Freshwater plumes and brackish lakes: Integrated microfossil and O-C-Sr isotopic evidence from the late Miocene and early Pliocene Bouse Formation (California-Arizona) supports a lake overflow model for the integration of the lower Colorado River corridor

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ABSTRACT

Uncertainty over the depositional environment of the late Miocene and early Pliocene Bouse Formation hinders our understanding of the evolution of the lower Colorado River corridor. Competing marine and lacustrine models for the origin of the southern Bouse Formation remain extremely difficult to reconcile after nearly 60 yr of study. This paper compares new microfossil data, inorganic and biologic carbonate δ¹⁸O and δ¹³C values, and Sr isotopes from northern and southern outcrops of the Bouse Formation. The lacustrine northern Bouse Formation and the contested southern Bouse Formation share a core Cyprideis (mixed marginal marine), Limnocythere (continental), and Candona (continental) ostracode assemblage, indicating similar environmental conditions. Micrite and ostracode valves from both areas yield nearly identical δ¹⁸O and δ¹³C values, suggesting similar origins. Ostracode valves from both areas document a large and abrupt shift from high δ¹⁸O values (–2‰) to low values (–10‰), consistent with fill-and-spill lacustrine origins. Tests of the planktic foraminifer Streptochilus from a southern outcrop yielded δ¹⁸O and δ¹³C values that are nearly identical to benthic ostracode δ¹⁸O and δ¹³C values. Recognition of benthic Streptochilus weakens a categorically marine interpretation for the southern Bouse Formation. Barnacle shell fragments at a key outcrop of the southern Bouse Formation that preserves sigmoidal bedding with possible spring-to-neap tidal bundling yielded low δ¹⁸O values (–8‰ ± 1‰) that are incompatible with calcification in seawater. The δ¹⁸O/δ¹³C ratios from co-occurring fish bones (0.71104) and ostracode valves (0.71100) and the surrounding micrite (0.71088) reveal an isotopically complex lacustrine depositional environment for the southern Bouse Formation. A model invoking freshwater plumes from the lower Colorado River into either a terminal or a tidally influenced, mildly brackish lake followed by an abrupt transition to a freshwater lake provides a comprehensive and internally consistent explanation for the microfossil and isotopic complexities observed in this southern Bouse Formation data set. A freshwater plume model is entirely consistent with fill-and-spill models for the downstream integration of the early Colorado River.

INTRODUCTION

The evolution of the Colorado River is a crucial link in understanding the late Cenozoic tectonic evolution of the southwestern United States (e.g., Blackwelder, 1934). Large portions of the Colorado River watershed were at sea level ~65 m.y. ago, but since then, the Rocky Mountains and the Colorado Plateau have been uplifted ~3 and 1.5 km, respectively (Pederson et al., 2002; Karlstrom et al., 2012). Late Cenozoic extension in the western United States lowered the Basin and Range Province, and strike-slip displacement along the San Andreas fault opened the Gulf of California (e.g., Karig and Jensky, 1972; Dickinson, 2002). The late Cenozoic saw a significant reorganization of watersheds in the southwestern United States (e.g., Cather et al., 2012) in response to a change in regional base level. Exotic Colorado River gravels appear beyond the western end of the modern Grand Canyon after 6 Ma (Lucchitta, 1972; Spencer et al., 2001), providing clear evidence that an early Colorado River had reached the southwestern United States. The river then integrated a southward course across the chaotic landscape of the southeastern Basin and Range Province (Faulds et al., 2008) before eventually reaching the Gulf of California ca. 5 Ma (Dorsey et al., 2007, 2011; Crow et al., 2017). Understanding the evolution of the Colorado River system is of critical importance when considering the timing and rates of uplift, faulting, and subsidence within its watershed (Lucchitta, 1979; Karlstrom et al., 2007); it is the common thread that ties much of the late Cenozoic tectonic history of the southwestern United States together. How the Colorado River system evolved is the topic of intense and long-standing debate (Ranney, 2014), particularly the sequence of...
events that finally led to the lower Colorado River reaching the Gulf of California (e.g., Spencer and Patchett, 1997; McDougall and Miranda Martinez, 2014; Pearthree and House, 2014; Dorsey et al., 2018). A highly contested record of the processes that integrated the lower Colorado River corridor is preserved as the Bouse Formation.

In this paper, we present microfossil assemblages, δ¹⁸O-δ¹³C values from inorganic calcite and biologic calcite (ostreocodes, barnacles, foraminifers), and ⁸⁷Sr/⁸⁶Sr ratios from inorganic calcite, ostracode valves, and fish bone. The data presented were collected at one outcrop of Bouse Formation in the Chemehuevi basin (i.e., northern Bouse Formation, hereafter “NBF”; Fig. 1), and at four Bouse Formation locations in the Blythe basin (i.e., southern Bouse Formation, hereafter “SBF”; Fig. 1). This sampling strategy took advantage of the uncontested lacustrine origin of the NBF to provide a template against which to compare the contested marine, estuarine, or lacustrine origin for the SBF. The nearly 100 km between the northernmost and southernmost SBF outcrops allowed us to test for an isotopic gradient along a possible Bouse estuary (e.g., Smith, 1970; Crossey et al., 2015). This multi-indicator approach combined a robust sampling scheme embedded within an improved understanding of Bouse Formation stratigraphy (e.g., Homan, 2014; Gooette et al., 2016b; O’Connell, 2017) to further test the marine, estuary, and lacustrine models for the origin of the SBF.

## BACKGROUND

### Geology and Geography of the Bouse Formation

The late Miocene and early Pliocene Bouse Formation (House et al., 2008; Sarna-Wojicki et al., 2011; Harvey, 2014; McDougall and Miranda Martinez, 2014, 2016; Crow et al., 2017) is discontinuously exposed in four sequential basins located along a nearly 300-km-long stretch of the lower Colorado River corridor (Fig. 1; Spencer et al., 2013). The Bouse Formation was initially described in detail by Metzger (1968), Metzger and Loeltz (1973), and Metzger et al. (1973) and can be broken out into three main units: (1) dense tufa deposits found around basin edges that directly and conformably overlie coarse-grained Miocene fanglomerates or bedrock; (2) a fossiliferous basal limestone unit that varies from 1 to 30 m thick; and (3) an overlying siliciclastic-dominated interbedded unit composed of clay, silt, and fine sand that can be a few tens of meters thick to over 200 m thick. A newly named “upper limestone member” (Dorsey et al., 2018) or “upper bioclastic unit” (Gooette et al., 2016b) caps the Bouse Formation. This capping unit corresponds to one of Metzger et al.’s (1973) older aluvium bodies, “Unit A.” The origin of this capping unit is beyond the scope of this paper, but it is thought to represent either a second, preferably marine flooding event (Dorsey et al., 2018) or an offlapping mixture of stranded Bouse Formation sediments and local tributary sediments that prograded toward the basin axis as the water level of the southern Bouse basin lowered (Gooette et al., 2016b).

Several of the southernmost exposures of the basal limestone unit in the Blythe basin have been described in detail by Homan (2014) and O’Connell (2017), who documented a variety of carbonate sediments, including limestone, calcarenite, and several marl variations. We prefer to use a broader “basal carbonate unit” designation for these sediments, following Homan (2014). Dorsey et al. (2018) referred to these sediments as the “basal carbonate member.” The thickness and details of the Bouse Formation vary from outcrop to outcrop, with the most conspicuous differences being that the capping bioclastic unit and an enigmatic mixture of marginal marine and freshwater fossils (e.g., Reynolds and Berry, 2008; McDougall and Miranda Martinez, 2014) have been found only in the Blythe basin. Otherwise, the overall physical characteristics of the Bouse sedimentary package are noticeably similar throughout lower Colorado River corridor.

Outcrops of the NBF are discontinuously exposed in the Cottonwood, Mohave, and Chemehuevi basins (Fig. 1; Spencer et al., 2008) at elevations as high as 550–560 m above sea level (m asl) in the Mohave and Cottonwood basins before stepping down across the Topock bedrock paleodivide (Fig. 1) to a maximum elevation of ~370 m asl in the Chemehuevi basin (Fig. 1; Pearthree and House, 2014). The SBF is confined to the Blythe basin and has a maximum elevation of ~330 m asl (Fig. 1; Spencer et al., 2008, 2013). Sediments described as Bouse Formation have been encountered in the subsurface of the Blythe basin to depths of at least 200 m below sea level (Metzger and Loeltz, 1973; Metzger et al., 1973), and the Bouse Formation may also be present several hundred meters below the surface near Yuma, Arizona (Fig. 1), where it overlies older Miocene marine rocks (Olmssted et al., 1973; McDougall, 2008).

### Age of the Bouse Formation

Several volcanic ashes provide a maximum age estimate for the NBF. The 5.51 ± 0.08 Ma Conant Creek Tuff (or possibly the 5.59 ± 0.05 Ma Wolverine Creek tephra) is exposed ~60 m below NBF sediments in Cottonwood basin (Fig. 1; House et al., 2008; Pearthree and House, 2014). Preliminary dating of another volcanic ash higher in the section but still below NBF sediments in Cottonwood basin produced a younger maximum age estimate of 5.35 ± 0.07 Ma (Crow et al., 2017). Thus, the early Colorado River was forming NBF lakes in the Cottonwood basin no earlier than 5.6 Ma, and most likely Bouse deposition there began after 5.3 Ma.

The age of the SBF is a topic of debate. Radiometric ages on basalt flows found in late Miocene Miocene gravels stratigraphically below the SBF show deposition started after 8.2 ± 0.3 Ma (Reynolds et al., 1986; Buisin and Beratan, 1993). A foraminifer-based biochronology implies that the basal carbonate unit was being deposited before 6 Ma in marine or estuarine conditions (McDougall and Miranda Martinez, 2014, 2016; Miranda-Martinez et al., 2017). The minimum age of 6 Ma is derived from a single test of the planktic foraminifer Globorotalia lenguensis, which has a global last appearance datum of 6 Ma (Wade et al., 2011). McDougall and Miranda Martinez (2014, 2016) hypothesized that marine
Figure 1. Regional map showing the relationship of Cottonwood, Mohave, Chemehuevi, and Blythe basins along the lower Colorado River corridor and in relation to North America (inset). Extents of paleolakes that deposited the northern Bouse Formation are shown in darker blue. Extent of water body that deposited the southern Bouse Formation is shown in light blue. Yellow dots are outcrop locations used in this study. Black bars with two-letter identifiers are paleodivides: Py—Pyramid, Tk—Topock, Ay—Aubrey, Ch—Chocolate Mountains. White stars are other locations mentioned in the text or figures: A—Amboy, BP—Buzzard Peak, BM—Big Maria quarry, C—Cibola, IM—Isla Montague, LS—Laguna Salada, SMG—Split Mountain Gorge. MX—Mexico, GWT—generalized location of Grand Wash Trough. Base figure is from the Global Multi-Resolution Topography (GMRT) Synthesis (Ryan et al., 2009), http://www.geomapapp.org.
conditions persisted until the arrival of the early Colorado River water and sediments ca. 5.3 Ma, when the overlying siliciclastic interbedded unit was deposited in a saline lake environment. This hybrid marine to lake interpretation will be referred to as a “partial marine model.” The 4.83 ± 0.01 Ma Lawlor Tuff (Sarna-Wojcicki et al., 2011) provides the youngest age constraint on the SBF. The Lawlor Tuff outcrops are within the basal carbonate unit but are high on the landscape at ~300 m asl (e.g., Spencer et al., 2013; Miller et al., 2014), which would imply that the SBF persisted until 4.8 Ma. This interpretation is supported by the upstream age constraints for the NBF (e.g., House et al., 2008; Crow et al., 2017), if the SBF was lacustrine. Dorsey et al. (2018) proposed that the Lawlor Tuff is within or just beneath their upper bioclastic member and, thus, postdates the deposition of the basal carbonate unit in the valley axis by ~0.5 m.y. In Dorsey et al.’s (2018) interpretation, the Lawlor Tuff was deposited during a younger marine inundation of the Blythe basin.

Similar discrepancies between radiometric age estimates and paleontological-based age estimates for the onset of basin flooding exist at other locations in the Gulf of California. For example, Carreño and Smith (2007) and Helenes et al. (2009) hypothesized that the earliest marine rocks in the central Gulf of California (not shown on Fig.1) are middle Miocene in age, based on microfossil content. Biostratigraphy coupled with radiometric dating of volcanic rocks on Isla Tiberón showed that marine deposition in the central Gulf of California started after 6.2 ± 0.2 Ma (Bennett et al., 2015). The middle Miocene microfossils were apparently reworked into younger marine strata (Bennett et al., 2015).

