[American Journal of Science, Vol. 314, December, 2014, P. 1436–1461, DOI 10.2475/10.2014.02]

ISOTOPE HYDROLOGY OF EARLY PALEOGENE LAKE FLAGSTAFF, CENTRAL UTAH: IMPLICATIONS FOR CORDILLERAN EVOLUTION

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ABSTRACT. Lacustrine sediments from Paleogene foreland basins of the North American Cordillera serve as powerful integrators of information on climatic and tectonic conditions, but episodes of hydrologic connection between basins may confound straightforward interpretations. We use stable isotope ratios of carbon and oxygen in authigenic carbonates of the Flagstaff Formation in central Utah to investigate the history of the lake system which produced these sediments. Flagstaff limestone and dolostone exhibit isotopic compositions consistent with having formed from the same fluid at surface conditions, indicating high evaporative concentration was unnecessary to drive dolomite precipitation. Isotopic data are consistent with the Flagstaff Formation having been produced by a large lake which occupied its basin from the late Paleocene into the early Eocene. Lake Flagstaff was hydrologically distinct from Lake Uinta to the north, but may have been connected to Lake Claron in the south. Paleoaltimetry estimates suggest that the mean elevation of Lake Flagstaff's catchment in the Sevier Highlands did not exceed 1.5 km. The growing evidence from stable isotope data that Cordilleran elevations were relatively low in the early Paleogene is in stark contrast to structural and sedimentological evidence for a high Nevadaplano and a topographically rugged Sevier thrust belt.

Key words: Flagstaff Formation, Eocene, stable isotopes, North American Cordillera, Sevier orogeny, dolomite, paleoaltimetry, paleolimnology

INTRODUCTION

Lacustrine deposits are common in the rock record of continental environments. Because they integrate information on hydrologic and environmental conditions for their surrounding drainage, authigenic lacustrine sediments have proven a particularly useful tool in the study of ancient and modern landscapes (Talbot, 1990; Arenas and others, 1997; Li and Ku, 1997; Saez and Cabrera, 2002; Paz and Rossetti, 2006; Leng and Henderson, 2013). A multitude of recent studies have used stable isotope analysis of lacustrine carbonates to investigate the history of the large lake systems which evolved along the Western Cordillera in association with the Laramide orogeny (fig. 1A) (Bowen and others, 2008; Carroll and others, 2008; Davis and others, 2008; Gierlowski-Kordesch and others, 2008; Smith and others, 2008; Davis and others, 2009a; Davis and others, 2009b; Doebbert and others, 2010; Bristow and others, 2012). These studies reveal a complex history of changing landscapes in response to the dynamic tectonics and climate of the early Paleogene. The Flagstaff Formation of central Utah provides one such record, deposited by a lake that formed in an intermontane basin bounded on the west by the frontal thrust of the Sevier orogenic belt, and on the east by the Laramide San Rafael Swell (fig. 1A) (Stanley and Collinson, 1979). Flagstaff Formation carbonates have been the subject of previous isotopic study (Weber, 1964; Neat and others, 1979; Bowen and others, 2008; Gierlowski-Kordesch and others, 2008) focused at sites on the Wasatch Plateau, where the best exposures are located.

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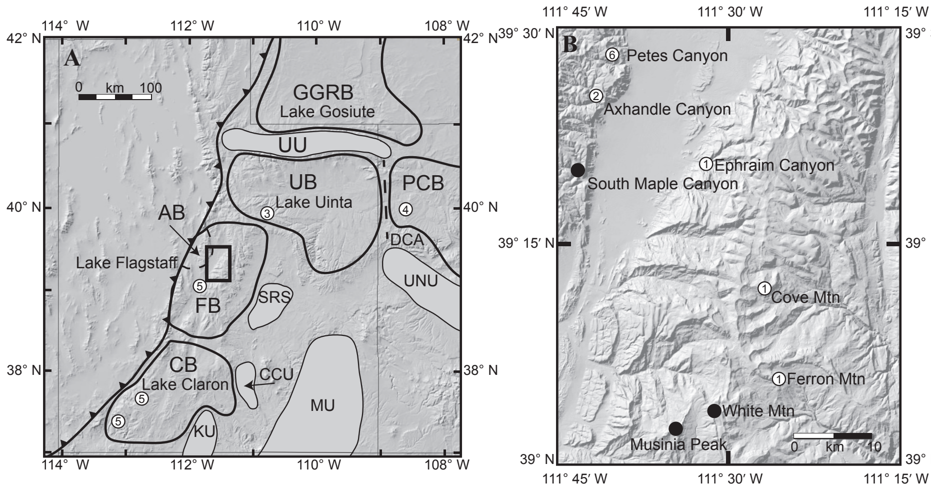

Fig. 1. (A) Digital elevation map of Utah and surrounding states, with Paleogene topographic features superimposed, after Lawton (2008) and Davis and others (2008). Structures are labeled as follows: AB, Axhandle Basin; CB, Claron Basin; CCU, Circle Cliffs Uplift; DCA, Douglas Creek Arch; FB, Flagstaff Basin; GGRB, Greater Green River Basin; KU, Kaibab Uplift; MU, Monument Uplift; PCB, Piceance Creek Basin; SRS, San Rafael Swell; UB, Uinta Basin; UNU, Uncompahgre Uplift; UU, Uinta Uplift. Bold black rectangle shows extent of (B), sampled localities referred to in text. Black circles are sites from this study, white, numbered circles are from (1) Bowen and others (2008), (2) Bowen and Bowen (2008), (3) Davis and others (2008), (4) Davis and others (2009a), (5) Davis and others (2009b), and (6) Chamberlain and others (2012).

The recent study of Bowen and others (2008) reported new, high-resolution isotopic and stratigraphic data relevant to reconstructing the paleoenvironmental geochemistry of Lake Flagstaff. This previous study was somewhat limited stratigraphically and spatially, in that it reported data only from the lower two members of the Flagstaff Formation, and did not investigate exposures beyond the Wasatch Plateau. It therefore provided an incomplete picture of the lake's evolution. Here, we report new data on stable isotope geochemistry, mineralogy, stratigraphy and petrology from the upper Flagstaff Formation, including sites on the southern Wasatch Plateau and in the San Pitch Mountains (Gunnison Plateau) situated ~40 km to the northwest in the syntectonic Axhandle Basin (fig. 1B) (Lawton and Trexler, 1991; Talling and others, 1995). We integrate this new data with previous studies to produce a paleohydrological record of Lake Flagstaff, which can be compared with records from other Paleogene lacustrine basins situated to the north and south along strike of the Sevier thrust. These comparisons allow us to test previously hypothesized episodes of connection between Lake Flagstaff and the adjacent basins, and to better understand the topographic setting in which the lake evolved.

SETTING

Beginning in the latest Cretaceous, drainage reorganizations as a result of the Sevier and Laramide orogenies were conducive to the development of large and well-developed lacustrine systems along the central Western Cordillera of North America (Dickinson and others, 1986; Dickinson and others, 1988; Yuretich, 1989; Lawton, 2008) (fig. 1A). In the Flagstaff Basin this transition occurs where the fluvial sandstones, siltstones and paleosols of the North Horn Formation are replaced by and locally interfinger with lacustrine carbonates and mudstones of the Flagstaff Formation (fig. 2) (Stanley and Collinson, 1979; Wells, 1983; Bowen and others, 2008). Non-marine limestones on the Wasatch Plateau were first described as a member of the

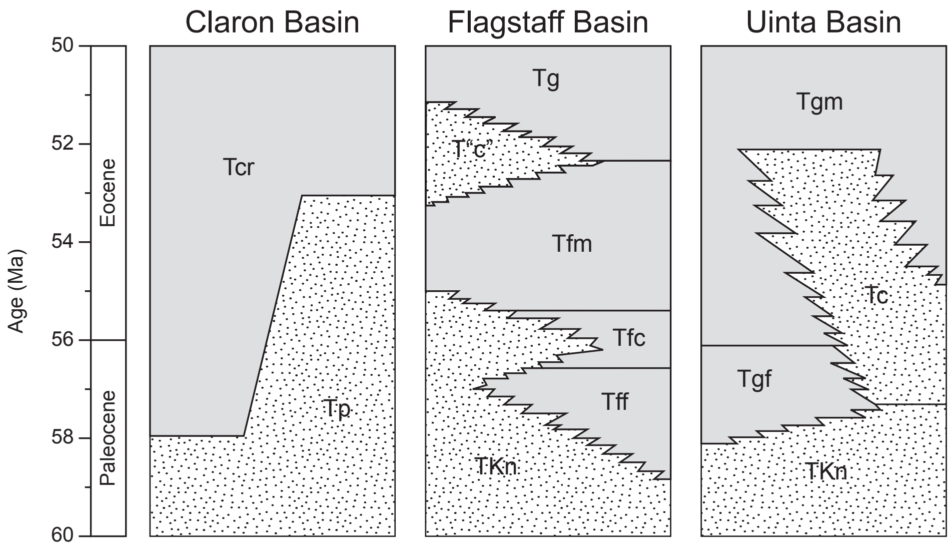

Fig. 2. Generalized chronostratigraphy of the major basins referred to in the text. Units are plotted from west to east within basins. Lacustrine units are shaded gray and fluvial units are patterned. Tp, Pine Hollow Fm; Tcr, Red Mbr of the Claron Fm; TKn, North Horn Fm; Tff, Tfc, Tfm, Ferron Mtn, Cove Mtn and Musinia Pk Mbrs of the Flagstaff Fm; Tc, Colton Fm and correlative in Flagstaff Basin; Tgf, Tgm, Tg, Flagstaff Mbr, Main Body and Undivided Green River Fm.

Wasatch Formation by Spieker and Reeside (1925); Spieker (1946) later elevated the Flagstaff Limestone to formation status. Subsequent investigations extended mapping of the Flagstaff Formation as far north as the Uinta Basin, where the Flagstaff was later demoted to a member of the Green River Formation (Ryder and others, 1976), and to the west and south, including the Gunnison Plateau (also called the San Pitch Mountains), the Valley and Pavant ranges, and the Fish Lake Plateau (Stanley and Collinson, 1979).

On the Wasatch Plateau, Stanley and Collinson (1979) divided the Flagstaff Formation into three members. The lower Ferron Mountain Member is distinguished by the dominance of carbonate deposition, and calcitic limestones within the Ferron Mountain Member contain abundant freshwater gastropod fossils. The dominant carbonate mineralogy shifts from calcite to dolomite up-section, and invertebrate fossils become markedly less common. The Ferron Mountain Member gradually transitions into the middle Cove Mountain Member, consisting predominantly of a fossiliferous mudstone and thin dolomite beds. Evaporative concentration of the lake during deposition of the Cove Mountain Member is indicated by mud cracking, paleosol formation (Bowen and others, 2008), and a ~20 meter package of massive bedded gypsum present at the type locality of Cove Mountain. The uppermost Musinia Peak Member is identified based on the reoccurrence of fossiliferous, calcitic micrites. On the Wasatch Plateau, the Musinia Peak Member is only preserved both in significant thickness and in stratigraphic continuity at its type locality. Elsewhere, it occurs only as thin, isolated remnants of gastropod-bearing limestone lying on the Cove Mountain Member, or where localized faulting has obscured stratigraphic relations.

Significant lacustrine deposits also exist within the Axhandle piggyback basin to the west, although deposition remained primarily fluvial through most of the early Flagstaff interval, with the Paleocene-Eocene boundary in this basin occurring in the

North Horn Formation (Bowen and Bowen, 2008). An interval of lacustrine carbonate interfingering with the North Horn Formation below the Paleocene-Eocene boundary is recognized as the Wales Tongue of the Flagstaff Formation (Talling and others, 1995), and several tens of meters of thickly bedded, fossiliferous lacustrine carbonate overlie the North Horn Formation. Previous work has suggested that these carbonates were deposited during highstand intervals of the Ferron Mountain and Musinia Peak Members, when the lake was able to overtop the Sanpete Valley Antiform separating the Axhandle and Flagstaff basins (Stanley and Collinson, 1979; Lawton and others, 1993; Talling and others, 1995). Most sediments overlying the Flagstaff Formation have been removed by subsequent erosion, but where they are preserved the Flagstaff Formation is typically separated from the lacustrine Green River Formation deposits by fluvial sandstone (Marcantel and Weiss, 1968). Though these fluvial deposits have historically been described as the Colton Formation, recent petrographic study shows these sands have a highly distinct provenance from the main body of the Colton Formation in the Uinta Basin (Dickinson and others, 2012), with those authors suggesting the Colton in this area merits reclassification.

Lake Flagstaff was only part of the first pulse of widespread lake systems existing along the Sevier thrust front during the early Paleogene. The Green River Lake system occupied the Uinta, Piceance Creek and Greater Green River Basins of northern Utah, northwestern Colorado, and southwestern Wyoming. As discussed above, in the Uinta Basin the Flagstaff is the basal member of the Green River Formation overlying the North Horn Formation, and in this initial stage the lake has been considered continuous with Lake Flagstaff during the Ferron Mountain Member highstand (Stanley and Collinson, 1979; Davis and others, 2008). Fluvial sands of the Colton Formation interfinger northward with the Flagstaff Member and the Main Body of the Green River Formation, deposits which are now regarded as the delta of a fluvial system integrating drainages as far south as the Mojave region and Mogollon Highlands (Davis and others, 2010; Dickinson and others, 2012). South of the Flagstaff Basin, Lake Claron occupied a similar basin situated between the Sevier front and a number of Laramide structures (Goldstrand, 1994). The Claron Formation has historically been divided into two informal members: the heavily pedogenically modified Red Member, and the more lacustrine White Member (Davis and others, 2009b). While the lake appears to have transgressed east-northeast over its history (Goldstrand, 1992; Taylor, 1993; Goldstrand, 1994), available age constraints are poor, particularly for the base of the Claron, where a depositional hiatus of unknown length separates it from the underlying Pine Hollow Formation.

