PALEOENVIRONMENTAL AND PALEOCLIMATIC INTERPRETATIONS OF THE LATE PALEOCENE GOLER FORMATION, SOUTHERN CALIFORNIA, U.S.A., BASED ON PALEOSOL GECHEMISTRY

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ABSTRACT: The Paleocene was characterized by high global temperatures and elevated concentrations of atmospheric CO2. Paleoclimatic records of this epoch are of particular interest to understanding the effects of anthropogenic climate forcing as well as linkages between the Paleocene climate and major tectonic and evolutionary events. However, terrestrial paleoclimatic records of the Paleocene are limited in their spatial extent, and few such records are known from western North America. In this study, paleosols from two members in the late Paleocene Goler Formation of Southern California were used to investigate local paleoclimatic conditions. The paleosols were described in detail and analyzed for morphology, clay mineralogy, and bulk geochemistry. Carbonate nodules present in paleosols from the upper member (member 4a) were characterized using petrographic and cathodoluminescence microscopy and were analyzed for carbon and oxygen stable-isotope composition. Enrichment and retention of soluble elements (Ca, Mg, and Na) and the mobilization of Fe and Mn in the paleosol profiles suggest a climate with strongly seasonal precipitation. This interpretation is further supported by the presence of pedogenic slickensides and the clay minerals smectite and illite. The appearance of pedogenic carbonate nodules in paleosols from the upper member is suggestive of a shift to more arid or more seasonal conditions that is coincident with a prominent change in sediment provenance and an inferred increase in sedimentation rate. Uplift of the ancestral Sierra Nevada during the late Paleocene was likely an important control on regional paleoclimate due to the effects of elevated topography on mean annual precipitation and seasonal distribution of precipitation. The correlation of a shift in pedogenic processes with changes in provenance and sedimentation rate suggests that the evolution of regional tectonic activity was an important control on these processes during the deposition of the Goler Formation.

INTRODUCTION

Paleoclimatic studies of the Paleocene Epoch (65.5–55.8 Ma) based on the marine record have revealed globally warm temperatures and elevated concentrations of atmospheric CO2 (Pearson and Palmer 2000; Zachos et al. 2001). While these aspects of the Paleocene are important to understanding long-term trends in Earth’s climate and their controls, analysis of regional paleoclimatic conditions during this period can be used to address additional questions. Particularly in terrestrial environments, regional conditions may be influenced by changes in topography caused by tectonic processes (Norris et al. 2000) and furthermore may influence the evolution and radiation of fauna and flora (e.g., Secord et al. 2012). However, the paucity of terrestrial deposits, difficulties associated with their dating, and the rarity of reliable paleoclimate proxies present significant challenges. In the case of the western United States, which experienced regional tectonic uplift as well as important mammalian radiation events during the Paleocene, there are few well-dated formations from which paleoclimatic records have been generated. In this study, newly discovered paleosols are described from the Goler Formation, a well-dated terrestrial late Paleocene deposit in Southern California, and are used to construct the westernmost terrestrial paleoclimate record of this age from the continental United States. In addition to recording the regional paleoclimate associated with the uplift of the Sierra Nevada, the Goler Formation also contains California’s oldest mammalian fossils (Albright et al. 2010).

Geologic Setting

The Goler Formation, exposed in the El Paso Mountains near Ridgecrest, California (Fig. 1), was deposited in a southwest-trending rift basin that formed as a result of widespread oblique extension along the Sierra Nevada magmatic arc and arc flank during the latest Cretaceous–early Paleogene (Wood and Saleeby 1997). The formation is over 3 km thick and is composed mainly of siliciclastic sediments deposited in three main depositional phases. Initial deposition occurred as canyon-fill, alluvial-fan, and distal alluvialPLAIN deposits (members 1–2 of Cox 1982), sourced from local igneous and metamorphic rocks. As the alluvial plain expanded northward, the deposition from a westward-flowing system of braided and meandering streams occurred within the basin (members 3–4c of Cox 1982). During this depositional stage, a change in sediment provenance from locally sourced (member 3) to more distally sourced (member 4) material occurred. Finally, a marine transgression submerged the stream system and deposited marine deltaic sediments (member 4d of Cox 1982, 1987). The formation unconformably overlies Paleozoic meta-sedimentary and metavolcanic rocks of the Mesquite Schist and the Garlock Formation as well as Paleozoic and
Mesozoic granitic rocks. The Goler Formation is unconformably overlain by the Miocene Ricardo Group (Cox 1982). This study focuses on newly identified paleosols in members 3 and 4 that were developed on siliciclastic deposits and experienced prolonged periods of subaerial exposure.

**Age Determination**

A Tiffanian North American land mammal age (56.8–60.2 Ma) for members 4a–c of the Goler Formation is suggested by the presence of mammalian fossils belonging to the genera *Plesiadapis*, *Tetraclaenodon*, and *Phenacodus* (Lofgren et al. 2008; Albright et al. 2010). Paleomagnetic analysis of intervals in members 4a–c has identified chronos 27r–25r, which confirm the Tiffanian age and has dated members 4a–c at 62–57.5 Ma (Albright et al. 2010).

Age assessment of the marine interval (member 4d) of the Goler Formation has yielded conflicting results. Analyses of coccoliths (Reid and Cox 1989), marine mollusks (Squires et al. 1988), and foraminifera (McDougall 1987) yield ages ranging from late Paleocene to early Eocene. The discovery of a single mammal tooth (genus *Phenacodus*) from the marine interval in conjunction with paleomagnetic analysis has suggested a late Paleocene age (58.4–57.2 Ma) for the marine interval (Albright et al. 2010).

Age determinations for members 1–3 are hindered due to lack of diagnostic fossil material. A clast of a silicic metavolcanic rock collected near the base of member 1a yielded a K-Ar age of 91.3 ± 2.7 Ma from biotite (Cox 1982); however, the absence of tuffaceous strata in the Goler provides indirect evidence that deposition began after a significant decline in local magmatism at about 80 Ma (Chen and Moore 1982).

**Regional Paleoclimate**

Analysis of plant and vertebrate fossil assemblages from the Goler Formation suggests that a warm and humid climate prevailed during deposition (Axelrod 1949; Cox 1982; Lofgren et al. 2008). Axelrod (1949) inferred the presence of a seasonal drought during the deposition of member 3 based on the size of fossil leaves. Conversely, petrographic analysis of petrified angiosperm logs from member 4 suggested no marked seasonality, although seasonal variations in water availability could have been suppressed in fluvial lowlands with a high water table (Cox 1982).

