

DID A CATASTROPHIC LAKE SPILLOVER INTEGRATE THE LATE MIOCENE EARLY PLIOCENE COLORADO RIVER AND THE GULF OF CALIFORNIA?: MICROFAUNAL AND STABLE ISOTOPE EVIDENCE FROM BLYTHE BASIN, CALIFORNIA-ARIZONA, USA

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ABSTRACT: The southern Bouse Formation (late Miocene–early Pliocene) in Blythe basin, CA-AZ, contains a controversial record of the events that preceded the integration of the developing Colorado River with the Gulf of California. High resolution microfaunal and stable isotope ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$; VPDB) data from a key outcrop of marl and claystone record an abrupt change in water chemistry that we interpret to be the result of a catastrophic sill breach. Basal marl contains a mix of brackish-water ostracodes and marine foraminifers. Ostracode $\delta^{18}\text{O}$ values are slightly negative and 6‰ higher than the host sediment carbonate precipitated in the upper water column, indicating isotopically stratified hydrologic conditions during deposition. Freshwater ostracodes abruptly appear in the overlying sediments in association with marine foraminifers, in conjunction with an abrupt change in the isotopic composition of ostracode and host sediment carbonate. The $\delta^{18}\text{O}$ values from brackish and freshwater ostracodes and the host sediment carbonate are similar ($\sim -10\text{‰}$), indicating an isotopically well-mixed water body during deposition. Sediment $\delta^{13}\text{C}$ values decrease by 4.5‰ across this transition but ostracode $\delta^{13}\text{C}$ values remain unchanged. We infer that the transition from stratified to well-mixed conditions likely took less than 300 years. The abruptness of this transition is best explained by catastrophic failure of a paleodam that rapidly altered the isotopic composition and salinity of a lake in Blythe basin. A marine or estuarine environment is unlikely. Our interpretation is consistent with other evidence for catastrophic breaching of another lacustrine Bouse Formation-bearing basin to the north.

INTRODUCTION

Controversy over the late Miocene–early Pliocene southern Bouse Formation and its role in the integration of the Colorado River and the Gulf of California continues after nearly 50 years of study (Metzger 1968; Smith 1970; Lucchitta 1979; Busing 1990; Spencer et al. 2008; House et al. 2008; McDougall and Miranda Martinez 2014; Pearthree and House 2014; Howard et al. 2015; Crossey et al. 2015). Prior to the late Miocene, drainages in the Grand Canyon region flowed dominantly to the northeast (e.g., Young and McKee 1978). By roughly 6 Ma, exotic Colorado River gravels appear west of the Grand Canyon for the first time (Lucchitta 1972). Downstream from Grand Canyon, along what is now the lower Colorado River corridor, a southward developing Colorado River integrated several topographically closed extensional basins (e.g., House et al. 2008) before making its final descent to the Gulf of California by about 5 Ma (Dorsey et al. 2007, 2011). The record of the processes that integrated these basins is preserved as the Bouse Formation.

The Bouse Formation comprises a basal limestone unit, ranging in thickness from less than 1 m to tens of meters thick, overlain by an interbedded unit consisting of up to two hundred meters of fine grained siliciclastic deposits (Metzger 1968). The physical properties of the Bouse Formation are conspicuously similar wherever it is exposed. The southern Bouse Formation exposed in Blythe basin (Fig. 1), and only in Blythe basin, is perplexing because there the basal limestone unit contains a faunal assemblage that is more characteristic of a nearshore marine environment than a continental lake (Metzger 1968; Smith

1970; McDougall and Miranda Martinez 2014). The depositional environment of the southern Bouse Formation, and the origin and significance of its marine fauna, remains unresolved. In order to explain the marine fauna, the southern Bouse Formation has been variously interpreted as representing a marine-estuarine (Smith 1970; Lucchitta 1979, 2001; Busing 1990; McDougall 2008; McDougall and Miranda Martinez 2014) or saline lacustrine (Spencer et al. 2008; Roskowski et al. 2010) depositional environment. Recently, McDougall and Miranda Martinez (2014) used foraminifer assemblages to provocatively bridge the gap between these end-member interpretations by suggesting that a key outcrop of the southern Bouse Formation in Blythe basin (Fig. 1) records a transition from marine-estuarine to saline-lacustrine conditions. To convert a basin filled with sea water to a saline lake would require exposing a previously submerged sill (i.e., tectonic uplift or lowering of sea level) in order to confine the lake. Although no detailed mechanism for isolating this postulated arm of the sea has been proposed, presumably it could have involved uplift along the developing San Andreas fault system and would likely have required tens to hundreds of thousands of years to complete. A transition from an open marine to estuarine environment due to uplift along the San Andreas fault system is another alternative. We tested the marine-estuary to saline-lake interpretation by analyzing the same Hart Mine Wash outcrop as McDougall and Miranda Martinez (2014) using ostracode faunas coupled with stable isotope ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$) data from both marl (inorganic calcite) and associated ostracode (biologic) calcite.