METHODS

Stratigraphically resolved carbonate hand samples were obtained from three locations; the summit area of Musinia Peak (also called Mary's Nipple; 39.0364° N, 111.5777° W), the northern ridge extending from White Mountain (39.0603° N, 111.5208° W), and the north wall of South Maple Canyon (39.3345 °N, 111.7197 °W) (fig. 1B). Hand samples were tied to sections measured using a standard 1.5 m staff and field logs including lithologic descriptions. Petrographic descriptions were made from slabbed and polished faces of hand samples, as well as representative thin sections stained with alizarin red (Dickson, 1965).

Carbonate microsamples (100 μg) for stable carbon and oxygen isotope analysis were drilled from slabbed and polished hand samples using a dental drill under a binocular microscope. Generally, multiple micrite microsamples were analyzed from each hand sample, including different textures and colors where present. Secondary carbonate spar was also sampled where present to determine offsets in isotopic composition from co-occurring micrites, which would indicate whether sparry carbonate precipitated from different fluids and/or at different temperatures from co-

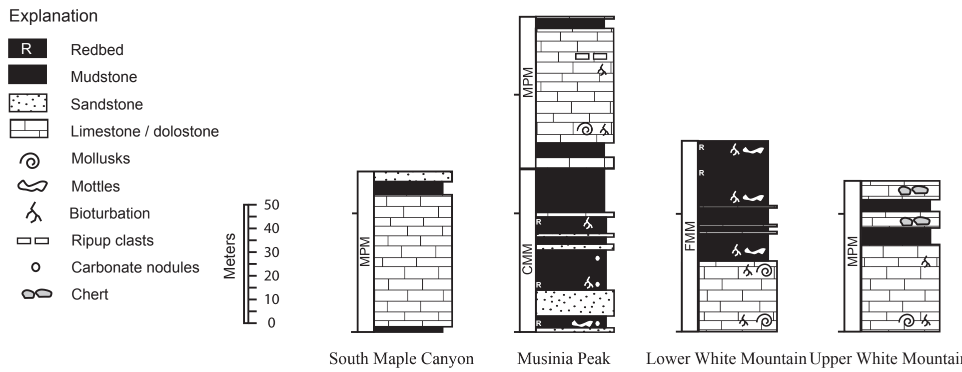

Fig. 3. Lithostratigraphy of measured sections.

existing micrites. Where they were observed, gastropod shell fragments showed evidence of pervasive recrystallization, in most cases being completely replaced by coarse calcite spar. Ostracode shells, while not displaying textural evidence for recrystallization, were generally too small and well-cemented in micrite to be sampled separately. For these reasons, shells were avoided when sampling for isotopic analyses. Samples were reacted with 104 percent orthophosphoric acid for 1 hour at 70 °C, with the resultant CO_2 being chromatographically purified using a Gasbench II and analyzed on a Delta V mass spectrometer (both ThermoFinnigan) at the Purdue Stable Isotope Lab, West Lafayette, Indiana. Ratios were standardized to the Vienna Pee Dee Belemnite reference scale using an in-house Carrara Marble reference calcite, with an analytical precision of 0.1 permil for both $\delta^{13}C$ and $\delta^{18}O$ (1σ) based on replicate analyses of an independent check standard.

Determination of major mineralogic components was performed on slabbed and polished hand samples using a Fieldspec 3 Analytical Spectral Device (PANalytical) with a visible/near-infrared-shortwave infrared spectral range from 350 to 2500 nm and wavelength sampling ranging from 3 to 10 nm. The contact probe is self-illuminated and has a 1 cm diameter spot size. Analysis of the spectral data focused on the shortwave infrared carbonate absorption feature in the 2300 to 2350 nm band (Gaffey, 1986) in order to distinguish calcite (~2338nm) from dolomite (~2320nm).

RESULTS

Lithologic Descriptions

The lower 50 m of the Musinia Peak section consists predominantly of clastic lithologies (fig. 3). A sequence of 10 to 30 cm thick, cross-bedded sandstones with gravel lags occurs from 7 to 18 m, and a thick package of clastic mudstone containing redbeds and carbonate nodules occurs from 18 to 35 m. Pedogenic modification decreases up-section, while sand content increases, with lenticular sand bodies common throughout the mudstone interval. The first limestone bed in the section occurs at the 50 m level, directly above a sequence of two 50 cm thick, reddish pedogenically modified clastic mudstones. This limestone is a single, 1 m thick bed, with abundant gastropod fragments and vugs filled with coarse calcite spar. Deposition changes abruptly from clastic mudstone to bedded limestone and dolostone at the 71 m level, and bedded carbonates are the dominant lithology to the top of the section. Carbonate beds vary in thickness from 10 cm to almost 2 m, with the thickest beds occurring around the 100 m level. Carbonate beds are well bioturbated and lack internal laminations, with 1 to 5 cm mudstones separating most limestone beds. The base of the Musinia Peak Member was defined by Stanley and Collinson (1979) as the lowermost calcitic, fossiliferous limestone beds overlying the dolomitic, afossiliferous Cove Mountain Member. In both of our upper Flagstaff sections, dolomite remains a major carbonate component throughout the upper carbonate package and though present, fossils are considerably rarer than in the lower Ferron Mountain Member, with many of the massive limestone and dolostone beds being largely afossiliferous. We place the contact between the Cove Mountain Member and Musinia Peak Member at the transition between mudstone and bedded, fossiliferous carbonate deposition at the 71 m level of the Musinia Peak section, for a total thickness of the Musinia Peak Member at this site of ~58 m, slightly thicker than the 37 m described by Stanley and Collinson (1979).

The lowermost 30 m of the White Mountain section is comprised of 30 to 50 cm thick, gastropod-bearing bioturbated limestone beds typical of the Ferron Mountain Member. Above 30 m, carbonate beds are sparse and much thinner, with the dominant lithology being gray-brown, bioturbated mudstone. Above 60 m, mudstones show evidence of pedogenic modification, including well-developed redbeds. A fault zone of undetermined thickness, largely covered, separates our two sections at this locality. Visual observation of the unfaulted northern face of the ridge shows that the limestone-mudstone-limestone succession typical of the Flagstaff Formation on the Wasatch Plateau is present at White Mountain, with roughly the same thicknesses observed at Musinia Peak; however, the fault obscures exact stratigraphic relationships within the Cove Mountain Member; for this reason we divided the White Mountain section into Lower and Upper sections. Above the fault zone, the upper section is dominated by carbonate beds of 20 to 80 cm thickness. Gastropod fossils are present in some beds, but rare. The uppermost 20 m of section also contains abundant chert, occurring both as nodules within carbonate beds as well as thin (5-10 cm), discrete beds where it is the dominant component. The fault zone present between our Lower and Upper White Mountain sections renders exact correlation between these sections and Musinia Peak difficult, however given the evidence of limited vertical offset on this fault suggested by the field relationships and previous mapping (Baughman, ms, 1959), total preservation of the Musinia Peak Member at both sites is similar.

The South Maple Canyon section is dominated by massive, silty carbonate beds, with thin 1 to 5 cm mudstones separating carbonate beds of 1 to 2 m thickness. Sand content increases up-section, with carbonate deposition replaced by sandstone and clastic mudstone above the top of the section at 64 m.

Petrography

The primary constituent of upper Flagstaff limestone is micritic carbonate, and most samples are almost entirely micrite. Siliciclastic content is low in bedded carbonates from Musinia Peak and Upper White Mountain; samples acid-digested to remove carbonate contain only 3 to 5 percent insoluble residue by weight ($n = 10$), and no correlation is found between clay content and carbonate mineralogy. Carbonates at all sites are variably fossiliferous. In the Lower White Mountain section gastropod fragments are a ubiquitous component of the lower beds, with many largely intact shells filled with blocky calcite spar. Shell fragments stain a deep red color, indicating a calcite or aragonite composition, whether they are preserved in micrite of calcitic or dolomitic composition. The Upper White Mountain, Musinia Peak and South Maple Canyon sections are considerably less fossiliferous. The most commonly occurring fossils at these sites are ostracode shells, while gastropod shells are present but rare.

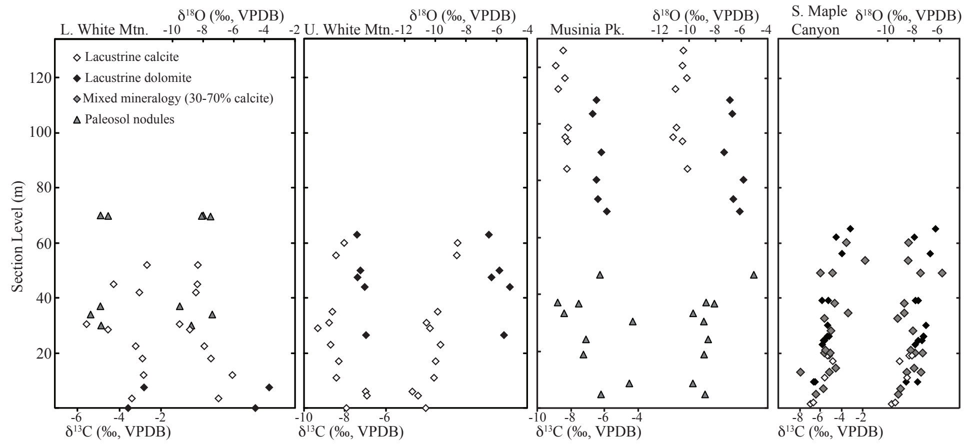

Fig. 4. Isotope stratigraphy of measured sections.

IR Spectroscopy

All samples from the Musinia Peak and Upper White Mountain sections exhibit a single absorption minimum in the 2300 to 2350 nm band, centered near 2320 or 2338 nm, indicating either a purely dolomitic or calcitic composition respectively. The changes in mineralogy exhibit coherent, through-section changes, with multiple alternating packages of calcitic or dolomitic carbonate beds occurring through each section (fig. 4). In contrast, samples from the Lower White Mountain section are almost entirely calcitic, with the exception of two thin dolomite beds. The South Maple Canyon section exhibits alternating packages of carbonate mineralogy, similar to the Musinia Peak and Upper White Mountain sections, however most samples display carbonate absorption features intermediate between the dolomite and calcite end-members, indicating mixed mineralogies. Comparison of IR spectroscopy of lower Flagstaff samples with X-ray diffraction (XRD) measurements from the sections of Bowen and others (2008) indicate that spectroscopy accurately records the composition of single-mineral samples and confirms that intermediate spectral absorption features are observed for mixed-mineralogy samples (as identified by XRD). The correlation between absorption wavelength and calcite/dolomite ratio from XRD is only moderate ($r^2 = .6$) for mixed-mineralogy samples. This is likely at least partly due to the different sampling of the analytical methods. XRD analyses used 2 mg sub-samples obtained by drilling targeted areas of hand samples, whereas the IR spectrometer integrates reflectance data evenly across a ~1 cm^2 area on the surface of a polished hand sample.

Stable Isotope Ratios

We measured carbon and oxygen isotope ratios from 183 carbonate microsamples within the Flagstaff Formation. Values of $\delta^{13}C$ for the entire sample set range from −10.7 to −1.8 permil, and $\delta^{18}O$ from −16.8 to −3.6 permil relative to V-PDB (fig. 5, table A1). The lowest values of $\delta^{13}C$ and $\delta^{18}O$ occur in sparry cements (mean $\delta^{13}C = -8.6$ ‰, $\sigma = 0.6$ ‰; mean $\delta^{18}O = -12.4$, $\sigma = 2.1$ ‰). Lacustrine micrite and nodular carbonate from pedogenically modified clastic mudstones have similar values for both $\delta^{13}C$ (mean = −6.0 ‰, $\sigma = 1.9$ ‰) and $\delta^{18}O$ (mean = −8.4 ‰, $\sigma = 1.8$ ‰).

Strong covariance exists between $\delta^{13}C$ and $\delta^{18}O$ of micrite in both the Musinia Peak ($r^2 = 0.92$, $p < 0.001$, t-test) and Upper White Mountain sections ($r^2 = 0.90$, $p <$

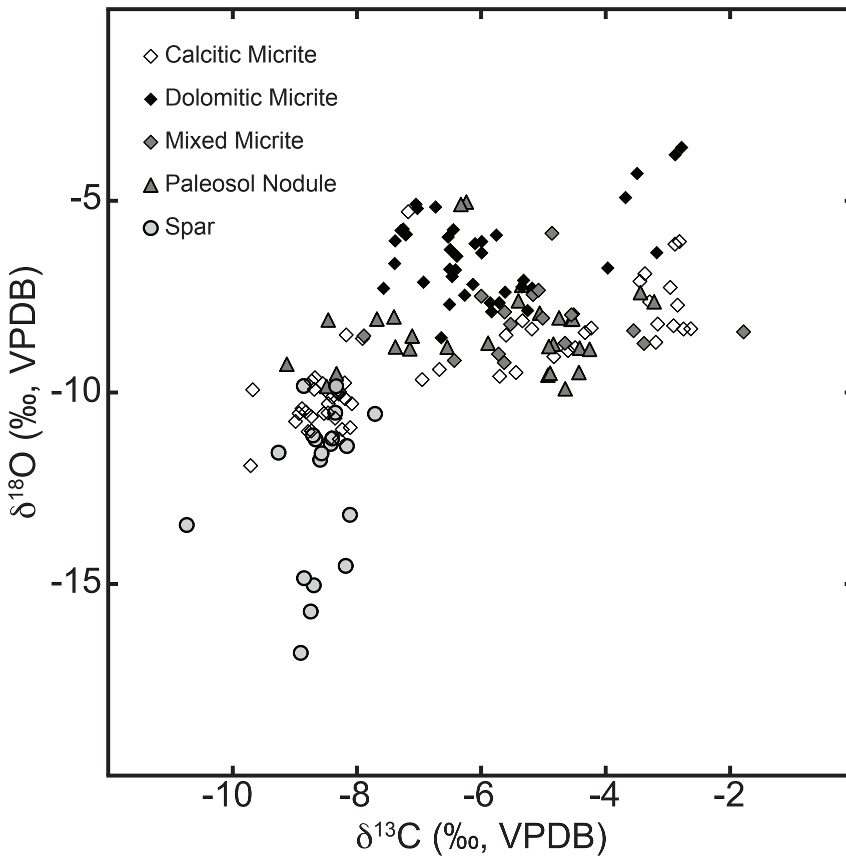

Fig. 5. Stable carbon and oxygen isotope ratios for all analyzed carbonate samples.