Cox (1982) identified a single paleosol profile in member 4a of the Goler Formation as well as a paleosol at the basal unconformity of the formation. Cox (1982) concluded that paleosols were largely absent from the formation and that the red color and pisolitic texture of the profile identified in member 4a corroborated the interpretation of a warm and humid climate. This study focuses on numerous new paleosol profiles identified in members 3c and 4a that record paleoclimatic and paleoenvironmental information (Torres and Gaines 2011).

**Paleosols as Paleoclimate Archives**

The distribution of major and trace elements in paleosol profiles can provide detailed information about weathering processes that were controlled by past climates. In non-eroding landscapes, the intensity of weathering, which is the extent to which soluble cations (e.g., Ca, Mg, Na, and K) are mobilized and removed from the soil profile, is controlled by the duration of subaerial exposure, mean annual temperature (MAT), and mean annual precipitation (MAP). With reduced intensity of weathering, soluble cations can accumulate in soil pore waters and lead...
to the formation of alkaline and alkaline earth-rich secondary minerals (e.g., smectite and carbonates; Sheldon and Tabor 2009). With higher weathering intensity, soluble elements are completely removed from the soil profile, resulting in the precipitation of secondary phases rich in Al and Si (e.g., kaolinite) and enrichment of refractory primary minerals (e.g., zircon). The distribution of redox-sensitive elements (e.g., iron and manganese) in soil profiles is also sensitive to climatic conditions. The diffusion of atmospheric O\(_2\) into the soil profile and its consumption by respiration are influenced by climatic factors via their impact on soil hydrology and biological activity (Armstrong et al. 1991; Yanful 1993; Silver et al. 1999).

The accumulation, depletion, and redistribution of elements in paleosol profiles can be assessed by normalizing the concentration of weathering-sensitive elements to the concentration of an index element presumed to be immobile during weathering (e.g., Nb, Ti, and Zr; Sheldon and Tabor 2009). Changes in the value of this ratio relative to the parent material reflect losses and gains associated with weathering processes and are insensitive to dilution/concentration effects. However, the assumption that index elements are immobile in soils is not always valid (e.g., Colin et al. 1993; Cornu et al. 1999), and their mobility must be assessed in order to accurately evaluate the role of weathering processes in the redistribution of elements in paleosol profiles. Similarly, the effects of sedimentation during pedogenesis must also be taken into account, because they may juxtapose fresh, minimally weathering sediment with deeply weathered material (Kraus 1999).

In addition to tracking the redistribution of individual elements, multielement ratios, like the chemical index of alteration (CIA),

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\text{CIA} = \frac{(Al)/(Al+Ca+Na+K)) \times 100}
\]

(1)
can be used to assess the intensity of weathering owing to the enrichment of Al and depletion of base cations during primary-mineral weathering and clay-mineral formation, with higher values of the CIA index indicating increased weathering intensity (Nesbitt and Young 1982). Due to diagenetic effects that may alter the concentration of potassium in paleosols, it has been proposed that potassium be omitted from the CIA calculation (CIA-K; Maynard 1992). This convention has been followed by more recent workers using paleosol geochemical data for paleoclimatic interpretations (e.g., Sheldon et al. 2002; Driese et al. 2005; Kraus and Riggins 2007).

The redox conditions during pedogenesis can be assessed using the sum of the molar concentrations of Fe and Mn divided by the Al concentration. The basis for this proxy is that under reducing conditions, Fe\(^{2+}\) and Mn\(^{2+}\) are stable as dissolved species and are leached from the soil solid phase, resulting in a low (Fe + Mn)/Al ratio, because Al is not mobilized significantly under most soil geochemical conditions (Retallack 2001).

The release of divalent cations (e.g., Ca, Mg, and Fe) into soil pore water can result in the formation of pedogenic carbonates under various climatic and soil conditions (Sheldon and Tabor 2009 and references therein). However, pedogenic calcium carbonate is typically found in environments with less than 100 cm MAP, because higher levels of precipitation generally favor carbonate dissolution (Royer 1999). In paleosols, the stable-isotope ratios of carbon (\(\delta^{13}C\)) and oxygen (\(\delta^{18}O\)) in pedogenic calcium carbonate can be used to infer the environment of carbonate precipitation and the associated pedoclimatic conditions.

In soils developed on noncalcareous parent material, dissolved inorganic carbon (DIC) for carbonate formation is sourced mainly from atmospheric CO\(_2\) and biologically respired carbon (Cerling 1984; Sheldon and Tabor 2009). Because these two sources have distinct carbon isotope ratios, the \(\delta^{13}C\) of pedogenic carbonate reflects mixing between these two components (Cerling 1984; Sheldon and Tabor 2009). The relative contributions of atmospheric and respired CO\(_2\) to soil DIC can vary as a function of atmospheric pCO\(_2\), primary productivity, carbonate precipitation depth, and soil hydrology (Cerling 1984; Mintz et al. 2011). Because these parameters are sensitive to climate conditions, the carbon isotope composition of pedogenic carbonate can be used to interpret paleoclimatic conditions. While there is also carbon isotopic fractionation associated with the dissolution and diffusion of CO\(_2\) in water, these fractionations are relatively invariant and minimally affect the \(\delta^{13}C\) of carbonates relative to mixing between sources (Cerling 1984; Sheldon and Tabor 2009).

The oxygen isotope ratio of pedogenic carbonate reflects contributions from the soil DIC as well as a temperature-dependent fractionation during carbonate mineral precipitation (Cerling 1984). The oxygen isotope ratio of soil DIC is determined largely by the isotopic exchange between DIC and pore water, which has an isotopic composition influenced by climatic processes such as precipitation and evapotranspiration (Hsieh et al. 1998; Stern et al. 1999). During carbonate precipitation, an additional temperature-dependent thermodynamic fractionation increases the \(\delta^{18}O\) of the mineral relative to the \(\delta^{18}O\) of DIC (O’Neill et al. 1969). Because diagenetic processes can influence both the carbon and oxygen isotope ratios of pedogenic carbonate, the use of supplementary textural and chemical data is necessary to evaluate the reliability of paleoclimatic interpretations based on the isotopic analysis of pedogenic carbonate (Budd et al. 2002; Dworkin et al. 2005).

**METHODS**

**Field Methods**

Seven paleosol horizons were identified in the Goler Formation (3 in mbr 3; 4 in mbr 4) by the presence of distinct horizons characterized by root traces, color mottling, and lack of primary depositional features (e.g., bedding). These were described in the field in detail and incorporated into two stratigraphic sections, one focusing on member 3 and the other on member 4, which were measured and described at the centimeter scale. The presence of pedogenic features in each unit (i.e., slickensides) was recorded during measuring, and paleosol horizons were sampled intensively. The base of the measured section in the member 4 locality corresponds to Raymond Alf Museum of Paleontology (RAM) locality V200510, which was determined to be within chron 26r and to have a maximum age of 61 Ma by Albright et al. (2010).

