



A stable isotope record from paleosols and groundwater carbonate of the Plio-Pleistocene Camp Rice Formation, Hatch-Rincon Basin, southern New Mexico

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A STABLE ISOTOPE RECORD FROM PALEOSOLS AND GROUNDWATER CARBONATE OF THE PLIO-PLEISTOCENE CAMP RICE FORMATION, HATCH-RINCON BASIN, SOUTHERN NEW MEXICO

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ABSTRACT—Stable oxygen and carbon isotope data from paleosols and shallow groundwater carbonates of the Plio-Pleistocene Camp Rice Formation in the western Hatch-Rincon Basin supplement an existing dataset from Neogene basin-fill in southern New Mexico and southeastern Arizona. In addition to their utility as proxies for paleoclimate and paleoenvironment, these data highlight local controls on the isotope chemistry of authigenic carbonate, such as depositional setting and hydrology. Oxygen isotope values for carbonates from Camp Rice piedmont deposits are higher on average (-6.8‰) than those from axial-fluvial parent material (-7.4‰), but mean carbon isotope values are identical (-4.3‰). The high carbon isotope values of soil carbonates formed in the ancestral Rio Grande floodplain differ from landscape mosaicism of C₃ versus C₄ plants observed in the Mangas Basin of southwestern New Mexico. This could be due to the influence of a shallow and perhaps saline water table. Mean δ¹⁸O values increase from -7.6‰ for >3.1 Ma to -6.7‰ for <3.1 Ma samples and mean δ¹³C values increase from -4.9‰ to -3.8‰. Our data generally support a latest Pliocene-early Pleistocene transition to a warmer, drier climate with increased summer precipitation. This interpretation is consistent with stable isotope records from correlative deposits in the neighboring Palomas and eastern Hatch-Rincon Basins as well as southeastern Arizona.

INTRODUCTION

Stable carbon and oxygen isotopes of pedogenic carbonate are well-established proxies for interpreting paleoclimate and paleoenvironment (e.g., Cerling, 1984; Cerling and Quade, 1993; Koch, 1998). Stable isotope data have been previously acquired for calcic paleosols and shallow groundwater carbonate deposits in Neogene basin-fill of the southern Rio Grande rift and Basin and Range of southeastern Arizona and southwestern New Mexico (Wang et al., 1993; Mack et al., 1994a; Mack et al., 2000; Mack and Tabor, 2008). These datasets demonstrate significant climatic transitions during the latest Pliocene and early Pleistocene that can be tied to regional proxy records as well as local controls on the isotopic composition of soils and groundwater.

The Plio-Pleistocene Camp Rice Formation in the Hatch-Rincon Basin of southern New Mexico includes deposits of the ancestral Rio Grande, surrounding floodplains and alluvial flats, and piedmonts extending from nearby uplifts (Mack et al., 1994b; Seager, 1995; Jochems, 2017). A stable isotope record spanning these depositional environments in space and time could indicate changing vegetation communities and perhaps variability in temperature and water composition.

The purpose of this study is to assess new stable isotope data from calcic paleosols and non-soil authigenic carbonate of the Camp Rice Formation to better understand local paleoenvironment of the western Hatch-Rincon Basin. We place our results in a stratigraphic context with existing magnetostratigraphic and radiometric age constraints as well as biostratigraphy. We then make inferences regarding local paleoclimate and paleoenvironment and offer comparisons to other locales in the southwestern U.S. (Wang et al., 1993; Mack et al., 1994a; Mack et al., 2000;

Mack and Tabor, 2008). This study adds to an ample set of stable isotope data for the southern Rio Grande rift and includes data from both axial-fluvial and piedmont depositional settings.

Geologic setting

The Hatch-Rincon Basin (Fig. 1) is located in the southern Rio Grande rift, a series of en echelon basins stretching from Colorado to northern Mexico that filled with sediment and lesser volcanic deposits; these comprise the Santa Fe Group (Chapin and Cather, 1994). The basin is a symmetric to asymmetric (half) graben superimposed on older basins formed by episodic subsidence beginning in the late Oligocene or early Miocene (Mack et al., 1994c). In the study area southwest of the Rio Grande, the oldest exposed sediment consists of mudstones and rare fine sandstones of the mid- to upper Miocene Rincon Valley Formation.

The Hatch-Rincon Basin narrowed following deposition of the Rincon Valley Formation and began receiving externally sourced fluvial sediment deposited by the ancestral Rio Grande (Mack et al., 1994b; Seager and Mack, 2003). This sediment belongs to the Plio-Pleistocene Camp Rice Formation and also includes subordinate piedmont gravel, sand, and silt deposited in tributary alluvial fan systems. The Camp Rice is correlative with the Palomas Formation in the Palomas Basin to the north (Lozinsky and Hawley, 1986). It overlies the Rincon Valley Formation with slight to moderate angular unconformity. Magnetostratigraphic and radiometric data constrain the base of the Camp Rice Formation to <3.6 Ma in the Hatch-Rincon Basin; it is capped in many places by La Mesa geomorphic surface, which slightly predates the 0.78–0 Ma Brunhes chron (Mack et al., 1993, 1998).