0.001, t-test) (fig. 6). This covariance is closely tied to mineralogy, as for both sections carbon and oxygen isotope ratios correlate strongly with the wavelength of the carbonate absorption feature ($\delta^{13}C$ $r^2 = 0.74$, $p < 0.001$; $\delta^{18}O$ $r^2 = 0.81$, $p < 0.001$, t-test) (fig. 7). As with the mineralogical changes, variation in isotope ratios oscillates repeatedly up-section. Carbonates at these two sites are thus divided into two populations, with dolomite having higher $\delta^{13}C$ (mean = −6.7 ‰, σ = 0.5 ‰) and $\delta^{18}O$ values (mean = −6.3 ‰, σ = 0.7 ‰) than calcite (mean $\delta^{13}C$ = −8.5 ‰, σ = 0.5 ‰; mean $\delta^{18}O$ = −10.2 ‰, σ = 1.1 ‰). Covariance between $\delta^{13}C$ and $\delta^{18}O$ is not significant for the South Maple Canyon section ($r^2 = 0.08$, $p = 0.09$, t-test), and the relationship between mineralogy and isotope ratios is weak for oxygen ($r^2 = 0.28$, $p = 0.01$, t-test), and not significant for carbon ($r^2 = 0.005$, $p = 0.68$, t-test).

DISCUSSION

Sedimentologic Interpretation

The resumption of carbonate deposition in the Musinia Peak Member has been interpreted to result from the re-expansion of Lake Flagstaff following the low-stand interval of the Cove Mountain Member (Stanley and Collinson, 1979). The pervasive pedogenic modification of mudstone within the Cove Mountain Member at these two sites, as well as common channelized sandstones within the mudstone-dominated

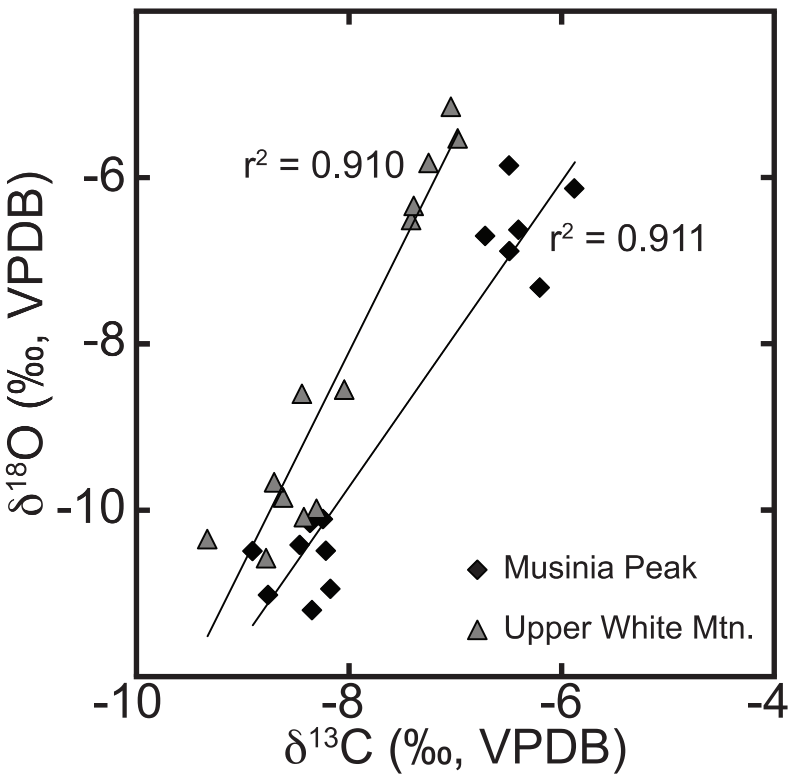

Fig. 6. Isotope ratios of Musinia Pk. Mbr. carbonates.

interval at Musinia Peak, suggests that these sites were subaerially exposed for significant intervals during Cove Mountain Member deposition. Within the Musinia Peak Member at these sites, indicators of subaerial exposure are absent, suggesting that Lake Flagstaff continuously occupied this portion of the basin during Musinia Peak Member deposition. The low proportions of siliciclastic sediment within carbonate beds, as well as their non-laminated, bioturbated character suggest that the lake was shallow and well oxygenated. Resumption of calcite deposition in the lacustrine sediments indicates significant changes in lake water chemistry occurred between the Cove Mountain Member and Musinia Peak Member, although the oscillation between calcite and dolomite beds suggests these changes were dissimilar to the secular transition from calcite to dolomite that occurred between the Ferron Mountain Member and Cove Mountain Member. Similarly, the renewed occurrence of freshwater gastropods in the Musinia Peak Member indicates a return to less saline conditions, but the sparseness of fossils compared to their abundance in the Ferron Mountain Member suggests either that return to less saline conditions was not complete, or that the depositional and burial environment were not conducive to fossil preservation. The general absence of gastropod shell molds or otherwise altered mollusk fossils suggests that they were originally rare rather than being destroyed during burial.

Mineralogic Controls on Carbonate Isotopes

The lacustrine micrites and dolomicrites of the Musinia Peak Member in the Musinia Peak and Upper White Mountain sections exhibit a bimodal distribution that

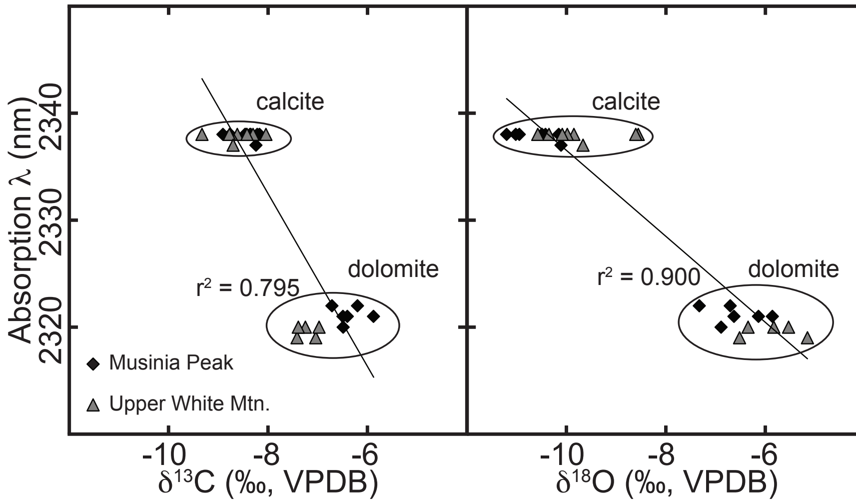

Fig. 7. Relationship between carbonate isotope ratios and mineralogy, determined by IR spectroscopy.

is strongly covariant between carbon and oxygen isotopic ratios (r^2 = ~0.9) (fig. 6). In other lake systems, isotope covariance has been regarded as evidence of hydrologic closure, (Talbot, 1990), due to the effects of atmospheric equilibration and evaporative enrichment on carbon and oxygen isotopes, respectively. Confounding this straightforward interpretation, Musinia Peak Member carbonates occur as two distinct mineralogies, which have differing isotopic fractionation factors for oxygen, and possibly carbon (Sheppard and Schwarcz, 1970; Vasconcelos and others, 2005). These differing fractionations must be considered before any environmental interpretation can be made using the isotope data. Mineralogic variation also occurs in paleosol carbonates associated with Lake Flagstaff (VanDeVelde and others, 2013), but we focus here on the isotopic interpretation of bedded micrite of clear lacustrine origin.

For oxygen, there are two separate fractionation effects to consider. The CO_2 gas obtained by phosphoric acid digestion experiences oxygen isotope fractionation relative to the original carbonate, an effect which is dependent on mineralogy, reaction temperature and analytical technique. For the sealed vessel technique at 70 °C, CO_2 derived from dolomite will be enriched in ^{18}O by 1.2 permil relative to that from calcite (Rosenbaum and Sheppard, 1986; Swart and others, 1991). Oxygen is also fractionated differently between dolomite and calcite at the time of formation. Over a range of plausible precipitation temperatures of 15 to 30 °C, dolomite will be enriched in ^{18}O by ~2.8 permil relative to co-precipitating calcite (Kim and O'Neil, 1997; Vasconcelos and others, 2005). As a result of these two contributing effects, oxygen isotope ratios measured from dolomite will be ~4 permil higher than ratios from calcite formed in equilibrium with the same fluid.

Carbon isotope ratios are not subject to a phosphoric acid fractionation, but differences in carbon isotope fractionation effects for the precipitation of dolomite and calcite are less well constrained than those for oxygen. A $\Delta^{13}C_{dol\text{-}cal}$ value of 2.4 permil at 20 °C was obtained as an extrapolation from coexisting metamorphic dolomites and calcites formed at temperatures >100 °C (Sheppard and Schwarcz, 1970). Experimental results have suggested a much larger equilibrium offset (Chacko and others, 2001), while coexisting lacustrine dolomite and calcite from the Miocene Ebro Basin suggest little or no differences in carbon isotope fractionation (Arenas and others, 1997).

Bowen and others (2008) demonstrated that in the lower Flagstaff Formation, oxygen isotope ratios of mixed-mineralogy carbonates define a mixing relationship with an offset close to that for dolomite and calcite forming in equilibrium at 15 °C. While this relationship was strong and consistent for oxygen isotopes and mineralogy in their data, it was weak and inconsistent for carbon isotopes. In our new data from the Musinia Peak Member, the ~2 permil offset in $\delta^{13}C$ between dolomite and calcite is consistent with the extrapolated fractionation factor of Sheppard and Schwarcz (1970), but given the uncertainty in our knowledge of carbon isotope fractionation for dolomite our subsequent discussion assumes $\delta^{13}C$ values which are uncorrected for any potential mineralogic effects.

When adjusted for the differing oxygen isotope fractionation factors, $\delta^{18}O$ values of the lacustrine calcites and dolomites at Musinia Peak and Upper White Mountain are each statistically indistinguishable (single-factor ANOVA, $p = 0.7$), and are stable throughout both sections (average: −10.2 ‰, $\sigma = 0.7$ ‰), with differences in $\delta^{18}O$ between sites being minimal (fig. 7A). Measured $\delta^{13}C$ values are not distinct between the two sites for calcite ($p = 0.4$), but are distinct for dolomite ($p << 0.001$), with Musinia Peak dolomites enriched in ^{13}C by ~1 permil relative to Upper White Mountain. This suggests that while lake water oxygen composition remained relatively constant throughout Musinia Peak Member deposition, during phases of dolomite precipitation the lake's biological systems developed sufficient spatial heterogeneity to impact the composition of the local dissolved inorganic carbon pool, which has potential implications for the selection of a preferred dolomitization model for Flagstaff Formation carbonates, as discussed below.

Flagstaff Dolomite Formation

The processes leading to dolomite formation remain controversial, given that dolomite is found far more commonly in ancient sediments than modern environments. In the marine realm, dolomite is typically considered a product of burial diagenesis. A multitude of recent dolomite-forming lakes have been documented (Last, 1990; Rosen and Coshell, 1992; Vasconcelos and McKenzie, 1997; Sinha and others, 2006; Wacey and others, 2007; Meister and others, 2011; Last and others, 2012), suggesting that in lacustrine environments dolomite is frequently primary in origin. This recognition has introduced its own complication, in that these dolomitic lakes occur in a wide variety of environments, with highly variable chemistries (Armenteros, 2010). Given this variety, several models have been proposed to explain the formation of dolomite in lacustrine environments, which have varying applicability to the Flagstaff Formation generally, and the Musinia Peak Member specifically.