To avoid the effects of modern weathering, outcrops were excavated until fresh rock was exposed. For three separate paleosol intervals in member 3, samples were taken along vertical transsects from each distinguishable horizon. In member 4, carbonate nodules associated with paleosols and representative samples of the surrounding matrix were collected from four paleosol units. Due to the poor induration of the fine-grained units in member 4a, these nodules were collected mainly as float on steep, actively eroding outcrop surfaces where there was no modern soil formation. In this region, modern soil carbonate occurs as white surface coatings (cf. caliche) that are texturally distinct from Goler carbonate nodules.

**Laboratory Methods**

Clay mineralogy was determined by X-ray diffractometry of clay separates at Pomona College. The clay-mineral fraction was isolated from the samples collected using the methods outlined by Moore and Reynolds (1997). Clay-mineral isolates were suspended in deionized water and applied to glass slides. After application to glass slides, the samples were dried using three separate treatments: air-drying, 12-hour ethylene glycol solvation at 60°C, and heating at 500°C in a Nabertherm muffle furnace for 1 hour. Clay samples were scanned with a Rigaku Ultima IV X-ray diffractometer using Cu-K\(\alpha\) radiation with a running voltage of 40 kV and 44 mA over a range of 4 to 40° 2theta.
The bulk mineralogy of the member 4a carbonate nodules was determined by X-ray diffractometry of randomly oriented whole-rock powders using the same analytical conditions as those used for clay-mineral analyses, and the samples were scanned over a range of 10 to 60° 2theta. The spectra produced were analyzed using the MDI Jade 8 software program.

Whole-rock geochemical analyses were performed at Acme labs in Vancouver, British Columbia. Sample digestion was completed by heating a 0.25 g split in HNO₃, HClO₄, HF, and HCl. The solutions produced were analyzed by inductively coupled plasma atomic emission spectrometry (ICP-AES) for major and trace elements. A total of 24 whole-rock geochemical analyses from three paleosol profiles were performed with analytical uncertainty (2σ) of ≤ 10% for the reported elemental concentrations. In order to assess the accumulation/depletion of elements in the mudstone horizons, the molar element/Ti ratio of each horizon was normalized to the element/Ti ratio of the stratigraphically lowest horizon in each of the sampled mudstones. The multi-element weathering ratios Fe⁺Mn/Al and CIA-K were calculated using molar concentrations.

Thin sections of selected carbonate nodules (n = 7) were prepared and examined using a petrographic microscope. Cathodoluminescence (CL) microscopy was performed on thin (~ 0.5 cm) slabs as well as thin sections of carbonate nodules using a Reliotron cathodoluminescence instrument connected to a binocular microscope. Images from both methods were taken using a digital camera and merged using Adobe Photoshop CS.

Approximately 50 mg of powder for δ¹³C and δ¹⁸O analysis was drilled from polished slabs prepared from 20 carbonate nodules using a Dremel tool. The relatively large sample size for isotopic analyses was selected due to the large content of detrital silicates in the nodules. To avoid sampling sparry phases, which are present in some nodules as macroscopically distinct veins, drilling was completed while visually monitoring the sample site with a binocular microscope. Powders were extracted from both the cores and rims of selected samples from each nodule-bearing horizon identified. Isotopic analysis of a total of 25 samples was completed at the University of California, Davis. Samples were dissolved in phosphoric acid, and the CO₂ evolved was analyzed online by isotope-ratio mass spectrometry. The average precision of the analyses was reported to be ± 0.08‰ for δ¹³C and ± 0.12‰ for δ¹⁸O (2σ).

**RESULTS**

**Member 3 Outcrop Description and Paleosol Identification**

The section measured through part of member 3 (BPS) is composed of a succession of interbedded sandstones, siltstones, and varicolored mudstones typical of deposits of member 3c (Figs. 2, 3; Cox 1982).
Three mudstone units (BPS-01, BPS-02, and BPS-03) were observed in the two sections (Figs. 2, 3). Section 1 contains mudstone units BPS-01 and BPS-02, with BPS-02 being stratigraphically highest. Section 2, which contains BPS-03, is stratigraphically above section 1 by a few meters. The mudstone units vary from 0.16 to 1.13 m in thickness, lack primary depositional features, display horizonation with respect to color and texture, and contain features associated with pedogenesis such as root traces, slickensides, and color mottling (Fig. 4A–D). Based on these characteristics, these mudstone units are interpreted as paleosols.

**Paleosol Unit 1 (BPS-01).—** The base of the member 3c stratigraphic section contains a paleosol profile (BPS-01; Figs. 2, 3). The lowermost horizon is gray colored and displays a vertical variation in texture with abundant fine- to medium-sand-sized grains of quartz and feldspar at its base (BPS-01-MS). The upper portion of the gray horizon lacks abundant sand-size grains and contains two zones of color mottling, a lower zone of red-colored mottles (BPS-01-G) and an upper zone of gray-green-colored mottles with red margins (BPS-01-M). Carbonaceous fragments are sometimes present in the gray-green portion of the color mottles in BPS-01-M. The middle horizon is dominantly red colored (BPS-01-R) and contains orange color mottles and abundant root traces. The root traces are characterized by carbonaceous cores surrounded by gray-green-colored haloes (Fig. 4A). At the top of the profile is a thin (30 cm) orange-colored horizon (BPS-01-O) that contains red color mottles and root traces similar to those in the underlying red horizon.

**Paleosol Unit 2 (BPS-02).—** Above BPS-01 is another paleosol profile (BPS-02) that contains three distinctly colored horizons similar to those in the underlying BPS-01, but these horizons do not occur in the same stratigraphic order (Figs. 2, 3). At the base of the profile is a red horizon (BPS-02-R) that displays orange and gray-green color mottling, gray-green halosed root traces, and large (up to 25 cm) slickensides (Fig. 4B). In the gray-green-colored mottles, fragmented carbonaceous material is often present. Overlying the red horizon is a gray horizon that contains slickensides and orange color mottles at its base (BPS-02-T/G) and in the uppermost portion (BPS-02-G), contains millimeter-size iron oxide nodules and centimeter-size carbonaceous fragments (Fig. 4C). The carbonaceous fragments in BPS-02-G are distinct in that they do not occur within gray-green-colored portions of the paleosol and are not associated with root traces. At the top of the profile is a red horizon (BPS-02-T) that displays the same textures as BPS-02-R.