Surface exposures of the Bouse Formation, regardless of the basin, are deeply incised and overlain by sands and gravels of the early Pliocene (ca. 4.5 Ma to 3.5 Ma) Bullhead Alluvium (Howard et al., 2015). These deposits provide the first unequivocal sedimentological evidence for a through-flowing early Colorado River (House et al., 2008; Pearthree and House, 2014; Howard et al., 2015).

Depositional Environment of the Bouse Formation: Previous Interpretations

The present consensus is that the NBF represents a series of freshwater to slightly brackish-water lakes that were sequentially filled-and-spilled in a southward progression by the advancing terminus of the early Colorado River (e.g., House et al., 2008; Pearthree and House, 2014). This lacustrine interpretation is based largely on carbonate δ18O-δ13C values and radiogenic carbonate 87Sr/86Sr ratios (Buiseng, 1988; Spencer and Patchett, 1997; Poulsou and John, 2003; Spencer et al., 2008; Crosseay et al., 2015), by a freshwater faunal assemblage that lacks marine fossils (Reynolds, 2008; McDougall and Miranda Martinez, 2014; Reynolds et al., 2016; Roeder and Smith, 2016), and by abrupt decreases in maximum paleowater surface elevations between basins (Spencer et al., 2008; Pearthree and House, 2014). The fill-and-spill lacustrine interpretation is convincingly supported by catastrophic paleoflood deposits to the south of the Pyramid paleodivide (Fig. 1; House et al., 2008). Cloos (2014) and Kimbrough et al. (2015) demonstrated that Laramide-aged detrital zircons from the upper Colorado River watershed were present in the first-arriving Colorado River sands deposited in the early Gulf of California, further supporting a “top-down” fill-and-spill model for lower Colorado River integration.

The depositional environment of the SBF continues to be debated, with the competing marine, estuarine, and lacustrine interpretations carrying important consequences for the geomorphic and tectonic evolution of the surrounding area (e.g., Lucchita, 1979; Pearthree and House, 2014; Miller et al., 2014; Dorsey et al., 2018). Southern Bouse Formation carbonates share numerous similarities in their isotopic and trace-element compositions with the NBF (Table 1). Several tracers, such as their radiogenic 87Sr/86Sr ratios, are statistically indistinguishable (Table 1), which is consistent with similar origins. Previous workers have argued that these similarities are entirely the result of postdepositional alteration (e.g., Lucchita et al., 2001). Efforts to determine the degree of diagenetic alteration of Bouse carbonates are ongoing ( Ferguson et al., 2017).

The partial marine model of McDougall and Miranda Martinez (2014, 2016), and as featured in Dorsey et al. (2018), relies heavily on foraminifer assemblages to hypothesize that the basal carbonate unit was deposited in a marine embayment or estuarine environment that was in existence for at least 0.7 m.y. before the arrival of the early Colorado River. This model raises two critical concerns. Late Miocene marine 87Sr/86Sr ratios (0.709) have not been found in SBF carbonates (Spencer and Patchett, 1997; Spencer et al., 2013; Crosseay et al., 2015), and a source of water capable of creating a Bouse estuary in the absence of the early Colorado River has not been identified.

### Table 1. Comparison of Averaged Oxygen (δ18O) and Carbon (δ13C) Values, 87Sr/86Sr Ratios, and Trace-Element Concentrations from Northern and Southern Bouse Formation Carbonates

|                      | Northern Bouse Formation | Southern Bouse Formation | p value | n | p value |
|----------------------|-------------------------|--------------------------|---------|---|---------|
| δ18O (%) VPDB        | ~7.9                    | ~6.6                     | ~0.034  | 0.940 | 0.0129  |
| δ13C (%) VPDB        | ~1.6                    | ~1.7                     | ~0.790  | 0.790 | 0.240   |
| δ13C (%) VPDB        | ~7.1084                 | ~7.1095                  | ~0.7109 | 0.7109 | 0.57    |
| 87Sr/86Sr             | ~1.7                    | ~0.57                    | ~0.029  | 0.029 | 0.079   |
| MgCO3 (mol %)        | ~300                    | ~394                     | ~394    | 394 | ~394    |
| Mn (ppm)             | ~933                    | ~1017                    | ~1017   | 1017 | ~1017   |
| Fe (ppm)             | ~429                    | ~325                     | ~325    | 325 | ~325    |
| Sr (ppm)             | ~136                    | ~341                     | ~341    | 341 | ~341    |
| Ba (ppm)             | ~4.2                    | ~3.0                     | ~3.0    | 3.0 | ~3.0    |
| Zn (ppm)             | ~197                    | ~197                     | ~197    | 197 | ~197    |

Note: Statistically significant (p value < 0.05) results are highlighted in bold. Data were compiled from Spencer and Patchett (1997), Poulsou and John (2003), Roskowski et al. (2010), Crosseay et al. (2015), and Bright et al. (2016). VPDB—Vienna Pee Dee Belemnite.
Crossey et al.’s (2015) estuary model uses carbonate δ18O-δ13C values and 87Sr/86Sr ratios to hypothesize that a downstream-propagating and progressively evolving mixture of radiogenic, strontium-rich, upstream water sources, including the early Colorado River, entered the Blythe basin, where it may have encountered seawater. Crossey et al. (2015) showed that this water could produce the radiogenic 87Sr/86Sr ratios (0.710–0.712) observed in the basal carbonate unit, even if 5% to 75% of the water in the Blythe basin was seawater. In this model, the early Colorado River was present to deliver that water. The hypothesis that the Colorado River was present during deposition of the basal carbonate unit is a significant departure from the partial marine model hypothesis that the Colorado River was present during deposition of the basal carbonate unit, even if 5% to 75% of the water in the Blythe basin was seawater. The marine and estuary models require that the Blythe basin was initially at sea level and was connected in some manner to the early Gulf of California before experiencing ~330 m of post-Miocene uplift (Lucchitta, 1979; O’Connell et al., 2017), acknowledging that late Miocene global sea level may have fluctuated within ~30 m of the modern level (Miller et al., 2005; De Schepper et al., 2014). The presence of a variety of marine and marginal marine organisms, including abundant benthic (e.g., Ammonia beccarii) and rare planktic foraminifers (e.g., Globorotalia sp., Neogloboquadrina sp.; McDougall and Miranda Martinez, 2014), large numbers of the barnacle Amphibalanus subalbidus (Zullo and Buising, 1988; Van Syoc, 1992; Pitombo, 2004), low numbers of marine diatoms (e.g., Actinocyclus octonarius, Terpinoë americana; Miller et al., 2014), and rare fossils of the marine fish Colpichthys regis (Todd, 1976), coupled with sedimentary structures that are interpreted as evidence for a tidal influence, such as possible tidal rhythmites and sigmoidal bedding (O’Connell et al., 2017), strengthen the marine and estuarine interpretations.

The lacustrine model of Spencer and Patchett (1997) and Spencer et al. (2008, 2013) relies heavily on strontium isotopes to hypothesize that the Blythe basin has been at or near its current elevation since the late Miocene, where it was the last in a sequence of basins to be flooded by the southward advance of early Colorado River (Spencer et al., 2013; Pearthree and House, 2014). The lacustrine model is supported by continental isotopic signatures (Spencer and Patchett, 1997; Poulson and John, 2003; Roskowski et al., 2010; Crossey et al., 2015) and by the presence of a variety of continental fossils, such as freshwater clams (e.g., Sphaerium californica, Pisidium sp.) and snails (e.g., Amnicola longinaqua, Fluminicola sp.; Reynolds and Berry, 2008); freshwater fish (e.g., Gila cypha; Roeder and Smith, 2016), freshwater ostracodes (e.g., Candonia spp.; Bright et al., 2016), and freshwater diatoms (e.g., Cocconeis diminuta, Pseudostaurosira brevistriata; Miller et al., 2014). The morphology of the abundant basin-margin travertine (tufa) in the Blythe basin has recently drawn comparison with lacustrine travertine at Pyramid Lake, Nevada, (Crossey et al., 2017). Abrupt downstream decreases in the maximum elevations of Bouse deposits across paleodivides between the Mohave, Chemehuevi, and Blythe basins (Pearthree and House, 2014) lend regional support to the overspilling lacustrine model. Physical evidence for a lake spill-over event at the Pyramid paleoland (Fig. 1), in the form of a southward-directed fluvial-boulder conglomerate containing boulders sourced in the Pyramid hills some 7 km to the north (House et al., 2008; Pearthree and House, 2014), provides an example of a paleoland overtopping in the lower Colorado River integration story.

The broad range of seemingly conflicting data sets leads to equally conflicting interpretational paradigms for the origin of the SBF. The marine and estuary models argue that the freshwater fossil assemblage was simply reworked from neighboring continental environments (e.g., Lucchitta et al., 2001; Miller et al., 2014). The lacustrine model argues that the marginal marine organisms were transported by birds (e.g., Figuerola and Green, 2002) or some other vector into a saline lake (Spencer and Patchett, 1997). Poulson and John (2003) reported higher carbonate δ18O-δ13C values and higher trace metal concentrations at the southern end of Blythe basin. The number of southern Blythe samples (n = 11) is higher than the number of northern Blythe samples (n = 3) in that particular study, making it difficult to robustly compare data from the northern and southern parts of the basin. Foraminifer species richness is also highest at the southern end of the basin (McDougall, 2008; McDougall and Miranda Martinez, 2014). Collectively, these data sets might suggest the presence of a Bouse estuary (e.g., Crossey et al., 2015), but other viable explanations are also available. For example, Dumont (1998) documented an estuary-like salinity gradient in the surface waters of the landlocked Caspian Sea. Salinities are lowest (< 1 ppt) near where the Volga River enters the sea, and values gradually increase to ~13 ppt at the southern end of the sea. Salinity-driven zonation of pelagic and benthic organisms within the sea is common (Dumont, 1998). Thus, faunal distributions and isotopic gradients that might mimic those found in estuaries can also occur in lakes. Similarly, Blair (1978) and Blair and Bradbury (1979) noted that the Hualapai Limestone of the Muddy Creek Formation, exposed in the Grand Wash Trough area (Fig. 1), contains a modestly diverse (n = 24 species) and very Bouse-like mix of euryhaline coastal marine, estuarine, and freshwater to brackish-water lacustrine diatoms. The marine and estuarine diatoms include several species of Melosira and Amphora (Blair and Bradbury, 1979). The 12–6 Ma Hualapai Limestone is a spring-fed lacustrine deposit; it is not a coastal marine or estuarine deposit (Crossey et al., 2015). Thus, the small numbers of marine and estuarine diatoms in the Bouse Formation at Amboy (Fig. 1) could represent a similar lacustrine environment (Miller et al., 2014). The marine and estuarine Bouse models can readily account for sedimentary structures in the Blythe basin that are interpreted to be marine in origin (i.e., tidal rhythmites, sigmoidal bedding; O’Connell et al., 2017). The lacustrine model requires these sedimentary structures to be explainable by lacustrine
processes, and there are numerous examples where sedimentary structures typically considered to be marine have been recognized in lacustrine environments (e.g., Davis et al., 1972; Ainsworth et al., 2012; Fraser et al., 2012).

Rationale Behind an Integrated Microfossil and Stable Isotope Approach

Many organisms are sensitive to a variety of ecological factors, such as temperature and overall salinity, but many are also sensitive to the relationship between overall salinity and individual ionic concentrations or ionic ratios. This has been demonstrated for continental ostracodes (Forester and Brouwers, 1985; Forester, 1986), diatoms (Blinn, 1993), brine flies (Bowen et al., 1985), molluscs (Sharpe and Forester, 2008), and other organisms (Plaziat, 1993; Rasmussen, 1988; Herbst, 2001). The distribution of marine organisms, on the other hand, is often presented in terms of salinity, temperature, and substrate characteristics (e.g., Arnal et al., 1980; Chin et al., 2010; McGann et al., 2013). Ionic ratios in seawater are less important because they are essentially stable. Numerous marginal marine organisms such as ostracodes (i.e., Cyprideis, Cytheromorpha), benthic foraminifers (i.e., Ammonia beccarii, Elphidium gunteri), and marine molluscs (i.e., Cardium, Potamides) are able to colonize continental lakes where the ratio of calcium to carbonate alkalinity (Ca/ALK; Forester, 1986) and the relative proportions of sodium and chloride converge on a “marine like” composition (e.g., Gasse et al., 1987; Anadón et al., 1986; Anadón, 1989, 1992; Plaziat, 1993; Wennrich et al., 2007). Some marginal marine ostracodes, like Cyprideis torosa and Cytheromorpha fuscata, are known to tolerate sulfate- and chloride-rich continental waters (Nielsen et al., 1987; Mezquita et al., 1999). Similarly, continental organisms can persist in marginal marine and estuarine environments, but they are restricted to dilute environments where salinity typically remains below ~5–10 ppt (Forester and Brouwers, 1985; Bulger et al., 1993). Here, the Bouse Formation microfossil assemblage is interpreted within the context of a well-established ecological framework that is augmented by various types of isotopic data contained within the microfossil assemblages.