Many modern dolomite-producing lakes occur in arid regions, and are saline to hypersaline (Last, 1990). These saline and playa lakes are characterized by high carbonate alkalinity (>5000 mg/L) and Mg/Ca ratios >10. Dolomite precipitation is also frequently associated with evaporative concentration, as suggested by higher $\delta^{18}O$ in lacustrine dolomite, relative to equilibrium values expected based on co-occurring calcite (Arenas and others, 1997). Despite this, in other cases dolomite has been documented to precipitate with little to no enrichment in ^{18}O (García del Cura and others, 2001). Bowen and others (2008) demonstrated that within the Ferron and Cove Mountain Members, higher $\delta^{18}O$ in dolomite generally does not exceed that expected for calcite-dolomite equilibrium at surface temperatures. Differences in $\delta^{18}O$ between dolomite and calcite from our new sites in the Musinia Peak Member are similarly no greater than those produced by equilibrium fractionation at surface temperatures. Carbonates in the Musinia Peak Member also exhibit relatively low and stable $\delta^{18}O$, relative both to the rest of the Flagstaff Formation (fig. 8) and to modern closed lakes (Talbot, 1990), suggesting the waters from which they precipitated were not subject to extreme evaporative concentration. Finally, even moderately evaporative

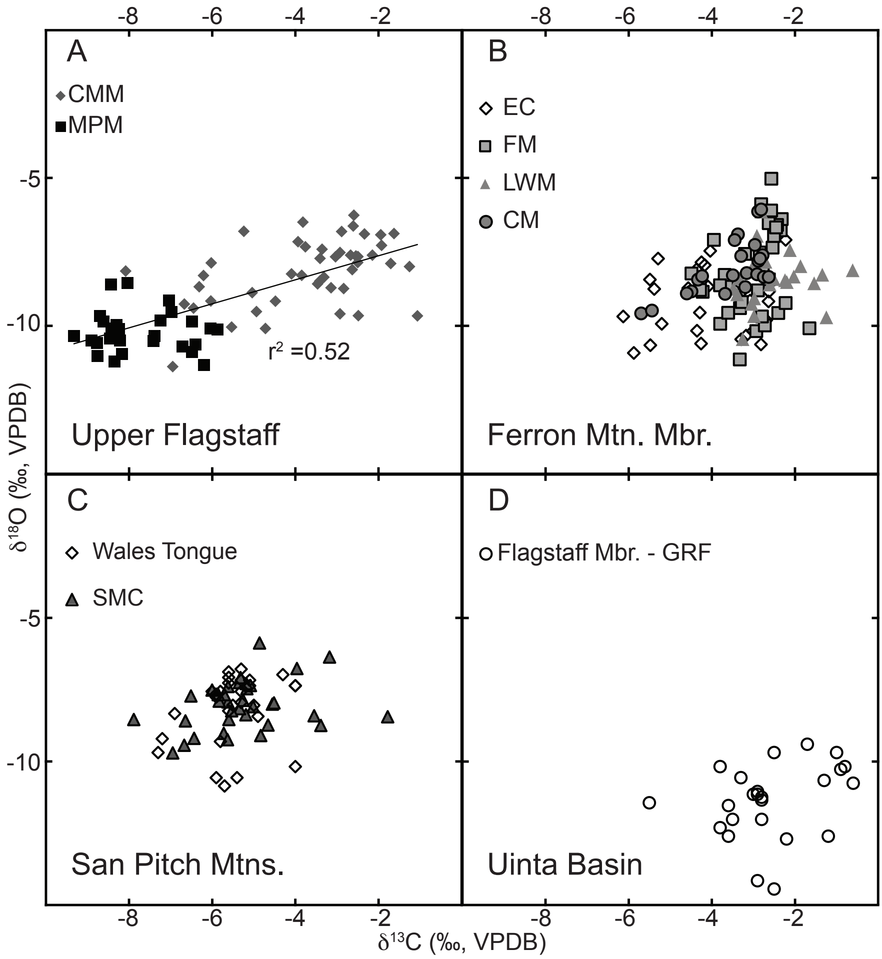

Fig. 8. Carbon and oxygen isotope cross-plots for Flagstaff lacustrine micrites. (A) Cove Mtn Mbr (CMM) and Musinia Pk Mbr (MPM) samples. (B) Ferron Mtn Mbr, grouped by location: EC – Ephraim Canyon; FM – Ferron Mountain; LWM – Lower White Mountain; CM – Cove Mountain (C) Axhandle Basin samples from the Wales Tongue and South Maple Canyon (SMC) in the San Pitch Mtns. (D) Flagstaff Mbr. of the Green River Fm. (GRF) in the Uinta Basin. Ferron Mtn Mbr and Cove Mtn Mbr data are from Bowen and others (2008), Wales Tongue data from Chamberlain and others (2012), and Uinta Basin data from Davis and others (2008).

lake waters are typically supersaturated with respect to dolomite (Deocampo, 2010), and experiments have demonstrated that dolomite fails to precipitate at reasonable surface temperatures from even extremely supersaturated waters (Land, 1998). We therefore regard evaporative concentration in Lake Flagstaff as insufficient on its own to induce precipitation of dolomite.

Trace elements in lacustrine carbonates are a potential independent proxy for the degree of evaporative concentration in lake waters. Magnesium-calcium and strontium-calcium ratios were used by Davis and others (2008, 2009a, 2009b) to assess the effects of evaporation in the Green River lake system, as these elements are typically conserved in evaporatively concentrated lakes waters before being incorporated into authigenic carbonates (Müller and others, 1972). Trace elements are somewhat problematic as indicators of evaporation, however. While carbonates take up Mg and Sr in proportion to their concentrations in lake water, their accommodation is also dependent on mineralogy (Cohen, 2003). Microbial mediation of dolomite precipitation has also been shown to obscure the relationship of Mg/Ca in carbonates to evaporative concentration in lake water (Vasconcelos and McKenzie, 1997). Finally, a recent study of strontium in Flagstaff carbonates showed relatively low Sr/Ca ratios throughout, even in the evaporite-containing sediments of the Cove Mountain Member (Gierlowski-Kordesch and others, 2008). Future study of trace elements in Flagstaff carbonates may prove useful in developing an improved understanding of the relationship between lake water evaporation, trace element concentration, and carbonate mineralogy.

Dolomitic lake sediments are frequently found in association with magnesium-rich authigenic clay minerals, such as sepiolite, palygorskite, and trioctahedral smectites (Hay and Kyser, 2001; Deocampo, 2004; Bustillo and Alonso-Zarza, 2007). Bristow and others (2012) proposed that the formation of these magnesium-rich clays could serve as a control on the mineralogy of co-occurring carbonate, as lake water magnesium is variably partitioned between clay minerals and dolomite. They demonstrate that Mg-rich authigenic smectite content in oil shale of the Green River Formation in the Uinta Basin varies cyclically between 9 and 39 percent of the sediment, and that authigenic smectite content correlates inversely with dolomite/calcite ratios. Further, they suggest that this cyclical variation in dolomite content can lead to covariation between carbon and oxygen isotope ratios in mixed-mineralogy carbonate sediments, due to isotopic offsets in dolomite and calcite precipitating from the same fluid. Given the previously discussed uncertainty in the carbon isotope fractionation for dolomite, we feel that this mineralogic effect on the $\delta^{13}C$ of carbonate should be regarded cautiously, though the bimodal distribution of Musinia Peak Member carbonates in both $\delta^{13}C$ and $\delta^{18}O$ is supportive of this interpretation (figs. 6 and 7). We consider clay formation to be an unlikely controlling factor for the mineralogy of Musinia Peak Member carbonates however, due to the low amounts of non-carbonate sediments present in the limestone and dolostone beds analyzed (<5 % by weight), and the absence of correlation between carbonate mineralogy and clay content.

Microbes have been increasingly regarded as playing a critical role in the precipitation of primary dolomite (Deocampo, 2010). The first microbes that were identified to mediate dolomite precipitation were sulfate-reducing bacteria from a modern coastal lagoon (Vasconcelos and McKenzie, 1997; Vasconcelos and others, 2005). These microbes are anaerobes, but subsequent studies have demonstrated that dolomite precipitation can be induced by aerobic, non-sulfate reducing bacteria (Sánchez-Román and others, 2009), and dolomite inferred to be microbially mediated has been observed in oxic sediments and waters of alkaline lakes (Meister and others, 2011), as well as caves (Alonso-Zarza and Martín-Pérez, 2008; Jones, 2010). Similar microbial activity has been suggested as the source of dolomite within Pliocene lake sediments of central Spain (García del Cura and others, 2001). The relatively low $\delta^{13}C$ of Musinia Peak Member carbonates suggest a strong contribution of organic-derived carbon to the dissolved inorganic carbon pool, though isotopic studies of some lacustrine dolomites indicate that $\delta^{13}C$ is not a conclusive indicator for the presence of microbial mediation, as microbially influenced dolomite has been shown to sometimes form with $\delta^{13}C$ in near-equilibrium with atmospheric CO_2 (Wacey and others, 2007). Meister and others (2011) suggest that variations in dolomite $\delta^{13}C$ may instead be related to precipitation in the water column versus shallow sediments.

Our data allow for the evaluation of the various dolomitization models in regards to Flagstaff Formation carbonates. Dolomite is unlikely to have formed as a result of persistent chemical stratification of lake water (Desborough, 1978), as the types of freshwater gastropods present (La Rocque, 1960) suggest the lake was too shallow for development of permanent stratification. The playa-lake model, in which carbonates are dolomitized through evaporative pumping on subaerial mudflats (Eugster and Surdam, 1973) is also unlikely for the Musinia Peak Member, due to the absence of indicators for either widespread subaerial exposure or transportation of carbonate such as ooids, and the similar $\delta^{18}O$ values for calcite and dolomite after accounting for equilibrium fractionation. The low clay content in bedded calcite and dolomite suggests formation of Mg-rich smectites did not control carbonate mineralogy, as Bristow and others (2012) suggest for the Green River Formation in the Uinta Basin. Dolomite precipitation in the Musinia Peak Member was probably most similar to modern alkaline lakes (Warren, 1990; Meister and others, 2011; Last and others, 2012) where microbial activity induces the primary precipitation of dolomite, either from the water column or in shallow subsurface sediments. The spatial variation in dolomite $\delta^{13}C$ suggests localized effects on the composition of the DIC pool within the sediment-water interface, perhaps due to varying contributions of respiration-derived carbon. The absence of similar spatial variance in $\delta^{18}O$ indicates that the lake remained a continuous water body rather than separating into individual pools.

Environmental Controls on Flagstaff Formation Isotopes

When adjusted for differences in equilibrium fractionation of oxygen isotopes between dolomite and calcite, the bedded carbonates of the Musinia Peak Member do not display covariance between carbon and oxygen isotope ratios ($r^2 = 0.09$) and exhibit remarkable stability in $\delta^{18}O$ over 60 m of section, which likely represents more than 1 million years of deposition, based on regional sedimentation rates (Talling and others, 1994; Bowen and others, 2008). Similarly, isotopic covariance is weak for the Ferron Mountain ($r^2 = 0.06$) and Cove Mountain Members ($r^2 = 0.17$) individually. When considered together, Cove Mountain and Musinia Peak Member carbonates define a single, moderately strong covariant trend ($r^2 = 0.52$, $p << 0.001$, t-test) (fig. 8A), suggesting a relatively isotopically constant water source to the lake basin throughout the time of deposition of these members. Most Ferron Mountain Member carbonates fall within the range of isotope ratios defined by the Cove Mountain–Musinia Peak Member trend, suggesting that the source-water isotopic stability may have spanned the deposition of all three members. Some dolomicrites at the Ferron Mountain site, however, exhibit elevated $\delta^{18}O$ values which place them outside the primary covariant trend (fig. 8B). These ^{18}O-enriched dolomites were noted by Bowen and others (2008), who suggested that they were the product of extreme oscillations in water chemistry at this basin-center site during an interval in which carbonate was not deposited at the Ephraim Canyon or Cove Mountain sites. Dolomicrites at the Ferron Mountain site assigned to the Cove Mountain Member do not exhibit similarly elevated $\delta^{18}O$, which supports the hypothesis that the enrichment was a localized, transient effect, and not related to basin-wide changes in source-water isotopic composition.

Global temperatures increased by 4 to 5 °C during the interval of Flagstaff Formation deposition (Zachos and others, 2001), and paleofloral records from southern Wyoming suggest large regional temperature increases as well as significant decreases in precipitation (Wilf, 2000). Flagstaff sediments reflect similarly major changes in lake hydrology, shifting from a calcite lake with abundant freshwater gastropods, to a saline playa precipitating dolomite and evaporates (Stanley and Collinson, 1979; Bowen and others, 2008), and then back to freshwater calcite. Despite these climate shifts, variations in lake $\delta^{18}O$ after excluding mineralogic effects do not exceed ~4 permil. Covariance of isotope ratios and the lack of plausible outflows for Lake Flagstaff (see below) suggest that this feature cannot be explained solely as the product of hydrologically open conditions. Comparison with modern data suggest this stability may be the product of buffering by large water volume (Leng and Marshall, 2004), though most large closed lakes also have more positive isotope ratios.

Closed-basin lakes can maintain a uniform covariant trend over millions of years, despite changes in climate, as long as major changes in basin morphology or drainage configuration do not occur (Talbot, 1990). Strontium isotope ratios are stable throughout the Flagstaff Formation, maintaining the $^{87}Sr/^{86}Sr$ composition of hypothesized source areas along the Sevier thrust front (Gierlowski-Kordesch and others, 2008) and suggesting limited reorganization of the Lake Flagstaff drainage basin during the time of Flagstaff Formation deposition. Taken together with the stable, uniform covariant trend in the carbon and oxygen isotope data, this suggests that through-section changes in Flagstaff carbonate $\delta^{13}C$ and $\delta^{18}O$ values most likely reflected basin-scale changes in the water balance of Lake Flagstaff over time.

The presence of a distinctive isotopic signature in Flagstaff Formation carbonates on the Wasatch Plateau allows for potential correlation with deposits mapped as Flagstaff elsewhere. The carbon and oxygen isotope composition of micrites from our South Maple Canyon site, which have previously been correlated with the Musinia Peak Member on the basis of mollusk biostratigraphy (La Rocque, 1960; Marcantel and Weiss, 1968; Stanley and Collinson, 1979) fall well within the covariant trend (fig. 8C). South Maple Canyon samples have higher $\delta^{13}C$ and $\delta^{18}O$ values than are found in Musinia Peak Member carbonates at Musinia Peak and White Mountain, as well as a greater range of variability. However, the mixed mineralogy of South Maple Canyon carbonates supports the previously proposed correlation with the Musinia Peak Member rather than the Cove Mountain Member, which is entirely dolomitic (Stanley and Collinson, 1979; Bowen and others, 2008). Given the limited remaining exposure of the Musinia Peak Member on the Wasatch Plateau, we consider it likely that the two sites we have sampled there do not capture the full variability of Musinia Peak Member geochemistry, more of which may be documented at South Maple Canyon.