**Paleosol Unit 3 (BPS-03).—** The third paleosol profile (BPS-03) contains three distinctly colored horizons that display a pattern similar to BPS-02. At its base is a red horizon (BPS-03-B) that displays orange and gray-green color mottles. Above the red horizon is a gray horizon (BPS-03-DG). Overlying the gray horizon is a fine- to medium-grained sandstone (BPS-03-SS) that, unlike other sandstones in the measured section, contains carbonaceous root traces as well as orange and red color mottles. Above the sandstone layer is a red horizon (BPS-03-T) that contains orange- and gray-green-colored mottles with carbonaceous fragments present in the gray-green mottles.

**Member 4 Outcrop Description and Paleosol Identification**

In member 4, the sampled locality consists of a sequence of interbedded mudstones and conglomeratic sandstones typical of member 4a (Figs. 2, 5; Cox 1982). The mudstones in the member 4 section are significantly thicker (2.5–52 m) than those sampled in from the locality in member 3. The member 4 mudstones display both gray-green and red coloration and are typically massive, with bedding preserved only in some units (Fig. 6A). In units where bedding is absent, features such as centimeter-scale slickensides (Fig. 6B), root traces (Fig. 6C), color mottling, and carbonate nodules (Fig. 6D–F) are present and are indicative of pedogenic alteration. Based on these characteristics, the mudstone units without bedding features are interpreted as paleosols. Compared to the paleosols of member 3c, color mottling, slickensides, and root traces in member 4a paleosols are not as prevalent or intense.

Sampled carbonate nodules associated with member 4a paleosols range from 1 to 10 cm in size (maximum dimension). These nodules occur mainly as float on the surfaces of outcrops, but they also occur in situ in the mudstones. The carbonate nodules display two distinct morphologies...
(spherical–subspherical) and are colored either red or a mixed brown and gray; Figs. 6D–F, 7A). For the brown and gray-colored nodules, the different colors often occur as distinct domains (see below). Despite differences in morphology and color, the carbonate cement in all of the nodule types is dominantly micrite and was determined to be low-magnesium calcite by XRD. In addition to micritic carbonate, minor amounts of sparry calcite as well as silt- to sand-sized detrital grains are also present in many of the nodules. Fluorapatite and quartz were also identified in the nodules by XRD.

Present in some of the brown and gray-colored carbonate nodules are centimeter-size micrite domains characterized by changes in cement color (i.e., a gray domain in a brown nodule), crystal size, and oxide mineral content (Fig. 7A–C). Fragments of organic carbon are often associated with the micrite domains and either occur as isolated fragments or display a root-like morphology (Fig. 7B). In thin section, a cell-like microstructure is visible in both types of organic-carbon inclusions.

Cathodoluminescence petrography of the carbonate nodules reveals up to three distinct phases of cements characterized by bright orange luminescence, dull orange luminescence, and nonluminescence. In the samples analyzed, the gray-colored micrite domains were associated with the bright orange luminescence (Fig. 7D) whereas the brown cements either showed nonluminescence or displayed a complex mix of fine (micrometer-scale) brightly luminescent veins in a non-luminescent matrix (Fig. 7D). Sparry calcite displayed both bright and dull luminescence.

**Clay Mineralogy**

In all paleosol samples analyzed from the member 3c locality, kaolinite was identified based on diagnostic peaks at 12° and 25° 20 in the air-dried slide that collapsed with heating at 550°C. Smectite was identified in all samples except BPS-01-MS based on diagnostic peaks at 6.25°, 18.63°, and 26.72° 20 in the air-dried slide. With ethylene glycol solvation, both the 001 and 003 peaks for smectite shifted to lower 20 values of 5.30° and 15.94° respectively. The 002 reflection of smectite, which should appear with ethylene glycol solvation at ~ 10.5° 20, was observed only in one of
the solvated samples (BPS-01-M). Illite was also identified in all samples analyzed except BPS-01-MS based on diagnostic peaks at 8.73°, 17.73°, and 26.72°. Interlaying of smectite and illite in the member 3c paleosols is not supported by comparisons of the ethylene glycol solvated and air-dried samples, as ethylene glycol solvation of interlayered smectite–illite results in the shift of the smectite–illite 002/003 peak to lower 2θ values. In all the analyzed samples, the illite 002 peak remains in place as the smectite 003 shifts to a lower 2θ angle. Although the absence of the smectite 002 can be associated with interlayered smectite–illite, it is also the weakest reflection, and its absence within most of the analyzed samples may be related to low concentrations of smectite (Moore and Reynolds 1997).

In samples taken from the member 4a locality, kaolinite was identified based on characteristic peaks at 12° and 25° 2θ in the air-dried slide that collapsed with heating at 550°C. Illite was identified in all samples based on characteristic peaks at 8.81° and 17.71° 2θ in the air-dried slide. However, the low relative intensity of the peaks suggests that the illite is present in low concentrations compared to kaolinite.

**Immobility Index**

For the BPS-01 paleosol profile, variability (2σ) in the Zr/Ti, Nb/Zr, and Nb/Ti ratios of each horizon are equal to 13%, 15%, and 4% of the average value of the paleosol unit, respectively (Table 1). For the BPS-02 paleosol profile, variability (2σ) in the Zr/Ti, Nb/Zr, and Nb/Ti ratios of each horizon are equal to 18%, 14%, and 6% of the average value of the paleosol unit, respectively (Table 1). For the BPS-03 paleosol profile, variability (2σ) in the Zr/Ti, Nb/Zr, and Nb/Ti ratios of each horizon are equal to 13%, 6%, and 8% of the average value of the paleosol unit, respectively (Table 1). Due to the lower variability of the Nb/Ti ratio relative to the Zr/Ti and Nb/Zr ratios, Ti is used to normalize elemental concentrations in the following sections.

**Whole-Rock Geochemistry—Alkaline and Alkaline-Earth Trends (Member 3c)**

Whole-rock geochemical analyses of samples from the Goler Formation reveal multiple trends in element/Ti ratios associated with the sampled paleosols from the member 3c locality (Table 2, Fig. 8). In the first paleosol profile (BPS-01), Mg/Ti is depleted in the top horizons (BPS-01-O and BPS-01-R) and enriched the underlying gray horizons (BPS-01-M and BPS-01-G) relative to the lowest horizon (BPS-01-MS). The Na profile shows a similar pattern except that it is depleted in BPS-01-G relative to the lowermost horizon. The Ca profile shows a slight accumulation in BPS-01-R and a slight depletion in BPS-01-G. In all of the horizons, K is depleted relative to the base of the profile.

In the second paleosol profile (BPS-02), Mg, Ca, and Na are enriched throughout the profile when normalized to the lowermost horizon. For each element, the enrichment is greatest at the top of the profile (BPS-02-T). Potassium is also enriched at the top of the profile, whereas it is depleted in the lower horizons. Analysis of differently colored portions (red, yellow, and gray-green) of the BPS-02-R horizon reveals low element/Ti ratios in the gray-green portion and high ratios in yellow portions. While this variability can potentially affect the shape of the Ca and Na profiles (especially for BPS-02-T and BPS-02-TG), the magnitude of the whole-profile Mg enrichment and K depletion is much greater than the observed variability between differently colored portions of the paleosol matrix.

