Rio Grande. The subsequent cessation of downcutting resulted in formation of the modern floodplain (Connell et al., 2007). Alluvial fans also formed on the floodplain at the mouths of arroyos incised into the lower terrace, but there is little age control for these deposits.

The stratigraphic and depositional record for Chupadera Draw in the Jornada del Muerto of central New Mexico is broadly similar to

that described for the draws of the Southern High Plains (Holliday, 1997). In Chupadera Draw, thick, stratified valley fill was discovered immediately adjacent to the extensive scatter of Clovis artifacts that define the Mockingbird Gap site (Holliday et al., 2009). Between ~5–10 m below the floor of Chupadera Draw are alluvial sands and gravels that grade vertically and laterally into black muds. A significant change in depositional environment occurred ~10.5 ka¹⁴C BP (~12.5 ka cal BP), when the flow in the draw was somehow obstructed, resulting in the development of paludal conditions. These conditions would have been attractive to foragers to the area. By ~10 ka¹⁴C BP, drying of marsh muds saturated with sulfate-enriched waters resulted in the accumulation of gypsum until 9.5 ka¹⁴C BP (~11.5–10.7 ka cal BP).

West of the Rio Grande Rift, in the San Juan Basin, Smith and McFaul (1997) describe the Tohatchi I paleosol as a widespread late Pleistocene-early Holocene eolian deposit, but its age is limited to a single radiocarbon date of ~13 ka¹⁴C BP. The Tohatchi I paleosol is best interpreted as a Mollisol, indicating that elevations below 1800 m supported a moist late Pleistocene-early Holocene grassland (Smith and McFaul, 1997). East of the Rio Grande Rift, in the upper headwaters of the Cimarron River in northeastern New Mexico (2109 m), Mann and Meltzer (2007) document stream aggradation during a cooler YDC based on the presence of temperature-dependent land snails at the Folsom site (Balakrishnan et al., 2005; Meltzer, 2006). In contrast to studies in the Southwest that correlate the expansion of C₄ grasses with summer rains (Liu et al., 1996; Connin et al., 1998), Mann and Meltzer (2007) interpret greater-than-modern proportions of C₄ plants at the Folsom site to indicate a weaker summer monsoon and drier conditions during the YDC.

The complex relationship between widespread environmental changes and the cut-and-fill stratigraphy of ephemeral streams has a rich history of research in the Southern Basin and Range (Bryan, 1925; Antevs, 1952, 1962; Cooke and Reeves, 1976; Balling and Wells, 1990; Waters, 1991; Webb and Betancourt, 1992; Bull, 1997; Ely, 1997; Elliot et al., 1999; Etheredge et al., 2004; Haynes, 1968b, 2007; Karlstrom, 2005; Huckleberry and Duff, 2008). A fundamental obstacle for attributing cut-and-fill alluvial cycles to climate change is the role of intrinsic geomorphic controls and complex responses (Patton and Schumm, 1981), but the occurrence of groundwater discharge deposits is a key indicator of increased effective moisture in modern arid landscapes (Quade et al., 2007; Pigati et al., 2008).

A distinctive type of wetland deposit described in arid and semi-arid environments are “black mats.” Antevs (1959) originally distinguished the black, organic stratum covering the Lehner Clovis site as a “swamp soil,” but the term “black mat” is now conventional (Quade et al., 1998; Firestone et al., 2007; Haynes, 2008; Holliday and Meltzer, 2010). Black mats are defined here as any type of organic or inorganic accumulation associated with spring-fed ponds, marshes, and meadows (Quade et al., 1998), including carbonate precipitants and accumulations of dark organic matter (marls, black mats, *ciénegas*, Histosols). Although the chemical and physical nature of wetland deposits is controlled by groundwater composition, water temperature, permanence and mobility of surface water, vegetation, and preservation, they are generally tied to greater effective moisture, higher water tables, and spring discharge (Haynes, 2008; Pigati et al., 2008).

In southeastern Arizona, alluvial deposits in Whitewater Draw span the late Pleistocene to early Holocene. Alluviation (largely in the form of braided streams) is dated between 15 and 8 ka¹⁴C BP, but most deposits occurred between 10 and 8 ka¹⁴C BP and the YDC is poorly distinguished in the alluvial chronology (Waters, 1986, 1991). In the middle Santa Cruz River around Tucson, erosion removed most late Pleistocene and early Holocene deposits in

exposed arroyo sections (Waters and Haynes, 2001; Ballenger and Mabry, 2011). Based on more than 700 ¹⁴C dates from fluvial deposits, Harden et al. (2010) infer increased flooding in Southwest streams between ~11.3 and 10.4 ka¹⁴C BP (13.2–12.3 ka cal BP), but their stratigraphic criteria for paleoflood events in high-order streams is not precise (Ballenger and Mabry, 2011).

A key source of information for the YDC is the stratigraphy of the Upper San Pedro River Valley in southeastern Arizona (Fig. 1). The San Pedro River is a low-gradient stream originating in oak-grassland hills of northern Sonora. The river flows north approximately 190 km before reaching the Gila River, draining ~11,800 km². The badlands and arroyos of the San Pedro River Valley in southeastern Arizona expose one of the most complete late Cenozoic chronostratigraphic sequences in the United States and northern Mexico that includes the Gauss/Matuyama boundary, the Olduvai subchron, the appearance of Clovis, late Pleistocene extinctions, the Younger Dryas-age “black mat,” and a detailed record of Holocene arroyo cutting-and-filling (Bryan, 1925; Johnson et al., 1975; Lindsay et al., 1990; Waters and Haynes, 2001; Bell et al., 2004; Haynes, 2007; Pigati et al., 2008; Ballenger, 2010a).