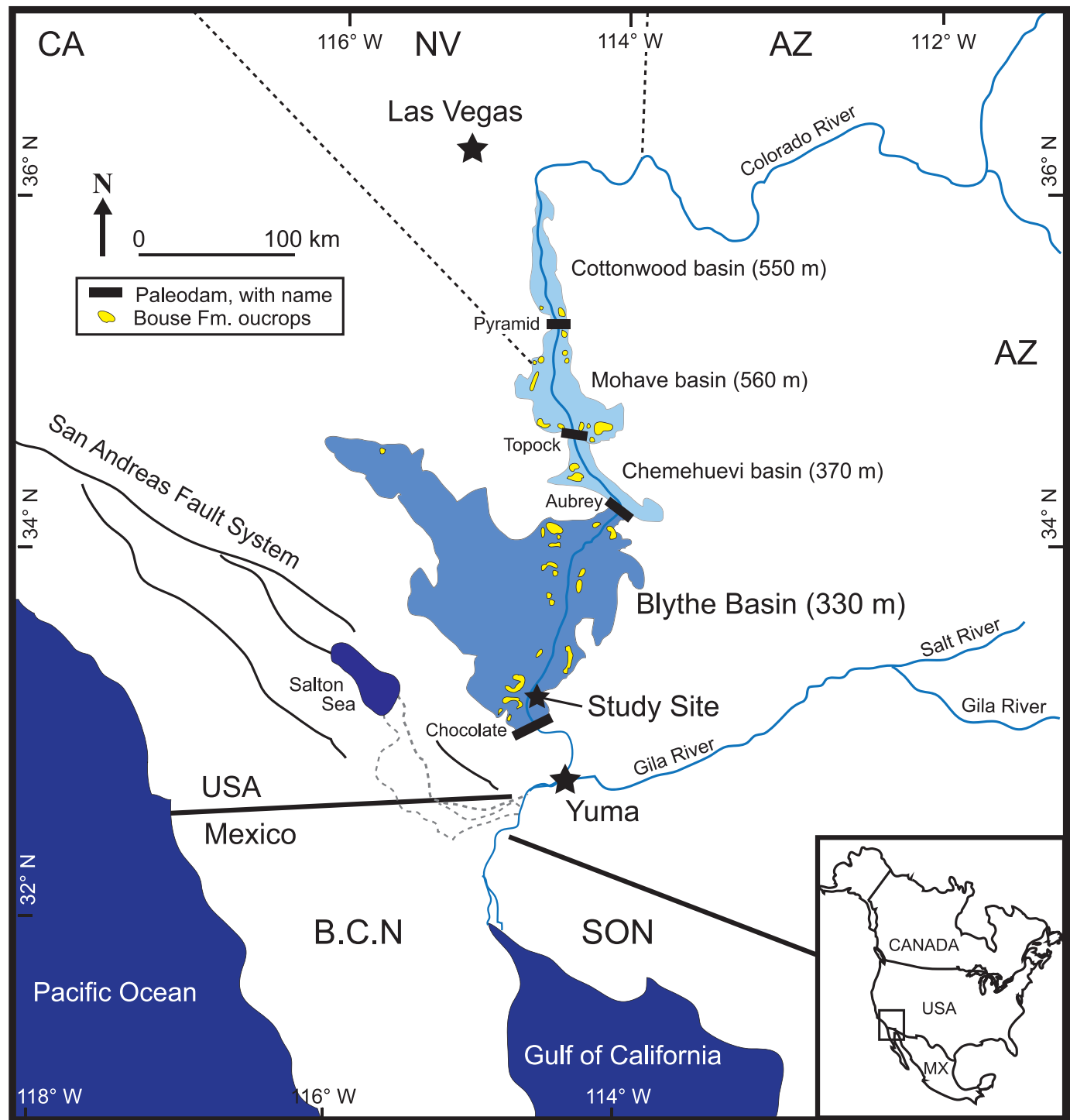


FIG. 1.—Regional map showing the relationship between the course of the modern lower Colorado River, Blythe basin (intermediate blue), basins to the north that contain the Bouse Formation (light blue), proposed paleodams (black bars with names), and the study location (solid black star). Numbers in parentheses denote highest elevation (masl) of Bouse Formation in the respective basins. Figure modified from image provided by R. Dorsey.

The premise behind our approach is that marine-estuarine environments and saline-lacustrine environments are typically defined by very different faunal associations and should have water masses with very different stable isotope characteristics. Marine faunas are rarely found in association with lacustrine faunas, primarily because seawater is characterized by a very stable and saline ($\sim 35,000 \text{ mg L}^{-1}$) sodium- and chloride-dominated chemistry, with

comparatively little calcium or carbonate alkalinity (CO_3^{2-} , HCO_3^{-}) (e.g., Forester and Brouwers 1985). Saline lakes, in contrast, encompass a wide range of chemical compositions (e.g., Eugster and Jones 1979) and are prone to seasonal and longer duration variability in their chemical compositions. The resulting faunal assemblages in marine and saline lacustrine environments are very different from each other and are largely mutually exclusive (e.g.,

Gray 1988). Similarly, seawater differs from lake water by having very stable and high $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values, which are faithfully recorded in both the inorganic and biologic carbonate precipitated from that water (e.g., Prokoph et al. 2008). In contrast, a saline lake fed by the ancestral Colorado River would have been supported by runoff and groundwater sourced from local mountain ranges and by seasonal snowmelt pulses sourced in the Rocky Mountains of Colorado, all of which would have had much lower and much more variable $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values than seawater (e.g., Guay et al. 2004). The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of saline lakes may be altered by evaporation and biologic processes to the extent that the resulting marl and biologic carbonates have marine-like $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values, although this is relatively rare (Talbot 1990). If the southern Bouse Formation in Blythe basin represents an estuarine environment, this should be evident by testing for a strong positive correlation between carbonate $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. This correlation is governed by the variable mixing of ^{18}O - and ^{13}C -enriched seawater and ^{18}O - and ^{13}C -depleted freshwater that characterizes estuarine environments (e.g., Ingram 1996).

We highlight a clear and sharp transition in the ostracode faunal content and the stable isotope composition of both marl and biologic carbonate in the southern Bouse Formation that was abrupt enough to confidently reject a marine or estuarine interpretation. We suggest that Blythe basin was filled with a hydrologically complex Colorado River-fed lake that was abruptly breached in an over-spilling event at Chocolate Mountain paleodam near its southern margin (Fig. 1). Our results have important implications for the evolution of the lower Colorado River corridor and dramatically improve our understanding of the events that took place just before the ancestral Colorado River was finally integrated with the Gulf of California.

BACKGROUND

The Colorado River is the seventh longest river corridor (~2,300 km) and has the seventh largest watershed (642,000 km²) in North America, despite draining a dominantly arid and desert landscape (Blinn and Poff 2005). From its headwaters in the Rocky Mountains of Colorado to its terminus at the Gulf of California, the Colorado River includes tributaries from seven western U.S. and two northern Mexican states (Blinn and Poff 2005) as it cuts across a tectonically active landscape, including the Colorado Plateau and the southern Basin and Range province. Belying its serene majesty, the evolution of the Colorado River system is the topic of intense debate (e.g., Lucchitta et al. 2001; Meek and Douglass 2001; Spencer and Pearthree 2001; Karlstrom et al. 2008; Polyak et al. 2008; Flowers and Farley 2012). Prior to about 15 Ma, rivers in the southwestern U.S. flowed largely to the north-east (Young and McKee 1978; Cather et al. 2012, and references therein). Between 15 and 6 Ma significant reorganization of the watersheds in the southwestern U.S. occurred in response to extension and subsidence in the Basin and Range province and to the opening of the Gulf of California (Potochnik and Faulds 1998; Potochnik 2001; Cather et al. 2012; Dickinson 2015). By about 6 Ma, the developing Colorado River appears for the first time west of the Grand Canyon (e.g., Lucchitta 1972; Karlstrom et al. 2008), and through a series of poorly understood integration events had reached the Gulf of California by about 5 Ma (Dorsey et al. 2007, 2011). The Colorado River now cuts through several prominent topographic features such as the Kaibab Plateau, the northern margin of the Colorado Plateau (i.e., the Grand Canyon), and several previously topographically closed extensional basins of the Basin and Range province located along its lower reaches (Fig. 1). Evidence for the geologic and geomorphic events that led to the developing Colorado River finally reaching the Gulf of California is poorly preserved or absent in canyon reaches, and this hinders our understanding of the evolution of this major continental river.

An enigmatic series of ~5 Ma-old carbonate and siliciclastic deposits that are discontinuously exposed in a series of adjacent basins along the lower Colorado River corridor (Fig. 1) preserves a potentially detailed record of the events

that occurred just prior to the Colorado River becoming fully integrated with the Gulf of California. These sediments are collectively called the Bouse Formation (Metzger 1968). The maximum elevations of Bouse Formation outcrops progressively decrease from ~550–560 masl (meters above sea level) in Cottonwood and Mohave basins, to ~370 masl in Chemehuevi basin, and to ~330 masl in Blythe basin (Fig. 1; Pearthree and House 2014). The Bouse Formation is also found as deep as about –200 masl in wells in Blythe basin (Metzger and Loeltz 1973; Metzger et al. 1973). From this point forward, the term “Bouse Formation” refers to all of the ~5 Ma-old limestone, marl and siliciclastic sediments exposed along the lower Colorado River corridor. The Bouse Formation preserved specifically in Blythe basin (Fig. 1) will be called the “southern Bouse Formation” and the Bouse Formation preserved specifically in the northern basins (Chemehuevi, Mohave, Cottonwood; Fig. 1) will be called the “northern Bouse Formation”.

Controversy over the Bouse Formation and its role in the integration of the ancestral Colorado River and the Gulf of California continues after nearly 50 years of study (Metzger 1968; Smith 1970; Lucchitta 1979; Buising 1990; Spencer et al. 2008; House et al. 2008; McDougall and Miranda Martinez 2014; Pearthree and House 2014; Howard et al. 2015; Crossey et al. 2015). The most recent interpretations, based on paleontologic, stratigraphic, and geochemical analyses (e.g., $^{87}\text{Sr}/^{86}\text{Sr}$ ratios), suggest that the northern Bouse Formation (Fig. 1) was deposited in a series of ancestral Colorado River-fed freshwater lakes, each with successively lower maximum elevations, that were sequentially filled and breached as the terminus of the river migrated southward towards the Gulf of California (Spencer and Patchett 1997; Poulson and John 2003; Spencer et al. 2008, 2013; Roskowski et al. 2010; Pearthree and House 2014; Crossey et al. 2015). Flood deposits downstream of the Pyramid paleodam (Fig. 1) provide direct evidence for catastrophic spillover and strongly support the chain-of-lakes interpretation for the northern basins (House et al. 2008; Pearthree and House 2014). All of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from the Bouse Formation are slightly higher and more variable than the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the modern Colorado River, however (Spencer et al. 2013), suggesting the Bouse Formation has a more complicated hydrologic origin than a simple river-fed lacustrine interpretation (e.g., Crossey et al. 2015).