In the third paleosol profile (BPS-03), the alkaline and alkaline-earth elements show trends that are dissimilar to those in the other two profiles. The upper two layers (BPS-03-T and BPS-03-SS) are enriched in Na and K relative to the lower two layers (BPS-03-DG and BPS-03-B). The Ca profile shows a similar trend, with the exception of a lack of accumulation in the uppermost layer (BPS-03-T). The Mg profile is dissimilar from the other elemental profiles in that Mg is enriched in the lower portion of the profile (BPS-03-DG) and depleted in the upper portion (BPS-03-SS).

**Whole-Rock Geochemistry—Iron and Manganese Trends (Member 3c)**

In all of the sampled paleosol profiles, Fe/Ti ratios vary with matrix color (Table 1, Fig. 9). Orange- and red-colored portions of the paleosol matrix are enriched in Fe relative to gray- and gray-green-colored portions (Fig. 9). A similar chemical variation with paleosol color is exhibited by lower Fe + Mn/Al ratios in the gray- and gray-green-colored portions relative to the red- and orange-colored portions (Table 2, Fig. 9).
Whole-Rock Geochemistry—Weathering Ratios (Member 3c)

Values for the CIA-K of samples from member 3c range from 63–84 (Table 2). Each paleosol unit shows a distinct CIA profile with depth. The BPS-01 profile has minima at the base (BPS-01-MS) and in the middle portion (BPS-01-M) of the profile, and higher values throughout the rest of the profile. For the BPS-02 profile, the base of the profile has the highest CIA-K value, with the middle two layers of the profile having similar values and the top of the profile having a much lower value. Multiple samples of the base of the BPS-02 profile (BPS-02-R) reveal a range of CIA-K of 81–84 between differently colored portions of the paleosol matrix. For the BPS-03 profile, the two lowest paleosol horizons have higher values than the overlying layers.

Carbon and Oxygen Isotope Data

In the four nodule-bearing paleosols sampled, $\delta^{13}C$ ranges from $-8.61\%$ to $-20.72\%$ VPDB (Table 3, Fig. 10). Stratigraphic variation of $\delta^{13}C$ is much greater than $\delta^{13}C$ variation within individual horizons sampled. $\delta^{18}O$ ranges from $-8.46$ to $-14.16$ VPDB, with more significant variation within individual horizons (Table 3).

In the stratigraphically lowest paleosol sampled (GC), nodules display the least negative $\delta^{13}C$ ($-8.47\%$ to $-11.07\%$ VPDB). $\delta^{18}O$ ranges from $-12.06\%$ to $-12.84\%$ VPDB (Table 3, Fig. 10). For nodules GC-4 and GC-5, from which both core and rim material were analyzed, lower $\delta^{13}C$ values are seen in the brown-colored cores ($-10.60\%$ and $-10.43\%$) relative to the gray-colored rims ($-8.47\%$ and $-9.82\%$). The gray-colored cores exhibit slightly lower $\delta^{18}O$ ($-12.58\%$ and $-12.84\%$) relative to the brown-colored rims ($-12.37\%$ and $-12.09\%$).

In the second member 4a paleosol unit, nodules have lower in $\delta^{13}C$ and $\delta^{18}O$ values relative to other nodules analyzed ($\delta^{13}C$, $-16.52\%$ to $-20.72\%$ VPDB, $\delta^{18}O$, 13.87\% to $-14.16\%$ VPDB; Table 3, Fig. 10). Both core and rim material was analyzed from a single nodule in this horizon (2-1), revealing a difference in $\delta^{13}C$ of 3.47\% between the core and the rim, whereas a difference in $\delta^{18}O$ of only 0.15\% is present.

In the third member 4a paleosol unit, $\delta^{13}C$ ranges from $-12.32\%$ to $-18.10\%$ VPDB and $\delta^{18}O$ ranges from $-11.49\%$ to $-12.75\%$ VPDB (Table 3, Fig. 10). Both core and rim samples were analyzed from a single nodule in this horizon (3-4). The gray-colored core is strongly $^{13}C$-depleted ($-18.10\%$) compared to the brown-colored rim ($-12.91\%$), a difference of 5.19\% (Fig. 7A). A difference in $\delta^{18}O$ of 0.74\% is present between the core and the rim of nodule 3-4.
FIG. 7.—Member 4a carbonate nodule textures. A) Nodule 3–4 showing gray-colored micrite in a brown-colored micrite matrix. An offset of 5.19% in $\delta^{13}C$ is seen between the brown matrix and the gray micrite. The small dashed white square shows the area figured in Part C. The large white square shows the area figured in Part D. B) Root trace in carbonate nodule enclosed in a gray-colored micrite. The gray-colored region bears a lower concentration of iron oxide minerals relative to the surrounding matrix. C) Plane-polarized-light photomicrograph of the area enclosed in the dashed white box of Part A showing the distribution of iron oxide minerals between the gray micrite and the brown nodule matrix. D) Cathodoluminescence photomicrograph of the area enclosed in the large white square in Part A. The gray micrite corresponds to a domain of bright luminescence whereas the surrounding matrix is characterized by a mixture of dull and bright luminescence.

<table>
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<tr>
<th></th>
<th>Na/Ti</th>
<th>K/Ti</th>
<th>Ca/Ti</th>
<th>Mg/Ti</th>
<th>Nb/Ti</th>
<th>Zr/Ti</th>
<th>Fe/Ti</th>
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<td>4.52E-03</td>
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<td>4.38</td>
<td>1.68E-03</td>
<td>3.00E-03</td>
<td>9.71</td>
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In the fourth and stratigraphically highest member 4a paleosol unit, 
$\delta^{13}$C ranges from $-8.61\%$ to $-13.94\%$ VPDB and $\delta^{18}$O ranges from $-8.46\%$ to $12.90\%$ VPDB (Table 3, Fig. 10). Both core and rim material was analyzed from a single nodule in this horizon (4–16). $\delta^{13}$C of the brown-colored core ($-8.61\%$) is significantly enriched compared to the gray-colored rim ($-13.94\%$). There is a greater difference in $\delta^{18}$O between the core ($-8.46\%$) and the rim ($-11.16\%$) than was observed in nodules belonging to other intervals.