As a discrete lithostratigraphic entity, the YDC is no better manifested in the Southwest than it is in the upper San Pedro Valley (Jull et al., 1999; Firestone et al., 2007; Haynes, 2008). Late Pleistocene deposits there are characterized by a lacustrine carbonate (the Coro marl), dated between 23 and 13.1 ka¹⁴C BP (~28–15.2 ka cal BP) at Murray Springs and between ~37–12.4 ka¹⁴C BP (~42–14 ka cal BP) near the Escapule Clovis kill site in Horsethief Draw (Pigati et al., 2008). Locally inset into the marl at these sites is a sandy and gravelly channel deposit (F₁) representing stream incision across the top of the marl. Radiocarbon dates on detrital charcoal from the F₁ alluvium at Murray Springs indicate incision before 13 ka¹⁴C BP (15 ka cal BP) (Haynes, 2007). The Coro marl also occurs upstream (south) at the Lehner site, where it was incised by a deep channel also before about 13 ka¹⁴C BP (Haynes, 2007).

Wetland formation resumed in the San Pedro Valley very soon after the appearance/disappearance of Clovis and the last appearance of several megafaunal species, depositing a widespread black mat (Stratum F₂) in spring-fed tributary streams between 10.8 and 9.8 ka¹⁴C BP (12.8 and 11.2 ka cal BP) (Haynes, 2007) (Fig. 4). Black mats occur as early as 11.8 ka¹⁴C BP (13.65 ka cal BP) in the southern Great Basin (Quade et al., 1998), but a black mat “maxima” is identified during the YDC (Haynes, 2008; Karlstrom, 2005; Quade



Fig. 4. A Late Pleistocene Coro marl scoured by alluvium underneath a thick YDC black mat in tributary stream of the upper San Pedro Valley, southeastern Arizona. Phil Pearthree pictured in photo.

et al., 1998). Pigati et al. (2008) document spring discharge between >42 and $12.4 \text{ ka}^{14}\text{C BP}$, as evidenced by deposition the Coro marl (Stratum E).

Owing to a hydrologic mechanism for black mat formation (Haynes, 2007; Haynes et al., 2010), gauging the balance between seasonal recharge and discharge is important for reconstructing paleoclimatic conditions. The widespread stratigraphic sequence of the San Pedro Valley indicates that black mats formed in response to renewed spring discharge during the YDC (Haynes, 2007), but residence times and the contribution of summer and winter rains to the local groundwater reservoirs feeding these paleosprings have not been directly studied (i.e., Baillie et al., 2007; Wahi et al., 2008). Clovis-mammoth associations also occur in the absence of Murray Springs-type (F_2) black mat deposits (Haynes and Huckell, 2007). In upper Greenbush Draw, mammoth bones rested on channel alluvium, but the bones were encased in either fine-grained alluvium (Leikem) or laminated muds (Naco). However, the primary mammoth at Leikem was buried by a locally thick layer of dark organic clays (F_2 equivalent) (Haynes, in preparation). Elsewhere in the valley, marl deposition indicating spring discharge but increasing evaporation (based on $\delta^{18}\text{O}$ values) may have persisted until approximately $9 \text{ ka}^{14}\text{C BP}$ (10 ka cal BP) at the Horsethief Draw (HD) and Seff (SF) localities (Pigati et al., 2008).

A unique look at late Pleistocene and early Holocene stratigraphy in the upper San Pedro River was provided by coring at the mouth of Palominas Arroyo, located just north of the international border. Wetland sediments were discovered 12–14 m below the top of the channel fan (Fig. 5), which is part of the modern sacaton grassland in the inner valley. These deposits include channel alluvium with peaty lamina (stratum IIa) that aggraded between 11.1 and $9.9 \text{ ka}^{14}\text{C BP}$ (12.9 – 11.3 ka cal BP) covered by black organic silty clays and peat (stratum IIb) that accumulated before approximately 8.9 until $8.5 \text{ ka}^{14}\text{C BP}$ (>9.9 – 9.5 ka cal BP). Stratum IIa is contemporaneous with deposition of the Clovis channel sands (F_1) and overlying black mat (F_2) at the Murray Springs and Lehner sites, but owing to its elevation relative to the floodplain the sequence does not include marl or black mat deposits typical of late Pleistocene springs in the basin. The lower boundary of Stratum III is marked by coarse alluvium that buried the wetland around $8.4 \text{ ka}^{14}\text{C BP}$ (9.4 ka

cal BP). Pedogenic carbonates from stratum IIa do not show a systematic shift in $\delta^{18}\text{O}$ values indicative of a dramatic change in temperature or precipitation during the YDC, but rather an isotopic composition that is similar to that of modern local groundwater, with minor exceptions (Fig. 6). $\delta^{13}\text{C}$ values of soil organic matter increase significantly (increase in C_4 grasses) beginning around $8.9 \text{ ka}^{14}\text{C BP}$ ($\sim 10 \text{ ka cal BP}$) (Ballenger, 2010a).

In the Black Mesa area of west-central Arizona, Karlstrom (2005) found a uniquely preserved sequence of late Pleistocene and early Holocene rhythmites, including “black mats,” that accumulated during the YDC between approximately 11.1 and $10 \text{ ka}^{14}\text{C BP}$, and again between 9.7 and $9 \text{ ka}^{14}\text{C BP}$. The wetland sequence was interrupted by arroyo entrenchment sometime between 11.1 and $10.5 \text{ ka}^{14}\text{C BP}$ (13 – 11.5 ka cal BP), and just prior to $9.7 \text{ ka}^{14}\text{C BP}$, but the YDC is not associated with a specific deposit in the sequence.

South of the international border in northwestern Mexico, Nordt (2003) describes the fluvial landscape evolution of the Rio Casas Grandes/Rio San Pedro system in northwest Chihuahua. The oldest late Quaternary unit is alluvium deposited by a mix-load, meandering stream prior to about $14.4 \text{ ka}^{14}\text{C BP}$. Sometime after that, the Rio San Pedro incised, formed a new floodplain meander belt channel, and then avulsed and formed wetland deposits by $9 \text{ ka}^{14}\text{C BP}$ (10.1 ka cal BP). Nordt (2003) correlates channel avulsion and wetland formation with a transition from warmer/drier to cooler/wetter conditions during the YDC based on the Lake Babí-cora stable carbon isotope record (Huang et al., 2001) and the maximum formation of black mats (Haynes, 2008; Quade et al., 1998).