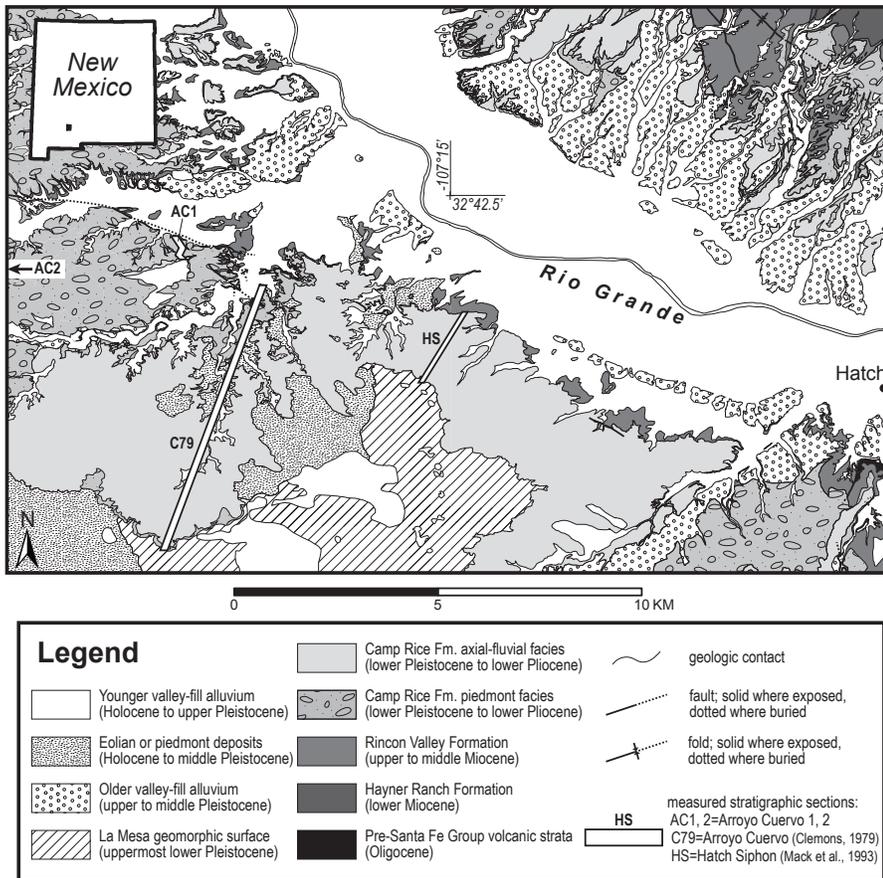


FIGURE 1. Simplified geologic map of study area showing locations of stratigraphic sections discussed in text. Section AC2 is ~2 km west of map boundary. Locality shown on inset map of New Mexico. Geology modified from Seager (1995), NM Bureau of Geology & Mineral Resources (2003), and Jochems (2017).

In the study area, the Camp Rice includes sand and pebble gravel/conglomerate deposited in channels as well as silt and mud deposited in floodplains, alluvial flats, and distal alluvial fan settings (Figs. 1, 2; Seager, 1995; Jochems, 2017). Quartzite and possible Pedernal chert clasts in axial-fluvial sediment indicate an extra-basinal source area as far away as northern New Mexico and southern Colorado. This sediment is interpreted as having been deposited at the head of a fluvio-deltaic system extending southward into west Texas and northern Chihuahua, Mexico (Mack et al., 2006).

Stable isotope samples were collected from three measured stratigraphic sections (Figs. 1, 2): section C79 (Arroyo Cuervo) of Clemons (1979) and sections AC1 and AC2 of Jochems (2017). These sections can be correlated to the nearby Hatch Siphon (HS) magnetostratigraphic section measured by Mack et al. (1993). Sections HS and C79 begin at the Camp Rice-Rincon Valley contact and end at carbonate caprock underlying La Mesa; the lower part of section C79 crosses a southwest-down fault with <1 m of displacement (Fig. 1). The sections are dominated by axial-fluvial channel sands and floodplain muds/shales with common paleosols containing stage II-IV carbonate accumulation (Gile et al., 1966), typically overlain by cambic or argillic soil horizons. At 101 m thick, section C79 exceeds local topographic relief, so we assume that Clemons (1979) measured this section using up to 1° of dip toward the south-southwest. Com-

bined with a gentle southwest-to-northeast gradient underlying La Mesa, this accounts for the thickness discrepancy between section C79 and the ~70-m section HS of Mack et al. (1993).

Sections AC1 and AC2 mostly contain piedmont alluvial-fan sediment of the Camp Rice Formation (Jochems, 2017); a tongue of pebbly axial-fluvial sand is found at the top of section AC1 (Fig. 2). Paleosols in these sections are similar to those in C79 but with weaker carbonate accumulation (stage II). Non-pedogenic calcite (micrite) in section AC1 has sharp contacts, horizontal bedding without soil horizons, and is interpreted as having been precipitated from shallow groundwater, perhaps in *ciénegas* (springs and marshes) at the distal toes of alluvial fans (Mack et al., 2000). Sections AC1 and C79 are correlated based on elevation, although tracing of individual beds is uncertain given the lenticular nature of sandy tongues within the Camp Rice axial-fluvial facies. Sections AC1 and AC2 are correlated based on physical mapping of geologic contacts between member-rank units of the Camp Rice Formation (Jochems, 2017).

Section HS includes a pumice conglomerate dated to 3.1 Ma (Mack et al., 1996). Although this bed was not found

in any of the other three sections, frothy pumice clasts with sanidine phenocrysts were recovered from float at ~45 m in section C79 (Fig. 2). Above the location of these clasts is <5 m of Holocene windblown sand draping a low-lying ridge. Figure 2 shows a 10-m interval from which the clasts could have been derived. Importantly, this interval spans the elevation at which the pumice conglomerate in section HS is found.