The origin of the southern Bouse Formation is more controversial because it contains a moderately diverse and apparently marginal-marine fauna, including barnacles (Zullo and Buising 1989), benthic and planktic foraminifers (McDougall and Miranda Martinez 2014), and one species of fish that is restricted today to the Gulf of California (Todd 1976). The marginal-marine fauna is found in limestones and marls that have geochemical signatures (e.g., $^{87}\text{Sr}/^{86}\text{Sr}$ ratios) similar to the northern Bouse Formation, and that are permissive of a similar Colorado River-fed origin (Spencer and Patchett 1997; Spencer et al. 2008, 2013; Poulson and John 2003; Roskowski et al. 2010; Crossey et al. 2015). Current debate over the southern Bouse Formation focuses on the faunal assemblage, a few high trace element concentrations, and several carbonate stable isotope values that approach a marine-like value of ($\delta^{18}\text{O} \approx 0\text{‰}$ and $\delta^{13}\text{C} \approx 0\text{‰}$); indicators that could suggest a marine influence during deposition (e.g., Poulson and John 2003). The lack of a definitive marine $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in southern Bouse Formation carbonates could result from input of groundwater with high Sr concentrations and high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios that masked the marine signal (Crossey et al. 2015).

Anchoring the argument for a marine influence in the southern Bouse Formation is the rare occurrence of several genera of planktic foraminifers whose known distributions are largely restricted to normal marine environments (e.g., Kucera 2007). However, fresh- to brackish-water organisms are often found in association with the marginal-marine fauna (Smith 1970; Reynolds and Berry 2008). The freshwater organisms have been interpreted as being reworked from the surrounding watersheds into a more saline estuarine or marine environment (Turak 2000; Miller et al. 2014). It is unclear if the presumed marginal-marine fauna, especially planktic foraminifers, required an open

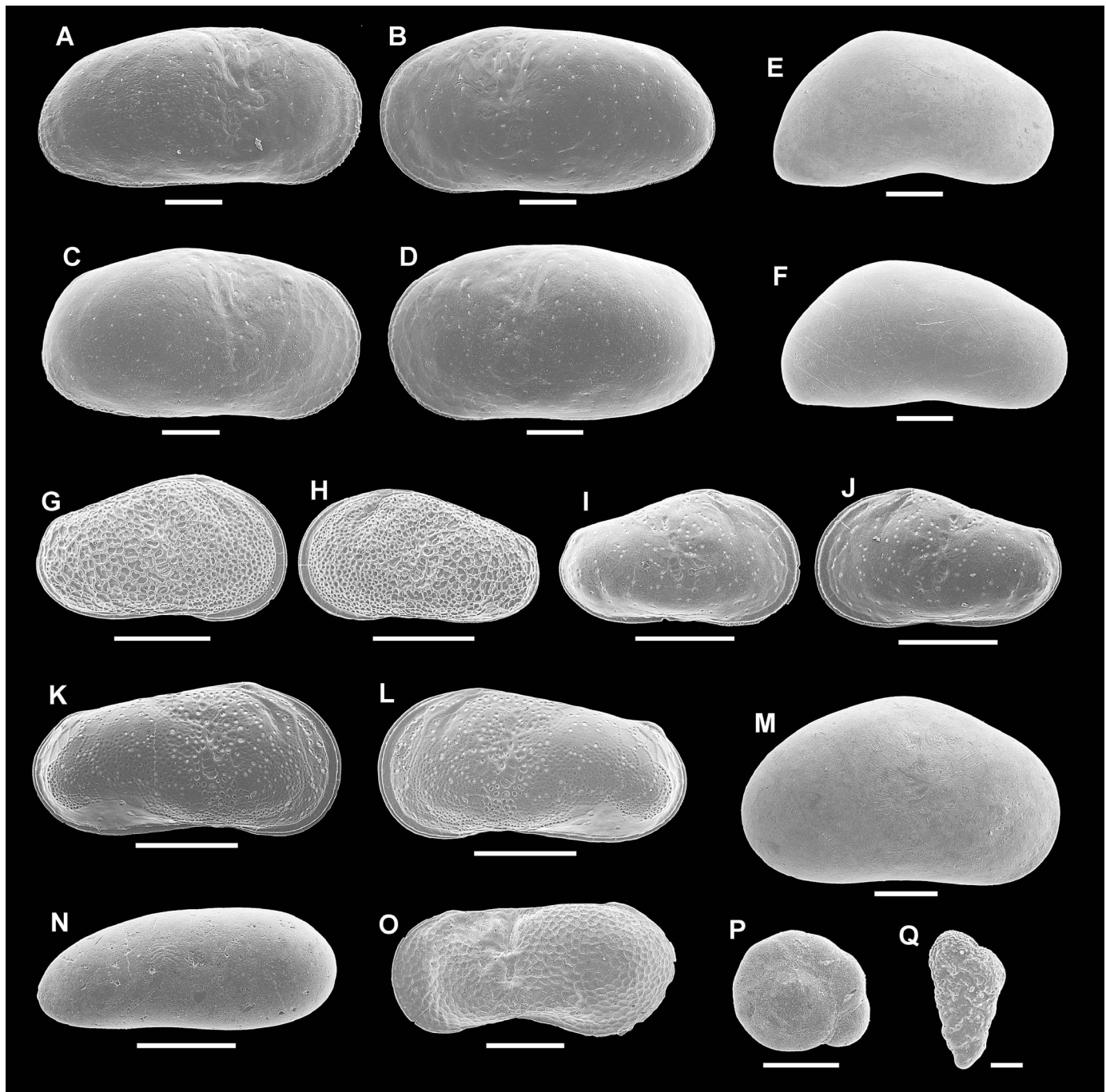


FIG. 2.—Scanning electron microscope images of ostracode valves from the southern Bouse Formation at Hart Mine Wash. All images are external lateral views. **A)** *Cyprideis* sp., male right valve, sample HM49. **B)** *Cyprideis* sp., male left valve, sample HM49. **C)** *Cyprideis* sp., female right valve, sample HM50. **D)** *Cyprideis* sp., female left valve, sample HM50. **E)** *Candona* sp. A, female right valve, sample HM53. **F)** *Candona* sp. B, female right valve, sample HM54. **G)** *Cytheromorpha* sp. A, female right valve, sample HM49 (common). **H)** *Cytheromorpha* sp. A, female left valve, sample HM49 (common). **I)** *Cytheromorpha* sp. B, female right valve, sample HM49 (rare). **J)** *Cytheromorpha* sp. B, female left valve, sample HM49 (rare). **K)** *Cytheromorpha* sp. B, male right valve, sample HM45 (rare). **L)** *Cytheromorpha* sp. B, male left valve, sample HM45 (rare). **M)** *Heterocypris* sp., right valve, sample HM53. **N)** *Darwinula* cf. *D. stvensoni*, left valve, sample HM55. **O)** *Limnocythere* sp., male left valve, sample HM55. **P)** Example of “spiraled” foraminifer, sample HM55. **Q)** Example of biserial foraminifer, sample HM55. See online Supplemental Data for sample information. Images were modified for contrast and clarity using Photoshop CS6. Scale bars = 200 μm for views A–P; scale bar = 50 μm for view Q.

connection between Blythe basin and the early Gulf of California (e.g., McDougall 2008; McDougall and Miranda Martinez 2014) or if the assemblage could have survived in a saline lake environment after being introduced by migrating birds (Spencer et al. 2008, 2013).