A late Pleistocene paleoenvironmental record is being developed at a desiccated pond at El Fin del Mundo, a Clovis site in the low mountains of northern Sonora (Sanchez et al., 2009). This sequence seems to be a small-scale mirror of the stratigraphy at Clovis and Lubbock Lake on the High Plains. A single radiocarbon age of approximately $11.04 \pm 0.58 \text{ ka}^{14}\text{C BP}$ (~ 13.6 – 12.1 ka cal BP) indicates that a spring-fed creek was draining into the Rio Bao-coachi at that time. Stream deposits (Stratum 3) abruptly shift to diatomite pond deposits (Stratum 4) sometime after Clovis and before $9.7 \text{ ka}^{14}\text{C BP}$ (Fig. 7). How water became impounded in the former stream is not clear, but the diatomite beds grade upward

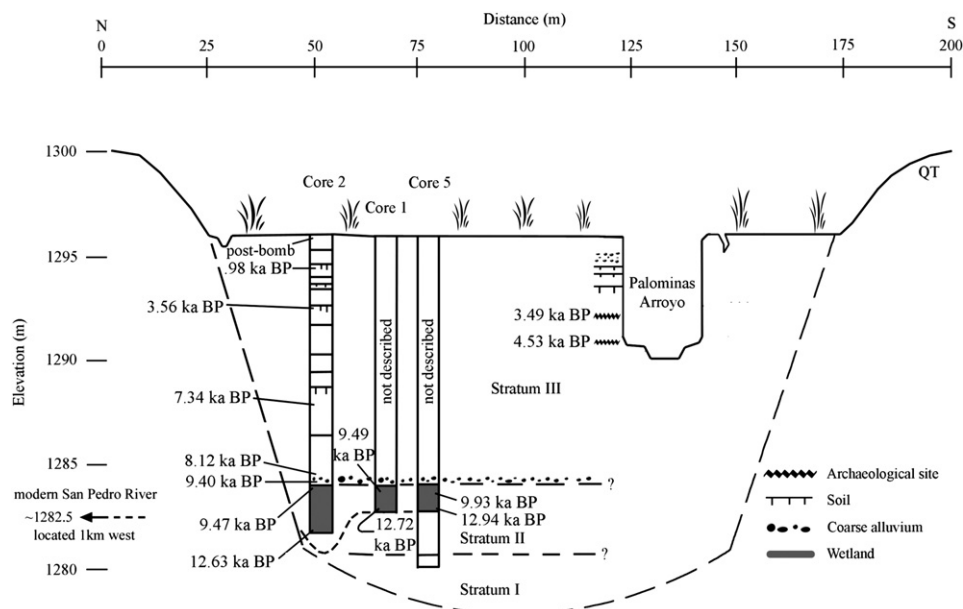


Fig. 5. Idealized cross-section of Palominas Arroyo showing core locations, select radiocarbon dates and stratigraphy in relation to modern arroyo channel.

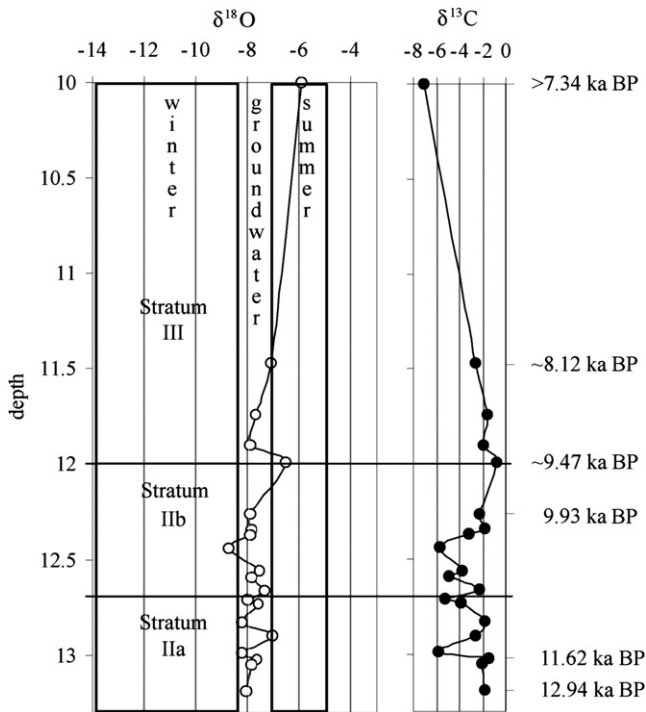


Fig. 6. Palominas Arroyo, $\delta^{18}\text{O}$ (open) and $\delta^{13}\text{C}$ (solid) values from Core 5 soil carbonates (SC) in stratigraphic sequence 10–13.3 m below surface compared to modern summer runoff ($-6.3 \pm 1.0\text{‰}$), groundwater (-7.7‰ – 7.8‰), and winter runoff ($-11.2 \pm 2.8\text{‰}$) from Baillie et al. (2007) at 15 °C. $\delta^{13}\text{C}$ value at 10 mbs is anomalously negative and is not supported by the $\delta^{13}\text{C}$ value of paired soil organic matter (Ballenger, 2010a).

into early Holocene diatomaceous earth. The YDC pond likely developed into a marsh as the small basin filled.

2.7. Dune stratigraphy

Eolian sand sheets do not preserve well-resolved paleoenvironmental records, but they do expose large numbers of Paleoindean artifacts. Eolian deposits are common east of the Pecos River valley, occurring as massive dune fields or sand sheets or as lunettes associated with small playa basins (Holliday, 1997, 2001). The dune chronology of the Southwest has only recently benefited from systematic geochronological research. In the Albuquerque Basin, not far from the Boca Negra Wash Folsom site, 12 OSL ages indicate that

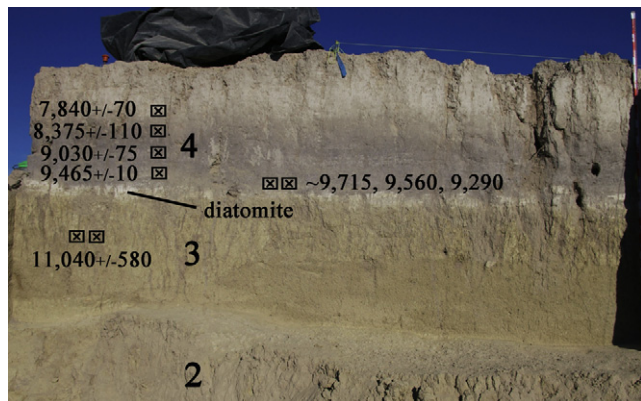


Fig. 7. Stratigraphic section showing location of radiocarbon samples, El Fin del Mundo, Sonora.

at least 1 m of eolian sands mantled a possible mid-Pleistocene sand sheet without interruption between 16 and 10 ka BP, forming a red calcic paleosol before experiencing widespread erosion throughout the Holocene (Hall et al., 2008).