Late Pliocene vertebrate fossils from the Arroyo Cuervo and Hatch local faunas (LF), referred to the Blancan North American land mammal age, were collected from axial-fluvial facies of the Camp Rice Formation (Morgan and Lucas, 2003; Morgan et al., in press). The early Blancan Arroyo Cuervo LF is derived from the lower part of the local Camp Rice section below the 3.1 Ma pumice bed; the younger late Blancan Hatch LF occurs above this bed (Fig. 2). The Arroyo Cuervo LF includes the zebrine horse *Equus simplicidens*, the large peccary *Platygonus*, the small camel *Hemiauchenia*, the large camel *Camelops*, the deer *Navahoceros*, the gomphothere *Rhynchotherium falconeri*, the large land tortoises *Gopherus* and *Hesperotestudo*, the mud turtle *Kinosternon*, and an emydid turtle. The Hatch LF includes the glyptodont *Glyptotherium texanum*, the three-toed horse *Nannippus peninsulatus*, the one-toed horses *Equus cummingsii* and *E. simplicidens*, the camels *Hemiauchenia* and *Camelops*, the antilocaprid *Capromeryx*, the tortoises *Gopherus* and *Hesperotestudo*, and the turtle *Kinosternon*. The

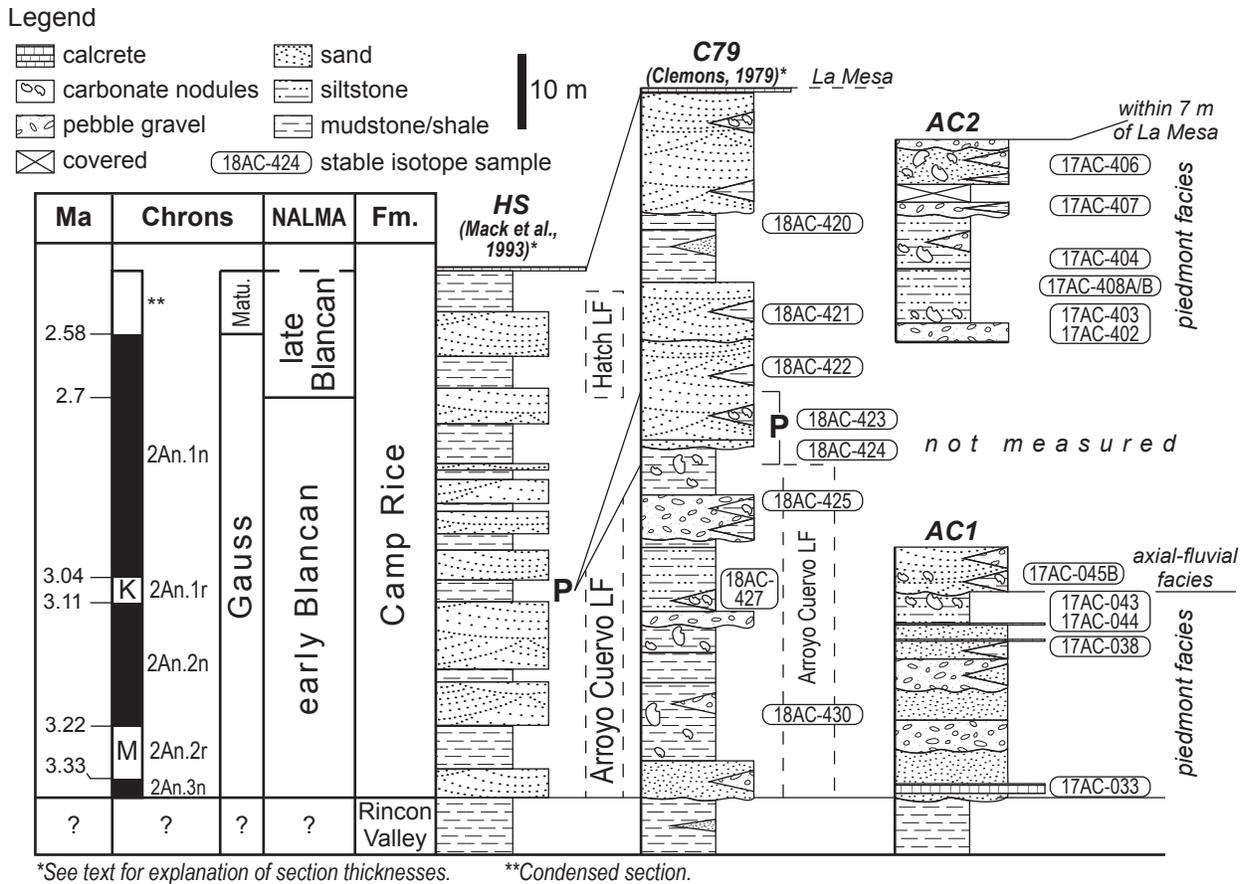


FIGURE 2. Stratigraphic sections in study area. Sections AC1 and AC2 are modified from Jochems (2017). Polarity reversal ages are from Berggren et al. (1995) and subchrons correspond to Geomagnetic Polarity Time Scale of Gradstein et al. (2012). Magnetostratigraphic section modified from Mack et al. (1993). K = Kaena subchron; LF = local fauna; M = Mammoth subchron; Matu. = Matuyama chron; NALMA = North American Land Mammal Age; P = pumice conglomerate bed or clasts.

co-occurrence in the Arroyo Cuervo LF of *Equus simplicidens*, *Camelops*, and *Platygonus* indicates an age younger than 3.6 Ma and the absence of South American Interchange mammals suggests an age older than 2.7 Ma. The presence in the Hatch LF of *Glyptotherium*, a South American Interchange mammal, indicates an age younger than 2.7 Ma and the presence of *Nannippus* suggests an age older than 2.4 Ma. In addition to mammalian biochronology, the 3.1 Ma pumice near the top of the Arroyo Cuervo LF (Mack et al., 1996) and magnetostratigraphy of the entire section (Mack et al., 1993) help to further constrain the ages of these two faunas as follows: Arroyo Cuervo LF ~3.6–3.0 Ma (late early Blancan) and Hatch LF ~2.7–2.6 Ma (early late Blancan). Most fossils were recovered from fine- to medium-grained, massive to cross-stratified or laminated, quartzose sand that was likely deposited at low-energy channel margins.