This study focuses more on δ18O values because of the strong correlation between δ18O and salinity in marine and estuarine environments (e.g., Rodríguez et al., 2000; Dettman et al., 2004), and among δ18O, salinity, inflow, and evaporation rates (e.g., Talbot, 1990; Leng and Marshall, 2004). Carbon isotopes will be discussed less frequently because numerous influential variables (degasging, organic respiration, etc.) reduce the correlation with salinity. The discussion of δ18O values of a variety of Bouse Formation carbonates is grounded on assumption that carbonate produced in similar environments will have similar δ18O values. Possible comparisons will be among aragonitic mollusc shells, and calcitic micrite, barnacle plates, ostracode valves, and foraminifer tests. Although these different materials have slightly different δ18O values when precipitated in the same environment, the difference between them is small. Low-temperature δ18O values of aragonite are roughly 1.2‰ higher than calcite formed in the same water and at the same temperature (Kim et al., 2007). Biologic carbonates often incorporate a variable “vital effect” that can increase their δ18O values by typically 0% to 2% above inorganic calcite precipitated in the same water and at the same temperature (e.g., Killingly and Newman, 1982; Grossman, 1984; von Grafenstein et al., 1999). Because the range in δ18O values in our study is very large, distinctions in δ18O values less than 2‰ will be treated as arising from similar environments. All stable isotope data in this study were measured values (relative to Vienna Pee Dee Belemnite [VPDB]) and were not corrected for mineralogy or vital effects.

Frequent use of δ18O-δ13C cross-plots is made because they often provide important clues about the source of water and potential environments of deposition (e.g., Talbot, 1990; Gross et al., 2013) and because they facilitate comparison with previous Bouse studies (e.g., Roskowski et al., 2010; McDougall and Miranda Martinez, 2014; Crossey et al., 2015). Fully marine environments within the past 10 m.y. have typically produced carbonates that group near the origin of a δ18O-δ13C cross-plot (Prokoph et al., 2008). Estuarine environments are typically characterized by a strong correlation between salinity and isotope composition, and, thus, they typically generate carbonates that are positively covariant, where lower-salinity environments near the head of the estuary plot with lower δ18O-δ13C values, and more saline environments near the mouth of the estuary plot with higher δ18O-δ13C values (e.g., Ingram et al., 1996a; Reinhardt et al., 2003; Sampe et al., 2005). The relationships between salinity and carbonate δ18O-δ13C values can be fairly complex in lakes, and at times, they may even be counterintuitive (Li and Ku, 1997). It is not uncommon to find examples where terminal but freshwater or mildly brackish lakes generate carbonate δ18O-δ13C values that overlap with or even exceed those found in marine carbonates (e.g., Anadón et al., 2008; Sharpe and Bright, 2014). Thus, lakes can produce carbonates that display a wide range of δ18O-δ13C relationships (e.g., Talbot, 1990). The fundamental differences in the isotopic compositions of marine, estuarine, and lacustrine environments, coupled with the fundamental differences in the ecology of strictly marine, marine but continentally invasive, and fully continental organisms, provide the foundational framework for our interpretations.

STUDY AREAS

The principal study areas included five Bouse Formation outcrops that were strategically located along a north-south transect from the Chemehuwini basin to Marl Wash (Fig. 1). One outcrop of the lacustrine NBF in the Chemehuwini basin (34.4477°N, 114.4010°W; Fig. 1) was sampled at roughly meter-scale resolution (Fig. 2). Four SBF outcrops were located at Parker, Arizona (34.1623°N, 114.3021°W), in Milpitas Wash (33.2599°N, 114.7303°W), in Hart Mine Wash (33.2907°N, 114.6352°W; Fig. 1), and in Marl Wash (33.2634°N, 114.6393°W; Fig. 1). The outcrops at Parker (Fig. 3), Milpitas Wash (Fig. 4), and Hart Mine Wash (Fig. 5) were sampled at roughly 10–40 cm resolution. The SBF in Marl Wash...
Figure 6 was coarsely sampled (>meter scale) at prominent changes in lithology. Four additional spot collection sites targeted barnacle-rich sediments of the basal carbonate unit. Two of these sites were in the vicinity of Hart Mine Wash (Fig. 1), and two sites were located in the Palo Verde Mountains, slightly to the north of Milpitas Wash (Fig. 1). Modern barnacle shells from the Salton Sea (Fig. 1) and from Isla Montague (Fig. 1) were also analyzed. Detailed descriptions of each of the Bouse sampling locations are available in the Supplemental Materials.1

This sampling strategy was intended to quantify the microfossil content and the δ18O-δ13C values in micrite and biologic calcite from a lacustrine environment in the Chemehuevi basin and then compare those values against equivalent data sets from the contested environment in the Blythe basin. The outcrop of SBF at Parker, Arizona, is located near the area where the early Colorado River would have entered the Blythe basin (Fig. 1). The outcrops at Milpitas Wash, Hart Mine Wash, and Marl Wash are located near the southern margin of the basin (Fig. 1), far from any major river input, and would have been near the mouth of a possible Bouse estuary. If the SBF formed in a normal marine environment, then the microfossil assemblages and the carbonate δ18O-δ13C values should show a clear transition from lacustrine conditions in the Chemehuevi basin, to mildly saline conditions at the head of the Bouse estuary near Parker, to more marine-like conditions near the mouth of the Bouse estuary at the southern end of the basin. A modern estuary analogy might be found in the freshwater influence of the pre-dam-era Colorado River on the δ18O values in *Mulinia* shells from the northern Gulf of California. There, the influence of the river can be measured up to 60 km away from its mouth (Rodriguez et al., 2001).

**METHODS**

**Microfossil Analysis**

Sediment samples were disaggregated for microfossil analysis (*n* = 140) by soaking them in a weak sodium bicarbonate and sodium hexametaphosphate solution for up to 1 wk. The most indurated samples were subjected to a single freeze-thaw cycle. Repeated freeze-thaw cycling was avoided to

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1Supplemental Materials. Detailed outcrop descriptions and supplemental tables. Please visit https://doi.org/10.1130/GES01610.S1 or the full-text article on www.gsapubs.org to view the Supplemental Materials.
reduce breaking foraminifer tests. Once disaggregated, each sediment slurry was washed over a 45 µm sieve. Microfossils in the >120 µm size fraction were identified and counted. The foraminifers in a small concentrated aliquot of each of the 45–120 µm size fractions were counted. The foraminifers were not identified but were grouped by appearance. “Spiraled” foraminifers included predominantly benthic genera such as *Rosalina* and *Ammonia*, and “biserial” foraminifers included the benthic genus *Bolivina* and the planktic genus *Streptochilus* (McDougall and Miranda Martínez, 2014). Microfossil counts were normalized on a “per gram sediment” basis. Standardizing the microfossil content by sample mass allows for direct comparison of the actual abundance of each component in any sample. This is the first Bouse study that normalized microfossil content. Other Bouse studies have presented microfossil data as either raw counts (Smith, 1970; McDougall and Miranda Martínez, 2014) or as species percentage plots (Dorsey et al., 2018). Microfossil counts are available in Tables SI-1 to SI-5 (footnote 1). Microfossil data for Outcrop E were provided in Bright et al. (2016).

**Stable Isotope (δ¹⁸O, δ¹³C) Analysis**

Previous studies of Bouse Formation sediments often reported bulk carbonate δ¹⁸O-δ¹³C values (Poulson and John, 2003; Roskowski et al., 2010; Crossey et al., 2015). This study is the first effort to analyze the isotopic composition of specific components within a bulk sediment sample.

Stable isotope (δ¹⁸O, δ¹³C) values were measured on fine-grained micrite (<45 µm; n = 131), aliquots of ostracode valves (n = 238) and foraminifer tests (n = 21), individual barnacle fragments (n = 32), unidentified individual shell fragments (n = 4), and secondary calcite (n = 4).
Fresh sediment from each sample was disaggregated in reverse osmosis (RO) water and sieved at 45 µm to remove biogenic carbonate. Deflocculants or other additives were not used. The <45 µm micrite slurries were dried overnight in a 40 °C oven and then gently homogenized into powders with a porcelain mortar and pestle.

The best-preserved ostracode valves (Cyprideis sp., Cytheromorpha sp., Candona spp., Heterocypris sp.; Bright et al., 2016) were picked out of the >250 µm size fraction. Adult valves were preferentially chosen, but large juvenile valves were used when suitable adult valves were not available. The ostracode valves were manually cleaned with a fine paintbrush. If secondary carbonate growths or other resistant debris was present, it was removed with a fine needle or by quickly brushing the valves with dilute (0.5 N) hydrochloric acid followed by a thorough rinsing with RO water. Analysis of the larger ostracode species (Cyprideis, Candona, Heterocypris) used two valves per aliquot, and 3–6 valves per aliquot were required for the Cytheromorpha analyses.

At Hart Mine Wash, six sediment samples near the top of the basal carbonate unit and at the base of the interbedded unit contained high concentrations of spiraled and biserial foraminifers (Fig. 6). These sediments correspond to the “Streptochilus biofacies” of McDougall and Miranda Martinez (2016) and Dorsey et al. (2018). The most abundant spiraled benthic foraminifers in this interval were Rosalina columbiensis, with less common Ammonia beccarii (McDougall and Miranda Martínez, 2014; Dorsey et al., 2018). The most abundant biserial foraminifer present was Streptochilus spp. (> 90%), but small numbers of Bolivina spp. were also present (e.g., McDougall and Miranda Martinez, 2014). The “spiraled” and biserial foraminifer isotopic analyses were each dominated by a single genus of foraminifer. The limited number of foraminifer genera present in each analysis will reduce the influence of environmental variability on the resulting δ¹⁸O-δ¹³C values. Foraminifers from the 45–120 µm size fractions from each of the six samples were picked for analysis under a binocular microscope using a two-step process. First, each residue was examined under transmitted light, and only translucent foraminifer tests that were free of visible encrustations or infilling sediments were selected. The picked foraminifers were then subjected to a second selection process using reflected light, in which any stained or discolored tests were removed. A minimum sample mass of ~40 µg was required for analysis, but the number of foraminifers in each sample was not counted. Scanning electron microscope (SEM) images...
Figure 5. Microfossil results and $^{87}$Sr/$^{86}$Sr values in micrite, fish bone, and ostracode valves from multiple outcrops of southern Bouse Formation at Hart Mine Wash, Blythe basin. Cyp—Cyprideis plotted as square root values to highlight low abundances low in the section, Cyth—Cytheromorpha, Per—juvenile Perissocytheridea, Lim—Limnocythere, Het—Heterocypris, Cand—Candona, Dar—Darwinula, B+S—fish bone and scale per gram sediment, Sf45—spiraled (benthic) foraminifers in a concentrated aliquot of 45–120 µm residue (raw counts, not as per gram sediment), Bf45—biserial foraminifers in a concentrated aliquot of 45–120 µm residue (raw counts, not as per gram sediment), b.o.g.—barnacle oncoid grainstone, DCL—distinctive clay layer, TM—transitional marl. Note differences in scales on x axis for Het, Dar, Bf45, and B+S. See Figure 1 for location and Supplemental Materials (text footnote 1) for additional details about the sampling scheme.
of representative “spiraled” and biserial tests from Hart Mine Wash are presented in Figure 7.