Determining whether the southern Bouse Formation was deposited in a marine embayment or estuary (Lucchitta et al. 2001; McDougall and Miranda Martinez 2014) or a saline lake (Spencer et al. 2013) has direct bearing on our understanding of regional post-Miocene tectonics, most notably because

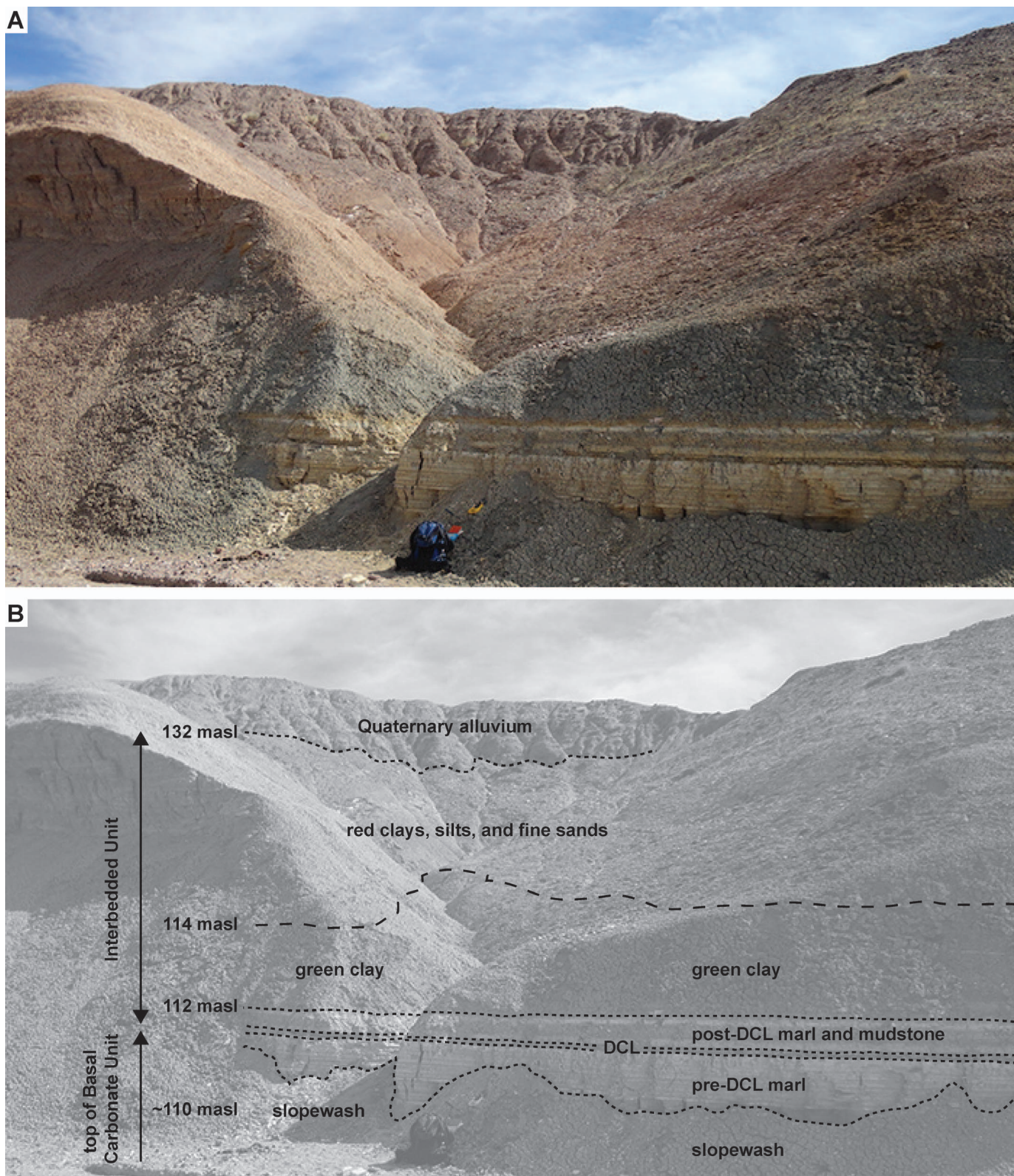


FIG. 3.—Photographs of the southern Bouse Formation outcrop featured in this study, with backpack for scale. **A)** Unaltered photograph. The elevation of the backpack is about 110 masl. **B)** Same photograph with important sedimentary packages highlighted. See text for discussion. The elevation of the Quaternary alluvium contact near the top of the photo is about 132 masl. Short dashed lines=clear contact; long dashed line=approximate contact.

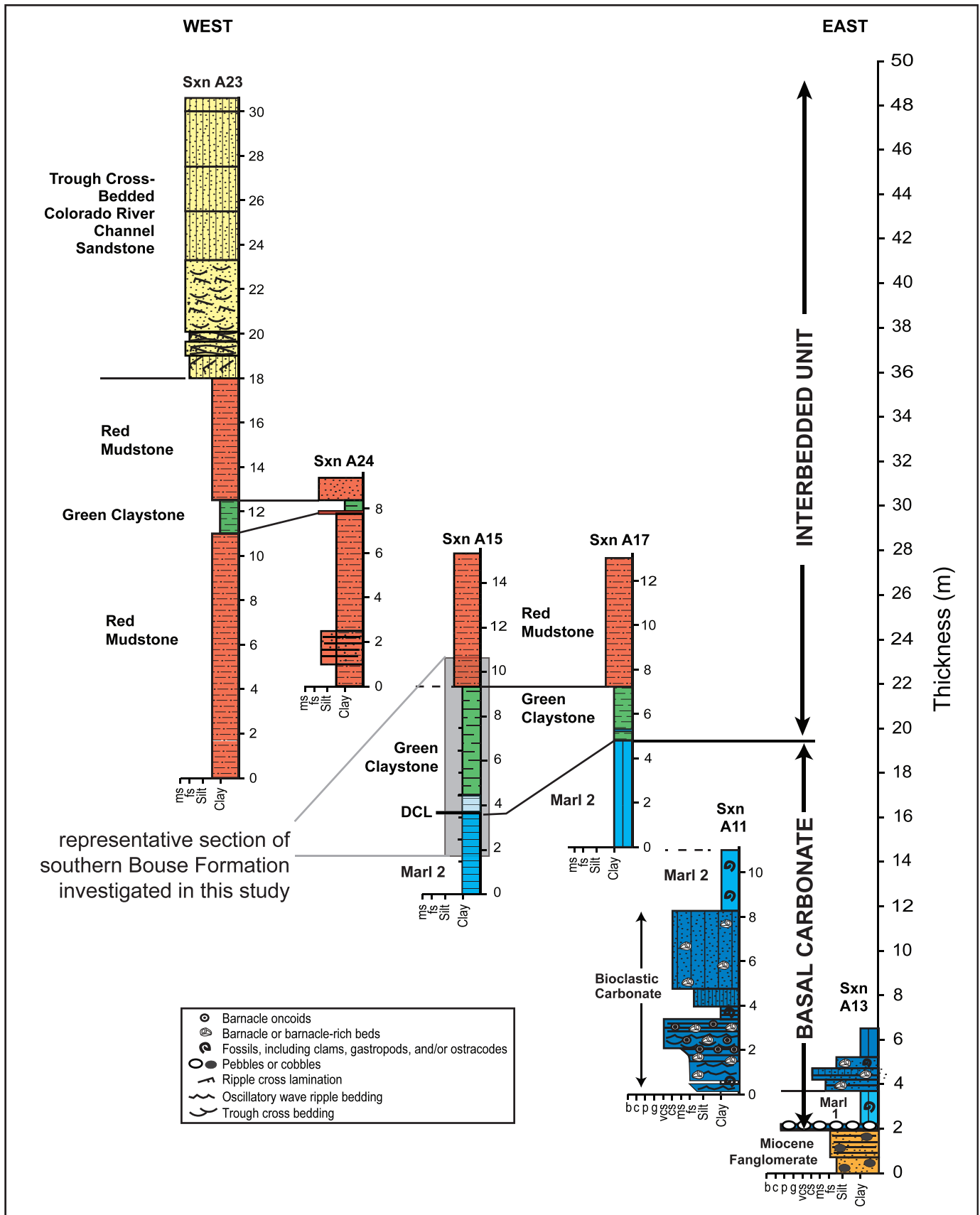


FIG. 4.—Stratigraphic sections of the southern Bouse Formation exposed in Hart Mine Wash. Gray box highlights the stratigraphic position of the outcrop highlighted in this study. Compiled from Homan (2014).