Stable carbon and oxygen isotopes

Calcareous paleosols in the Camp Rice Formation are typical of arid or semiarid soils where carbonate leaches from the surface and progressively precipitates in subsurface horizons (Gile et al., 1966; Machette, 1985). The ¹³C content of soil

carbonate is a function of soil and atmospheric CO₂ and the relative abundance of vegetative cover using the C₃ versus C₄ photosynthetic pathways (Cerling, 1984; Cerling and Quade, 1993). C₃ plants include trees, shrubs, and grasses that are generally adapted to cooler and more humid conditions, whereas C₄ plants include grasses that evolved under lower atmospheric p(CO₂) and are therefore better suited to warmer and drier conditions (Deines, 1980; Ehleringer et al., 1991). Due to diffusion and isotopic fractionation, average δ¹³C values of pedogenic carbonate formed in the presence of pure C₃ or C₄ biomass are -11‰ and 4‰, respectively (Cerling, 1984, 1991; Cerling and Hay, 1986; Cerling et al., 1988).

The oxygen isotope composition of pedogenic carbonate is controlled by the δ¹⁸O value of meteoric water that falls as precipitation onto the ground surface, which in turn relates to latitude and mean annual temperature (Cerling, 1984; Fricke and O’Neil, 1999). Soil temperature, infiltration, and evaporation also affect the isotopic composition of meteoric water (Cerling and Quade, 1993; Mack and Cole, 2005). Thus, relationships between oxygen isotope composition and climate can be estimated from vadose zone paleosols. However, carbonates that formed in proximity to the water table can have more variable δ¹⁸O values due to mixing of groundwater with soil water and

gases (Cerling, 1984; Slate et al., 1996; Mack et al., 2000), and may not offer a direct means of assessing paleoclimate.

METHODS

Twenty carbonate samples were analyzed. Paleosol carbonate samples were taken as nodules ≤ 2 cm in diameter from outcrops where possible, but samples 17AC-408A, 17AC-408B, 18AC-420, and 18AC-422 are laminar carbonate from strongly developed calcic horizons or groundwater deposits. Sample 17AC-402 consists of ~ 1 -mm-thick calcite rinds from piedmont gravel clasts and 17AC-033 is a massive micrite similar to spring deposits observed in the Palomas Basin (Mack et al., 2000). All soil carbonates were sampled at least 30 cm below the inferred top of the paleosols to avoid down-profile diffusion of atmospheric CO_2 that occurs in modern soils (Quade et al., 1989). Samples were filed, drilled, or ground to a fine powder and sieved with 53 or 63 μm sieves.

Samples were measured at the Center for Stable Isotopes, University of New Mexico, using the method described by Spötl and Vennemann (2003). The samples were loaded in 12 mL borosilicate vials that were then flushed with He and reacted for 12 hr with H_3PO_4 at 50° C. The evolved CO_2 gas was measured by continuous-flow isotope ratio mass spectrometry using a Gasbench device coupled to a Thermo Finnigan Delta Plus mass spectrometer.

Reproducibility was better than 0.1‰ for both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ based on repeat measurements of a laboratory standard (Carrara Marble). The laboratory standard was calibrated against NBS-19, for which $\delta^{13}\text{C} = 1.95\text{‰}$ and $\delta^{18}\text{O} = -2.2\text{‰}$. Isotopic values are reported in standard notation:

$$\delta^{13}\text{C} \text{ or } \delta^{18}\text{O} = \left(\frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) * 1000,$$

where R_{sample} is the ratio of heavy to light isotope in the sample and R_{standard} is the ratio of heavy to light isotope in the PeeDee Belemnite reference standard.

For temporal relationships discussed below, magnetostratigraphic ages from Wang et al. (1993) (who used geopolarity data from Johnson et al., 1975) and Mack et al. (1993, 1994a) were reconfigured to updated geopolarity ages of Berggren et al. (1995). With the possible exception of the Reunion subchron in the data of Wang et al. (1993), all subchrons from 2Ar through 1r.1r were present in measured magnetostratigraphic sections; these were used to interpolate the ages of isotope samples from their stratigraphic positions in relation to polarity reversals. We then compared their data with the stratigraphic intervals of our isotope data that are constrained by biostratigraphy and the 3.1 Ma pumice conglomerate (Mack et al., 1996; Morgan et al., in press). We binned our data into the age categories of >3.1 Ma, ~ 3.1 Ma, and <3.1 Ma.

RESULTS

Stable isotope samples from the study area display several trends by both parent-material type and inferred age (Table 1). Oxygen isotope values for soil and groundwater carbon-

ates from piedmont deposits of the Camp Rice Formation are higher on average (-6.8‰ ; standard deviation, s.d. = 0.9‰) than those from axial-fluvial parent material (mean = -7.4‰ ; s.d. = 0.6‰). On the other hand, axial-fluvial and piedmont carbonate have identical mean carbon isotope values (-4.3‰), although axial-fluvial samples are more variable (s.d. = 1.7‰). Isotopic values of samples from the western Hatch-Rincon Basin broadly overlap those from both axial-fluvial and hanging wall- and footwall-derived piedmont deposits in the Palomas Basin (Fig. 3; Mack et al., 2000). Our samples tend to have higher $\delta^{18}\text{O}$ values than other axial-fluvial Camp Rice Formation deposits in the central and eastern Hatch-Rincon Basin (Fig. 3; Mack et al., 1994a).

Pedogenic and groundwater carbonate in Camp Rice deposits of the study area are more similar in their oxygen isotope values than in their carbon isotope values, but average differences among the latter are relatively minor. Paleosols, including both nodular and laminar carbonate, have average $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of -7.2‰ and -4.1‰ , respectively. Groundwater carbonates, including a massive micrite interpreted as a shallow groundwater carbonate deposited at a spring, have average $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of -7.0‰ and -4.8‰ , respectively. Among paleosol samples, carbonate nodules and laminar carbonate yield average $\delta^{18}\text{O}$ values within 0.3‰ (-7.1 and -7.4‰ , respectively) with identical average $\delta^{13}\text{C}$ values (-4.1‰). Regardless of parent-material type, carbon isotope values consistently exhibit more variability (s.d. = 1.1 – 1.7‰) than oxygen isotope values (s.d. = 0.2 – 0.9‰). One groundwater carbonate sample (18AC-430) forms an outlier from the rest of our dataset ($\delta^{13}\text{C} = -5.8\text{‰}$; $\delta^{18}\text{O} = -6.0\text{‰}$; Fig. 3). This carbonate was collected just above the base of an arroyo and probably formed in association with the water table as it fell during late Pleistocene-Holocene incision.