Barnacle shell fragments were chosen for isotopic analysis based on their appearance under a microscope using reflected light. Only smooth, pearlescent material was selected. Barnacle fragments were soaked in 0.5 N HCl and visually inspected until the majority of their mass was removed. The acid leaching step ensured that any surface contaminants were removed. The remaining fragments were thoroughly rinsed under RO water. Single barnacle fragments were analyzed. Digital images of typical barnacle fragments are available in the Supplemental Material (footnote 1).

Two fragments of clean and pearlescent shell material and two fragments of shell material that was clearly overgrown and contaminated with secondary carbonate were selected from the >250 µm size fraction of the Parker basal carbonate unit residues. In addition, four large pieces of purely secondary calcite were collected from the Hart Mine Wash basal carbonate unit residues.

The carbonate δ18O-δ13C values were measured 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 was calibrated based on repeated...
Figure 7. Scanning electron microscope (SEM) images of foraminifers from Hart Mine Wash. (A–H) Biserial (Streptochilus spp.) foraminifer tests, (I–K) biserial (Bolivina spp.) foraminifer tests, (L–M) “spiraled” benthic foraminifer tests, (N–O) broken chambers on tests A and H, respectively. Images A to O are from transitional marl sample HM-55 at the base of the interbedded unit (see Supplemental Materials for stratigraphic context [text footnote 1]). (P–W) Biserial (Streptochilus spp.) foraminifer tests, (X–Bb) biserial (Bolivina spp.) foraminifer tests, (Cc–Ff) “spiraled” benthic foraminifer test. Images P to Ff are from oxidized marl sample HM-38 near the top of the basal carbonate unit (see Supplemental Materials for stratigraphic context [text footnote 1]). Note jagged edges on tests Cc and Dd suggesting they broke off of something larger. Scale bars = 50 µm, except N and O, where scale bars = 25 µm.
measurements of standards NBS-18 and NBS-19, and acid fractionation factors were assumed to be identical for both aragonite and calcite. The stable isotope results are reported in standard delta (δ) notation, where: δ (%) = [(Rsample/Rref) − 1] × 10^3, and R is ratio of δ18O or δ13C. Rref refers to the standard VPDB. Precision for δ18O and δ13C measurements were ±0.1‰ and ±0.08‰ (1σ), respectively. Stable isotope values are available in Supplemental Tables SI-6 to SI-10 (footnote 1). Additional stable isotope data for Outcrop E were provided in Bright et al. (2016).

Strontium Isotope Ratio (87Sr/86Sr) Analysis

Previous studies of Bouse Formation sediments often reported the bulk carbonate 87Sr/86Sr ratio (Spencer and Patchett, 1997; Roskowski et al., 2010; Crossey et al., 2015). As with the stable isotope analyses, this study is the first effort to analyze the 87Sr/86Sr ratio of specific components within a bulk sediment sample.

Strontium isotope ratios (87Sr/86Sr) were measured on <45 µm micrite powders (n = 5), >250 µm fish bone fragments (n = 9), and aliquots of ostracode (Cyprideis sp.) valves (n = 2) from the upper portion of the basal carbonate unit and from one sample near the bottom of the interbedded unit in Hart Mine Wash (Fig. 5). The fish bones were sonicated in RO water for several minutes and examined under a binocular microscope, and only the most pristine fragments were chosen for analysis. Two aliquots of adult Cyprideis sp. valves were subjected to a brief 0.5 N hydrochloric acid bath to remove the outermost surfaces of the valves. The valves were rinsed in RO water and air dried, and the most pristine valves were selected for analysis.

Strontium isotope ratios were measured at Geochron Laboratories in Chelmsford, Massachusetts. The samples were dissolved in 1 N hydrochloric acid, centrifuged, and evaporated. If organic material was present, the sample was treated with concentrated nitric acid and hydrogen peroxide. The samples were converted to nitrate with concentrated nitric acid and finally dissolved in 500 µL of 3.5 N nitric acid in preparation for strontium separation by ion exchange. Strontium was separated from other cations by using 50 µL columns filled with Eichrome Sr-Spec resin. The samples, dissolved in 500 µL of 3.5 N nitric acid, were centrifuged and loaded onto the columns. The columns were rinsed with 1200 µL of 3.5 N nitric acid. Strontium was then eluted with 800 µL of deionized water. A drop of 0.1 M phosphoric acid was added to each sample, and then they were dried down. For carbonate samples, this process was repeated twice to improve separation of strontium from calcium, which interferes with the mass spectrometry.

Isotope ratio measurements were made on a GV IsoProbe T multicollector mass spectrometer. Strontium was loaded on Re filaments in a solution of tantalum chloride (TaCl5) and phosphoric acid, which were analyzed by dynamic multicollection. The data were corrected for 87Rb interference based on measured 87Rb abundance, fractionation corrected to 87Sr/86Sr = 0.1194, and reported with respect to a value of 0.710250 for NBS-987. All 87Sr/86Sr ratios are available in Table SI-11 (footnote 1).

RESULTS

Northern Bouse Formation

Sediments from the NBF contain the marginal marine but continentally invasive ostracode Cyprideis sp. in association with the continental ostracodes Limnocythere sp., Candona sp., and Darwiniula stevensoni (Fig. 2). Rare spired and bivalve foraminifers were present at the bottom of the section (Fig. 2). Digital photographs of several of the more common bivalve foraminifers that are tentatively identified as Bolivina are presented in Figure 8. This is the first occurrence of foraminifers reported from the NBF, which is a significant departure from previous studies (Reynolds, 2008; McDougall and Miranda Martinez, 2014; Reynolds et al., 2016). To rule out the possibility of cross-contamination, raw sediments from three of the foraminifer-bearing samples were processed a second time but using different beaters and sieves. Similar numbers and types of foraminifers were found. Ostracode abundances conspicuously decline and both Limnocythere sp. and the foraminifer disappear across an abrupt transition within the marl near the bottom of the outcrop (Fig. 2).

The micrite and ostracode valves displayed a wide range of δ18O values (~18‰ to +1‰) and δ13C values (~4‰ to +3‰; Figs. 9 and 10A). The oxygen isotope values in micrite (δ18Omic) and ostracode valves (δ18Oost) were noticeably higher at the bottom of the outcrop in association with the foraminifer-bearing sediments (Fig. 9).

Southern Bouse Formation

Sediments from the SBF contained an ostracode faunal assemblage that is similar to, but more diverse than, the NBF. SEM images of SBF ostracodes are available in Bright et al. (2016), and additional images are presented in Figure 11. A comparable Cyprideis sp., Limnocythere sp., Candona spp., and Darwiniula stevensoni assemblage was observed to be variably present throughout the SBF (Figs. 3–6). Additional marginal marine but continentally invasive ostracodes Cytheromorpha spp. and Perissocytheridea sp. were also present, but the Perissocytheridea sp. were rare and occurred primarily as juveniles. One additional continental ostracode, Heterocypris sp., also appeared (Figs. 4–6). No strictly marine ostracodes were found. We define “strictly marine ostracodes” as those genera that are not known to colonize continental lakes. Representative genera include Caudities, Aurila, or Kostoleberis, all of which are common ostracodes in nearshore environments in the modern Gulf of California (Swain, 1967), and all of which have been identified from fully marine sediments of the age-equivalent late Miocene and early Pliocene Imperial
bonate analyses of SBF carbonates (Poulson and John, 2003; Roskowski et al., 2010; Croseay et al., 2015). Bioclastic-rich and bedded limestones at the bottom of the basal carbonate unit exposed at Parker and at Hart Mine Wash yielded micrite and ostracode δ18O and δ13C values that are nearly identical, with values of about −6‰ and −2‰, respectively (Figs. 10B and 10D). Softer marls deposited above the bioclastic-rich sediments consistently yielded δ18O values up to 12‰ higher than the associated δ18O values (Figs. 9 and 10B–10D). Several Cytheromorpha sp. samples from the top of the basal carbonate unit had δ14COST values that were nearly identical to δ18OOST values from the bottom of the unit (Fig. 10D).

Spiraled foraminifers at the top of the basal carbonate unit and at the bottom of the overlying interbedded unit in Hart Mine Wash had δ18O and δ13C values that were nearly identical to δ18O-δ13C values from the bottom of the basal carbonate unit (Fig. 10D), much like several of the Cytheromorpha sp. noted previously. The biserial foraminifers had δ18O-δ13C values similar to the benthic marginal marine ostracodes (Cyprideis sp. and Cytheromorpha sp.) with which they were associated (Figs. 9 and 10D). Notably, the Streptochilus δ18O-δ13C values at Hart Mine Wash were virtually identical to the Cyprideis sp. δ18O-δ13C values at Parker (Figs. 10B and 10D).

The transition from the top of the basal carbonate unit to the overlying interbedded unit at both Hart Mine Wash and Milpitas Wash is associated with an increase in the abundance of continental ostracodes (i.e., Candoninae, Darwinula stevensoni; Figs. 4 and 5; Bright et al., 2016) and with an abrupt −8‰ decrease in δ18O values (Figs. 9, 10C, and 10D; Bright et al., 2016). The δ18O values in the associated foraminifers did not change across this boundary (Fig. 10D). An equivalent decrease in δ18O values was not found at the Parker outcrop (Fig. 9).

Clean shell fragments from the basal carbonate unit at Parker yielded δ18O values (−9.0‰ and −8.7‰) that were fairly similar to the δ18O values (−9.5‰ and −7.4‰) of shell fragments encrusted in secondary carbonate. Four large pieces of purely secondary calcite from the bottom of the basal carbonate unit in Hart Mine Wash yielded an average δ18O value of −9.1‰ ± 0.3‰.

Modern barnacle shells from the Salton Sea and from Isla Montague (Fig. 1) yielded a comparatively narrow range of δ18O values (−1‰ to +3‰) and δ13C values (−2‰ to +2‰; Fig. 12). The fossil barnacle shells from the basal carbonate unit at SBF localities displayed a wider range of δ18O (−12‰ to −5‰) and δ13C (−6‰ to +1‰) values (Fig. 12). The isotopic compositions of the fossil barnacles tended to cluster by location (Fig. 12).

Material-specific 87Sr/86Sr ratios from the oxidized marl at the top of the basal carbonate unit and from the overlying transitional carbonate at the base of the interbedded unit in Hart Mine Wash (Fig. 1) were often found to be dissimilar (Fig. 5). The oxidized marl samples yielded 87Sr/86Sr ratios from fish bone (0.7104 ± 0.0006; n = 8) and Cyprideis calcite (0.71096; n = 1) that were typically higher than 87Sr/86Sr ratios from the encasing micrite (0.71084 ± 0.00013; n = 4; Fig. 5). One sample of transitional marl yielded 87Sr/86Sr ratios in fish bone (n = 1), Cyprideis calcite (n = 1), and micrite (n = 1) of 0.71084, 0.71104, and 0.71097, respectively (Fig. 5).

Formation in the Salton Trough (Jefferson, 2001; Reynolds et al., 2008). Foraminifers were variably abundant throughout the basal carbonate unit (Figs. 3–6). In Hart Mine Wash, notable peaks in foraminifer abundances occurred in the bioclastic-rich and bedded limestone horizon at the bottom of the basal carbonate unit and again in the soft oxidized marls at the top of the basal carbonate unit (Fig. 5).