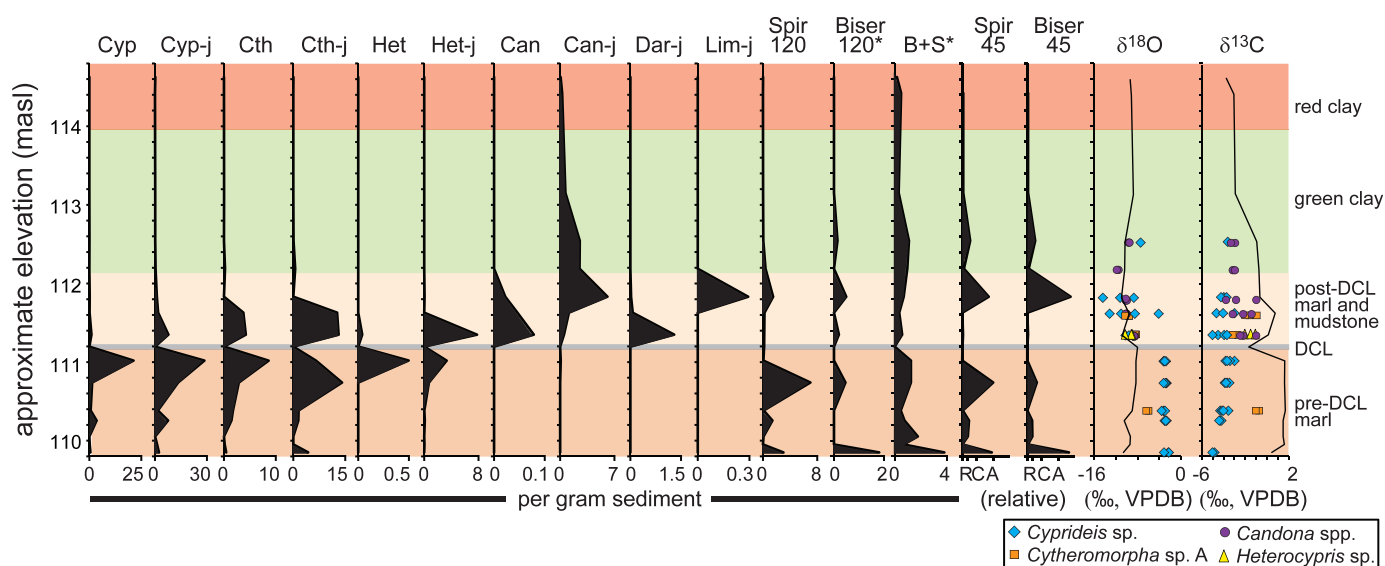


FIG. 5.—Microfaunal and sediment stable isotope data from Hart Mine Wash. Abbreviations: Cyp=adult *Cyprideis* sp.; Cyp-j=juvenile *Cyprideis* sp.; Cth=adult *Cytheromorpha* spp.; Cth-j=juvenile *Cytheromorpha* spp.; Het=adult *Heterocypris* sp.; Het-j=juvenile *Heterocypris* sp.; Can=adult *Candona* spp.; Can-j=juvenile *Candona* spp.; Dar-j=juvenile and fragmented *Darwinula* cf *D. stevensoni*; Lim-j=juvenile and fragmented *Limnocythere* sp.; Spir 120=spiraled foraminifers in 120 μ m; Biser 120*=biserial foraminifers in 120 μ m plotted as square root of concentration; B+S*=combined bone and scale plotted as square root of concentration; Spir 45=spiraled foraminifers in 45–120 μ m; Biser 45=biserial foraminifers in 45–120 μ m; $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values from < 45 μ m sediment fraction. Relative foraminifer abundances, as specimens observed in a concentrated aliquot of 45–120 μ m sediment residue; R=rare (20); C=common (60); A=abundant (120).

330 m of uplift would be required to explain the highest elevation of the southern Bouse Formation outcrops if these deposits were originally deposited at sea level. The southern Bouse Formation may also provide important context for how unexpected faunal associations (e.g., mixtures of marine and freshwater faunas) that have been found at other locations around the world might be interpreted (e.g., Keller et al. 2009; Linhares et al. 2011; Ruskin et al. 2011; Gross et al. 2013). Efforts to determine the depositional environment of the southern Bouse Formation have been hindered by the relatively small number of samples that have been placed in a detailed stratigraphic framework and analyzed for both faunal and isotopic content. This paper presents the results of the first continuous high-resolution study of a key outcrop that significantly advances our understanding of the southern Bouse Formation.

Our study site is an outcrop of southern Bouse Formation in Hart Mine Wash (Fig. 1; 33.2907° N, 114.6364° W) that has been interpreted, based on foraminifer assemblages, to record a transition from marine-estuarine to saline-lake conditions (McDougall and Miranda Martinez 2014). This location may offer critical new information that could support or refute the competing marine and saline lake interpretations. We acknowledge that a portion of the marine-estuary to saline-lake hypothesis is based on the analysis of foraminifer faunas in the subsurface of Blythe basin (e.g., Smith 1970; McDougall and Miranda Martinez 2014); sediments that at face value may not be exposed in surface outcrops, and sediments that we have not analyzed. However, the foraminifer assemblage at Hart Mine Wash over the interval 110–126 masl has been described as “shallow inner neritic” (McDougall and Miranda Martinez 2014, p. 846), or representing marine water depths from < 50 m deep to the low tide mark. Additional details about the significance of the planktic foraminifers include “...hyposaline tolerant species are absent in the Bouse Formation in Blythe basin, whereas species common in normal marine salinities like *Streptochilus* are present...” (McDougall and Miranda Martinez 2014, p. 849), and “The portion of the Bouse Formation that contains evidence of marine deposition is today at elevations of 123 masl and lower. Planktic foraminifers in the Hart Mine Wash section were found at an elevation of 112–123 masl....” (McDougall and Miranda Martinez 2014; p. 859). We interpret these statements

to imply that the presence of planktic foraminifers (e.g., *Streptochilus*) in sediments below 126 masl at this particular outcrop in Hart Mine Wash indicate deposition under marine conditions. Here we test the marine-estuary to saline-lacustrine interpretation through a high-resolution (20–40 cm) analysis of microfaunal elements (ostracodes, fish remains, foraminifers) and stable isotope ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$) values (VPDB) from paired micrite (< 45 μ m; $\delta^{18}\text{O}_{\text{SED}}$, $\delta^{13}\text{C}_{\text{SED}}$) and ostracode valves ($\delta^{18}\text{O}_{\text{OST}}$, $\delta^{13}\text{C}_{\text{OST}}$) from this outcrop.

METHODS

Microfaunal Analysis

Sediment samples were disaggregated by soaking in a dilute sodium hexametaphosphate and sodium bicarbonate solution for up to one week. Resistant samples were subjected to no more than three freeze-thaw cycles. Sediment slurries were washed over 45 μ m sieves using hot distilled water. Microfauna were counted from the > 120 μ m size fractions under a binocular microscope. Ostracode carapaces were counted as two valves. Small aliquots of the 45–120 μ m size fractions were qualitatively scanned for foraminifer abundances.

Scanning electron microscope (SEM) images of representative ostracode valves and foraminifers are presented in Figure 2. Ostracode valves and foraminifer tests were coated with carbon and imaged using a Hitachi 3400N Variable Pressure SEM at the University of Arizona’s LaserChron SEM Facility.