### DISCUSSION

#### Classification of Paleosols

Based on the paleosol classification scheme outlined in Mack et al. (1993), the paleosols from member 3c correspond to the paleosol order of gleyed vertisols based on the presence of gray-green color motting, slickensides, and smectite clays. Both field observations and analysis of whole-rock geochemical data suggest that pulses of sedimentation occurred during pedogenesis, resulting in the formation of both compound and composite paleosol profiles. Compound and composite paleosol profiles are vertically stacked profiles that are either separated by minimally weathered sediment (compound) or partially overlap (composite; Kraus 1999). The presence of pedogenically altered sandstone in the middle of the BPS-03 sequence is indicative of the formation of a compound paleosol profile and is suggestive of rapid and episodic sedimentation during pedogenesis (Fig. 2; Kraus 1999). Similarly, the repetition of horizons and the upper-profile enrichment in alkaline and alkaline-earth elements in the BPS-02 profile (Fig. 8) are consistent with the formation of a composite paleosol profile and suggest that the rate of pedogenesis greatly exceeded the rate of sedimentation, allowing vertically successive profiles to overlap (Kraus 1999).

Using the classification scheme of Mack et al. (1993), the paleosols of member 4a are interpreted as calcic protosols based on the paucity of pedogenic features observed (Fig. 6A–C) and the presence of carbonate nodules (Figs. 6D–F, 7A–D). The carbonate nodules present in the member 4a paleosols are interpreted as pedogenic in origin based on petrographic analysis, mineralogy, and the stable-isotope values of carbon and oxygen. The dominance of micritic low-magnesium calcite and the presence of organic-carbon inclusions with root-like morphologies and cell-like microstructures are consistent with carbonate precipitation within an active soil profile (Fig. 7B).

In member 4a, the poor development and greater thickness of the paleosol profiles relative to member 3a is indicative of the development of cumulative paleosol profiles. Cumulative paleosol profiles occur when the sedimentation rate is steady, allowing the successive deposition of thin increments of sediment during pedogenesis, ultimately forming a thick and poorly developed paleosol profile (Kraus 1999).

#### Paleoenvironmental and Paleoclimatic Interpretations-Member 3c

Based on evidence from whole-rock geochemistry, clay mineralogy, and morphology of the studied paleosol profiles, the paleoclimate that...
existed during the deposition of member 3c of the Goler Formation was characterized by alternating wet and dry cycles likely associated with highly seasonal precipitation. The red, yellow, and orange portions of the paleosols reflect periodic dry conditions due to their relative enrichment in Fe and Mn (Fig. 9). Dry soil conditions allow more rapid diffusion of atmospheric O\textsubscript{2} into the soil, creating an oxidizing environment (Yanful 1993; Silver et al. 1999), which at circumneutral pH favors the formation of secondary Fe\textsuperscript{3+} and Mn\textsuperscript{4+} oxide minerals, resulting in the retention or enrichment of these elements in the soil solid phase (Hem 1972). The gray horizons and the gray-green color mottles reflect periodic water-saturated conditions judging by the presence of organic matter and the relative depletion in Fe and Mn (Fig. 9). In water-saturated soils, the diffusion of atmospheric O\textsubscript{2} into the soil is slow and oxygen is consumed by respiration faster than it is replenished, which results in anoxic conditions (Armstrong et al. 1991; Yanful 1993; Silver et al. 1999). This depletion of oxygen in turn limits the aerobic decomposition of organic matter, allowing greater preservation efficiency (LaRowe and VanCappellen 2011). Under low-oxygen conditions, both Fe\textsuperscript{2+} and Mn\textsuperscript{2+} are stable in solution and may be leached from the soil solid phase, resulting in lower concentrations of both elements in the gray-green portions of the paleosol (Hem 1972).

Instead of forming as a result of pedogenic processes, the gray-green color mottles could have formed after burial of paleosol horizons, which would also have limited the flux of atmospheric O\textsubscript{2} into the soil. The presence of oxide-mineral nodules in low-iron intervals of the member 3c paleosols (BPS-02-G, Figs. 2, 4C) as well as oxide-mineral coatings on gray-green color mottles (BPS-01-M, Fig. 2) indicate reprecipitation of iron that was mobilized under reducing conditions. In order for iron reprecipitation to occur, it is necessary that soil redox conditions change from reducing to oxidizing, a shift that is inconsistent with formation under burial conditions.

The clay-mineral assemblage of the member 3c paleosols is interpreted to have resulted from \textit{in situ} pedogenesis as opposed to either detrital inheritance or later burial diagenetic alteration. The basis for this interpretation is the enrichment of Mg\textsuperscript{2+} in each of the paleosols profiles, which suggests the retention of Mg\textsuperscript{2+} in the solid phase by the precipitation of secondary minerals (e.g., smectite and illite). The lack of smectite and illite in the lowermost horizon of the BPS-01 (BPS-01-MS), which corresponds to a minimally weathered horizon of the profile based on the low CIA-K value (Table 2) and presence of sand-size grains, also suggests that these minerals were \textit{in situ} as a result of

Fig. 9.—Plot of Fe/Ti and (Fe + Mn)/Al for the member 3c paleosols by matrix color. Lower values of both ratios reflect greater mobilization of elements soluble under reducing conditions.
Table 3.—Carbon and oxygen isotopic ratios of member 4a paleosol carbonate nodules. Core and rim refer to the sampling of the inner and outer portions of individual nodules. The label “R” indicates samples taken from red-colored matrix.

<table>
<thead>
<tr>
<th>Sample</th>
<th>( \delta^{13}C ) (VPDB)</th>
<th>( \delta^{18}O ) (VPDB)</th>
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<tr>
<td>GC-3</td>
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<tr>
<td>GC-4 Core</td>
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<tr>
<td>GC-4 Rim</td>
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<tr>
<td>GC-5 Core</td>
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<td>-12.84</td>
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<tr>
<td>4-16 Core</td>
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<td>4-16 Rim</td>
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<td>4-17</td>
<td>-13.93</td>
<td>-11.99</td>
</tr>
<tr>
<td>4-19</td>
<td>-13.04</td>
<td>-12.28</td>
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</table>

Pedogenic alteration. The presence of kaolinite in the BPS-01-MS horizon may reflect some detrital supply of kaolinite or incipient silicate-mineral weathering in this lowest horizon. In modern soils from the Sierra Nevada region, kaolinite is found across a wide range of MAP conditions (Barshad 1966), which suggests that kaolinite likely formed in situ in the member 3c paleosols along with smectite and illite.

The pedogenic formation of smectite requires elevated pH and high activities of base cations and silica (Borchardt 1989; Wilson 1999). These geochemical conditions can be provided by climates with low to moderate precipitation, which act to concentrate these elements in soil porewaters by limiting the flux of water through the soil profile. In the presence of water, smectite clays expand as water is taken up into the mineral interlayer space (Moore and Reynolds 1997). Alternation between wet and dry conditions can result in the repeated shrinking and swelling of smectite and can produce slickensides, which are present in the member 3c paleosols (Figs. 2, 4B; Dixon 1991). The pedogenic formation of smectite together with evidence for its repeated shrinking and swelling suggests a paleoclimate with highly seasonal precipitation.