A larger sample of dates is available from the Bolson sand sheet, a nearly 4500 km² eolian deposit in the Tularosa Basin of New Mexico and West Texas. A series of 26 OSL dates and stratigraphic discontinuity indicate that the Bolson sand sheet is composed of two eolian units: a Pleistocene-age (~45 ka BP) massive sand (Q₂) with a red Bt upper boundary that is commonly truncated, and a nearly equally thick but more widespread massive sand unit (Q₃) with a weak Bk and sometimes Bw soil and stage I carbonate morphology that accumulated between ~24–5.2 ka BP (Hall et al., 2011). The Bolson chronology is replicated at the Mescalero sand sheet, where 17 OSL dates and stratigraphy distinguish lower (90–50 ka BP) and upper (18–5 ka BP) sand units, the latter forming a Bw soil during the late Holocene (Hall and Goble, 2008). However, OSL ages on eolian gypsum sands at White Sands, New Mexico, indicate that local dune formation commenced ~7 ka BP (Kocurek et al., 2007).

2.8. Summary of YDC paleoenvironmental records

High-resolution records of the BA/YDC/EH transition are rare, and consensus does not exist regarding the magnitude or direction of YD climate change in the Southwest or elsewhere in the mid-continent (Holliday, 2000; Dorale et al., 2010; Haynes, 2008; Meltzer and Holliday, 2010). Well-dated paleoenvironmental records that are coincident with the YDC indicate abrupt but inconsistent changes. However, both COB-01-02 and FS-2, the only records that allow direct comparison to the Greenland ice cores and that cross the BA/YDC transition, are interpreted to show the onset of increased winter precipitation and/or cooler temperatures around 12.9 ka BP (Fig. 8). YDC $\delta^{18}\text{O}$ values from these records do not approach their LGM minimums.

Atmospheric circulation models indicate that the westerly jet stream and associated low pressure cells migrated south of 30°N during the late Pleistocene, causing an increase in winter

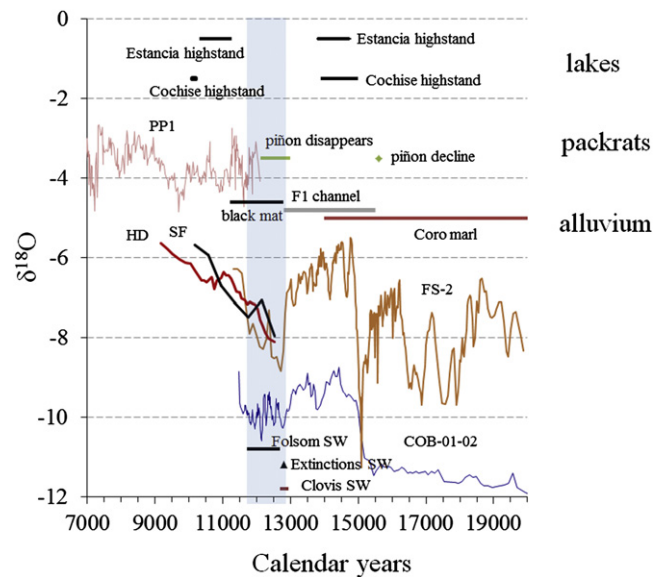


Fig. 8. Speleothem $\delta^{18}\text{O}$ values from Pink Panther (PP), Cave-of-the-Bells (COB-01-02), and Fort Stanton Cave (FS-2), and Earp marl $\delta^{18}\text{O}$ values from Horsethief Draw (HD), and Seff Locality (SF) (Pigati et al., 2008) in relation to major changes detected in Paleolake, Packrat midden, alluvial, paleontological, and archaeological records in the Southwest.

precipitation and the virtual elimination of summer monsoons (Kutzbach, 1987; COHMAP; Members, 1988; Spaulding, 1991). However, cooler/wetter YDC conditions are not supported by packrat records that indicate increasingly xeric conditions beginning around 13 ka cal BP (Van Devender, 1990a, 1990b). Van Devender (1980) describes a Clovis landscape in the vicinity of the Guadalupe Mountains dominated by piñon-juniper woodlands and an understory of C₄ grasses, followed by more xeric Folsom-age environments characterized by juniper–oak associations and grasses at mid-elevations.

The onset of Holocene warming as measured by the loss of piñon is correlated by Van Devender and Spaulding (1979) with pollen evidence for extensive grasslands at the Lehner site. However, more recent pollen studies in the San Pedro Valley actually indicate both desert grasslands (Mehring and Haynes, 1965) and piñon-juniper-oak woodlands (Ballenger, 2010a) at mid-elevations between 12.9 and 12.6 ka cal BP. Stable carbon isotope values from bison tooth enamel indicate reliable summer rains and C₄ grasslands through the BA/YDC transition, and summer annuals and grasses are documented in YD-age packrat middens (Holmgren et al., 2003), but Folsom-age bison are not reported from Arizona. Millennial-scale lake level chronologies from the Southwest collectively indicate that lake stands during the YDC were lower than (1) during the LGM, (2) the period of deglaciation just before the BA interstadial, and (3) the EH. The stratigraphic records of paleosprings in the Southwest indicate a switch to groundwater discharge and the formation of wetlands in former channels at the onset of the YDC (Haynes, 2007, 2008; Holliday et al., 2009), but the paleoclimatic interpretation of the transition is debated (Haynes, 1991; Holliday, 2000).