The correlation of lacustrine sediments deposited within the Axhandle piggyback basin with the greater Flagstaff Basin is particularly critical in understanding the evolution of Lake Flagstaff, as several existing age constraints are from deposits in the Axhandle Basin, including magnetostratigraphy (Talling and others, 1994; Talling and others, 1995), the negative carbon isotope excursion associated with the Paleocene-Eocene boundary (Bowen and Bowen, 2008), and one of two vertebrate fossils identified from Flagstaff deposits, a miacid jawbone of Wasatchian age collected from South Maple Canyon (Rich and Collinson, 1973). Axhandle Basin carbonates from the Wales Tongue of the Flagstaff Formation (Chamberlain and others, 2012), previously correlated to the Ferron Mountain Member (Talling and others, 1994), also fall within the range expressed by the broader Flagstaff Formation covariant trend (fig. 8C). This is consistent with carbonates at both sites in the Axhandle Basin being deposited by a lake that was contiguous with the greater Flagstaff Basin.

The Axhandle Basin constitutes a wedge-top depozone carried atop the Gunnison Thrust duplex (DeCelles and Coogan, 2006), which had ceased motion by middle Paleocene time (Lawton and Trexler, 1991; Lawton and others, 1993; Talling and others, 1994; Talling and others, 1995). Following the end of uplift, continuing sedimentation within both the Axhandle and Flagstaff Basins would have allowed Lake Flagstaff to eventually breach and overtop the San Pete Valley Anticline, first during the interval represented by the Wales Tongue and Ferron Mountain Member, and later during Musinia Peak Member deposition. The similarity in isotope ratios between Axhandle and Flagstaff Basin limestones is consistent with that interpretation. Continuity is further supported by the presence of arkosic sandstones in upper Flagstaff sediments of both the Flagstaff and Axhandle Basins (Stanley and Collinson, 1979), with paleocurrent data indicating a northwesterly dispersal pattern. These arkosic sands are petrographically distinct from those derived from both the Sevier Thrust Belt to the west, and local structural features such as the San Pete Valley Anticline. The pattern of both westerly and easterly sediment dispersal across both structural basins therefore indicates that they eventually merged into a single sedimentary basin. The provenance of these arkosic sands is a particularly intriguing question, given their petrographic similarity to similar sediments of the Colton Formation in the Uinta Basin that were shown to be the product of a California paleoriver, originating in the Mogollon Highlands of California and Arizona (Davis and others, 2010; Dickinson and others, 2012).

Isotope Hydrology of Paleogene Foreland Lakes

Lake Flagstaff was one of a series of large, persistent lakes that occupied basins along the North American Cordillera throughout the Cenozoic, the result of changing drainage patterns produced by Sevier and Laramide tectonic deformation. Large rivers draining the Sevier highlands (DeCelles, 1994) were increasingly impounded by the growth of Laramide uplifts beginning in the late Cretaceous. By the late Paleocene, this deformation had produced a series of internally drained basins along a line trending northeast-southwest from southwestern Wyoming to southern Utah and Nevada, roughly paralleling strike of the Sevier thrust belt (fig. 1A). The lakes that occupied these basins precipitated large volumes of authigenic carbonate sediments, the composition of which integrates the effects of tectonic and climatic changes upon each individual lake.

At various points, many or all of these lakes may have overtopped their basin margins, as climate change and drainage reorganizations impacted their water balance. In the northern basins, intervals of major interconnection have been suggested on the basis of changes in lacustrine carbonate $\delta^{18}O$ and $^{87}Sr/^{86}Sr$ for Lakes Gosiute, Piceance and Uinta (Davis and others, 2008; Davis and others, 2009a). In the Uinta Basin, the Flagstaff Member of the Green River Formation (Fouch, 1976) reflects the proposed late Paleocene expansion of Lake Flagstaff northward into the Uinta Basin which has been correlated to the highstand of the Ferron Mountain Member (Stanley and Collinson, 1979). Micritic carbonates of the Flagstaff Member in the Uinta Basin exhibit distinct carbon and oxygen isotopic ratios from those in the Flagstaff Basin (fig. 8D), indicating that if hydrologic connection did occur between the Flagstaff and Uinta Basins, it was too intermittent to be reflected in the composition of authigenic carbonates. Thus, while sediments mapped as Flagstaff in both basins are potentially time-equivalent, they are unlikely to be products of the same lake. Recent investigations of petrofacies and detrital zircon ages in the overlying Colton Formation (Davis and others, 2010; Dickinson and others, 2012) suggest that these basins had markedly different sediment inputs during this interval, with the authors suggesting that the Colton Formation in the Flagstaff Basin should be redesignated. While additional investigation is needed, we suggest that the existence of separate lacustrine systems in the Flagstaff and Uinta Basins may warrant further revisions in the stratigraphic terminology.

Carbonates from the main body of the Green River Formation in the Uinta Basin (fig. 9A) (Davis and others, 2008), and in the Piceance Creek Basin (fig. 9B) (Davis and others, 2009a) do partially overlap the isotopic trend of the Flagstaff Formation, although the overall patterns are quite different, as the Uinta and Piceance basin carbonates exhibit much greater variability in both carbon and oxygen isotopes. The Uinta and Piceance Creek Basin carbonates which overlap with the Flagstaff Formation in carbon and oxygen isotope composition have differing strontium isotope ratios (Davis and others, 2008) from Flagstaff Formation samples (Gierlowski-Kordesch and others, 2008). This interval of high $^{87}Sr/^{86}Sr$ in the Uinta and Piceance Basins is attributed to a period of connection between these lakes and Lake Gosiute in the Greater Green River Basin to the north, which received inflows draining volcanic fields in northern Wyoming and central Idaho (Carroll and others, 2008).

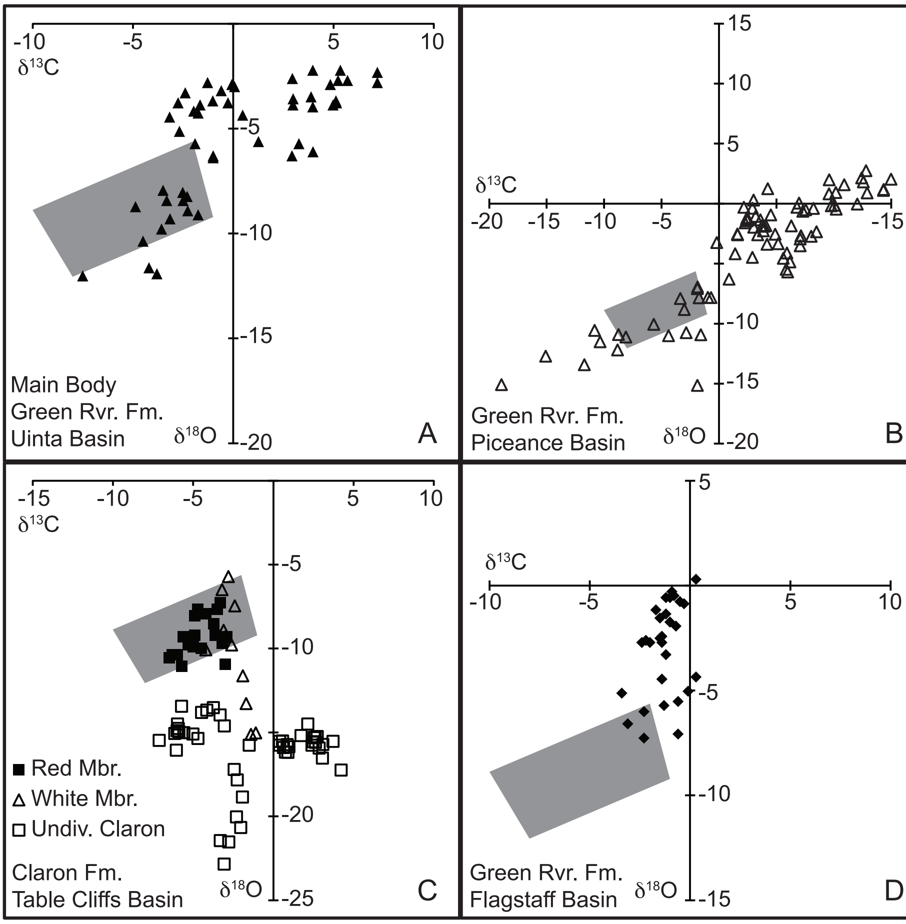

Fig. 9. Carbon and oxygen isotopic cross-plots for lacustrine micrites from Paleogene basins of the central Cordillera. Gray shaded boxes denote the isotope space occupied by the Flagstaff Fm. Non- Flagstaff Fm data from Davis and others (2008, 2009a, 2009b).

An expansion of Lake Flagstaff southward into the vicinity of the Table Cliffs (Claron) Basin was suggested for the Musinia Peak Member high-stand (Stanley and Collinson, 1979), though an early Paleogene east-west trending highland ~100 km south of Musinia Peak was proposed as an alternate southerly limit. The isotopic composition of the Flagstaff Formation overlaps with the Red Member of the Claron Formation on the Markagunt Plateau (fig. 9C) (Davis and others, 2009b), but not with undivided Claron Formation samples from sites to the southwest. Current age constraints suggest that these units do not temporally overlap, but given the overall paucity of age constraints throughout this entire region prior to 50 Ma (Goldstrand, 1994; Davis and others, 2009b), further investigation is needed.

Only thin and spatially patchy remnants of the Musinia Peak Member remain on the Wasatch Plateau, but large thicknesses of lacustrine carbonate to the west have been previously correlated to this member on the basis of mollusk biostratigraphy (La Rocque, 1960), with as much as 900 meters of sediment attributed to the Flagstaff Formation in the Canyon Range (Stanley and Collinson, 1979). While our results do not resolve the correlation of the upper Flagstaff Formation to its westernmost limit, our isotope data do indicate that Flagstaff carbonates west of the San Pitch Valley Anticline can be attributed to the same lake system as those exposed along the Wasatch Plateau. In the Axhandle Basin and elsewhere, the Flagstaff Formation at some sites is directly overlain by lacustrine carbonates assigned to the Green River Formation (Stanley and Collinson, 1979; Davis and others, 2009b). Green River limestones exhibit distinctly higher carbon and oxygen isotope ratios compared to the Flagstaff Formation, with most samples falling well above the Flagstaff covariant trend (fig. 9D). This suggests that when lacustrine deposition resumed in this area, basin morphology and/or climate conditions had changed dramatically.

The isotopic data presented here and in other recent studies present a great challenge to understanding the regional hydrology along the Sevier thrust-front during the early Paleogene. Sedimentologic evidence strongly supports the interpretation that paleoflows were southwest to northeast. The similarity of isotope ratios between the lower Claron Formation and the Flagstaff Formation could be evidence of an upstream-downstream relationship, though if this were the case the expectation would be a covariant trend in Lake Flagstaff that is an extension of Lake Claron's, rather than overlapping. The absence of a relationship between Lakes Flagstaff and Uinta is more puzzling. The relatively weak covariance in Flagstaff Formation carbonates, taken on its own implies a balanced-fill lake, with at least periodic episodes of outflow. If Lake Flagstaff was open, the San Rafael Swell would have impeded eastward outflow, leaving northeastward flow into Lake Uinta as the only apparent possibility, and yet the isotopic data seem to rule out such a connection. This paradox is only highlighted by the apparent existence of a California paleoriver flowing northward from the Mogollon Highlands into Lake Uinta (Davis and others, 2010; Dickinson and others, 2012). Petrographic and detrital zircon data from fluvial sands overlying the Flagstaff Formation in the Axhandle Basin are distinct from the California river delta deposits of the Colton Formation in the Uinta Basin (Dickinson and others, 2012), so that the preferred course of this paleoriver is east of the San Rafael Swell rather than across the Flagstaff Basin. A similar river is suggested to have carved a segment of the western Grand Canyon in the early Paleogene (Karlstrom and others, 2014) before flowing northward into the Claron Basin, however fluvial sandstones from the apparent paleochannel, the Music Mountain Formation of northern Arizona, show that it cannot be the same river that flowed into Lake Uinta (Dickinson and others, 2012). With the Kaibab, Circle Cliffs and Monument uplifts blocking eastward flow just as the San Rafael Swell would, the only apparent exit from the Claron Basin would be northward into the Flagstaff Basin. Further study of fluvial and lacustrine sediments in the region, particularly in the Claron and Flagstaff Basins, will be necessary to resolve this discrepancy.

Elevation of the Lake Flagstaff Catchment

A principal driving force behind the study of the Paleogene foreland lake systems has been to understand the tectonic evolution of the North American Cordillera, in particular its change in elevation over time. The northern Green River lakes (Uinta, Piceance Creek and Gosiute) were at times fed by integrated fluvial systems originating from as far north as central Idaho (Davis and others, 2008; Chetel and others, 2011), and as far south as the Mogollon highlands of southern California and Arizona (Davis and others, 2010; Dickinson and others, 2012). These distant hydrological connections make it difficult to relate the isotopic histories of the northern lakes to more proximal tectonic and climatic events. Our data show that Lake Flagstaff was isolated from the larger Green River lake system through the late Paleocene and early Eocene, and was fed by drainages originating in the Sevier hinterland, making it a more faithful recorder of regional elevations in the central Cordillera.

To investigate the elevation of the Sevier thrust front, we employ a recently developed method referred to as the "δ^{13}C-excess" parameter (Horton and Oze, 2012) to determine the mean hypsometric altitude for Lake Flagstaff's catchment. Briefly, the δ^{13}C-excess proxy uses the covariance between δ^{13}C and δ^{18}O in lake waters and authigenic lacustrine sediments to calculate lake inflow δ^{18}O, which is correlated with catchment elevation after the influence of latitude, continentality and carbonate formation temperature are controlled for. The necessary temperature estimates for Lake Flagstaff are provided by clumped-isotope (Δ_{47}) thermometry of paleosol carbonate from the North Horn Formation in the Axhandle Basin (VanDeVelde and others, 2013). Studies of modern soils have shown that in mid-latitude arid environments, such as we expect for the Eocene Flagstaff Basin, Δ_{47} temperatures typically exceed mean annual air temperature (MAAT) by 12 to 15 °C (Passey and others, 2010; Quade and others, 2013). Average early Eocene Δ_{47} temperature of 38 °C from the Axhandle Basin would therefore correspond to a MAAT of 23 to 26 °C. Although the exact relationship between soil carbonate Δ_{47} and MAAT remains uncertain, δ^{13}C-excess is relatively insensitive to temperature, as even a 25 °C temperature error alters the resulting elevation estimate by ~300 meters. With this caveat in mind, stable isotope data from the Flagstaff Formation suggest catchment elevations of 1 to 1.5 km, similar to estimates for eastern Nevada from δ^{13}C-excess (Horton and Oze, 2012) and Δ_{47} (Lechler and others, 2013), but differing somewhat from an estimate for the late Cretaceous of >2 km (Snell and others, 2014). These elevation estimates are in stark contrast to the abundant geologic evidence for a high elevation (>3 km), low relief Nevadaplano and topographically rugged Sevier thrust-front from the late Cretaceous to early Eocene (DeCelles, 2004).