The micrite and ostracode values from the SBF displayed a wide range of δ18O values (−15‰ to −1‰) and δ13C values (−7‰ to +2‰; Figs. 9 and 10B–10D), which are comparable to those measured in the NBF (Figs. 9 and 10A). Similar ranges of δ18O and δ13C values have been previously reported from bulk carbonate analyses of SBF carbonates (Poulson and John, 2003; Roskowski et al., 2010; Croseay et al., 2015).
Figure 9. Compilation of δ¹⁸O values in micrite and ostracode valves from one section of northern Bouse Formation in Chemhuevi basin and three sections of southern Bouse Formation in Blythe basin. See Figure 1 for locations. MM—abundance of marginal-marine (Cyprideis + Cytheromorpha + Perissocytheridea) ostracode valves (juvenile + adult) in valves per gram (vpg). C—total abundance of continental (Candona + Heterocypris + Limnocythere + Darwinula) ostracode valves (juvenile + adult) in valves per gram (vpg). Micrite and ostracode δ¹⁸O values are similar in the shallow-water bedded limestone and lower marls because of enhanced mixing of river and lake water by waves. Micrite δ¹⁸O values in the deeper-water marls are consistently several per mil lower than the ostracode valve δ¹⁸O values because the epilimnic micrite formed in an isotopically stratified and poorly mixed seasonal freshwater cap derived from spring snowmelt flood discharge on the early Colorado River. The ostracode valves calcified in the benthos and represent the isotopic composition of deeper water. Stylized stratigraphic columns are aligned at the contact between the basal carbonate unit and the interbedded unit simply for convenience. No temporal correlations are implied. Dashed line in Chemhuevi panel correlates an abrupt decrease in ostracode δ¹⁸O values in the Chemhuevi basin with similar abrupt decreases at Hart Mine Wash and Milpitas Wash. VPDB—Vienna Pee Dee Belemnite. Microfossil and isotopic data are available in the Data Repository (see text footnote 1).
Figure 10. Cross-plots of stable oxygen ($\delta^{18}O$) and carbon ($\delta^{13}C$) isotope values in micrite and various forms of biologic calcite from the Bouse Formation. (A) Micrite and multiple genera ostracode results from the northern Bouse Formation, Chemehuevi basin. (B) Micrite, Cyprideis, and shell fragment results from the southern Bouse Formation, Parker. (C) Micrite and multiple genera ostracode results from the southern Bouse Formation, Milpitas Wash. (D) Micrite and multiple genera ostracode results from the southern Bouse Formation, Hart Mine Wash. Results are grouped by stratigraphic horizon. DCL—distinctive clay layer. Ostracode and foraminifer $\delta^{18}O$ values have not been corrected for vital effects. VPDB—Vienna Pee Dee Belemnite. See Figure 1 for locations.
Isotopic Resetting and Evidence for the Reworking of Microfossils

Bouse carbonates have undoubtedly been subjected to postdepositional influences since their deposition ca. 5 Ma. The question is whether or not the available isotopic information is a useful archive capable of informing us of past environmental conditions. Several observations suggest that the isotopic information is useful, and they provide two compelling arguments against significant isotopic resetting of (at least some) Bouse carbonates. First, Roskowski et al. (2010) analyzed the δ¹⁸O and δ¹³C values from a micromilled SBF barnacle shell and found that the shell retained clear cyclic patterns in the δ¹⁸O and the δ¹³C values, similar to that observed in modern (e.g., Surge et al., 2013) and fossil (e.g., Wesselingh et al., 2006) shell materials. The barnacle δ¹⁸O values ranged from −9‰ to −10.5‰ (Roskowski et al., 2010) and clearly do not reflect a marine origin. The isotopic integrity of the shell would have been destroyed if the shell had experienced significant postdepositional isotopic resetting or recrystallization. Second, this study documents a surprisingly high degree of reproducible isotopic heterogeneity from a variety of carbonate sources within a single sample. For example, sample HM-38 (see Supplemental Materials [footnote 1]) yielded a δ¹⁸Osed value of about −9‰, multiple Cyprideis δ¹⁸Oost values of about −2‰, multiple benthic foraminifer δ¹⁸O values of about −6‰, and multiple planktic foraminifer δ¹⁸O values of about −3‰. Similar items in adjacent and distantly spaced samples showed a similar degree of isotopic heterogeneity (Fig. 9). If significant isotopic resetting or recrystallization of the various carbonate sources had occurred, then the repetitive and persistent differences in δ¹⁸O values between each carbonate type within any given sample would have been destroyed. Bouse carbonates are probably not pristine, but carefully selected items clearly retain a degree of internally consistent isotopic heterogeneity that does not support claims of significant isotopic resetting or recrystallization (e.g., Lucchitta et al., 2001).

The detailed nature of this data set allows us to address for the first time the issue of reworking of fossils within the Bouse carbonates. Earlier studies have suggested that the freshwater, continental fossils might have been reworked from the surrounding watershed into a marine or estuarine environment (Turak, 2000; Miller et al., 2014). In this study, two genera of marginal marine ostracodes and two genera of continental ostracodes consistently occurred together and showed persistently similar δ¹⁸O values (Figs. 9 and 13B).

Figure 11. Scanning electron images of ostracode valves from the southern Bouse Formation. All images are external lateral views. (A) Candona sp. C, female right valve, sample HM73. (B) Candona sp. C, female left valve, sample HM73. (C) Candona sp. C, male left valve, sample HM73. (D) Candona sp. D, female right valve, sample HM73. (E) Candona sp. E, female right valve, sample EARP26. (F) Perissocythereidea sp., juvenile right valve, sample HM33. (G) Perissocythereidea sp., juvenile left valve, sample HM33. For stratigraphic information, refer to Figure 3 and the Supplemental Materials (text footnote 1) on Hart Mine Wash. Ostracode valves were coated with carbon and imaged on a Hitachi 3400N SEM. Original images were modified for contrast and clarity using Photoshop CS6. See Bright et al. (2016) for additional ostracode valve images.
Figure 12. Cross-plot comparing δ¹⁸O and δ¹³C values from barnacle fragments (large colored circles and gray diamonds) and <45 µm micrite (colored triangles) in the Blythe basin against barnacle fragments from the modern Salton Sea (red boxes), Isla Montague (orange stars), and marine barnacles from previously published sources (white circles). Published marine barnacle data were compiled from Smith et al. (1988), Craven et al. (2008), and Detjen et al. (2015). The δ¹⁸O and δ¹³C values from pre-dam-era Mulinia specimens from Isla Montague and from other locations up to 65 km away from the mouth of the Colorado River (small black and gray circles, respectively; Rodriguez et al., 2001; data provided by D. Dettman), as well as δ¹⁸O and δ¹³C values from late Pleistocene and Holocene (17–2 ka) planktic foraminifers from the Guaymas basin (small purple circles; Keigwin, 2002; data provided by L. Keigwin), are plotted for context. Symbols and respective colors for Blythe basin barnacle values match locations outlined in the Supplemental Materials (text footnote 1). VPDB—Vienna Pee Dee Belemnite. No δ¹⁸O values were corrected for mineralogical or vital effects.

- Late Pleistocene to Holocene (17-2 ka) planktic foraminifera, Guaymas Basin, Gulf of California
- Pre-dam era *Mulinia*, 20-65 km from mouth of the Colorado River, Gulf of California
- Pre-dam era *Mulinia*, Isla Montague, mouth of the Colorado River, Gulf of California
- Modern barnacles, Isla Montague, mouth of the Colorado River, Gulf of California
- Modern barnacles, Salton Sea
- Marine barnacles

- HM126 tidal bundle section (top)
- HM125 tidal bundle section
- HM124 tidal bundle section
- HM123 tidal bundle section (base)
- HM120
- PV6
- PV6 micrite
- PVc
- HM109
- HM110
- HM111
- HM113
- Roskowski et al. (2010) Milpitas
- Roskowski et al. (2010) Cibola
- Crossey et al. (2015) Cibola
Figure 13. Cross-plots of composite stable oxygen ($\delta^{18}$O) and carbon ($\delta^{13}$C) isotope values from micrite, bulk sediment, and biologic calcite from outcrops of northern and southern Bouse Formation. (A) Comparison of northern Bouse Formation (NBF) and southern Bouse Formation (SBF) micrite (this study) and late Miocene and early Pliocene planktic foraminifers and nannofossils from the Pacific Ocean. (B) Comparison of multiple genera of northern and southern Bouse Formation ostracodes and foraminifers (this study) and late Miocene and early Pliocene marine carbonate. CHEM—Chemehuevi, PKR—Parker, MILP—Milpitas Wash, HMW—Hart Mine Wash. Pacific Ocean planktic foraminifer and nannofossil data (small white circles) were compiled from Douglas and Savin (1971, 1975, 1978), Keigwin (1979, 1982), Barrera et al. (1985), and Cannariato and Ravelo (1997). When written descriptions were given, the data were confined to “late Miocene” and “early Pliocene” samples, which approximates foraminifer-based age estimates for the southern Bouse Formation (McDougall and Miranda Martínez, 2014; Miranda-Martínez et al., 2017). When numerical ages were provided, the data were confined to the interval 5.8 Ma to 4.5 Ma. The western Pacific Ocean Streptochilus ages, isotopic values, and nomenclature are as in Resig and Kroopnick (1983) and Resig (1989). VPDB—Vienna Pee Dee Belemnite.
This indicates that both types of ostracodes were calcified in water with similar $\delta^{18}O$ values. The continental ostracodes do not have the low $\delta^{18}O$ values that would be expected if they calcified in meteoric water (e.g., streams) before being reworked into a marine environment. The large offset in $\delta^{13}C_{\text{ost}}$ values between the continental and marginal marine ostracodes from the softer marls (Figs. 10B–10D) is consistent with a deeper-water environment (Homan, 2014) and seasonal differences in ostracode population dynamics. The continental ostracodes likely calcified during winter months, when the export of organic matter (i.e., algae) from the epilimnion and the subsequent decay of $^{13}C$-enriched organic matter were minimal. In contrast, the marginal marine ostracodes likely calcified during the summer, when the epilimnion was more productive, and the resulting benthic decay of $^{13}C$-enriched organic matter was higher (e.g., Decrouy et al., 2011a, 2011b). Alternatively, the marginal marine ostracodes could have burrowed into the sediment before calcifying, and so their valves might record the isotopic composition of $^{13}C$-enriched interstitial pore water rather than the composition of the overlying water body (e.g., Decrouy et al., 2011a, 2011b). The ostracode valves analyzed in this study do not support the notion that the SBF fossil assemblage is a mixture of continental- and marine-derived fossils (e.g., Cronin et al., 2012).

This study does demonstrate that the abundant spiraled (benthic) foraminifers (i.e., *Rosalina columbiensis*, *Ammonia beccari*) from the oxidized marls at the top of the basal carbonate unit and from the bottom of the interbedded unit at Hart Mine Wash (Fig. 5) were likely reworked. Multiple samples of spiraled foraminifer tests from these sediments consistently yielded displaced $\delta^{18}O$-$\delta^{13}C$ values that were dissimilar to the associated benthic ostracodes (Fig. 10D), but that were virtually identical to values observed in micrite and ostracodes from the bioclastic and bedded limestone (Fig. 10D) at the bottom of the section. Both *Rosalina* and *Ammonia* are common in the bioclastic and bedded limestone horizon (Dorsey et al., 2018), and these sediments are a likely source of the reworked tests. Several *Cytheromorpha* $\delta^{18}O$-$\delta^{13}C$ values from soft upper marl sediments displayed a similar, seemingly displaced, pattern (Fig. 10D). *Cytheromorpha* valves are the smallest and least massive ostracode valves analyzed in this data set. Their small size likely renders them more susceptible to reworking, an explanation that is extended to the even smaller foraminifer tests. In contrast, the biserial foraminifer tests from the oxidized marl have $\delta^{18}O$-$\delta^{13}C$ values similar to the associated benthic ostracodes (Fig. 10D), suggesting they were not reworked from older sediments. Biserial foraminifer tests from the oxidized marl showed nearly identical $\delta^{18}O$-$\delta^{13}C$ values as biserial foraminifer tests in the overlying interbedded unit (Fig. 10D). Biserial foraminifers from the interbedded unit were likely reworked from the oxidized marls (Fig. 10D). Previous SBF studies have acknowledged that foraminifer tests have been reworked into the interbedded unit (McDougall and Miranda Martinez, 2016), but the “spiraled” foraminifer $\delta^{18}O$-$\delta^{13}C$ values provide the first evidence that benthic foraminifer tests at the top of the basal carbonate unit were likely reworked as well. Foraminifer-based biostratigraphic or biochronologic interpretations for the SBF should be interpreted with caution.