Stable Isotope Analysis

Small subsamples of sediment were disaggregated in distilled water and washed over 45 μ m sieves to remove biogenic carbonate (foraminifers, ostracodes). The < 45 μ m fractions were retained in their entirety and dried overnight at 40°C. The dried residues were gently powdered using a porcelain mortar and pestle. The best preserved ostracode valves (*Cyprideis* sp., *Cytheromorpha* sp. A., *Candona* spp., *Heterocypris* sp.; Fig. 2) were picked from larger sediment subsamples that were disaggregated using only distilled water. Detritus was manually removed from the valves using

fine paintbrushes and surgical needles. Visible secondary carbonate was removed using surgical needles or by quickly brushing the valves with dilute (0.5 M) hydrochloric acid followed by a thorough rinsing with distilled water.

The sediment powders and ostracode valves were analyzed for their $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values at the University of Arizona's Environmental Isotope Laboratory using an automated KIEL-III carbonate preparation device coupled to a Finnigan MAT 252 gas-ratio mass spectrometer. The samples were reacted with dehydrated phosphoric acid under vacuum at 70°C. The isotope ratio measurement is calibrated based on repeated measurements of NBS-18 and NBS-19 standards. The stable isotope results are reported in standard delta (δ) notation where: $\delta\text{‰} = [(R_{\text{sample}}/R_{\text{std}}) - 1] \times 10^3$; and R = ratio of $^{18}\text{O}:^{16}\text{O}$ or $^{13}\text{C}:^{12}\text{C}$. R_{std} refers to the standard Vienna Pee Dee belemnite (VPDB). Precision for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ measurements are $\pm 0.1\text{‰}$ and $\pm 0.08\text{‰}$ (1 σ), respectively.

EVIDENCE FOR DRAMATIC HYDROCHEMICAL CHANGE DURING DEPOSITION OF THE SOUTHERN BOUSE FORMATION

The study section comprises roughly 2 m of iron oxide-stained marl (110–112 masl), of which the upper ~1 m is characterized by a transitional zone of alternating marl and mudstone beds (Fig. 3). The marl and marl-mudstone is overlain by about 2 m of dense green claystone (112–114 masl), followed by nearly 20 m of dense reddish claystone, siltstone, and fine sandstone (114–132 m asl; Fig. 3). A 3-cm-thick distinctive claystone layer (DCL) occurs at about 111 masl, roughly 1 m below the top of the alternating marl and mudstone beds (Fig. 3). The marl below the DCL corresponds to the top of the Bouse “basal limestone” unit and the overlying interval of alternating marl-mudstone beds and the green and red claystones are part of the predominantly siliciclastic “interbedded” unit (Fig. 4) of Metzger (1968). Metzger (1968) noted that the basal limestone unit contains marly intervals. Therefore, for simplicity, the basal limestone and marl horizons are included together here as a “basal carbonate” unit (Fig. 4). Placing the contact between the basal carbonate and the interbedded units at the DCL follows the recent work of Homan (2014). For this study we assume that the < 45 μm sediment fraction is dominated by calcite that was formed in epilimnetic water and the ostracode valves were calcified *in situ* in the benthic zone. The foraminifers we report were not identified, but the “spiraled” types (Fig. 2) include benthic genera like *Ammonia* and *Rosalina* and the “biserial” types (Fig. 2) include the benthic genus *Bolivina* and the planktic genus *Streptochilus*. Previous work concluded that *Streptochilus* comprise 99% of the biserial foraminifers at this particular exposure of southern Bouse Formation in Hart Mine Wash (McDougall and Miranda Martinez 2014). Critically, *Streptochilus* is planktic, and as such, we interpret it to represent normal marine or near-normal marine conditions (e.g., Kucera 2007; McDougall and Miranda Martinez 2014).

The lowest marl (~110 masl) that is exposed below the DCL contains few ostracodes and has the highest concentration of biserial foraminifers and fish remains (Fig. 5). Concentrations of the ostracode genera *Cyprideis* sp., *Cytheromorpha* sp., *Heterocypris* sp. and spiraled foraminifers peak at about 111 masl, just below the DCL (Fig. 5). The pre-DCL marl has $\delta^{18}\text{O}_{\text{SED}}$ values that increase from about -10.5 to -8‰ , and the $\delta^{13}\text{C}_{\text{SED}}$ values are constant at about $+2\text{‰}$ (Fig. 5). *Cyprideis* sp. and *Cytheromorpha* sp. A valves (Fig. 2) from this interval have stable $\delta^{18}\text{O}_{\text{OST}}$ and $\delta^{13}\text{C}_{\text{OST}}$ values of about $-3 \pm 1\text{‰}$ and $-4 \pm 1\text{‰}$, respectively (Figs. 5, 6). The DCL itself is devoid of fossils and has $\delta^{18}\text{O}_{\text{SED}}$ and $\delta^{13}\text{C}_{\text{SED}}$ values of -8‰ and -1.7‰ , respectively (Fig. 5).

Two samples from immediately above the DCL (~111.5 masl) contain low concentrations of *Cyprideis* sp., high concentrations of *Cytheromorpha* sp., and high concentrations of juvenile *Heterocypris* sp. The continental ostracodes *Darwinula* cf. *D. stvensoni*, *Candona* spp., and *Limnocythere* sp. (Fig. 2) abruptly appear at this level (Fig. 5). Spiraled foraminifers are

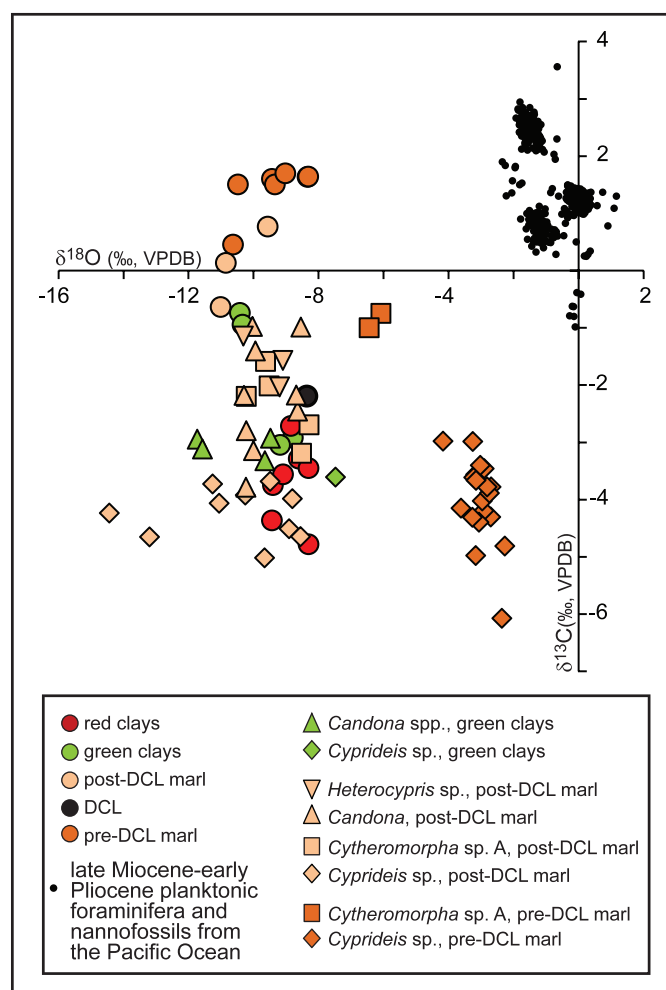


FIG. 6.—Cross-plot of $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values from ostracode valves and < 45 μm sediment fraction from pre-DCL marl (dark orange symbols), the DCL (black circle), post-DCL marl and mudstone (light orange symbols), and green and red clays (green and red symbols) at Hart Mine Wash. The small black dots are $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values from planktic foraminifera and nannofossils in Pacific Ocean sediments that span 4.2 to 6.4 Ma as compiled from Douglas and Savin (1971, 1975, 1978), Keigwin (1979, 1982), Barrera et al. (1985), Cannariato and Ravelo (1997), and Prokoph et al. (2008).