Noting that their reconstruction of regional δ^{18}O patterns do not show evidence for high elevations in most of the Cordillera prior to the middle Eocene, Chamberlain and others (2012) suggested that a high Nevadaplano was masked during the late Cretaceous to early Eocene by the combination of warm climate and a strong summer monsoon originating from the Mississippi Embayment to the southeast, effects which could also impact δ^{13}C-excess. Fricke and others (2010) combined climate modeling and stable isotope analysis to support the existence of a strong monsoon along the eastern flank of the Sevier Highlands during the Campanian. Carbonates from different environments suggest that local precipitation in the Sevier foreland had relatively high oxygen isotope ratios, while large trunk rivers were fed primarily by high-altitude precipitation with significantly lower δ^{18}O. While their study supported a strong monsoon along the entire front of the Sevier Highlands during the Campanian, they suggest that growth of Laramide uplifts in the foreland and the retreat of the Western Interior Seaway during the Maastrichtian would have subsequently limited the western penetration of monsoon precipitation. The weakening of the monsoon is supported both by climate models (Sewall and Sloan, 2006) and the loss of bimodality in oxygen isotope records by the early Eocene in the Bighorn and Powder River Basins (Dettman and Lohmann, 2000).

With the Sevier Highlands at the latitude of Lake Flagstaff largely shielded from monsoonal penetration by the Laramide Front Range Uplift (Sewall and Sloan, 2006), its effects on the oxygen isotope composition of inflow would likely be equally limited. Monsoonal effects on oxygen isotope ratios of precipitation would also not impact Δ_{47} elevation estimates, which are independent of changes in δ^{18}O. Two studies of the Sheep Pass Basin in east central Nevada (Lechler and others, 2013; Snell and others, 2014) produced elevation estimates for the Nevadaplano roughly overlapping at ~2 km. Lechler and others regarded their Eocene estimate as a maximum elevation, while Snell and others considered their Late Cretaceous estimate to be a minimum. Our results support the interpretation of Lechler and others, though we note that significant uncertainties remain in the use of both δ^{13}C-excess and Δ_{47} lapse rates in estimating paleoelevation.

These elevation estimates therefore support the hypothesis that maximum Cordilleran elevations were reached later in the Paleogene, during a north to south sweep of surface uplift (Horton and others, 2004; Kent-Corson and others, 2006; Mix and others, 2011; Chamberlain and others, 2012) that reached the Flagstaff Basin ~44 Ma (Davis and others, 2009b). While this interpretation is consistent with the paleoelevation data, there is abundant structural evidence to support the existence of a broad Nevadaplano with at least moderate elevation prior to the middle Eocene (DeCelles, 2004; DeCelles and Coogan, 2006; Ernst, 2010; Henry and others, 2012; Dickinson, 2013). The elevation of the Nevadaplano during the late Cretaceous and early Paleogene therefore remains a fundamental question for the evolution of the Cordillera, which will only be resolved with intensive further study on both the paleotopographic and geodynamic fronts.

CONCLUSIONS

Our study presents new petrographic and isotopic data for the upper Flagstaff Formation at multiple sites on the Wasatch and Gunnison Plateaus. These results integrate with previous studies of the Flagstaff Formation to provide a more comprehensive view of the evolution of the lake system. Our data suggest that dolomite in the Flagstaff Formation formed as a primary precipitate from fluids of a similar chemical composition and temperature to those that precipitated calcitic micrite. Significant evaporative enrichment of these fluids, beyond that required to induce authigenic carbonate formation generally, is not indicated by the stable isotope data. We suggest that microbial activity in the shallow sediments of Lake Flagstaff served as a primary control on the mineralogy of precipitating carbonates.

Our isotopic data, when corrected for differences in equilibrium fractionation of oxygen between dolomite and calcite, allow for the identification of a moderately strong covariant trend between carbon and oxygen in the Flagstaff Formation. In carbonate-producing lakes this covariance indicates closed-basin conditions, and typically forms a consistent trend for any individual lake. Our data therefore indicate that lacustrine carbonates of the Flagstaff Formation sampled from the southern Wasatch Plateau formed in a closed lake which occupied the Flagstaff Basin from the late Paleocene into the early Eocene. We use this isotopic trend to expand the correlation of the Flagstaff Formation beyond the Wasatch Plateau, indicating that carbonates in the Axhandle Basin were also deposited by Lake Flagstaff. Limestones assigned to the Flagstaff Member of the Green River Formation in the Uinta Basin possess a geochemical signature distinct from Flagstaff Formation carbonates exposed to the south and were likely the product of a separate Paleocene-age Lake Uinta. Lake Flagstaff was fed by drainages in what was apparently a relatively low-elevation Sevier thrust-belt, a discrepancy that only deepens questions concerning the evolution of topography in the Cordillera.

ACKNOWLEDGMENTS

The authors would like to thank C. Bataille and V. Srinivasaraghavan for assistance in the field, as well as B. Bowen, S. Oleynik and B. Thieling in the lab. Support was provided by U.S. National Science Foundation grants EAR-0628302 and OCE-0902882 to G. Bowen, and a Purdue University Geology Michael C. Gardner award to J. VanDeVelde. The quality of the manuscript was greatly improved through careful reviews by S.J. Davis and an anonymous reviewer.

APPENDIX

Table A1

Carbon and oxygen isotopes of Flagstaff Formation carbonates

Locality Sample number	$\delta^{13}C$ VPDB	$\delta^{18}O$ VPDB	% Calcite	Section Level(m)	Material
Musinia Peak					
FF09001	-8.46	-10.42	100	130	limestone
FF09002	-8.90	-10.49	100	124.5	limestone
FF09003	-8.36	-10.16	100	120	limestone
FF09004	-8.76	-11.02	100	116	limestone
FF09005	-6.49	-6.89	0	112	limestone
FF09006	-6.71	-6.70	0	107	limestone
FF09007	-8.17	-10.95	100	102	limestone
FF09008	-8.35	-11.20	100	98.5	limestone
FF09009	-8.21	-10.49	100	97	limestone
FF09010	-6.20	-7.33	0	93	limestone
FF09011	-8.24	-10.11	100	87	limestone
FF09012	-6.49	-5.86	0	83	limestone
FF09013	-6.40	-6.63	0	76	limestone
FF09014	-5.87	-6.14	0	71.5	limestone
FF09076	-6.04	-6.10	0	38.4	limestone
FF09067	-6.2	-8.8	n/a	5	paleosol
FF09068	-4.5	-9.7	n/a	9	paleosol
FF09069	-7.3	-8.8	n/a	19.5	paleosol
FF09070	-7.1	-8.5	n/a	25	paleosol
FF09071	-4.3	-8.9	n/a	31.5	paleosol
FF09072	-8.4	-9.7	n/a	34.5	paleosol
FF09073	-7.5	-8.1	n/a	38	paleosol
FF09074	-8.8	-8.7	n/a	38.4	paleosol
FF09075	-6.3	-5.1	n/a	48.5	paleosol
FF09001S	-8.9	-16.8	n/a	130	cement
FF09002S	-9.3	-11.6	n/a	124.5	cement
FF09003S	-8.6	-11.8	n/a	120	cement
FF09004S	-8.7	-15.0	n/a	116	cement
FF09005S	-8.7	-15.7	n/a	112	cement
FF09006S	-8.3	-9.8	n/a	107	cement
FF09007S	-8.4	-11.4	n/a	102	cement
FF09008S	-8.2	-11.4	n/a	98.5	cement
FF09009S	-10.7	-13.5	n/a	97	cement
FF09011S	-8.2	-14.5	n/a	87	cement
FF09012S	-7.7	-10.6	n/a	83	cement
FF09013S	-8.1	-13.2	n/a	76	cement
Upper White Mountain					
FF09015	-7.41	-6.52	0	63	limestone
FF09016	-8.04	-8.55	100	60	limestone
FF09017	-8.44	-8.60	100	55.5	limestone
FF09018	-7.25	-5.83	0	50	limestone
FF09019	-7.39	-6.34	0	47.5	limestone
FF09020	-7.04	-5.15	0	44	limestone
FF09021	-8.62	-9.85	100	35	limestone
FF09022	-8.78	-10.58	100	31	limestone
FF09023	-9.33	-10.35	100	29	limestone
FF09024	-6.97	-5.53	0	26.5	limestone
FF09025	-8.70	-9.67	100	23	limestone
FF09026	-8.30	-9.98	100	17	limestone

Table A1

(continued)

Locality Sample number	$\delta^{13}C$ VPDB	$\delta^{18}O$ VPDB	% Calcite	Section Level(m)	Material
Upper White Mountain					
FF09027	-8.42	-10.09	100	11	limestone
FF09022S	-8.8	-14.9	n/a	31	cement
FF09023S	-8.4	-11.2	n/a	29	cement
FF09024S	-8.3	-10.5	n/a	26.5	cement
FF09025S	-8.6	-11.6	n/a	23	cement
FF09026S	-8.9	-9.8	n/a	17	cement
FF09027S	-8.7	-11.2	n/a	11	cement
FF09030S	-8.7	-11.1	n/a	0	cement
Lower White Mountain					
FF09031	-3.49	-4.29	0	0	limestone
FF09032	-3.37	-6.91	100	3.5	limestone
FF09033	-2.78	-3.61	0	7.5	limestone
FF09034	-2.88	-6.14	100	12	limestone
FF09035	-2.84	-7.73	100	18	limestone
FF09036	-3.16	-8.22	100	22.5	limestone
FF09037	-4.49	-8.84	100	28.5	limestone
FF09039	-5.70	-9.59	100	30.5	limestone
FF09042	-2.91	-8.26	100	42	limestone
FF09043	-4.33	-8.44	100	45	limestone
FF09044	-2.75	-8.35	100	52	limestone
South Maple Canyon					
FF1017	-6.95	-9.67	11	1.5	limestone
FF1018	-6.67	-9.41	28	2	limestone
FF1019	-6.43	-9.17	39	5	limestone
FF1020	-5.72	-9.00	44	7	limestone
FF1021	-6.64	-8.58	78	9.5	limestone
FF1022	-5.60	-8.51	28	11	limestone
FF10023	-5.16	-7.45	61	13	limestone
FF10024	-4.55	-7.98	44	14.5	limestone
FF10025	-4.83	-9.07	22	17	limestone
FF10026	-5.19	-8.35	28	19	limestone
FF10027	-5.62	-7.90	39	20	limestone
FF10028	-5.53	-8.23	67	21	limestone
FF10029	-5.83	-7.90	94	23	limestone
FF10030	-5.61	-7.39	89	24.5	limestone
FF10031	-5.35	-7.25	83	26	limestone
FF10032	-5.01	-8.07	39	28	limestone
FF10033	-5.32	-7.08	89	30	limestone
FF10034	-5.63	-9.23	50	32.5	limestone
FF10035	-3.38	-8.73	44	34.5	limestone
FF10036	-4.65	-8.72	56	38	limestone
FF10037	-5.85	-7.67	89	39	limestone
FF10039	-6.00	-7.49	44	49	limestone
FF10040	-1.78	-8.43	56	53.5	limestone
FF10041	-3.96	-6.76	83	56	limestone
FF10042	-3.55	-8.40	56	60	limestone
FF10043	-4.51	-7.96	83	62	limestone
FF10044	-3.18	-6.36	78	65	limestone

References

Alonso-Zarza, A. M., and Martín-Pérez, A., 2008, Dolomite in caves: Recent dolomite formation in oxic, non-sulfate environments. Castañar Cave, Spain: Sedimentary Geology, v. 205, n. 3–4, p. 160–164, http://dx.doi.org/10.1016/j.sedgeo.2008.02.006

Arenas, C., Casanova, J., and Pardo, G., 1997, Stable-isotope characterization of the Miocene lacustrine systems of Los Monegros (Ebro Basin, Spain): palaeogeographic and palaeoclimatic implications: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 128, n. 1–4, p. 133–155, http://dx.doi.org/10.1016/S0031-0182(96)00052-1

Armenteros, I., 2010, Diagenesis of Carbonates in Continental Settings, *in* Alonso-Zarza, A. M., and Tanner, L. H., editors, Carbonates in Continental Settings: Geochemistry, Diagenesis and Applications: Elsevier, Developments in Sedimentology, v. 62, p. 61–151, http://dx.doi.org/10.1016/S0070-4571(09)06202-5

Baughman, R., ms, 1959, The geology of the Musinia graben area: Sevier and Sanpete Counties, Utah: Columbus, Ohio, Ohio State University, Masters Thesis, 112 p.