Potential Sources of the $\delta^{18}O_{\text{SED}}$ and $\delta^{13}C_{\text{SED}}$ Values in the <45 µm Sediment Fraction

The <45 µm $\delta^{18}O_{\text{SED}}$-$\delta^{13}C_{\text{SED}}$ values reflect autochthonous or detrital sources, or perhaps a mixture of the two sources. Geologic mapping (Bishop, 1963; Jennings, 1967; Richard et al., 2000) has revealed that there are no detrital carbonate sources in the study area other than limited outcrops of Paleozoic marine limestones in the Big Maria Mountains northwest of Blythe (Fig. 1) and in the Riverside Mountains near Parker (Fig. 1). The isotopic composition of these limestones is unknown, but the limited exposures suggest that they were not a significant source of detrital carbonate in this study, especially in the context of the meters worth of carbonate sediments that typically comprise the basal carbonate unit. A second potential source of detrital carbonate is Colorado Plateau sediment delivered by the early Colorado River. This source is unlikely for two reasons. First, river sediments were likely trapped in upstream basins as the river terminus worked its way south from the Grand Canyon area (Pearthree and House, 2014). Sediment could not be transferred to the next basin downstream until sedimentation in the upstream basin had accumulated to the over spill threshold. The first-arriving river water in any Bouse basin was probably relatively clear water, lacking abundant entrained sediment (e.g., Pearthree and House, 2014). The basal carbonate unit was likely deposited in a clear-water environment. Second, the SBF <45 µm $\delta^{18}O_{\text{SED}}$-$\delta^{13}C_{\text{SED}}$ values can be compared to the $\delta^{18}O$-$\delta^{13}C$ values of potential detrital carbonate sources on and near the Colorado Plateau. For example, the averaged $\delta^{18}O$ and $\delta^{13}C$ values from a variety of carbonate bedrock types on the Colorado Plateau and nearby areas (i.e., Supai Group, Redwall Limestone, Leadville Limestone) are $-5.3\% \pm 2.1\%$ and $-2.5\% \pm 1.8\%$, respectively ($n = 70$; McKee, 1982; Muller and Mayo, 1986; Chidsey et al., 2006). These values are distinctly different from the <45 µm $\delta^{18}O_{\text{SED}}$-$\delta^{13}C_{\text{SED}}$ values from the NSF and SBF (Fig. 14A). The more terrigenous interbedded unit sediments have $\delta^{18}O_{\text{SED}}$ values that are more similar to a potential detrital signal, but the interbedded unit $\delta^{18}O_{\text{SED}}$ values are several per mil lower than the detrital signal (Fig. 14A); thus, detrital carbonate cannot have been the dominant source of the interbedded unit $\delta^{18}O_{\text{SED}}$-$\delta^{13}C_{\text{SED}}$ values. Furthermore, the $\delta^{18}O_{\text{SED}}$-$\delta^{13}C_{\text{SED}}$ values from the interbedded unit are similar to those measured in associated ostracode valves, which would not have incorporated a terrigenous carbonate signal (Figs. 10B–10D). Collectively, these observations show that the Bouse $\delta^{18}O_{\text{SED}}$, $\delta^{13}C_{\text{SED}}$, $\delta^{18}O_{\text{OST}}$, and $\delta^{13}C_{\text{OST}}$ values all recorded the isotopic composition of the water body present during deposition. The influence of detrital carbonate was minimal to nonexistent.

Northern Bouse Formation—The Lacustrine Template

The NBF exposed in the Chemehuevi basin contains a mix of continental (**Limnocythere, Candona, Darwinula**) and typically marginal marine but continentally invasive (**Cyprideis**) ostracodes in association with rare foraminifers (Fig. 2).
Figure 14. Cross-plots of stable oxygen ($\delta^{18}$O) and carbon ($\delta^{13}$C) isotope values from micrite, bulk sediment, and biologic calcite from outcrops of northern Bouse Formation (NBF) and southern Bouse Formation (SBF). (A) Comparison of a potential detrital carbonate signal from the Colorado Plateau area and <45 µm micrite from this study. Colorado Plateau data were compiled from McKee (1982), Muller and Mayo (1986), and Chidsey et al. (2006). (B) Previously reported southern Bouse Formation bulk carbonate samples highlighting potential trends in $\delta^{18}$O-$\delta^{13}$C values. Data were compiled from Poulson and John (2003), Roskowski et al. (2010), and Crossey et al. (2015). See text for discussion. (C) Comparison of mollusc shell from the San Francisco Bay estuary and micrite and ostracode valves from this study. San Francisco Bay mollusc data are from Ingram et al. (1996a). (D) Previously published southern Bouse Formation biologic carbonate. Data were compiled from Roskowski et al. (2010) and Crossey et al. (2015). No $\delta^{18}$O values have been corrected for mineralogical or vital effects. C—Cibola, M—Milpitas Wash, BM—Big Maria quarry, NBF—northern Bouse Formation, SBF—southern Bouse Formation, VPDB—Vienna Pee Dee Belemnite. See Figure 1 for locations.
The marginal marine ostracode genus *Cyprides* is common in western U.S. lakes and wetlands (e.g., Stout, 1981; Jayko et al., 2008), as well as in nearshore marine and estuarine environments (e.g., McGann et al., 2013). The two species of *Cyprides* commonly found in North America have broad salinity tolerances that range from $-200$ to 40,000 mg L$^{-1}$ (Forester et al., 2005). *Darwinula stevensoni* specimens are rarely found in water where salinity exceeds 1000 mg L$^{-1}$ (Forester et al., 2005). However, Holmes et al. (2010) noted that *D. stevensoni* can tolerate higher salinities (to 15,000 mg L$^{-1}$), but only when the salinity value is stable; otherwise, they generally prefer less saline waters up to $-2000$ mg L$^{-1}$, similar to the upper limit reported by Forester et al. (2005). Although the *Limnocythere* sp. and *Candona* spp. are yet to be identified to species level, they do not belong to any of the common species of North American *Limnocythere* (i.e., *Limnocythere stipilinii*, *Limnocythere sappensa*) or candonid species (i.e., *Fabaeformiscandonia rawsoni*) that might indicate elevated salinity values above $-5000$ mg L$^{-1}$ (Forester et al., 2005). *Cyprides* co-occurring with *Limnocythere*, *Candona*, and *Darwinula* in a low-salinity lacustrine setting is not uncommon (e.g., Mischke and Wünnemann, 2006). Finding foraminifers associated with freshwater or mildly brackish species of *Candona* is less common, but it does occur (e.g., Willard et al., 2000; Petit-Marie et al., 2010; Primavera et al., 2011). The absence of any known salinity-tolerant candonids or limnocytherids and the initial presence of only rare foraminifers suggest that salinity during the deposition of the NBF in the Chemehuevi basin was probably on the order of $5000$ mg L$^{-1}$ or less. Although the salinity was relatively low, the *Candona* and *Cyprides* valves at the bottom of the section yielded marine-like $\delta^{18}O_{\text{SED}}-\delta^{13}C_{\text{SED}}$ values (Figs. 9 and 10A). These high values likely reflect some combination of the initial evaporation of water in Lake Mohave and additional evaporation of this water as it was impounded in the Chemehuevi basin. The initial *Cyprides-Limnocythere* ostracode faunal assemblage associated with rare foraminifers and high $\delta^{18}O_{\text{SED}}$ and high $\delta^{13}C_{\text{SED}}$ values (Fig. 2) is interpreted to represent a nearly freshwater, terminal lacustrine environment that formed as the upstream Mohave basin (Fig. 1) first began to drain southward, by either groundwater sapping (e.g., Crossley et al., 2015) or threshold overspilling (e.g., Pearthree and House, 2014), or both. The initial inflow to Chemehuevi basin may have been low compared to local evaporation rates, which would help elevate the initial $\delta^{13}C_{\text{SED}}$ and $\delta^{18}O_{\text{SED}}$ values.

The abrupt change in the microfossil assemblage coupled with an abrupt decrease in the isotopic composition of micrite and ostracode valves low in the Chemehuevi marl (Fig. 2) are interpreted to represent a south-directed overtopping event at the Aubrey paleodam (Fig. 1). This resulted in a transition from a closed-basin to a through-flowing basin configuration and thus, likely reflects the onset of truly freshwater conditions. An analogy can be drawn to a similar event that occurred in Pliocene Villarroya Lake, Spain, (Anadón et al., 2008). There, marine-like $\delta^{18}O-\delta^{13}C$ values in *Candona* valves and freshwater clam shells abruptly shifted to much lower values when the depositional environment changed from a terminal-basin to an open-basin configuration (Anadón et al., 2008). The brown claystone at the top of the Chemehuevi sequence yielded the lowest $\delta^{18}O_{\text{SED}}$ values in the data set (Fig. 2), probably recording a through-flowing, fluviually dominated environment feeding the developing paleolake Blythe to the south.

This interpretation supports previous interpretations that the NBF in Chemehuevi basin was deposited in a lacustrine environment (Spencer and Patchett, 1997; Poulson and John, 2003; Reynolds, 2008; Crosse et al., 2015; Reynolds et al., 2016). Now that the microfossil and stable isotope characteristics of a known fill-and-spill lacustrine setting in the Chemehuevi basin have been characterized, the NBF lacustrine model can be compared against the microfossil and stable isotope results from the contested SBF.

**Southern Bouse Formation—Testing the Marine Model**

The defining criteria of the partial marine model are the age and the source of the water that produced the SBF. The partial marine model hypothesizes that the basal carbonate unit was being deposited in a marine basin by at least 6 Ma, and at least 0.7 m.y. before the early Colorado River reached the basin (McDougall and Miranda Martinez, 2014, 2016; Dorsey et al., 2018). Thus, it would have been deposited in an environment that, from a faunal and isotopic perspective, should have been fundamentally different from the younger Colorado River–fed lacustrine environments of the NBF.

Numerous studies have suggested the SBF basal carbonate unit was deposited in a marine environment, based largely on the ecological constraints provided by its diverse foraminifer assemblage (McDougall and Miranda Martinez, 2014; Dorsey et al., 2018). This study reveals several important observations that make a marine origin for the SBF unlikely. First, micrite and ostracode valves from the limestone at the bottom of the formation, which should represent the first flooding of the land surface by seawater (e.g., McDougall and Miranda Martinez, 2014), lack a definitive marine isotopic signature (Figs. 13A and 13B). Second, the ostracode fauna in the SBF is similar to, but more diverse than, the ostracode fauna of the lacustrine NBF described herein. No strictly marine ostracode genera were found in this study, much like earlier studies (Smith, 1970; Metzger et al., 1973; Miller et al., 2014; Bright et al., 2016). Third, continental ostracodes (*Candona, Heterocypris*) yielding $\delta^{18}O_{\text{OST}}$ values that are virtually identical to their marginal marine ostracode counterparts are variably present throughout the basal carbonate unit (Fig. 9).

Thus, the continental and marginal marine ostracodes were calcifying in water with the same $\delta^{18}O$ value. Fourth, $\delta^{18}O_{\text{SED}}, \delta^{13}C_{\text{SED}}, \delta^{18}O_{\text{OST}}$ and $\delta^{13}C_{\text{OST}}$ values from the SBF can be compared against the $\delta^{18}O-\delta^{13}C$ values from a variety of mid-Miocene to early Pliocene marine planktic foraminifers and nanofossils from the Pacific Ocean (Figs. 13A and 13B). This comparison is grounded in the assumption that the isotopic composition of the late Miocene and early Pliocene seawater that inundated Blythe basin was comparable to late Miocene and early Pliocene seawater in the Pacific Ocean. Small differences might be expected, as illustrated in the modern Gulf of California, where the restricted environment and high evaporation rates cause surface-water $\delta^{18}O$ values to be slightly higher than open-marine $\delta^{18}O$ values (Keigwin, 2002). The late Miocene...
and early Pliocene planktic foraminifers cluster near the origin of the graph, \((\delta^{18}O = 0\%, \delta^{13}C = 0\%)\), as expected for marine carbonates (Figs. 13A and 13B). Multiple outcrops of the SBF yielded micrite and ostracode valves with isotopic compositions that are strongly dissimilar to the marine carbonate analogue, but that are virtually indistinguishable from each other and that are virtually indistinguishable from comparable material from the lacustrine NBF (Figs. 13A and 13B). Most importantly, the most marine-like \(\delta^{18}O-\delta^{13}C\) values in this study come from the valves of the continental freshwater to mildly brackish water ostracodes and from the lacustrine NBF (Figs. 13A and 13B). Most importantly, the most marine-like \(\delta^{18}O-\delta^{13}C\) values in this study come from the valves of the continental freshwater to mildly brackish water ostracodes and from the lacustrine NBF (Figs. 13A and 13B). Most importantly, the most marine-like \(\delta^{18}O-\delta^{13}C\) values in this study come from the valves of the continental freshwater to mildly brackish water ostracodes and from the lacustrine NBF (Figs. 13A and 13B). Most importantly, the most marine-like \(\delta^{18}O-\delta^{13}C\) values in this study come from the valves of the continental freshwater to mildly brackish water ostracodes and from the lacustrine NBF (Figs. 13A and 13B).