common, but fish remains and biserial foraminifers are rare (Fig. 5). The $\delta^{18}\text{O}_{\text{SED}}$ values from these two samples decrease to -10.9‰ and -9.6‰ , and $\delta^{13}\text{C}_{\text{SED}}$ values also decrease to $+0.1\text{‰}$ and $+0.8\text{‰}$ (Fig. 5). Valves from four genera of ostracodes, including *Cyprideis* sp. and *Cytheromorpha* sp. A., have $\delta^{18}\text{O}_{\text{OST}}$ and $\delta^{13}\text{C}_{\text{OST}}$ values of about $-9.5 \pm 2\text{‰}$ and $-3 \pm 1\text{‰}$, respectively (Figs. 5, 6). One sample from near the top of the marl-mudstone horizon (112 masl) contains relatively few valves of *Cyprideis* sp. and *Cytheromorpha* sp., but juvenile *Candona* spp. and both biserial and spiraled foraminifers are relatively abundant (Fig. 5). The $\delta^{18}\text{O}_{\text{SED}}$ and $\delta^{13}\text{C}_{\text{SED}}$ values from this sample are -11‰ and -0.6‰ , respectively (Figs. 5, 6). *Cyprideis* sp. and *Candona* spp. valves from this sample have $\delta^{18}\text{O}_{\text{OST}}$ and $\delta^{13}\text{C}_{\text{OST}}$ values of $-11 \pm 2\text{‰}$ and $-3 \pm 1\text{‰}$, respectively (Figs. 5, 6).

Green and red claystones above the marly sediments (> 112 masl; Figs. 3, 4) are largely devoid of fossils. Faunal remains in the lower green claystone are dominated by juveniles and fragments of *Candona* spp. and low concentrations of *Cyprideis* sp. and *Cytheromorpha* sp. A valves. Foraminifers are rare, except at 112.5 masl where both biserial and spiraled foraminifers (Fig. 2) are more common (Fig. 5). The $\delta^{18}\text{O}_{\text{SED}}$ and $\delta^{13}\text{C}_{\text{SED}}$ values from the green claystone

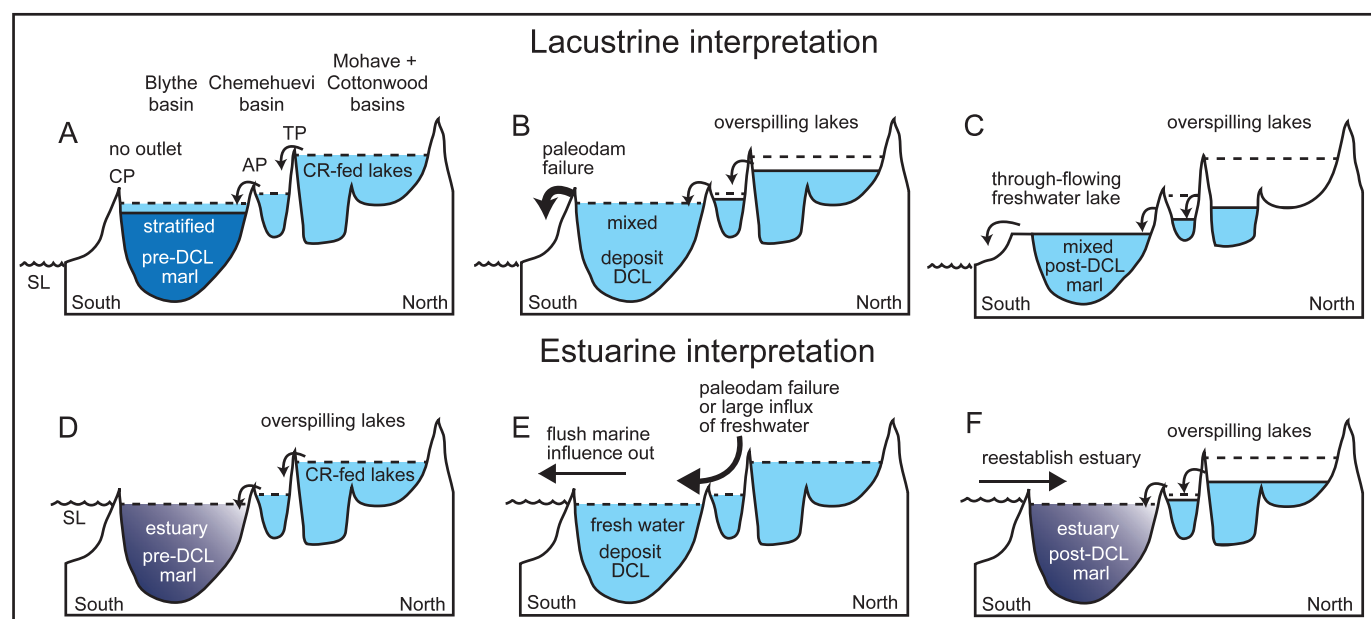


FIG. 7.—Cartoon schematics highlighting two end-member interpretations of the possible sequence of events that produced the pre- and post-DCL sediments at Hart Mine Wash. A–C) Lacustrine interpretation featuring an overspilling chain of lakes (A), an eventual sill breach (B), and establishment of through-flowing lake (C). D–F) Estuarine interpretation featuring a chain of overspilling lakes to the north of a Blythe basin estuary (D), a flood event (E), and the re-establishment of estuarine conditions in Blythe basin after passage of the flood event (F). Abbreviations: CP=Chocolate Mountain paleodam; AP=Aubrey paleodam; TP=Topock paleodam; CR=Colorado River; SL=sea level. See Figure 1 for a map view of basin and paleodam locations.

are about $-10 \pm 2\text{‰}$ and $-2 \pm 1\text{‰}$, respectively (Figs. 5, 6). The red claystones produced only a few isolated valves of *Cyprideis* sp., and have $\delta^{18}\text{O}_{\text{SED}}$ and $\delta^{13}\text{C}_{\text{SED}}$ values of about $-9 \pm 0.5\text{‰}$ and $-4 \pm 1\text{‰}$, respectively (Figs. 5, 6). Because of the lack of fossils in the red claystone, we focus our discussion on the marl and green claystone units. (All microfaunal concentrations, $\delta^{18}\text{O}_{\text{SED}}$, $\delta^{13}\text{C}_{\text{SED}}$, $\delta^{18}\text{O}_{\text{OST}}$ and $\delta^{13}\text{C}_{\text{OST}}$ values are available in online Supplementary Data Tables 1 and 2)

DISCUSSION AND CONCLUSIONS

Three key findings are evident in our study. First, a sharp change in the stable isotope characteristics and ostracode faunal assemblages occurs at the DCL. Specifically, $\delta^{18}\text{O}_{\text{OST}}$ values from below the DCL are consistently about 6‰ higher than $\delta^{18}\text{O}_{\text{SED}}$, whereas above the DCL the two values are nearly identical (Fig. 5). A coincident and similarly abrupt shift in the ostracode assemblages occurs across the DCL as well. Simple explanations that rely on ostracode vital effects and differences in the temperature of calcification between the epilimnion and the benthic zone cannot explain why the 6‰ offset should disappear after deposition of the DCL. The best explanation is that the pre-DCL marl was deposited in an isotopically stratified body of water, in which benthic ostracodes calcified in bottom water that was more ^{18}O -enriched than the epilimnion. In contrast, post-DCL marl was deposited in an isotopically well-mixed body of water. Also, $\delta^{18}\text{O}_{\text{SED}}$ and $\delta^{13}\text{C}_{\text{SED}}$ values decrease by several per mil above the DCL (Figs. 3, 4), which is consistent with a change in the residence time of the water body. The appearance of freshwater continental ostracodes is coincident with the isotopic shift, indicating a simultaneous reduction in salinity. The process that eliminated the isotopic stratification and initiated freshening occurred during, and persisted after deposition of the DCL. The duration of time represented by the DCL is unknown, but reasonable estimates of sedimentation rate ($0.1\text{--}10\text{ mm yr}^{-1}$; Cohen 2003) suggest a timeframe on the order of 3 to 300 years.