Although we lack the geochronologic precision of stable isotope studies by Wang et al. (1993) and Mack et al. (1994a), we can confidently assign general ages to our samples based on their stratigraphic positions relative to the 3.1 Ma pumice bed as well as the Arroyo Cuervo local fauna (Figs. 4, 5). Only data from paleosol samples were used in this analysis, because they are most likely to be representative of past climate (e.g., Cerling, 1984; Amundson et al., 1989; Slate et al., 1996; Wang et al., 1996). Note, however, that shallow groundwater carbonates have been shown to reflect paleoclimatic conditions where they form quickly and in equilibrium with the soil CO_2 -soil water reservoir (Quade and Roe, 1999). From our paleosol dataset, we observe the following: (1) average $\delta^{18}\text{O}$ values increase from -7.6‰ for >3.1 Ma to -6.7‰ for <3.1 Ma samples (Fig. 4); (2) average $\delta^{13}\text{C}$ values increase from -4.9‰ to -3.8‰ (Fig. 5); (3) $\delta^{18}\text{O}$ values are more variable after 3.1 Ma; and (4) average oxygen and carbon isotope values are -7.4‰ and -2.7‰ , respectively, for two samples inferred to be approximately 3.1 Ma in age (Table 1).

Among pedogenic carbonate, variability between stratigraphically adjacent samples ranges from 0–2.0‰ for oxygen isotope values and 0.4–3.8‰ for carbon isotope values. The highest shifts are $+2.0\text{‰}$ for $\delta^{18}\text{O}$ between samples 17AC-407 and 17AC-404 in section AC2, and $+3.8\text{‰}$ for $\delta^{13}\text{C}$ between

TABLE 1. Stable oxygen and carbon isotope data from the Camp Rice Formation.

Sample ID	$\delta^{18}\text{O}_{\text{PDB}}$ (‰)	$\delta^{13}\text{C}_{\text{PDB}}$ (‰)	Stratigraphic Section*	m Above Base of Section	Description	Inferred Age (Ma)
<i>Piedmont facies</i>						
17AC-033	-6.5	-5.2	AC1	0.5	paludal carbonate	>3.1
17AC-038	-7.9	-5.7	AC1	20	paleosol carbonate nodules	>3.1
17AC-402	-6.6	-4.6	AC2	0.5	non-paleosol carbonate rinds	<3.1
17AC-403	-5.6	-3.9	AC2	1	paleosol carbonate nodules	<3.1
17AC-404	-7.2	-4.5	AC2	8	paleosol carbonate nodules	<3.1
17AC-406	-6.9	-3.7	AC2	23	paleosol carbonate nodules	<3.1
17AC-407	-5.2	-2.3	AC2	19	paleosol carbonate nodules	<3.1
17AC-408A	-7.2	-5.1	AC2	7	paleosol laminar carbonate	<3.1
17AC-408B	-7.4	-5.0	AC2	7	paleosol laminar carbonate	<3.1
18AC-421	-7.5	-2.9	C79	64	groundwater carbonate	<3.1
<i>Axial-fluvial facies</i>						
17AC-043	-7.9	-6.3	AC1	25	paleosol carbonate nodules	>3.1
17AC-044	-7.8	-2.5	AC1	24	paleosol carbonate nodules	>3.1
17AC-045B	-7.3	-4.8	AC1	30	paleosol carbonate nodules	>3.1
18AC-420	-7.5	-2.1	C79	76	paleosol laminar carbonate	<3.1
18AC-422	-8.2	-5.7	C79	57	groundwater carbonate	<3.1
18AC-423	-7.7	-3.9	C79	49	paleosol carbonate nodules	~3.1
18AC-424	-7.1	-1.4	C79	46	paleosol carbonate nodules	~3.1
18AC-425	-6.9	-5.0	C79	40	paleosol carbonate nodules	>3.1
18AC-427	-8.0	-5.4	C79	24	paleosol carbonate nodules	>3.1
18AC-430	-6.0	-5.8	C79	11	groundwater carbonate	>3.1

*AC1, AC2 from Jochems (2017); C79 from Clemons (1979).

samples 17AC-044 and 17AC-043 in section AC1. Shifts of similar magnitudes are observed between stratigraphically adjacent stable isotope samples in the datasets of both Wang et al. (1993) and Mack et al. (1994a).