Collectively, these records have not been articulated to form a coherent model of YDC temperature, precipitation, floral, and geomorphological responses in the Southwest, and they highlight the importance of distinguishing the effects of global climate change in different systems and at local scales. Such records are critical for limiting paleoenvironmental conditions in ways that are relevant to human adaptations, but an important proxy is overlooked by studies that exclude people from the paleoenvironmental record. In the proper context, using the archaeological record to provide information about paleoenvironments is just as valid as using the paleoenvironmental record to provide information about artifacts. The following section provides a brief overview of the BA/YDC archaeological record, including opposing paleoenvironmental reconstructions that were tested using the distribution of time-sensitive Paleoindian projectile points.

3. Archeological record of the YDC in the Southwest

The earliest widely accepted evidence for human occupation in the Southwest is the Lehner Clovis site, radiocarbon dated 10.95 ± 0.04 ka¹⁴C BP (~12.8 ka cal BP). The Murray Springs site is dated only slightly later at 10.89 ± 0.05 ka¹⁴C BP, but Clovis artifacts there also occur directly beneath a widespread black mat indicating close contemporaneity (Haynes, 2007). The Clovis occupation at Blackwater Draw is dated to 10.91 ± 0.07 ka¹⁴C BP, overlapping with the age range of Lehner and Murray Springs. Folsom sites in the Southwest are also poorly dated, but in eastern New Mexico bison kills occurred around 10.5 ± 0.02 ka¹⁴C BP (12.5 ka cal BP) at the Folsom site (Meltzer, 2006), and at 10.3 ± 0.09 ka¹⁴C BP (12.2 ka cal BP) at Blackwater Draw (Taylor et al., 1996). More widely, Clovis kill sites are usually dated between 11.05 and 10.8 ka¹⁴C BP (~12.95–12.7 ka cal BP) (Waters and Stafford, 2007), whereas Folsom sites range between 10.8 and 10.1 ka¹⁴C BP (~12.7–11.7 ka cal BP) (Collard et al., 2010).

Other lanceolate and constricted stem projectile point types documented from the Southwest and radiocarbon dated earlier than 11.6 ka cal BP include Plainview-Goshen and Agate Basin (Faught and Freeman, 1998; Huckell and Judge, 2006), but considerable uncertainty exists in the age range of these broadly defined types, and they have not been found in an YDC-dated context in the Southwest. Likewise, Western Stemmed (Lake Mohave, Silver Lake) type points span the YDC in the Intermontane West (Beck and Jones, 2010), but overlap morphologically with early-middle Holocene types in the Southwest (Irwin-Williams, 1973). Bodily (2009:36) recovered several post-YDC stemmed points bracketed between ~10–9.5 ka¹⁴C BP at North Creek Shelter on the Southern Colorado Plateau of south-central Utah. At Ventana Cave in western Arizona, a Plainview-like point was recovered from deposits radiocarbon dated between 10.5 and 8.8 ka¹⁴C BP, but the occupation is suspected to post-date Folsom (Huckell and Haynes, 2003). The earliest post-Clovis sites in southeastern Arizona occur in Whitewater Draw, where Waters (1986) uncovered artifacts but no diagnostic tools probably deposited between 10 and 8 ka¹⁴C BP (Waters, 2000). At the Lehner site, a hearth situated near the upper boundary of the black mat was radiocarbon dated around 9.9 ka¹⁴C BP (Haynes, 1982, 2007). No projectile point type is described for the YDC in the southern Southwest.

The prevailing evidence of YDC adaptations in the Southwest points to large grazer-oriented hunting societies focused on water and high-quality lithic sources for their primary game and technologies (Amick, 1996; Hamilton, 2008; LeTourneau, 2000; Huckell, 2004; Huckell and Judge, 2006; Haynes and Huckell, 2007; Irwin-Williams and Haynes, 1970). This evidence is systematically biased in favor of hunting-related sites. Uncontested Clovis-mega faunal associations include mammoth, gomphothere, and bison (Haury et al., 1953, 1959; Haynes and Huckell, 2007; Sanchez de Carpenter, 2009; Saunders, 1992), supplemented to an unknown extent by common smaller animals such as rabbits and tortoises (Haynes, 1982).

Curiously, Clovis raw material use in the Southwest does not show the dominant use of extra-local lithic sources and the strategic emplacement of toolstone caches as seen in the mid-continent (Kilby, 2008), but redundant use of the best-quality local materials (Huckell, 2007), including obsidian marekanites (Shackley, 2007; Hamilton et al., 2009). A robust Clovis hunting tool kit at Murray Springs included bifaces, projectile points, retouched unifaces, end and side scrapers, graters, blades, and expedient tools (Huckell, 2007). Clovis points from the Southwest are generally smaller than those found on the Northern Plains. Hamilton and Buchanan (2007) attribute this to multigenerational copying errors, but an alternative explanation is that various raw material packages in the Southwest are relatively small (Shackley, 2005). Haynes (1991) proposed that Clovis times were droughty, as indicated by lower water tables and the aboriginal excavation of a well at the Clovis site (Haynes et al., 1999), possibly marking the height of BA environmental deterioration that prefaced sudden late Pleistocene extinctions at the beginning of a cooler/wetter YDC (Haynes, 2008).

Folsom groups are generally regarded as specialized bison hunters (Kelly and Todd, 1988; Hofman and Todd, 2001) whose technology is most readily distinguished from Clovis on the basis of projectile point manufacture and design and the loss of a robust blade technology (Boldurian and Cotter, 1999; Collins, 1999). Amick (1996) relies on raw material use and modern bison feeding behavior to reconstruct a Folsom settlement and subsistence strategy focused on the seasonal movement of bison between protected foothills and basins in the Rocky Mountains during winter and the open Plains during the growing season, but fall kills occurred in the foothills (Meltzer, 2006). The use of regional lithic

raw materials in the Rio Grande Valley indicates movements in residential territories covering 16,000 km² (Amick, 1996), whereas raw material use and discard on the Southern High Plains indicates a strong emphasis on the Edwards Plateau and vast hunting territories (Amick, 1996; Bement, 1999, but see Bamforth, 2011). The lack of large Folsom camps or kills in the Basin and Range is interpreted to imply small foraging groups in pursuit of small herds or individual bison, but preservation is poor at known sites (Amick, 1996; Holliday, 2005).