The second key finding is that biserial foraminifers, which we presume are dominantly planktic *Streptochilus* and an indicator of normal or near-normal

marine salinities, are present in pre- and post-DCL marl and in the lower green claystone (Fig. 5; McDougall and Miranda Martinez 2014). The isotopic stratification and $\delta^{18}\text{O}_{\text{OST}}$ values near -2‰ that define the pre-DCL marl could suggest deposition under stratified estuarine conditions, in which case *Streptochilus* would support a marine influence. *Streptochilus* tests in post-DCL marl and lower green claystone, however, are associated with fresh- and brackish water ostracodes and low $\delta^{18}\text{O}_{\text{SED}}$ and $\delta^{18}\text{O}_{\text{OST}}$ values that argue against a marine influence. Unless reworked from older sediments, our results challenge the assertion that *Streptochilus* definitively represents normal marine salinities. Neither the foraminifers nor the ostracodes show any obvious signs of reworking (e.g., broken, dissolved, encrusted with sediment, etc.), suggesting they coexisted in this same depositional environment. Additionally, we expect that carbonate formed at normal marine salinity would have marine-like $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. We compiled $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values in late Miocene-early Pliocene planktic foraminifera from the Pacific Ocean to represent $\sim 5\text{ Ma}$ marine carbonate (Fig. 6). Foraminifer $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values were selected based on available age determinations and were confined to written descriptions such as “early Pliocene” or “late Miocene”, or, when numerical ages were provided, were confined to the interval from 4.2 to 6.4 Ma. The $\delta^{18}\text{O}_{\text{SED}}$ and $\delta^{18}\text{O}_{\text{OST}}$ values from pre- and post-DCL sediments at Hart Mine Wash plot far away from our marine carbonate analog (Fig. 6).

The third key finding challenges the hypothesis that the freshwater organisms of the southern Bouse Formation were washed into a saline marine or estuarine environment. Although both *Cyprideis* sp. and *Cytheromorpha* spp. are commonly considered marginal-marine ostracodes, both have been found in brackish to almost freshwater lacustrine habitats (e.g., Oviatt 1988). Their presence in the southern Bouse Formation could suggest, but does not require, a marine influence. *Candona* spp. are mildly brackish to freshwater continental ostracodes that to our knowledge do not live in normal marine environments. Freshwater ostracodes can be found in estuaries but they are restricted to the freshwater and mildly brackish (e.g., $\leq 5\text{ ppt}$) portions of the estuary (e.g., Elliot et al. 1966; Forester and Brouwers 1985), at salinities that would be inhospitable to planktic foraminifers (e.g., Kucera 2007). The similarly low

$\delta^{18}\text{O}_{\text{OST}}$ values from *Cyprideis* sp. and *Candona* spp. valves above the DCL (Figs. 5, 6) indicate these ostracodes were living and calcifying together in an environment that was likely approaching freshwater conditions. The similarity of the *Candona* spp. $\delta^{18}\text{O}_{\text{OST}}$ values to those from brackish water ostracodes and to the surrounding marl (Fig. 6) indicates that the *Candona* spp. valves above the DCL were not washed in to a marine or estuarine environment as previously proposed (e.g., Turak 2000; Miller et al. 2014).

Lacustrine or Marine/Estuarine Environment?

The shift from an isotopically stratified water body to a less saline, well-mixed water body provides new insight into the history of the Blythe basin water mass. The isotopic stratification seen in the pre-DCL marl makes a marine interpretation unlikely. If we suppose the stratified water column existed in a marine or estuarine environment, the marine water was absent or its volume was significantly reduced during deposition of the post-DCL marl and green claystone. Reducing the volume of marine water would require a drop in sea level, uplift of Blythe basin, a substantial increase in water discharge of the ancestral Colorado River that pushed out a salt-water wedge, or some combination of all three options. Our data suggest that the isotopic change occurred over a decadal to century time scale (deposition of the 3-cm-thick DCL) which rules out slower millennial-scale processes such as changes in sea level or gradual crustal uplift or tilting. Neither the micrite nor the ostracode calcite has $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values that fall along a linear isotopic mixing line (Fig. 6) as might be expected when river water and sea water mixes in an estuarine or marginal marine environment (e.g., Reinhardt et al. 2003; Sampei et al. 2005). Thus, an estuarine interpretation (e.g., Smith 1970; McDougall and Miranda Martinez 2014) is not supported by our results.

We interpret the DCL as recording catastrophic failure of the Chocolate Mountain paleodam that formed the southern margin of a saline, stratified lake in Blythe basin (Fig. 7). Our interpretation is based in part on an analogous event in the Lake Bonneville basin of Utah-Idaho. Lake Bonneville catastrophically breached its confining sill and within a year lake level fell nearly 100 m. The Bonneville flood resulted in a distinctive sediment layer, bracketed by pre- and post-flood marl, which is easily identifiable in Lake Bonneville sediments (Oviatt 1987; Oviatt et al. 1994). By analogy, we suggest that the pre-DCL marl at our study location was deposited in a terminal, isotopically stratified saline lake fed by the ancestral Colorado River. The lake eventually breached the paleodam in the Chocolate Mountains (Figs. 1, 7) and the subsequent rapid drop in lake level remobilized fine-grained sediments from the surrounding basin margins and the delta at the northern end of Blythe basin and redeposited them as the DCL. Post-DCL sediments represent alternating carbonate and mudstone deposition in a hydrologically open lake that was isotopically well-mixed and less saline than the previous terminal lake. The persistent presence of the planktic foraminifer *Streptochilus* at Hart Mine Wash is puzzling and remains unresolved. Possible explanations for the marine faunal components of the southern Bouse Formation will be the focus of further study.

Alternatively, if the southern Bouse Formation accumulated in a marine or stratified estuarine environment, then a large and sudden influx of fresh water to Blythe basin such as an abrupt lowering of the Topock paleodam or a river capture event to the north might have temporarily flushed the estuary (Fig. 7). While possible, this interpretation seems unlikely to explain the persistent change in isotope values because the flood pulse would have passed through the estuary in a very short time. Dense seawater undoubtedly would have reentered the basin and stratified estuarine conditions would have been reestablished (Fig. 7). We do not favor this interpretation because the post-DCL sediments do not contain any evidence for a return to marine or stratified estuarine conditions.

When coupled with the geomorphic evidence for catastrophic spillover to the north (House et al. 2005; Pearthree and House 2015), our results support

the interpretation that the southern Bouse Formation was deposited in a lacustrine rather than a marine or estuarine setting.

ACKNOWLEDGMENTS

This research was partly funded by the U.S. Geological Survey, by the Geological Society of America's Kerry Kelts Research Award (JB), and by the University of Arizona's Maxwell Short Scholarship (JB). A portion of this research was supported by funds from NSF EAR-1545998. We thank Jacob Favela for assistance with the SEM imaging. The University of Arizona's LaserChron SEM Facility is supported by funds from NSF EAR-0929777. We appreciate conversations with Jon Spencer, Kristin McDougall, and Kyle House, and their willingness to share their ideas about the Bouse Formation. We appreciate the comments of two anonymous reviewers that helped improve the clarity of the manuscript.

SUPPLEMENTAL MATERIAL

Data are available from the PALAIOS Data Archive: <http://www.sepm.org/pages.aspx?pageid=332>.

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Received 4 June 2015; accepted 5 January 2016.