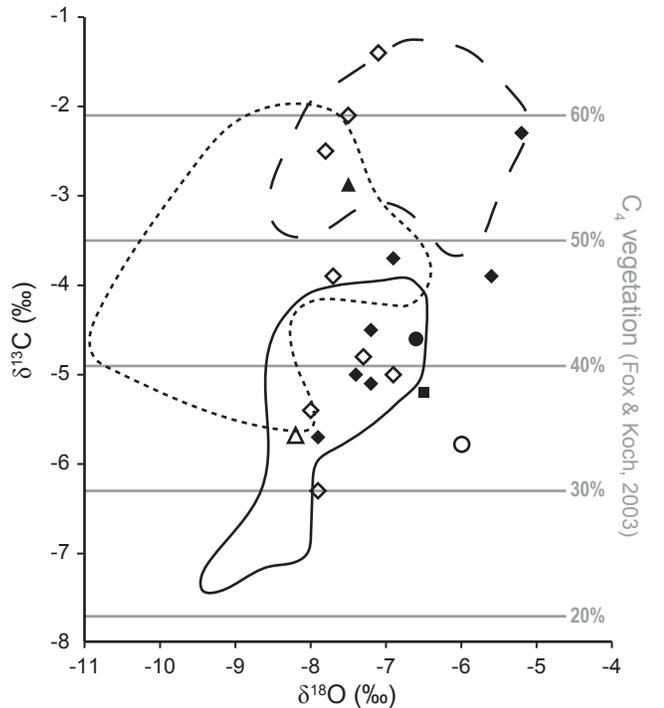
DISCUSSION

Because we did not sample organic matter in Camp Rice paleosols, we cannot directly assess the influence of soil productivity and atmospheric CO_2 on our $\delta^{13}\text{C}$ values (cf. Fox and Koch, 2003). However, the diversity of the Blancan local faunas (Morgan and Lucas, 2003; Morgan et al., in press) suggests that soil productivity was high compared to modern day, and that high $\delta^{13}\text{C}$ values are not related to organic-poor soils. The variability in $\delta^{13}\text{C}$ values for authigenic carbonates in the study area is unlikely to be the result of elevation differences, as the Camp Rice Formation exhibits no evidence for significant differential uplift or subsidence since the early Pleistocene, and because gradients of the ancestral Rio Grande and its tributaries were probably quite low (cf. Quade et al., 1989; Mack et al., 1994a). Diagenesis can be discounted as a control on $\delta^{13}\text{C}$ values because of small maximum (<328 ft/100 m) burial depth for the Camp Rice Formation (Mack et al., 1994a). Furthermore, petrographic examinations of samples 17AC-033, 17AC-038B, 17AC-044, and 17AC-045B show

no sparry calcite development, indicating little or no diagenetic alteration.

Isotopic values for shallow groundwater carbonate overlap with pedogenic carbonates in our dataset (Fig. 3). Mack and others (2000) observed a similar pattern in the Palomas Basin, which they ascribed to equilibrium conditions between groundwater carbonate and soil solution. Such conditions probably existed throughout the depositional interval of section AC1, where we interpret Camp Rice piedmont sediment as having been deposited at the distal toes of alluvial fans. Depositional gradients in this area would have been low and the water table shallow, with distal fans graded to the ancestral Rio Grande. Thus, groundwater was likely in equilibrium with soil water in the vicinity of the axial river. The lack of strongly ^{18}O -depleted shallow groundwater samples indicates that local aquifers were not connected to high-elevation source areas (cf. Mack et al., 2000).

Using the linear mixing model of Fox and Koch (2003, 2004), we infer that most Camp Rice paleosols sampled in this study supported <50% C_4 biomass, regardless of parent material (Fig. 3). However, most soil samples have $\delta^{13}\text{C}$ values indicative of >40% C_4 vegetation, with only two <3.1 Ma samples (duplicates 17AC-408A and 17AC-408B) having carbon isotope values below this threshold. Likewise, only two >3.1 Ma samples (17AC-044 and 17AC-045B) plot above the 40%



previous studies

- Palomas Basin, piedmont hanging-wall facies (Mack et al., 2000)
- - Palomas Basin, piedmont footwall and axial-fluvial facies (Mack et al., 2000)
- Hatch-Rincon Basin, axial-fluvial facies (Mack et al., 1994a)

this study

- ◆ piedmont paleosols
- ▲ piedmont groundwater
- piedmont carbonate rind
- piedmont cienega carbonate
- ◇ axial-fluvial paleosols
- △ axial-fluvial groundwater
- Holo-Pleist(?) groundwater

FIGURE 3. Stable isotope data with ranges of datasets from other local or regional studies (all reference standards are PDB). Percentage of C_4 plants using fractionation factor of 15.5‰ shown by gray horizontal lines (modified from Fox and Koch, 2003).

threshold. It seems that paleosols in the study area supported mixed vegetation regimes throughout the late Pliocene and early Pleistocene with a general (but not linear) trend towards greater C_4 abundance through time.

The relative abundance of Plio-Pleistocene C_3 vegetation in the study area suggests one of the following scenarios: (1) C_3 plants were common (30-70%) in the landscape throughout deposition of the Camp Rice Formation; or (2) C_3 vegetation consisted of forest grasses, shrubs, and trees before 3.1 Ma and transitioned to mixed C_3 desert shrubs and C_4 grasses (possibly with diminished vegetative cover/increased atmospheric CO_2 contribution) in the latest Pliocene and early Pleistocene. An analog for the latter situation is found in Holocene C_3 -dominated landscapes in southern New Mexico documented by Monger et al. (1998). There, isotopic and geomorphic evidence indicates that C_3 desert shrubs colonized a soil-poor land surface during the mid-Holocene.