Holliday (2000) provides evidence from the adjacent Southern High Plains indicating that Folsom times were increasingly warm and arid compared to Clovis. The Paleoindian occupation of the Rio Grande Rift and the adjacent Sacramento Section, including the Tularosa Basin, shows that Folsom sites occur in closer proximity to locally wet playas and paleolake basins than do earlier and later Paleoindian sites (Holliday, 2005; Judge, 1973; Wessel et al., 1997). Both Clovis and Folsom-age wetlands are inferred in the C-N Basin on the Plains of San Agustin (Holliday et al., 2006b), but a recent study by Hill and Holliday (2011) indicates that Folsom occurrences there are more strongly correlated with paleolakes than both Clovis and post-Folsom finds. Using a cooler/wetter model of the YD, others point to increased grass production and competitive release to explain prosperous bison herds and hunters during Folsom times (Huckell and Judge, 2006; Mullen, 2008).

4. Expectations and analysis

The opposing paleoenvironmental reconstructions outlined by Haynes (1991) for the North American continent, and Holliday (2000) for the Southern High Plains, carry implications that can be tested using the distribution of time-sensitive projectile point types. The matter is simplified by assuming that early Paleoindian projectile points were hunting technologies employed in the pursuit of large grazers (mammoth and bison by Clovis, and bison by Folsom), as the regional archaeological record indicates. Specifically, the focus is on the distribution of Clovis and Folsom/Midland projectile points to distinguish the BA/YDC transition in terms of a common unit of hunting technology. If Clovis-age grazers were faced with a landscape characterized by falling water tables, possibly leading to circumscribed refugia before late Pleistocene extinctions (Ballenger, 2010b; Haynes, 2002), then it is reasonable to expect Clovis hunting tools to be clustered in their distribution, possibly around reservoirs and other dwindling water holes (Haynes, 2002; Haynes, 1991; Haynes and Huckell, 2007). An argument could be made that forager populations become dispersed rather than concentrated if resources are scarce, an example of drought evasion rather than drought escape (Gould, 1991), but this model assumes that Clovis points reflect hunting and that late Pleistocene extinctions played out at the scale of regional refugia. Cooler/wetter conditions and rebounding water tables (spring discharge) during the YDC is expected to have allowed grazers and hunters to forage more widely. In contrast, if water tables fell during warmer/drier Folsom times (Holliday, 2000), then the opposite is expected in this model. However, it is important to point out that standing water and lacustrine wetlands (suitable water holes for animals) are interpreted to indicate warmer/drier conditions and less spring discharge in the Folsom-dry model (Holliday, 2000), and this caveat probably falsifies these expectations to an unknown extent.

To test these expectations, PIDBA was consulted to map the distribution of Clovis and Folsom/Midland projectile points in Arizona and New Mexico. Fig. 9 shows the number of Clovis and Folsom/Midland projectile points per 1000 km² at the county-level. Clovis projectile points appear to be widely distributed at this scale

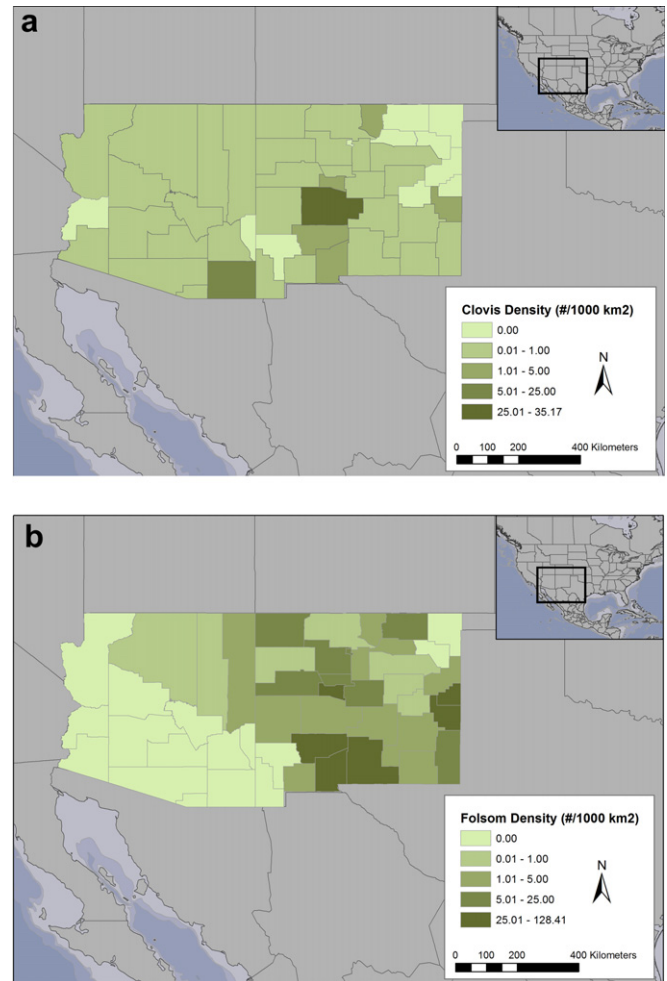


Fig. 9. Distribution of Clovis (a) and Folsom/Midland (b) projectile points in Arizona and New Mexico based on PIDBA dataset and supplemental data (see Methods).

of analysis (Fig. 9a), whereas Folsom points are documented throughout most of New Mexico and northern Arizona but are excluded from the northwestern Chihuahuan Desert and the Sonoran Desert (Fig. 9b). Quantification of the degree of autocorrelation for the Folsom and Clovis distributions utilized Moran's *I* (Moran, 1950), which provides a value between -1 (perfectly dispersed) and 1 (perfectly autocorrelation) with zero indicating a completely random distribution. ESRI ArcGIS 10 was used to calculate this value and the distribution of Clovis projectile points is indistinguishable from a random distribution (Moran's $I = -0.01$; $p = 0.77$), whereas the distribution of Folsom projectile points is significantly clustered (Moran's $I = 0.14$; $p = 0.038$).