Bowen, G. J., and Bowen, B. B., 2008, Mechanisms of PETM global change constrained by a new record from central Utah: Geology, v. 36, n. 5, p. 379–382, http://dx.doi.org/10.1130/G24597a.1

Bowen, G. J., Daniels, A. L., and Bowen, B. B., 2008, Paleoenvironmental isotope geochemistry and paragenesis of lacustrine and palustrine carbonates, Flagstaff Formation, Central Utah, U.S.A.: Journal of Sedimentary Research, v. 78, n. 3, p. 162–174, http://dx.doi.org/10.2110/Jsr.2008.021

Bristow, T. F., Kennedy, M. J., Morrison, K. D., and Mrofka, D. D., 2012, The influence of authigenic clay formation on the mineralogy and stable isotopic record of lacustrine carbonates: Geochimica et Cosmochimica Acta, v. 90, p. 64–82, http://dx.doi.org/10.1016/j.gca.2012.05.006

Bustillo, M. A., and Alonso-Zarza, A. M., 2007, Overlapping of pedogenesis and meteoric diagenesis in distal alluvial and shallow lacustrine deposits in the Madrid Miocene Basin, Spain: Sedimentary Geology, v. 198, n. 3–4, p. 255–271, http://dx.doi.org/10.1016/j.sedgeo.2006.12.006

Carroll, A. R., Doebbert, A. C., Booth, A. L., Chamberlain, C. P., Rhodes-Carson, M. K., Smith, M. E., Johnson, C. M., and Beard, B. L., 2008, Capture of high-altitude precipitation by a low-altitude Eocene lake, western U.S.: Geology, v. 36, n. 10, p. 791–794, http://dx.doi.org/10.1130/G24783a.1

Chacko, T., Cole, D. R., and Horita, J., 2001, Equilibrium Oxygen, Hydrogen and Carbon Isotope Fractionation Factors Applicable to Geologic Systems: Reviews in Mineralogy and Geochemistry, v. 43, p. 1–81, http://dx.doi.org/10.2138/gsrmg.43.1.1

Chamberlain, C. P., Mix, H. T., Mulch, A., Hren, M. T., Kent-Corson, M. L., Davis, S. J., Horton, T. W., and Graham, S. A., 2012, The Cenozoic climatic and topographic evolution of the Western North American Cordillera: American Journal of Science, v. 312, n. 2, p. 213–262, http://dx.doi.org/10.2475/02.2012.05

Chetel, L. M., Janecke, S. U., Carroll, A. R., Beard, B. L., Johnson, C. M., and Singer, B. S., 2011, Paleogeographic reconstruction of the Eocene Idaho River, North American Cordillera: Geological Society of America Bulletin, v. 123, n. 1–2, p. 71–88, http://dx.doi.org/10.1130/b30213.1

Cohen, A. S., 2003, Paleolimnology: The History and Evolution of Lake Systems: Oxford, Oxford University Press, 528 p.

Davis, S. J., Wiegand, B. A., Carroll, A. R., and Chamberlain, C. P., 2008, The effect of drainage reorganization on paleoaltimetry studies: An example from the Paleogene Laramide foreland: Earth and Planetary Science Letters, v. 275, n. 3–4, p. 258–268, http://dx.doi.org/10.1016/j.epsl.2008.08.009

Davis, S. J., Mix, H. T., Wiegand, B. A., Carroll, A. R., and Chamberlain, C. P., 2009a, Synorogenic evolution of large-scale drainage patterns: Isotope paleohydrology of sequential Laramide basins: American Journal of Science, v. 309, n. 7, p. 549–602, http://dx.doi.org/10.2475/07.2009.02

Davis, S. J., Mulch, A., Carroll, A. R., Horton, T. W., and Chamberlain, C. P., 2009b, Paleogene landscape evolution of the central North American Cordillera: Developing topography and hydrology in the Laramide foreland: Geological Society of America Bulletin, v. 121, n. 1–2, p. 100–116, http://dx.doi.org/10.1130/B26308.1

Davis, S. J., Dickinson, W. R., Gehrels, G. E., Spencer, J. E., Lawton, T. F., and Carroll, A. R., 2010, The Paleogene California River: Evidence of Mojave-Uinta paleodrainage from U-Pb ages of detrital zircons: Geology, v. 38, n. 10, p. 931–934, http://dx.doi.org/10.1130/G31250.1

DeCelles, P. G., 1994, Late Cretaceous-Paleocene synorogenic sedimentation and kinematic history of the Sevier thrust belt, northeast Utah and southwest Wyoming: Geological Society of America Bulletin, v. 106, n. 1, p. 32–56, http://dx.doi.org/10.1130/0016-7606(1994)106<0032:Lcpssa>2.3.Co;2

——— 2004, Late Jurassic to Eocene evolution of the Cordilleran thrust belt and foreland basin system, western U.S.A: American Journal of Science, v. 304, n. 2, p. 105–168, http://dx.doi.org/10.2475/ajs.304.2.105

DeCelles, P. G., and Coogan, J. C., 2006, Regional structure and kinematic history of the Sevier fold-and-thrust belt, central Utah: Geological Society of America Bulletin, v. 118, n. 7–8, p. 841–864, http://dx.doi.org/10.1130/B25759.1

Deocampo, D. M., 2004, Authigenic clays in East Africa: Regional trends and paleolimnology at the Plio–Pleistocene boundary, Olduvai Gorge, Tanzania: Journal of Paleolimnology, v. 31, n. 1, p. 1–9, http://dx.doi.org/10.1023/B:JOPL.0000013353.86120.9b

——— 2010, The Geochemistry of Continental Carbonates, *in* Alonso-Zarza, A. M., and Tanner, L. H., editors, Carbonates in Continental Settings: Geochemistry, Diagenesis and Applications: Elsevier, Developments in Sedimentology, v. 62, p. 1–59, http://dx.doi.org/10.1016/S0070-4571(09)06201-3

Desborough, G. A., 1978, A biogenic-chemical stratified lake model for the origin of oil-shale of the Green River Formation: An alternative to playa-lake model: Geological Society of America Bulletin, v. 89, n. 7, p. 961–971, http://dx.doi.org/10.1130/0016-7606(1978)89<961:Abslmf>2.0.Co;2

Dettman, D. L., and Lohmann, K. C., 2000, Oxygen isotope evidence for high-altitude snow in the Laramide

Rocky Mountains of North America during the Late Cretaceous and Paleogene: Geology, v. 28, n. 3, p. 243–246, http://dx.doi.org/10.1130/0091-7613(2000)28<243:oiefhs>2.0.co;2

Dickinson, W. R., 2013, Phanerozoic palinspastic reconstructions of Great Basin geotectonics (Nevada-Utah, USA): Geosphere, , v. 9, n. 5, http://dx.doi.org/10.1130/ges00888.1

Dickinson, W. R., Lawton, T. F., and Inman, K. F., 1986, Sandstone Detrital Modes, Central Utah Foreland Region: Stratigraphic Record of Cretaceous-Paleogene Tectonic Evolution: Journal of Sedimentary Petrology, v. 56, n. 2, p. 276–293, http://dx.doi.org/10.1306/212F88E6-2B24-11D7-8648000102C1865D

Dickinson, W. R., Klute, M. A., Hayes, M. J., Janecke, S. U., Lundin, E. R., McKittrick, M. A., and Olivares, M. D., 1988, Paleogeographic and Paleotectonic Setting of Laramide Sedimentary Basins in the Central Rocky-Mountain Region: Geological Society of America Bulletin, v. 100, n. 7, p. 1023–1039, http://dx.doi.org/10.1130/0016-7606(1988)100<1023:Papsol>2.3.Co;2

Dickinson, W. R., Lawton, T. F., Pecha, M., Davis, S. J., Gehrels, G. E., and Young, R. A., 2012, Provenance of the Paleogene Colton Formation (Uinta Basin) and Cretaceous-Paleogene provenance evolution in the Utah foreland: Evidence from U-Pb ages of detrital zircons, paleocurrent trends, and sandstone petrofacies: Geosphere, v. 8, n. 4, p. 854–880, http://dx.doi.org/10.1130/Ges00763.1

Dickson, J. A. D., 1965, A modified staining technique for carbonates in thin section: Nature, v. 205, p. 587, http://dx.doi.org/10.1038/205587a0

Doebbert, A. C., Carroll, A. R., Mulch, A., Chetel, L. M., and Chamberlain, C. P., 2010, Geomorphic controls on lacustrine isotopic compositions: Evidence from the Laney Member, Green River Formation, Wyoming: Geological Society of America Bulletin, v. 122, p. 236–252, http://dx.doi.org/10.1130/B26522.1

Ernst, W. G., 2010, Young convergent-margin orogens, climate, and crustal thickness—A Late Cretaceous–Paleogene Nevadaplano in the American Southwest?: Lithosphere, v. 2, n. 2, p. 67–75, http://dx.doi.org/10.1130/l84.1

Eugster, H. P., and Surdam, R. C., 1973, Depositional Environment of the Green River Formation of Wyoming: A Preliminary Report: Geological Society of America Bulletin, v. 84, n. 4, p. 1115–1120, http://dx.doi.org/10.1130/0016-7606(1973)84<1115:deotgr>2.0.co;2

Fouch, T. D., 1976, Revision of the lower part of the Tertiary system in the central and western Uinta Basin, Utah : correlation of lower Tertiary stratigraphic units recently penetrated in northeastern Utah: US Geological Survey Bulletin, v. 1405-C, p. 1–7.

Fricke, H. C., Foreman, B. Z., and Sewall, J. O., 2010, Integrated climate model-oxygen isotope evidence for a North American monsoon during the Late Cretaceous: Earth and Planetary Science Letters, v. 289, n. 1–2, p. 11–21, http://dx.doi.org/10.1016/j.epsl.2009.10.018

Gaffey, S. J., 1986, Spectral reflectance of carbonate minerals in the visible and near-infrared (0.35-2.55 microns) - calcite, aragonite, and dolomite: American Mineralogist, v. 71, n. 1–2, p. 151–162.

García del Cura, M. A., Calvo, J. P., Ordóñez, S., Jones, B. F., and Cañaveras, J. C., 2001, Petrographic and geochemical evidence for the formation of primary, bacterially induced lacustrine dolomite: La Roda 'white earth' (Pliocene, central Spain): Sedimentology, v. 48, n. 4, p. 897–915, http://dx.doi.org/10.1046/j.1365-3091.2001.00388.x

Gierlowski-Kordesch, E. H., Jacobson, A. D., Blum, J. D., and Garces, B. L. V., 2008, Watershed reconstruction of a Paleocene-Eocene lake basin using Sr isotopes in carbonate rocks: Geological Society of America Bulletin, v. 120, n. 1–2, p. 85–95, http://dx.doi.org/10.1130/B26070.1

Goldstrand, P. M., 1992, Evolution of Late Cretaceous and early Tertiary basins of Southwest Utah based on clastic petrology: Journal of Sedimentary Research, v. 62, n. 3, p. 495–507, http://dx.doi.org/10.1306/d4267933-2b26-11d7-8648000102c1865d

——— 1994, Tectonic development of Upper Cretaceous to Eocene strata of southwestern Utah: Geological Society of America Bulletin, v. 106, n. 1, p. 145–154, http://dx.doi.org/10.1130/0016-7606(1994)106<0145:Tdouct>2.3.Co;2

Hay, R. L., and Kyser, T. K., 2001, Chemical sedimentology and paleoenvironmental history of Lake Olduvai, a Pliocene lake in northern Tanzania: Geological Society of America Bulletin, v. 113, n. 12, p. 1505–1521, http://dx.doi.org/10.1130/0016-7606(2001)113<1505:Csapho>2.0.Co;2

Henry, C. D., Hinz, N. H., Faulds, J. E., Colgan, J. P., John, D. A., Brooks, E. R., Cassel, E. J., Garside, L. J., Davis, D. A., and Castor, S. B., 2012, Eocene–Early Miocene paleotopography of the Sierra Nevada–Great Basin–Nevadaplano based on widespread ash-flow tuffs and paleovalleys: Geosphere, v. 8, n. 1, p. 1–27, http://dx.doi.org/10.1130/ges00727.1

Horton, T. W., and Oze, C., 2012, Are two elements better than one? Dual isotope-ratio detrending of evaporative effects on lake carbonate paleoelevation proxies: Geochemistry, Geophysics, Geosystems, v. 13, n. 6, p. Q0AK05, http://dx.doi.org/10.1029/2012gc004132

Horton, T. W., Sjostrom, D. J., Abruzzese, M. J., Poage, M. A., Waldbauer, J. R., Hren, M., Wooden, J., and Chamberlain, C. P., 2004, Spatial and temporal variation of Cenozoic surface elevation in the Great Basin and Sierra Nevada: American Journal of Science, v. 304, n. 10, p. 862–888, http://dx.doi.org/10.2475/ajs.304.10.862

Jones, B., 2010, The preferential association of dolomite with microbes in stalactites from Cayman Brac, British West Indies: Sedimentary Geology, v. 226, n. 1–4, p. 94–109, http://dx.doi.org/10.1016/j.sedgeo.2010.03.004

Karlstrom, K. E., Lee, J. P., Kelley, S. A., Crow, R. S., Crossey, L. J., Young, R. A., Lazear, G., Beard, L. S., Ricketts, J. W., Fox, M., and Shuster, D. L., 2014, Formation of the Grand Canyon 5 to 6 million years ago through integration of older palaeocanyons: Nature Geoscience, v. 7, p. 239–244, http://dx.doi.org/10.1038/ngeo2065

Kent-Corson, M. L., Sherman, L. S., Mulch, A., and Chamberlain, C. P., 2006, Cenozoic topographic and climatic response to changing tectonic boundary conditions in Western North America: Earth and Planetary Science Letters, v. 252, n. 3–4, p. 453–466, http://dx.doi.org/10.1016/j.epsl.2006.09.049

Kim, S. T., and O'Neil, J. R., 1997, Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates: Geochimica et Cosmochimica Acta, v. 61, n. 16, p. 3461–3475, http://dx.doi.org/10.1016/S0016-7037(97)00169-5

La Rocque, A., 1960, Molluscan faunas of the Flagstaff Formation of Central Utah: Geological Society of America Memoirs, v. 78, 100 p., http://dx.doi.org/10.1130/MEM78-p1

Land, L. S., 1998, Failure to precipitate dolomite at 25 °C from dilute solution despite 1000-fold oversaturation after 32 years: Aquatic Geochemistry, v. 4, n. 3–4, p. 361–368, http://dx.doi.org/10.1023/A:1009688315854

Last, W. M., 1990, Lacustrine dolomite - an overview of modern, Holocene, and Pleistocene occurrences: Earth-Science Reviews, v. 27, n. 3, p. 221–263, http://dx.doi.org/10.1016/0012-8252(90)90004-F

Last, F. M., Last, W. M., and Halden, N. M., 2012, Modern and late Holocene dolomite formation: Manito Lake, Saskatchewan, Canada: Sedimentary Geology, v. 281, p. 222–237, http://dx.doi.org/10.1016/j.sedgeo.2012.09.012

Lawton, T. F., 2008, Laramide Sedimentary Basins, *in* Miall, A. D., editor, The Sedimentary Basins of the United States and Canada: Elsevier, Sedimentary Basins of the World, v. 5, p. 429–450, http://dx.doi.org/10.1016/S1874-5997(08)00012-9

Lawton, T. F., and Trexler, J. H., Jr., 1991, Piggyback basin in the Sevier orogenic Belt, Utah: Implications for development of the thrust wedge: Geology, v. 19, n. 8, p. 827–830, http://dx.doi.org/10.1130/0091-7613(1991)019<0827:Pbitso>2.3.Co;2

Lawton, T. F., Talling, P. J., Hobbs, R. S., Trexler, J. H., Jr., Weiss, M. P., and Burbank, D. W., 1993, Structure and stratigraphy of Upper Cretaceous and Paleogene strata (North Horn Formation), eastern San Pitch Mountains, Utah– Sedimentation at the front of the Sevier orogenic belt: US Geological Survey Bulletin, v. 1787-II, 43 p.