Southern Bouse Formation—Testing the Estuary Model

The Bouse estuary model has been expressed in two slightly different contexts by McDougall and Miranda Martínez (2014) and Crossey et al. (2015). McDougall and Miranda Martínez (2014) defined their use of the term estuary as a “coastal body of water with free communication to the ocean and within which ocean water is diluted by freshwater derived from land” (p. 845, and citing Valle-Levinson, 2010). They hypothesized that the basal carbonate unit was being deposited at least 0.7 m.y. prior to the arrival of the early Colorado River, and so any dilution of seawater in their Bouse estuary must, by necessity, have resulted from water sourced in the small tributaries that drained the local highlands surrounding Blythe basin. Crossey et al. (2015) provided a slightly different solution; they allowed water from the early Colorado River and other upstream sources to be present during the deposition of the basal carbonate unit. Crossey et al. (2015) drew comparisons between the \(\delta^{18}O-\delta^{13}C\) values of marl in the southern Blythe basin and of mollusc shells from the San Francisco Bay estuary. The \(\delta^{18}O-\delta^{13}C\) composition of the San Francisco Bay estuary results from seawater that is diluted, in both salinity and isotopic composition, by freshwater from the Sacramento and San Joaquin Rivers. The Sacramento–San Joaquin River watershed covers nearly 150,000 km², or nearly 40% of the state of California, and these rivers deliver an average annual inflow of 600 m³ s⁻¹ to the San Francisco Bay estuary (McGann et al., 2013, and references therein). During spring runoff season, these rivers drain voluminous amounts of snowpack with low isotopic values from the Sierra Nevada, the Cascade Range, the Klamath Mountains, and the Coast Range.

The Bouse estuary models feature a positive and highly covariant trend in the \(\delta^{18}O-\delta^{13}C\) values from the basal carbonate unit (Crossey et al., 2015). McDougall and Miranda Martínez (2014) featured data that encompass most of the Blythe basin, while Crossey et al. (2015) limited their discussion to marls restricted to the southernmost portion of the basin, where planktic foraminifers are most abundant. Both data sets overlap considerably. The proposed Bouse estuary trend is defined by bulk carbonate \(\delta^{18}O-\delta^{13}C\) values that range from modestly low values (\(\delta^{18}O = -8\%, \delta^{13}C = -6\%)\) to values that approach those expected in normal marine carbonates (\(\delta^{18}O = +1\%, \delta^{13}C = +1\%\), e.g., Crossey et al., 2015). The low values would presumably represent the most fluvially dominated water in a Bouse estuary, whereas the high values would presumably represent the most marine-dominated water in the estuary. Crossey et al. (2015) showed that a similar trend with similar \(\delta^{18}O-\delta^{13}C\) values is seen in the San Francisco Bay estuary, based on the analysis of mollusc shells (Ingram et al., 1996a, 1996b). Salinity and \(\delta^{18}O\) values in the San Francisco Bay display a strong positive covariance (\(R^2 = 0.98\), where dilute water (0.5 ppt) has low \(\delta^{18}O\) values (−10.5‰ relative to Vienna Standard Mean Ocean Water [VSMOW]), and saline water (29.7 ppt) has high \(\delta^{18}O\) values (−2.0‰ VSMOW; Ingram et al., 1996a). Strong covariance between low/high salinity and low/high \(\delta^{18}O-\delta^{13}C\) values is common in nearshore marine and estuarine environments (e.g., Dettman et al., 2004; Sampei et al., 2005). Mollusc shells collected from nearly the mouth of the San Francisco Bay estuary grew in water with an average salinity of ~28.2 ± 3.8 ppt (\(n = 17\)), and these shells yield the highest \(\delta^{18}O-\delta^{13}C\) values (approximately −1‰ and −1‰, respectively; Ingram et al., 1996a). Mollusc shells collected near the head of the San Francisco Bay estuary grew in water with a lower average salinity of ~16.6 ± 7.1 ppt (\(n = 18\)), and these shells yielded lower \(\delta^{18}O-\delta^{13}C\) values (approximately −6‰ and −5‰, respectively; Ingram et al., 1996a). The lowest \(\delta^{18}O-\delta^{13}C\) values in the inferred Bouse estuary trend (\(\delta^{18}O = -8\%, \delta^{13}C = -6\%)\) are comparable to the lowest values in the San Francisco Bay estuary, yet the estuary component of the partial marine model (McDougall and Miranda Martínez, 2014) requires these values be generated in the absence of the
early Colorado River—a river that would have drained voluminous amounts of high-elevation snowpack from the Rocky Mountains. The partial marine model requires that the lowest δ18O-δ13C values were generated by the isotopic dilution of seawater by small, local, and ephemeral tributaries that drained several low-elevation mountain ranges (<1000 m asl) adjacent to Blythe basin. There is no clear evidence, such as pre-Bouse terminal lakes in the Mohave or Blythe basins, that suggests that the study area was substantially wetter during the late Miocene and early Pliocene than it is today. It is highly unlikely that the volume and isotopic composition of local runoff could have been sufficient to isotopically dilute seawater in a Bouse estuary to the same degree as observed in the modern San Francisco Bay estuary. A simple thought experiment using local precipitation rates and the surface area of “paleolake Blythe” highlights this improbability.

The city of Blythe, California, is located just to the southeast of the Big Maria Mountains (Fig. 1) and receives roughly 0.1 m yr⁻¹ of precipitation (wrcdri.edu; station 040924). The evaporation rate for the local region is ~1.4 m yr⁻¹ (Spencer et al., 2013). There are no perennial streams in the Blythe basin today because the evaporation rate greatly exceeds the precipitation rate. In this exercise, the drainage basin feeding the Blythe basin will be approximated by the maximum surface area of “paleolake Blythe” (Fig. 1), which is roughly 10,400 km² (Spencer et al., 2013). A maximum of 1 km² yr⁻¹ of water could be generated if 100% of the precipitation falling on the Blythe basin is converted to runoff and if evaporation and infiltration are completely (and unrealistically) excluded from consideration. In contrast, the Sacramento–San Joaquin watershed delivers ~19 km³ yr⁻¹ (estimated from McGann et al., 2013), i.e., nearly 20 times the estimated (and unreasonable) maximum volume of water that can be generated in the Blythe basin by precipitation alone. We also consider that discharge from the Sacramento–San Joaquin watershed includes a large seasonal contribution from high-elevation snowmelt (i.e., Sierra Nevada) that would have low δ18O values (i.e., ~12‰ VSMOW; Ingram et al., 1996a). In contrast, the highest elevations around the Blythe basin are below 1000 m asl and would be incapable of generating runoff with similarly low, snowmelt-derived, δ18O values. Thus, it is inconceivable that precipitation falling on Blythe basin during the late Miocene could realistically provide enough water of the appropriate isotopic composition to create a Bouse estuary that could mimic the salinity and isotopic characteristics of the San Francisco Bay estuary. The estuarine component of the partial-marine model (McDouggall and Miranda Martínez, 2014, 2016), and as featured in Dorsey et al. (2018), is untenable. Crosssey et al.’s (2015) estuary model resolves this issue by allowing the early Colorado River to be present during deposition of the basal carbonate unit.

Careful analysis of the proposed Bouse estuary trend reveals that McDougall and Miranda Martínez (2014) and Crosssey et al. (2015) did not include the δ18O-δ13C values from a suite of marl slabs containing the impressions and a few bones of the marine fish Colpichthys regis (Roskowski et al., 2010). These marls yielded low, nonmarine δ18O values of approximately ~11‰ (Roskowski et al., 2010), even though they preserve the impressions and bones of a marine fish. These marls were identified in various earlier figures as “Colpichthys regis” (McDouggall and Miranda Martínez, 2014) or as “southern Blythe fish” (Crosssey et al., 2015) and are excluded from discussion. If these “fish” marls are included, then the proposed Bouse estuary-like trend becomes less impressive (Fig. 14B). The trend becomes even less convincing if data from the northern Blythe basin are also included in the discussion. Poulsen and John (2003) presented numerous basal carbonate unit δ18O-δ13C values from the northern part of the basin that do not conform to the proposed Bouse estuary trend. Their Blythe basin data trend in a horizontal fashion, from low values similar to those measured in the southern Blythe “fish” marls (δ18O = ~12‰, δ13C = ~1‰) to a single analysis that yielded values close to those expected for normal marine carbonate (δ18O = ~1‰, δ13C = 0‰; Fig. 14B).

Further analysis reveals that the proposed Bouse estuary trend (e.g., Crosssey et al., 2015) is dominated by data from a geographically limited portion of Blythe basin. Roskowski et al. (2010) analyzed 16 bulk samples of the basal carbonate unit along an ~2-km-long section of “Marl” or “Cibola” wash (Fig. 1). These samples yielded highly covariant δ18O-δ13C values (R² value = 0.69) that define a positive trend from low to high, marine-like, values (Fig. 14B). However, the remaining bulk carbonate samples from Blythe basin do not behave in this fashion. If the Marl Wash data are excluded, then the R² value drops to 0.01 (Fig. 14B). The dominance of the Marl Wash data is also evident in the estuarine trend proposed by McDougall and Miranda Martínez (2014). Furthermore, the few bulk sediment samples with the most marine-like δ18O-δ13C values (δ18O = 0‰, δ13C = 0‰; Fig. 14B) seemingly require calcification in 100% seawater, yet strontium isotope considerations suggest that the amount of seawater present in Blythe basin could have been as little as 5% to 15% to as much as 60% to 75% (Crosssey et al., 2015). Strontium isotope modeling does not suggest that fully marine conditions were ever present (Crosssey et al., 2015). Thus, previous discussions of estuarine isotopic trends in the SBF cannot adequately account for the most marine-like δ18O-δ13C values in SBF bulk carbonates, and they place a disproportional amount of interpretational leverage on 16 analyses of bulk carbonate from one 2-km-long transect in Marl Wash (Fig. 1).

The δ18O-δ13C values in this study were compared to δ18O-δ13C values from mollusc shells from the San Francisco Bay estuary (Fig. 14C) following Crosssey et al. (2015). The highest, most marine-influenced (highest salinity) δ18O-δ13C values from San Francisco Bay overlap with the Cyprideis δ18O – δ13C values from the lowest marls in the Chemehuevi basin (Fig. 14C). Several data points also overlap with the continental ostracode δ18O – δ13C values from the soft marls at Hart Mine Wash and Milpitas Wash and several of the continental ostracode δ18O – δ13C values from the lacustrine Chemehuevi basin (Fig. 14C).

The lowest, most fluvially influenced (lowest salinity) δ18O-δ13C values from San Francisco Bay are most similar to the marginal marine Cyprideis and Citharomorpha δ18O – δ13C values from the soft marls at Parker, Hart Mine Wash, and Milpitas Wash (Fig. 14C). The ostracode ecology and the δ18O – δ13C values in this study are, for all intents and purposes, inversed from what would be expected in a normal estuary. The continental ostracodes should have been calcifying in the most dilute, fluvially influenced portion of the estuary (e.g.,...
Elliott et al., 1966; Forester and Brouwers, 1985; Smith and Horne, 2002), and the marginal marine ostracodes would presumably be calcifying their valves in the most saline, marine-influenced portion of the estuary. This is not the case with the SBF (Fig. 14C). It is also unreasonable to suggest that the Candona and Cyprideis valves in the lacustrine Chemehuevi basin (Figs. 10A and 13B) calcified in normal seawater.