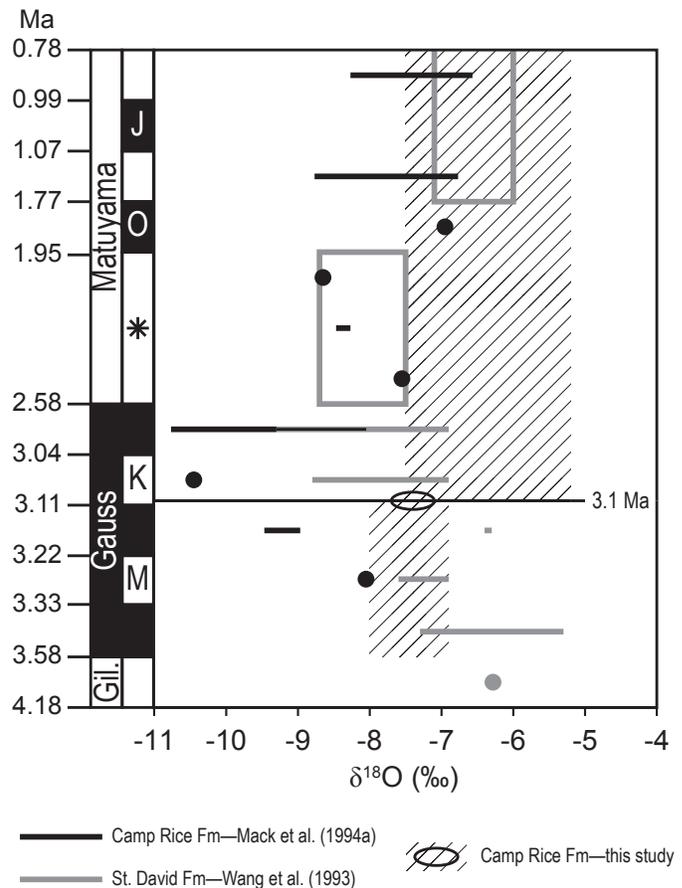


FIGURE 4. Pedogenic carbonate $\delta^{18}O$ data for local and regional studies through time. Lines show ranges of data from Wang et al. (1993) and Mack et al. (1994a); boxes and hachures show ranges of data from Wang et al. (1993) and this study, respectively. Circles are individual data points. Polarity reversal ages are from Berggren et al. (1995) and subchrons correspond to Geomagnetic Polarity Time Scale of Gradstein et al. (2012). Chrons: Gil. = Gilbert. Subchrons: M = Mammoth; K = Kaena; O = Olduvai; J = Jaramillo. Star represents Reunion subchron (2.15–2.14 Ma). Note non-linear age scale.

Interpretations of our carbon isotope data are complicated by the wide range (-6.3 to -1.4‰) in $\delta^{13}C$ values of soils formed in ancestral Rio Grande floodplain deposits. C_3 and perhaps some CAM (Crassulacean acid metabolism) vegetation were growing in the floodplain at the onset of Camp Rice deposition, as indicated by fossilized wood fragments of Leguminosae, Oleaceae, and/or Sapindaceae recovered from the lowest Camp Rice axial-fluvial beds (T. Dillhoff, pers. commun., 2017). The proportion of C_4 vegetation clearly increased through time in the floodplain (Figs. 3, 5). This result is consistent with other stable isotope records in the Hatch-Rincon Basin and southeastern Arizona (Wang et al., 1993; Mack et al., 1994a). However, all but one piedmont sample returned $\delta^{13}C$ values indicating <50% C_4 vegetation, regardless of inferred age. Based on the presence of browsing mammals (e.g., gomphotheres), it would be expected that woodlands replete with C_3 plants were more common along the river corridor than in piedmont settings (e.g., Lucas and Oakes, 1986). A possible explanation for these apparently contradictory results could be that increasing aridity in the early Pleistocene coupled with increasing salinity in shallow flood-

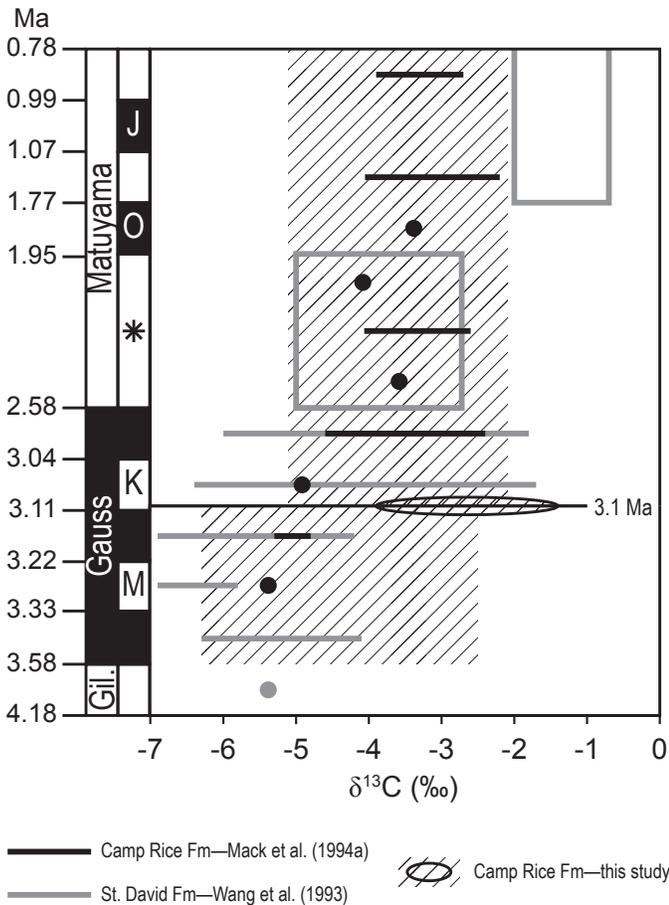


FIGURE 5. Pedogenic carbonate $\delta^{13}\text{C}$ data for local and regional studies through time. Lines show ranges of data from Wang et al. (1993) and Mack et al. (1994a); boxes and hachures show ranges of data from Wang et al. (1993) and this study, respectively. Circles are individual data points. Polarity reversal ages are from Berggren et al. (1995) and subchrons correspond to Geomagnetic Polarity Time Scale of Gradstein et al. (2012). Chrons: Gil. = Gilbert. Subchrons: M = Mammoth; K = Kaena; O = Olduvai; J = Jaramillo. Star represents Reunion subchron (2.15–2.14 Ma). Note non-linear age scale.

plain groundwater encouraged the growth of more resilient C_4 vegetation. Indeed, mud beds in the upper part of section C79 contain common gypsum casts and veinlets (Jochems, 2017).