Lechler, A. R., Niemi, N. A., Hren, M. T., and Lohmann, K. C., 2013, Paleoelevation estimates for the northern and central proto–Basin and Range from carbonate clumped isotope thermometry: Tectonics, v. 32, n. 3, p. 295–316, http://dx.doi.org/10.1002/tect.20016

Leng, M. J., and Henderson, A. C. G., 2013, Recent advances in isotopes as palaeolimnological proxies: Journal of Paleolimnology, v. 49, n. 3, p. 481–496, http://dx.doi.org/10.1007/s10933-012-9667-5

Leng, M. J., and Marshall, J. D., 2004, Palaeoclimate interpretation of stable isotope data from lake sediment archives: Quaternary Science Reviews, v. 23, n. 7–8, p. 811–831, http://dx.doi.org/10.1016/j.quascirev.2003.06.012

Li, H. C., and Ku, T. L., 1997, $\delta^{13}C$-$\delta^{18}O$ covariance as a paleohydrological indicator for closed-basin lakes: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 133, n. 1–2, p. 69–80, http://dx.doi.org/10.1016/S0031-0182(96)00153-8

Marcantel, E. L., and Weiss, M. P., 1968, Colton Formation (Eocene: Fluviatile) and Associated Lacustrine Beds, Gunnison Plateau, Central Utah: Ohio Journal of Science, v. 68, p. 40–49,

Meister, P., Reyes, C., Beaumont, W., Rincon, M., Collins, L., Berelson, W., Stott, L., Corsetti, F., and Nealson, K. H., 2011, Calcium and magnesium-limited dolomite precipitation at Deep Springs Lake, California: Sedimentology, v. 58, n. 7, p. 1810–1830, http://dx.doi.org/10.1111/j.1365-3091.2011.01240.x

Mix, H. T., Mulch, A., Kent-Corson, M. L., and Chamberlain, C. P., 2011, Cenozoic migration of topography in the North American Cordillera: Geology, v. 39, n. 1, p. 87–90, http://dx.doi.org/10.1130/G31450.1

Müller, G., Irion, G., and Förstner, U., 1972, Formation and diagenesis of inorganic Ca−Mg carbonates in the lacustrine environment: Naturwissenschaften, v. 59, n. 4, p. 158–164, http://dx.doi.org/10.1007/bf00637354

Neat, P. L., Faure, G., and Pegram, W. J., 1979, The isotopic composition of strontium in non-marine carbonate rocks: the Flagstaff Formation of Utah*: Sedimentology, v. 26, n. 2, p. 271–282, http://dx.doi.org/10.1111/j.1365-3091.1979.tb00354.x

Passey, B. H., Levin, N. E., Cerling, T. E., Brown, F. H., and Eiler, J. M., 2010, High-temperature environments of human evolution in East Africa based on bond ordering in paleosol carbonates: Proceedings of the National Academy of Sciences of the United States of America, v. 107, n. 25, p. 11245–11249, http://dx.doi.org/10.1073/pnas.1001824107

Paz, J. D. S., and Rossetti, D. F., 2006, Paleohydrology of an Upper Aptian lacustrine system from northeastern Brazil: Integration of facies and isotopic geochemistry: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 241, n. 2, p. 247–266, http://dx.doi.org/10.1016/j.palaeo.2006.03.040

Quade, J., Eiler, J., Daëron, M., and Achyuthan, H., 2013, The clumped isotope geothermometer in soil and paleosol carbonate: Geochimica et Cosmochimica Acta, v. 105, p. 92–107, http://dx.doi.org/10.1016/j.gca.2012.11.031

Rich, R. H. V., and Collinson, J. W., 1973, First mammalian fossil from Flagstaff Limestone, central Utah: *Vulpavus* australis (Carnivora: Miacidae): Journal of Paleontology, v. 47, n. 5, p. 854–860, http://www.jstor.org/stable/1303064

Rosen, M. R., and Coshell, L., 1992, A New Location of Holocene Dolomite Formation, Lake Hayward, Western Australia: Sedimentology, v. 39, n. 1, p. 161–166, http://dx.doi.org/10.1111/j.1365-3091.1992.tb01028.x

Rosenbaum, J., and Sheppard, S. M. F., 1986, An isotopic study of siderites, dolomites and ankerites at High-Temperatures: Geochimica et Cosmochimica Acta, v. 50, n. 6, p. 1147–1150, http://dx.doi.org/10.1016/0016-7037(86)90396-0

Ryder, R. T., Fouch, T. D., and Elison, J. H., 1976, Early Tertiary sedimentation in the western Uinta Basin, Utah: Geological Society of America Bulletin, v. 87, n. 4, p. 496–512, http://dx.doi.org/10.1130/0016-7606(1976)87<496:Etsitw>2.0.Co;2

Saez, A., and Cabrera, L., 2002, Sedimentological and palaeohydrological responses to tectonics and climate

in a small, closed, lacustrine system: Oligocene As Pontes Basin (Spain): Sedimentology, v. 49, n. 5, p. 1073–1094, http://dx.doi.org/10.1046/j.1365-3091.2002.00490.x
Sánchez-Román, M., McKenzie, J. A., de Luca Rebello Wagener, A., Rivadeneyra, M. A., and Vasconcelos, C., 2009, Presence of sulfate does not inhibit low-temperature dolomite precipitation: Earth and Planetary Science Letters, v. 285, n. 1–2, p. 131–139, http://dx.doi.org/10.1016/j.epsl.2009.06.003
Sewall, J. O., and Sloan, L. C., 2006, Come a little bit closer: A high-resolution climate study of the early Paleogene Laramide foreland: Geology, v. 34, n. 2, p. 81–84, http://dx.doi.org/10.1130/g22177.1
Sheppard, S. M. F., and Schwarcz, H. P., 1970, Fractionation of carbon and oxygen isotopes and magnesium between coexisting metamorphic calcite and dolomite: Contributions to Mineralogy and Petrology, v. 26, p. 161–198, http://dx.doi.org/10.1007/Bf00373200
Sinha, R., Smykatz-Kloss, W., Stüben, D., Harrison, S. P., Berner, Z., and Kramar, U., 2006, Late Quaternary palaeoclimatic reconstruction from the lacustrine sediments of the Sambhar playa core, Thar Desert margin, India: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 233, n. 3–4, p. 252–270, http://dx.doi.org/10.1016/j.palaeo.2005.09.012
Smith, M. E., Carroll, A. R., and Singer, B. S., 2008, Synoptic reconstruction of a major ancient lake system: Eocene Green River Formation, western United States: Geological Society of America Bulletin, v. 120, n. 1–2, p. 54–84, http://dx.doi.org/10.1130/B26073.1
Snell, K. E., Koch, P. L., Druschke, P., Foreman, B. Z., and Eiler, J. M., 2014, High elevation of the 'Nevadaplano' during the Late Cretaceous: Earth and Planetary Science Letters, v. 386, p. 52–63, http://dx.doi.org/10.1016/j.epsl.2013.10.046
Spieker, E. M., 1946, Late Mesozoic and early Cenozoic history of central Utah: US Geological Survey Professional Paper 205-D, p. 117–161.
Spieker, E. M., and Reeside, J. B., Jr., 1925, Cretaceous and Tertiary formations of the Wasatch Plateau, Utah: Geological Society of America Bulletin, v. 36, p. 435–454, http://dx.doi.org/10.1130/GSAB-36-435
Stanley, K. O., and Collinson, J. W., 1979, Depositional history of Paleocene-lower Eocene Flagstaff Limestone and coeval rocks, central Utah: American Association of Petroleum Geologists Bulletin, v. 63, n. 3, p. 311–323.
Swart, P. K., Burns, S. J., and Leder, J. J., 1991, Fractionation of the stable Isotopes of oxygen and carbon in carbon dioxide during the reaction of calcite with phosphoric acid as a function of temperature and technique: Chemical Geology, v. 86, n. 2, p. 89–96, http://dx.doi.org/10.1016/0168-9622(91)90055-2
Talbot, M. R., 1990, A review of the paleohydrological interpretation of carbon and oxygen isotopic ratios in primary lacustrine carbonates: Chemical Geology, v. 80, n. 4, p. 261–279, http://dx.doi.org/10.1016/0168-9622(90)90009-2
Talling, P. J., Burbank, D. W., Lawton, T. F., Hobbs, R. S., and Lund, S. P., 1994, Magnetostratigraphic chronology of Cretaceous-to-Eocene thrust belt evolution, central Utah, USA: The Journal of Geology, v. 102, n. 2, p. 181–196, http://dx.doi.org/10.1086/629662
Talling, P. J., Lawton, T. F., Burbank, D. W., and Hobbs, R. S., 1995, Evolution of Latest Cretaceous–Eocene Nonmarine Deposystems in the Axhandle Piggyback Basin of Central Utah: Geological Society of America Bulletin, v. 107, n. 3, p. 297–315, http://dx.doi.org/10.1130/0016-7606(1995)107<0297:Eolcen>2.3.Co;2
Taylor, W. J., 1993, Stratigraphic and lithologic analysis of the Claron Formation in southwestern Utah: Utah Geological Survey Miscellaneous Publication n. 93–1, p. 52.
VanDeVelde, J. H., Bowen, G. J., Passey, B. H., and Bowen, B. B., 2013, Climatic and diagenetic signals in the stable isotope geochemistry of dolomitic paleosols spanning the Paleocene–Eocene boundary: Geochimica et Cosmochimica Acta, v. 109, p. 254–267, http://dx.doi.org/10.1016/j.gca.2013.02.005
Vasconcelos, C., and McKenzie, J. A., 1997, Microbial mediation of modern dolomite precipitation and diagenesis under anoxic conditions (Lagoa Vermelha, Rio de Janeiro, Brazil): Journal of Sedimentary Research, v. 67, n. 3, p. 378–390, http://dx.doi.org/10.1306/D4268577-2B26-11D7-8648000102C1865D
Vasconcelos, C., McKenzie, J. A., Warthmann, R., and Bernasconi, S. M., 2005, Calibration of the $\delta^{18}O$ paleothermometer for dolomite precipitated in microbial cultures and natural environments: Geology, v. 33, n. 4, p. 317–320, http://dx.doi.org/10.1130/G20992.1
Wacey, D., Wright, D. T., and Boyce, A. J., 2007, A stable isotope study of microbial dolomite formation in the Coorong Region, South Australia: Chemical Geology, v. 244, n. 1–2, p. 155–174, http://dx.doi.org/10.1016/j.chemgeo.2007.06.032
Warren, J. K., 1990, Sedimentology and mineralogy of dolomitic Coorong lakes, South Australia: Journal of Sedimentary Research, v. 60, n. 6, p. 843–858, http://dx.doi.org/10.1306/212F929B-2B24-11D7-8648000102C1865D
Weber, J. N., 1964, Carbon-oxygen isotopic composition of Flagstaff carbonate rocks and its bearing on the history of Paleocene-Eocene Lake Flagstaff of central Utah: Geochimica et Cosmochimica Acta, v. 28, n. 8, p. 1219–1242, http://dx.doi.org/10.1016/0016-7037(64)90125-5
Wells, N. A., 1983, Carbonate deposition, physical limnology and environmentally controlled chert formation in Paleocene-Eocene Lake Flagstaff, central Utah: Sedimentary Geology, v. 35, n. 4, p. 263–296, http://dx.doi.org/10.1016/0037-0738(83)90062-3
Wilf, P., 2000, Late Paleocene-early Eocene climate changes in southwestern Wyoming: Paleobotanical analysis: Geological Society of America Bulletin, v. 112, n. 2, p. 292–307, http://dx.doi.org/10.1130/0016-7606(2000)112<0292:Lpeecc>2.3.Co;2
Yuretich, R., 1989, Paleocene lakes of the central Rocky Mountains, western United States: Palaeogeography, Palaeoclimatology, Palaeoecology, v. 70, n. 1–3, p. 53–63, http://dx.doi.org/10.1016/0031-0182(89)90079-5
Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K., 2001, Trends, rhythms, and aberrations in global climate 65 Ma to present: Science, v. 292, n. 5517, p. 686–693, http://dx.doi.org/10.1126/science.1059412