Pedogenic illite is reported to be a dominant clay phase in soils forming under arid climates (Singer 1989) and can be formed through the alteration of smectite during repeated wetting and drying cycles in the presence of K⁺ (Eberl et al. 1986). The presence of slickensides (Figs. 2, 4B) in member 3c paleosols indicates repeated shrinking and swelling of smectite in response to wetting and drying cycles, making this a possible mechanism for illite authigenesis. While the activity of K⁺ in these paleosols cannot be assessed directly, the depletion profile of K⁺ in all of member 3c paleosols suggests that it was significantly mobilized into the soil solution (Fig. 8). Additionally, K⁺ for the illitization of smectite in some modern coastal arid soils is sourced from marine aerosols (Jahn et al. 1987). The occurrence of a marine transgression at the top of the Goler Formation indicates that it was near the paleo-coastline (Cox 1987), which supports the possibility of K⁺ input from marine aerosols.

An additional control on the pedogenic formation of illite is the parent-material composition. In a study of pedogenic clay formation on igneous rocks in Sierra Nevada region of California, Barshad (1966) found that only felsic rocks give rise to the formation of illite regardless of MAP. Provenance determinations for member 3c by Cox (1982) suggest a granitic source, on the basis of conglomerate clast composition and the dominance of arkosic sandstones, which is consistent with pedogenic formation of illite.

In member 3c, the retention and accumulation of alkaline and alkaline-earth elements is shown by whole-rock geochemical data. Upper-profile depletion of magnesium, calcium, and sodium in the BPS-01 sequence (Fig. 8) reflects the removal of these elements from the upper part of the profile. Corresponding mid-profile enrichment of magnesium, calcium, and sodium indicates that the amount of precipitation was insufficient to remove these elements from the profile, and they instead accumulated in the B horizon. Similarly, the mid-profile enrichment of magnesium, calcium, and sodium in the BPS-02 sequence (Fig. 8) also reflects the retention of these elements in the paleosol profile. The lack of upper-profile depletion of calcium, magnesium, and sodium in the BPS-02 sequence is likely the result of the development of a composite paleosol profile. Under this model, we infer that the top horizon of the BPS-02 sequence (BPS-02-T) was deposited on top of an already developed profile and did not experience the same amount of weathering, resulting in elevated concentrations of alkaline and alkaline-earth elements relative to the rest of the profile. Further evidence for this interpretation is provided by the low CIA-K value of the BPS-02-T horizon relative to the rest of the BPS-02 succession, which is suggestive of less intense weathering (Table 2). The complex interplay of sedimentation and pedogenesis in the BPS-03 horizon, as evident in the complex element trends (Fig. 8) and the presence of pedogenically altered sandstone (Fig. 2) hinders determination of the effects of leaching in this profile.
Paleoenvironmental and Paleoclimatic Interpretations—Member 4a

Two distinct groups of nodules, high δ13C (n = 21, −14 to −8‰) and low δ13C (n = 4, −21 to −16‰), are distinguished from the isotopic data (Table 3). Nodules with high δ13C values likely formed from the mixing of DIC derived from atmospheric CO2 (δ13C ~ −4.5 to −6.5‰ PDB; Sinha and Stott 1994) with that derived from biological respiration of Corg (δ13C = −24 to −33‰ PDB for C3 plants; Deines 1980; Cerling and Quade 1993). Variation in the relative contribution of these two components is expected to result from varying depths of precipitation and amounts of biologic respiration (Cerling 1984).

The presence of nodules with low δ13C values (−21 to −16‰) reflects low δ13C values of soil DIC, which could have resulted from multiple soil processes. Potentially, seasonal changes in soil drainage conditions could result in large changes in the δ13C of soil DIC by limiting the diffusion of atmospheric gases (e.g., CO2 and O2) into the soil during water-saturated conditions. This condition results in the diminishment of isotopically enriched atmospheric CO2 contributions to soil DIC and promotes the formation of 13C-depleted DIC derived from anaerobic respiration of Corg (Tabor et al. 2007; Mintz et al. 2011). The correspondence of low δ13C values with a brightly luminescent carbonate cement in nodule 3–4 (Fig. 7A, D) supports the precipitation of the low-δ13C cements under suboxic or anoxic conditions, because Mn2+ causes luminescence in calcite (Hemming et al. 1989), is stable in solution under low-oxygen conditions (Hem 1972). Similarly, the lower concentration of oxide minerals in areas with lower δ13C values and luminescent cements relative to areas with higher δ13C values and nonluminescent cements further supports this interpretation, because these minerals can undergo reductive dissolution under low-oxygen conditions (Fig. 7A, C, D; Favre et al. 2002).

Seasonal changes in soil drainage conditions caused by either highly seasonal precipitation or changes in evapotranspiration should each result in changes in the δ18O of soil pore water. Thus, the lack of a clear difference in the oxygen isotope composition of the high-δ13C and low-δ13C nodules in the member 4a paleosols is seemingly inconsistent with a seasonal change in soil drainage conditions. However, many authors have reported that the effects of evaporation on the δ18O of soil pore water are most intense in the upper portions of the soil profile (−10 cm) and that the oxygen isotope composition of pore water at intermediate depths is less subject to evaporation and mainly reflects a mixture of evaporated water and precipitation from individual events and seasons (Breecker et al. 2009). Thus, if the Goler nodules precipitated at intermediate depths in the soil (i.e., 30 to 50 cm), they may have been buffered against seasonal changes in the δ18O of precipitation or evapotranspiration, but still have been influenced by seasonal changes in soil redox conditions.

Alternatively, the lack of variation in the oxygen isotope data could be the result of meteoric diagenesis. Mintz et al. (2011) found in modern Vertisols that carbonate nodules formed during different seasons displayed similar δ18O values despite large variations in their carbon isotope composition and formation at different temperatures and with differing intensities of evapotranspiration. To explain the lack of δ18O variation, the authors argue that the nodules had recrystallized in the active soil profile.

The presence of organic-carbon inclusions in some nodules suggests that the in situ oxidation of organic carbon in nodules may have caused the observed carbon isotopic variation. The oxidation of organic carbon in carbonate nodules would locally consume the available oxygen, leading to dysoxic or anoxic microenvironments and would produce DIC with a low δ13C value. This production of low-δ13C DIC could then lead to the alteration of the original carbonate, resulting in the lower measured δ13C value. While δ13C analyses were not preformed on any carbonate cements adjacent to organic-carbon fragments, these cements are characterized by the similar microfabrics (i.e., bright luminescence and mobilization of oxide minerals; Fig. 7B) to cements with low-δ13C values (Fig. 7C).