The densest concentrations of Clovis projectile points are associated with large, rare sites or clusters such as Mockingbird Gap, Blackwater Draw, and the San Pedro Valley, or extensively surveyed lands such as the southern Tularosa Basin (Fig. 9a), indicating either a strong research bias in the data (Huckell, 2004), the actual structure of the archaeological record (Andrews et al., 2008), or something in-between. Clovis projectile points are rarely reported from the Sonoran Desert west of Phoenix, and are lacking from the Southern High Plains of northeastern New Mexico. Information from these regions is lacking. Regardless, Clovis points occur in a variety of biotic settings, including rare sites in the western Sonoran Desert (Haynes, 2011; Huckell, 2004). Fig. 9b is the same analysis using the data available for Folsom/

Midland points. Folsom points are not widely scattered but rather constricted in their distribution, with rare occurrences in the northeastern Chihuahuan Desert and none yet documented from the Sonoran Desert (Huckell, 1982; Gaines et al., 2009; Sanchez de Carpenter, 2010), where Paleoindian research has been intensive.

5. Discussion

The question of whether the Clovis-to-Folsom transition was characterized by a switch from warmer/drier to cooler/wetter conditions, or vice-versa, is a key issue in Paleoindian studies focused on the Southwest and Southern Plains (Haynes, 2008; Holliday and Meltzer, 2010). This review of multiple paleoenvironmental records in the Southwest indicates that those records with the best resolution (speleothems) do record isotopic oscillations coincident with the BA/YDC transition; these changes are interpreted to reflect a shift from warmer/drier to cooler/wetter conditions and/or increased winter precipitation (Asmerom et al., 2010; Wagner et al., 2010). The speleothem record is supported by rare YDC pollen sequences that may indicate the transition from warmer/drier desert grasslands to cooler/wetter piñon-juniper-oak woodlands during the early YDC (Mehring and Haynes, 1965; Ballenger, 2010a), as well as alluvial chronologies that indicate the activation of springs and black mat formation centered around the YDC (Haynes, 2008; Quade et al., 1998). Other important paleoenvironmental and paleohydrological records do not support cooler/wetter YDC conditions. Numerous packrat middens in the Southwest show the loss of mesic piñon-juniper woodlands through the YDC; none of the lake records in New Mexico or Arizona show YDC highstands.

Highly mobile hunter/gatherers are a key indicator for resource availability (Winterhalder, 2001). However, the use of projectile point distributions to isolate food availability rather than favored raw material sources relies in part on the general patchiness of the arid Southwest, which limits the chances for food to co-occur with high-quality lithic sources and other permanent features (e.g., Anderson and Gillam, 2000). The anthropogenic proxy for paleoenvironments lacks details, but it provides an opportunity to predict and evaluate broad changes in human hunting across the BA/YDC landscape. The scale of this perspective is useful because it departs from regional reconstructions based on rare sites that receive careful study (Meltzer and Holliday, 2010).

The distribution of Clovis and Folsom/Midland projectile points does not live up to expectations for a “Clovis drought” followed by a cooler/wetter YD (Fig. 9). Clovis hunters pursued game far and wide relative to Folsom hunters, and their tools do not cluster around reservoir bottoms, but mid-elevation springs (Haynes, 2008; Huckell, 2004; Wessel et al., 1997). This pattern could record the different land use strategies between Clovis explorers and Folsom residents, but there is reason to believe that Clovis explorations were possible because of widespread large game during the late Pleistocene (Kelly and Todd, 1988; Waguespack and Surovell, 2003; Agenbroad, 2005; Haynes, 1966). The distribution of Clovis points also does not lend support to the hypothesis that megamammal hunting occurred in circumscribed refugia (Ballenger, 2010b; Haynes, 2002) because the resolution of this analysis is coarse, and it is difficult to tease apart research bias and the underlying structure of the archaeological record.

The distribution of Folsom points is constricted relative to Clovis (Fig. 9). Folsom points are not included in the data from several contiguous counties in the northwestern Chihuahuan Desert and the Sonoran Desert, and they appear to cluster around reservoirs (Judge, 1973; Hill and Holliday, 2011). However, neither does the archaeological record provide support for a “Folsom drought,” if YDC climates diminished local grasslands and other seasonal

resources. In fact, Folsom populations in the Southwest and Plains appear to have increased (Meltzer and Holliday, 2010). Over 60 Paleoindian sites have been recorded in proximity to playa or small lake basins in the central Rio Grande Valley, and although Folsom technologies endured nearly twice as long as Clovis, the number of Folsom sites is significantly more than twice as many as the number of Clovis sites (Holliday, 2005). Of the 197 Younger Dryas age sites documented in New Mexico by Mullen (2008:28), 172 are Folsom/Midland, eight are Agate Basin, two are Hell Gap, 10 are Plainview, two are Milnesand, and three are Meserve. Amick and Lukowski (2006) report only three Clovis points collected from the southern Tularosa Basin, but 227 Folsom/Midland projectile points that indicate abundant bison and intensive hunting.

Collard et al. (2010) characterize Folsom hunters in the Southwest as a population expansion from the Southern Plains. However, this is based on the temporal frequency distribution of only a few radiocarbon dates. The range of late Pleistocene hunters clearly contracted during Folsom times in relation to Clovis. Their model also implies a time gap between late Clovis groups who hunted in the southern Southwest and subsequent Folsom groups who did not. However, strong cultural continuity is seen where the Clovis-to-Folsom transition is well-dated and virtually complete, namely, the Jake Bluff and Cooper kill sites where the ^{14}C ages of the Clovis and latest Folsom bonebeds nearly overlap around 12.7 ka cal BP (Bement and Carter, 2010).

The absence of Folsom points in the Sonoran Desert is a long-standing problem, and the lack of a clearly defined Paleoindian presence after Clovis is widely acknowledged (Firestone et al., 2007; Holliday, 2005; Mabry and Faught, 1998; Sanchez de Carpenter, 2010). The apparent absence of Folsom sites in the southern Southwest is probably not a result of research bias in light of the number of fluted point surveys conducted there (Agenbroad, 1967; Huckell, 1982), and the preservation of YDC sediments in some local streams (Waters and Haynes, 2001). Rather, Huckell (2004) correlates Clovis projectile point distributions with late Pleistocene grasslands, and Irwin-Williams and Haynes (1970) attribute subsequent Paleoindian range “shrinkage” to the geographical limits of bison during the YDC. In northwestern Mexico, Sanchez de Carpenter (2010:78) identifies the Sierra Madre Occidental as a western barrier for Folsom projectile points, whereas Folsom points are rarely reported from the states of Chihuahua and Nuevo Leon.