Average $\delta^{18}\text{O}$ values for both groundwater and pedogenic carbonates (-7.0‰ vs. -7.2‰ , respectively) in the study area are 2–3.5‰ higher than for modern precipitation in southern New Mexico and southeastern Arizona (Bowen and Wilkinson, 2002; Bowen and Revenaugh, 2003; Bowen et al., 2005; Mack and Tabor, 2008). Most axial-fluvial paleosol oxygen isotope values are somewhat lower than those for piedmont soils (mean values -7.4‰ and -6.8‰ , respectively). Thus, soil and groundwater carbonates in the floodplain probably incorporated slightly ^{18}O -depleted water relative to their counterparts higher on the piedmont slope. Because precipitation falling on the study area would have had a nearly uniform oxygen isotope composition, this difference likely arises from groundwater mixing and/or evapotranspiration effects.

Similar to our $\delta^{13}\text{C}$ data, $\delta^{18}\text{O}$ values in study area soil carbonates increase somewhat after 3.1 Ma (Fig. 4) but in a non-linear manner. The increase could have occurred in con-

junction with a transition to some combination of warmer temperatures, enhanced evaporation, and/or an increase in the ratio of summer (monsoonal) to winter precipitation (Cerling, 1984). This interpretation is potentially consistent with our $\delta^{13}\text{C}$ data if the C_3 plants that continued to grow in the study area after 3.1 Ma consisted mostly of shrubs adapted to drier conditions.

The St. David Formation of southeastern Arizona also includes piedmont and axial stream sediment deposited in an intermontane basin. Up-section increases in pedogenic carbonate $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values in St. David paleosols have been attributed to increasing aridity and the transition to summer-dominated (monsoonal) precipitation; this transition followed a brief interval of wetter conditions at ~ 3 Ma (Smith et al., 1993; Wang et al., 1993). Mack and others (1994a) identify similar patterns among Camp Rice Formation paleosols in the central and eastern Hatch-Rincon Basin. One hypothesis invoked for increasingly dominant monsoonal moisture after ~ 3 Ma is an orographic effect created by the uplift of the Transverse Ranges and Sierra Nevada in California that blocked the influx of moisture from the Pacific Ocean to inland areas (Winograd et al., 1985; Smith et al., 1993). Unfortunately, we lack the geochronologic resolution to identify climate fluctuations on 10^5 yr timescales, with the possible exception of two samples indicative of transitional conditions at ~ 3.1 Ma (Figs. 4, 5).

Our dataset, particularly the subset of axial-fluvial paleosol data, generally support a transition to a drier and/or more seasonal climate through the early Pleistocene. However, two lines of evidence suggest that the climate of the late Pliocene in southern New Mexico was already somewhat warm and dry. First, the presence of the giant tortoise *Hesperotestudo* in the ≥ 3.0 Ma Arroyo Cuervo LF indicates frost-free conditions as this genus does not dig burrows and can therefore survive only in habitats with frost-free winters (Hibbard, 1960; Cassiliano, 1997). This and other large land tortoise genera disappear from the New Mexico fossil record after ~ 1.8 Ma, signifying a transition to colder, drier conditions (Morgan and Harris, 2015). Second, calcic soils form under dry conditions with mean annual precipitation of <50 – 60 cm/yr in the continental U.S. (Jenny, 1941; Birkeland, 1999). Although less common in the lower part of the Camp Rice Formation, stage II carbonate horizons suggest that paleoclimatic conditions were at least periodically arid to subhumid before ~ 3 Ma.

Although regional and global climatic changes can control depositional and isotopic evidence in the stratigraphic record (e.g., Zachos et al., 2001; Chapin, 2008), it is critical to assess whether patterns in isotopic data are affected by local factors, including the influence of groundwater or depositional setting. Although the St. David and Camp Rice records show a trend toward aridity through the early Pleistocene, Mack and Tabor (2008) found $\delta^{13}\text{C}$ values indicative of C_3 vegetation regimes in temporally correlative Gila Formation deposits in the Mangas Basin of southwestern New Mexico. These findings may relate in part to the higher elevation of the Mangas Basin (~ 150 m higher than the Rio Grande in the study area). However, those authors also identified distinctions between isotope values from piedmont and axial stream settings, where lower $\delta^{13}\text{C}$ values (indicating C_3 vegetation) were most common in flood-

plain soils. Depositional setting clearly plays a role in the isotopic composition of soils in the western Hatch-Rincon Basin, albeit in a way that is apparently opposite that of the Mangas Basin. Furthermore, intrabasin variability in factors such as groundwater chemistry are also important and may account for the disparity between the oxygen isotope values of our dataset and those of Mack et al. (1994a) from the central and eastern Hatch-Rincon Basin.

CONCLUSIONS

Carbon and oxygen stable isotope data from Camp Rice Formation paleosols in the western Hatch-Rincon Basin are generally consistent with increasing aridity and temperature through the latest Pliocene and early Pleistocene after ~3 Ma. However, vegetation remained mixed, particularly in piedmont settings; desert shrubs using the C_3 photosynthetic pathway may have taken over for more water-dependent vegetation but this hypothesis cannot be currently tested. In any case, local biomass was sufficient to support a relatively diverse faunal assemblage through the early Pleistocene. Oxygen isotope data indicate some combination of higher temperatures, greater evaporation, and/or greater summer precipitation through time.

Some factors preclude straightforward interpretations of paleoclimate in the study area. These include the close association of groundwater with soils formed in the ancestral Rio Grande floodplain. Additionally, the stronger tendency toward high carbon isotope values in floodplain paleosols is unusual given that more water-dependent vegetation is expected to grow in this setting. Thus, depositional setting and hydrologic controls potentially modify or obscure detailed interpretations of Plio-Pleistocene paleoclimate in the western Hatch-Rincon Basin. These factors should always be considered when inferring paleoclimate from terrestrial stable isotope records.

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View to the north of the valley of the Rio Grande in Selden Canyon, north of Radium Springs. Photograph by Greg H. Mack.