Both scenarios outlined above require anaerobic respiratory pathways that produce bicarbonate as opposed to CO2(g) (e.g., NO3−, Mn2+, and Fe3+ reduction, Soetaert et al. 2007). This is because DIC is enriched in δ13C relative to CO2(g) at equilibrium and would produce carbonates with carbon isotope ratios higher than those observed for the nodules in question if Corg was oxidized to CO2(g) and not directly to HCO3−. While there is some evidence for appropriate anaerobic respiratory pathways (i.e., enrichment of Mn2+ in the carbonates that is consistent with Mn2+ reduction), pedogenic carbonates with low carbon isotopic ratios may also be produced from the oxidation of methane diffused upwards from deeper, anaerobic portions of the soil (e.g., Bowen and Bloch 2002).

Distinguishing the effects of water saturation, organic-carbon oxidation pathways, and early diagenesis on the observed carbon and oxygen isotope compositions of the member 4a carbonate nodules is difficult based on the available data. Additionally, none of these processes are mutually exclusive, and it is possible that multiple controls on the isotopic composition of the member 4a carbonate nodules were operative. Despite the difficulty in assigning a specific process to the formation of pedogenic carbonate in the member 4a paleosols, their presence is suggestive of arid to semihumid conditions, because they require the accumulation of highly soluble Ca2+ ions, which are readily leached from the profile under higher-precipitation conditions (Royer 1999). Additionally, the supersaturation and precipitation of carbonate minerals in modern soils has been linked to seasonal changes in precipitation and evapotranspiration (Breecker et al. 2009). The presence of slickensides in member 4a paleosols is also consistent with seasonal changes in soil moisture conditions, in that the shrinking and swelling of clay minerals requires cycles of wetting and drying (Moore and Reynolds 1997).

Paleosols in members 3c and 4a are distinct in their morphology and geochemistry, but both indicate seasonal changes in soil moisture conditions during pedogenesis. The shift in pedogenesis seen between the two members and the lack of carbonate nodules in member 3c may be accounted for by changes in sedimentation rate, sediment provenance, or climatic conditions. While the sedimentation rate for each member has not been independently assessed, the presence of well-developed composite and compound paleosols in member 3c suggests a lower sedimentation rate relative to member 4, which contains poorly developed cumulative paleosols. While this change is consistent with the observed changes in many pedogenic features between the two paleosol types, it is inconsistent with the change in pedogenic carbonate accumulation, in that a lower sedimentation rate during the deposition of member 3c would result in a greater duration of subaerial exposure, which has been shown to result in greater accumulations of pedogenic carbonate (Gile et al. 1966; Machette 1985).

Differences in clast composition between conglomerates in members 3c and 4a identified by Cox (1982) suggests a change in provenance between the two members, in that sediments from member 3c contain a greater fraction of locally derived sediment. However, it is thought that the calcium required for the precipitation of pedogenic carbonate is usually derived from the deposition of continental dust and/or marine aerosols (Gardner 1972; Bachman and Machette 1977; Machette 1985). Thus, the lack of carbonate in the member 3c paleosols is likely not the result of the change in provenance between the two members.

Inasmuch as the observed changes in sedimentation rate and provenance likely do not account for the apparent absence of pedogenic carbonate in member 3c paleosols, it is possible that their appearance in member 4 reflects a paleoclimatic shift between the deposition of the two members. The dominant climatic controls on pedogenic carbonate precipitation are thought to be MAP (Royer 1999) and seasonality with respect to precipitation (Breecker et al. 2009). Thus, it is possible that the appearance of pedogenic carbonate in member 4a was the result of a decrease in MAP and/or an increase in seasonality with respect to precipitation relative to conditions during the deposition of member 3c.
The Role of Paleotopography in Controlling Precipitation

On continents, areas of elevated topography have the effect of limiting precipitation to adjacent areas of low topography by means of the rain-shadow effect. This effect is caused by forced release of moisture from air masses by cooling and expansion as they pass upward over elevated topography, which results in decreased precipitation on the lee side of a topographic barrier. Elevated topography can also cause seasonal changes in precipitation due to the heating and cooling of air masses over land. During the summer season, the heating of elevated topography leads to the heating of the overlying air mass. This produces a rising, low-density air mass that causes air flow towards the mountain front, ultimately resulting in increased precipitation. Conversely, winter cooling results in wind directions that blow from the elevated topography towards the ocean, preventing the flux of moist ocean air to the continent. This effect is exaggerated over areas of elevated topography relative to low topography due to the fact that there is less air mass above elevated topography, making it easier to heat (Norris et al. 2000).

Paleocene paleogeographic reconstructions of the western United States place the Goler Formation near the paleo-coastline with prominent mountain ranges, including the ancestral Sierra Nevada, separating it from the Pacific Ocean (Nilsen and McKee 1979; Nilsen 1987; Lechler and Niemi 2011). The presence of a topographic barrier may have been an important factor driving seasonal precipitation in the Goler Formation during the Paleocene. Estimates of paleo-elevation by Lechler and Niemi (2011) suggest 1–2 km of elevation for the southern Sierra Nevada during the Paleocene based on the low δ18O values of lacustrine (Lechler and Niemi 2011) and pedogenic (this study) carbonates from the Goler Formation relative to Eocene seawater δ18O estimates from pedogenic goethite (Yapp 2008) and organic biomarkers (Hren et al. 2010). At modern elevations, the Sierra Nevada has a profound impact on the arid climate of the Great Basin via the rain-shadow effect. Evidence for a correlation between mountain elevation and the intensity of its climatic effect (Kutzbach et al. 1989) supports the hypothesis that the lower 1–2 km elevation of the Paleocene Sierra Nevada was the cause for the highly seasonal, but not entirely arid, climate of the Goler Formation. Inasmuch as the observed change in paleoclimatic conditions between members 3c and members 4a is coincident with an apparent increase in paleo-elevation by Lechler and Niemi (2011), the dynamics of regional tectonic uplift were likely an important control on the regional paleoclimatic of Southern California during the late Paleocene.

Acknowledgments

This work was supported by funding from the Pomona College Geology Department, The Pomona College Dean’s Office, and the Picker College Dean of Faculty’s office. The authors would like to thank D. Logan, A. Farke, C. Windham, and D. Manning for their assistance in the field. Thorough reviews of this manuscript provided by J. Smith, G. Bowen, and G. Ludvigson were greatly appreciated.

References

LATE PALEOCENE PALEOClimAtE OF sOUTHERN CALIFORNIA BASED ON PALEOSOL GEOCHEMISTRY


Received 2 June 2012; accepted 19 March 2013.