Climatic reconstructions that explain paleohydrological and biogeographical differences between the Southwest and the Southern High Plains exist but lack systematic testing. For example, Higgins and MacFadden (2004) explain variation in the $\delta^{18}\text{O}$ of late Pleistocene bison and horse as a result of a moisture gradient between strong summer rains in the Chihuahuan Desert and dampened summer rains in the Sonoran Desert, but their samples predate the YDC. Connin et al. (1998) likewise hypothesized a steep gradient from summer- to winter-dominated rainfall between the Southwest and the western Southern High Plains to explain the strongly C_4 diets of Clovis-age bison at Blackwater Draw in contrast to C_3 -dominated early Holocene bison diets at Lubbock Lake (Stafford et al., 1994). More recently, Holmgren et al., (2007) have argued that summer rains supporting C_4 grasses persisted in portions of the Southwest throughout the last glacial–interglacial cycle (i.e., Betancourt et al., 2001; Liu et al., 1996), with a late Pleistocene grassland stretching from central Texas to southern Arizona south of 35° N. This reconstruction indicates that the Chihuahuan Desert may have been the northernmost grassland to “green-up” during late Pleistocene summers (Holmgren et al., 2007), possibly explaining the concentration of mammoth kill sites in the San Pedro Valley (Holmgren et al., 2006). On the other hand, Wilson et al. (2007) infer a winter bison kill at the Murray

Springs site, when C_4 grasses are at a minimum today (Van Devender, 1995).

Packrat middens provided early evidence that summer rains and grasses were limited in the Southwest during the late Pleistocene. Desert grasslands are thought to have experienced their greatest expanse during the early to mid-Holocene as the Bermuda high and associated storm tracks resumed an interglacial pattern (COHMAP Members, 1988; Van Devender and Spaulding, 1979), with the loss of juniper and increased grasses at mid elevations between 8.9 and 8.5ka ^{14}C BP (after 10.1 ka cal BP) in the Sonoran and Chihuahuan deserts (Van Devender, 1990a; 1995). By 8.85 ka ^{14}C BP (10 ka cal BP) C_4 grasses began to make large contributions to soil biomass and the pollen assemblage in Palominas Arroyo (Ballenger, 2010a).

Late Pleistocene bison of Clovis and pre-Clovis-age occur at Lehner, Murray Springs, El Fin del Mundo, and a number of paleontological localities (Haury et al., 1959; Haynes and Huckell, 2007; Morgan and Lucas, 2005; Sanchez de Carpenter, personal communication 2011, Saunders, 1970). The persistence of reliable summer rains in the Southwest during the YDC is equivocal based on the $\delta^{13}C$ values of bison bone and tooth enamel. At the Folsom site, situated on the western edge of the Southern Plains, Meltzer (2006) describes bone collagen values that indicate greater-than-modern proportions of C_4 grasses. However, Meltzer (2006) cites probable extra-local foraging by bison herds, as well as a documented YDC decrease in atmospheric pCO_2 levels that may have favored the growth of C_4 plants under decreased summer precipitation and/or cooler temperatures (Cerling et al., 1997). The consumption of greater-than-modern amounts of summer grasses is likewise recorded at the Boca Negra Wash Folsom site in the middle Rio Grande Valley (Mullen, 2008). However, YDC bison are not reported from the Sonoran Desert.

The post-Clovis archaeological record of the Southwest has been explained by the early development of a Desert Archaic adaptation (Cordell, 1997; Holliday, 2005; Irwin-Williams and Haynes, 1970; Sanchez de Carpenter, 2010), but those sites have not been identified and dated. The most significant post-YDC change in material culture is the appearance of groundstone and Pinto-type and notched projectile points that occur between 9 and 8 ka ^{14}C BP on the southern Colorado Plateau (Geib, 1996; Bodily, 2009). In southeastern Arizona, a noticeable 1400-year hiatus occurs in the archaeological record following Clovis and Pleistocene extinctions (Ballenger and Mabry, 2011). Western Stemmed projectile points are also rare in southern Arizona. Mabry and Faught (1998) argue that Great Basin populations did not flourish there due to highly variable moisture relative to the Intermontane West. In any case, there is presently no compelling evidence for hunters or gatherers in the northwestern Chihuahuan Desert or the Sonoran Desert interior during the YDC.

6. Conclusions

This review of multi-proxy paleoenvironmental records clearly indicates significant and widespread paleoenvironmental changes coincident with the YDC, but the amplitude and even the direction of those changes are inconsistent across and sometimes within proxies. As to why the various paleoenvironmental records reviewed are not in agreement, that problem deserves more discussion than can be provided here. As expressed in Haynes et al. (2010), an understanding of what happened at approximately 12.8 ka cal BP requires further research. However, the best-resolved records indicate that YDC conditions in the northern Chihuahuan and Sonoran deserts were characterized by cooler/wetter climates, and/or increased winter precipitation. On the other hand, YDC $\delta^{18}O$ values in those records do not show a return to LGM moisture,

temperature, or vegetation, consistent with global trends (Shakun and Carlson, 2010).

To isolate a human “response” to YDC climate changes requires controls that are not always possible using the types of evidence at hand, but our analysis does demonstrate unambiguous and large-scale changes in human land use. In conclusion, attention is drawn to chronological correlations that exist between the YDC as measured in the Greenland ice cores, regional paleoenvironmental changes, and major human adaptive changes during the BA/YD/EH transition, including:

- 1) the appearance/disappearance of Clovis technologies at approximately 12.8 ka cal BP;
- 2) a switch in large animal prey from mammoth and bison to primarily bison between approximately 12.8 and 12.7 ka cal BP;
- 3) Folsom regionalization and population growth between approximately 12.7 and 11.7 ka cal BP;
- 4) the apparent abandonment of the northwestern Chihuahuan and Sonoran deserts by hunter-gatherer populations after 12.8 ka cal BP; and
- 5) the reappearance of people in the stratigraphic record of the northwestern Chihuahuan and Sonoran deserts around 10 ka ^{14}C BP (~11.5ka cal BP).

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