Several other observations make an estuarine origin for the SBF unlikely. No evidence was found in this study for an isotopic gradient between the northern (Parker, Fig. 1) and southern (Hart Mine Wash or Milpitas Wash, Fig. 1) margins of the Blythe basin (Figs. 13A and 13B), as might be expected in a Bouse estuary (e.g., Ingram et al., 1996a; Sampei et al., 2005). No evidence for variable and evolving mixtures of marine and fluvial water masses that might be expected in an estuary was found either, presuming that sea level may have varied or that the early Colorado River experienced century- or millennial-scale (or longer) climatically induced variations in discharge that would be detectable as antithetic fluctuations in marine-brackish-freshwater ostracode spe-

Eventually, the ostracode fauna of the San Francisco Bay estuary was summarized in McGann et al. (2013). They found 19 species, of which only two were continental, Physocypris globula and Ilyocypris gibba. Ilyocypris gibba was found at only two stations where salinities at the time of collection were 0 practical salinity units (psu) and 10 psu (McGann et al., 2013). Ilyocypris gibba have limited salinity tolerance and are considered restricted to oligohaline (<5 ppt) environments (e.g., Wilkinson et al., 2005), and so its occurrence at 10 psu in the San Francisco Bay likely reflects transport into more brackish water (McGann et al., 2013). No specific salinity measurements were provided for the occurrences of Physocypris globula, but the overall salinity dynamics of San Francisco Bay suggest that P. globula individuals occur where salinities are typically less than 15 psu. Note that for the purposes of this study, “psu” and “ppt” can be considered equivalent (e.g., UNESCO, 1981). In contrast, the basal carbonate unit of the SBF contains a higher diversity of continental ostracodes, including Heterocypris, Darwinula, Limnothyrus, and multiple species of Candona (Bright et al., 2016; this study). Freshwater fish (Gila) and a variety of continental molluscs (Pisidium, Sphaerium, Fluviilicola, Amnicola, Physal[?]) are also variably present (Reynolds and Berry, 2008; Roeder and Smith, 2016). Two marginal marine ostracodes, Cyprideis beaconensis and Cytheromorpha grandwashensis, are present in San Francisco Bay, and both Cyprideis and Cytheromorpha are present in the SBF. Critically though, the San Francisco Bay estuary contains 16 genera of marine ostracodes that as a group are completely absent in the SBF. The marine ostracode fauna in the San Francisco Bay estuary consists of Acuminocythere, Ambastracon, Aurlia, Cytherura, Eusarsiella, Kangarina, Luxocepha, Palmoconcha, Paradoxostoma, Radimella, Rubustula, Sagmatocythere, Sahnicythere, Serrocysthera, Spinielberis, and Xeustoleberis (McGann et al., 2013). Some of the marine ostracode valves appear to be reworked from fully marine environments outside of the San Francisco Bay (McGann et al., 2013), but the majority of the genera are presumably living in the bay along with a diverse benthic foraminifer assemblage that has been cited as a potential analogue for the benthic foraminifer assemblage in the SBF (McDougall and Miranda Martinez, 2014). The complete absence of similar marine ostracode genera in the SBF is convincing evidence that the SBF is not a comparable estuarine deposit. Although previous Bouse studies have provided thorough discussions of estuary-like trends in the δ18O-δ13C values in SBF carbonates (McDougall and Miranda Martinez, 2014; Crosse et al., 2015), the combination of ostracode ecology and the δ18O-δ13C values in the ostracode valves themselves is incompatible with an estuarine environment.

**Southern Bouse Formation—Testing the Lacustrine Model**

In contrast to the marine and estuary models, the lake model of Spencer et al. (2008, 2013) hypothesizes that the entirety of the SBF was deposited in simply another Colorado River-fed lake and, thus, was deposited in an environment that, from a faunal and isotopic perspective, might be expected to possess fundamental similarities with the upstream NBF.

**Similarities in Ostracode Faunas with the Northern Bouse Formation**

The ostracode fauna in the SBF is similar to, but more diverse than, that of the lacustrine NBF described above. The SBF fauna includes a mix of continental ostracodes (Candona spp., Heterocypris spp., Darwinula stevensoni, Limnothyrus sp.) and typically marginal marine ostracodes (Cyprideis spp., Cytheromorpha spp., rare Perissocytheridea spp.; Figs. 3–6). The similarities in the ostracode faunas argue for similar lacustrine paleoenvironments, although the addition of Heterocypris spp., Cytheromorpha spp., and rare Perissocytheridea (Figs. 4 and 5) in the SBF may imply slightly different water composition—perhaps a composition slightly more similar to dilute seawater. Like the NBF, the SBF also lacks any of the saline-tolerant limnocytherid or candonid species (e.g., F. rawsoni) that might suggest elevated salinity. The fairly persistent presence of Candona spp. during deposition of most of the basal carbonate unit (Figs. 3–6) suggests that salinity likely remained below 10,000 mg L−1, and may have been lower (see NBF discussion).

Although the continental ostracode taxa in the SBF are found in association with a variety of marine foraminifers, they are incompatible with a fully marine or saline-estuarine environment (see previous discussion). A more comparable environment might be found in Lake Pontchartrain, where a modestly diverse benthic foraminifer fauna (Otvos, 1978) occurs alongside a very Bouse-like Cyprideis-Candona-Darwinula ostracode fauna in low-salinity (<10 ppt) water (Willard et al., 2000). The only typically marginal marine ostracodes so far recovered from the SBF (Cyprideis, Cytheromorpha, rare Perissocytheridea) are all genera that have also been reported from continental lakes (e.g., Stout, 1981; Pérez et al., 2011; Pint et al., 2015). This is in sharp contrast to
the numerous strictly marine ostracode genera, such as *Cushmania*, *Puriana*, *Aurila*, and others, that appeared in the nearby continental Laguna Salada basin (Fig. 1) when seawater apparently entered that basin during the early- to mid-Holocene (Romero Mayén, 2008), or to the 19 species of marine ostracodes present in the San Francisco Bay (McGann et al., 2013). Thus, the ostracode fauna from the SBF is better explained by a lacustrine origin than a marine or estuarine origin.

In addition to a salinity-based interpretation, the more abundant marine organisms in the SBF likely imply evolution of the overall ionic composition and the Ca/ALK ratio of paleolake Blythe to a more sodium-, chloride-, and sulfate-rich, marine-like, composition (e.g., Anadón, 1992). Recall that a marine-like water composition was initially present in the lacustrine Chemehuevi basin, as revealed by the early presence of *Cyprideis* sp. in association with rare foraminifers there (Fig. 2). Drawing on Spencer et al. (2008, 2013), the more diverse marginal marine ostracode fauna in the SBF is interpreted as representing a gradual downstream evolution in lake waters along a chain of Colorado River-fed lakes (Fig. 1), with paleolake Blythe being the largest, the most distal, and therefore the most compositionally evolved and most marine-like lake in the chain (e.g., Wood and Talling, 1988). In contrast to Spencer et al. (2008, 2013), the ostracode fauna implies that paleolake Blythe was not as saline as normal seawater and instead was only mildly brackish (~5000–10,000 mg L⁻¹). Higher-salinity phases of paleolake Blythe may have occurred at lower lake levels, but these would likely only be preserved in the subsurface closer to the depocenter of the basin. The ostracode fauna that has thus far been reported from the subsurface contains the same genera found in the surface outcrops (Smith, 1970; Metzger et al., 1973), suggesting that similar mildly brackish environments are preserved in the subsurface. A mildly brackish environmental interpretation echoes Taylor’s initial assessment from nearly 50 yr ago (in Smith, 1970), which stated that the impoverished marine fauna of the SBF implies a water composition that was different from normal seawater.

**Similarities in δ¹⁸O and δ¹³C Values with the Northern Bouse Formation**

Multiple outcrops of the SBF have yielded individual δ¹⁸OSED, δ¹³CSED, δ¹⁸OOST, and δ¹³COST values that are virtually indistinguishable from comparable lacustrine NBF values (Figs. 13A and 13B). Additional similarities with the NBF can be observed in the 8‰ to 10‰ difference between δ¹⁸OOST and δ¹³COST values at Parker, Milpitas Wash, and Hart Mine Wash (Fig. 9), and in an abrupt 8‰ decrease in δ¹⁸OOST values across the basal carbonate–interbedded unit contact (the distinctive clay layer or “DCL” of Bright et al., 2016) at Milpitas Wash and Hart Mine Wash (Fig. 9), and in a virtually identical abrupt decrease in δ¹³COST values in multiple ostracode genera throughout the thick Hart Mine Wash sedimentary sequence (Fig. 9), is more compatible with an evaporative terminal lake environment. However, the low-salinity (mildly brackish) constraints imposed by the SBF ostracode fauna is not compatible with the high, marine-like continental δ¹⁸OOST values encountered in this study (Fig. 13B).

A lacustrine interpretation can reconcile this apparent conflict because, unlike marine or estuarine systems, salinity and δ¹⁸O values can be decoupled in lakes. For example, Echo Lake, situated at nearly 3000 m asl in Colorado, has a low electrical conductivity that equates to a salinity of ~50 mg L⁻¹, yet its water has a δ¹⁸O value of about ~4‰ VSMOW (Henderson and Shuman, 2009). This decoupling between salinity and δ¹⁸O in some lakes readily explains how *Candona* valves with high, marine-like δ¹⁸OOST values could be produced in a low-salinity, closed-basin environment at Villarroya Lake, Spain (Anadón et al., 2008), and in a freshwater alpine pond at Snowmass Village, Colorado (Sharpe and Bright, 2014). A similar decoupling of salinity and δ¹⁸O values in both the Chemehuevi and Blythe basins effectively accounts for the marine-like *Candona* δ¹⁸OOST values in both basins.

Bright et al. (2016) noted that the abundance of continental ostracodes increases across the distinctive clay layer (e.g., Fig. 5) and interpreted that as evidence for a decrease in overall salinity. Their study was limited to a single outcrop in Hart Mine Wash. The larger data set featured in this study can now place that faunal transition into a broader context. The abundance of continental ostracodes (e.g., *Candona*) does increase across the distinctive clay layer (Figs. 4 and 5), but continental ostracodes, although rare, are variably present throughout the basal carbonate unit (Figs. 3–6). Ostracode abundances are noticeably low through several meters of sediments that were deposited prior to the distinctive clay layer (Figs. 4 and 5). These sediments are finely laminated, possibly indicating low-oxygen conditions in a stratified lake (e.g., Anadón et al., 1998; Sáez and Cabrera, 2002). Low-oxygen conditions are also implied by the high abundance of the biserial foraminifers *Streptochilus/Bolivina* in this same interval (Figs. 4 and 5). *Streptochilus/Bolivina* spp. often dominate foraminifer faunas in low-oxygen environments (e.g., Smart and Murray, 1994). The re-establishment of continental ostracodes above the distinctive clay layer may be explained by a return to a more oxygenated benthos as much as by a decrease in salinity. The fact that continental ostracodes are present throughout most of the basal carbonate unit suggests that any decrease in salinity across the distinctive clay layer was modest at best, and almost certainly could not be characterized as a transition from a fully marine environment to a river-fed environment (e.g., McDougall and Miranda Martinez, 2014, 2016; Dorsey et al., 2018).
Isotopic Evidence for the First Benthic and Lacustrine Occurrence of the Planktic Foraminifer Streptochilus

The SBF Streptochilus fauna is dominated by the planktic foraminifer *Streptochilus latum* (McDouggall and Miranda Martinez, 2014), which may indicate water depths of ~100–150 m (Nikolaev et al., 1998). However, the associated benthic foraminifers suggest inner neritic (<50 m) water depths (McDouggall and Miranda Martinez, 2014). Isotopic evidence presented in this study implies reworking of benthic foraminifers from nearshore sediments into deeper-water sediments, but the data do not constrain if the deeper-water environment was >100 m or <50 m deep. Miranda-Martinez et al. (2017) recently acknowledged that the Bouse *Streptochilus* might have been living in shallow water (presumably < 50 m) as benthic foraminifers. Rather than having δ18O-δ13C values that are similar to age-equivalent *Streptochilus* from normal seawater (Fig. 13B), the Bouse *Streptochilus* at Hart Mine Wash have δ18O-δ13C values that are similar to those measured in associated benthic Cyprideis valves (Figs. 10D and 13B) and that are virtually identical to those from *Cyprideis* valves at Parker (Figs. 10D and 13B). Regardless of the water depth, the Bouse *Streptochilus* specimens appear to have been calcifying alongside ostracodes in the benthos. This interpretation would fully support the assertion made by Miranda Martinez et al. (2017) that perhaps the Bouse *Streptochilus* species were benthic foraminifers (i.e., *Bolivina*). The broader context of this study would suggest that the Bouse *Streptochilus* calcified in a mildly brackish lake and in water with a δ18O value that was below that of normal seawater.

Invoking a benthic lacustrine *Streptochilus* population is a controversial interpretation, but a combination of nomenclatural and genetic evidence supports this conclusion. One of the more abundant benthic foraminifers in the SBF is *Bolivina subexcavata* (McDougall and Miranda Martinez, 2014; Bright, 2017). *Bolivina subexcavata* is synonymized with *Bolivina variabilis* (Hayward et al., 2010). Darling et al. (2009) demonstrated that modern benthic *Bolivina variabilis* and modern planktic *Streptochilus globigerus* are actually genetically identical. Additional work by Kucera et al. (2017) has identified five phylogenetic clades within *Bolivina*, one of which is populated by exclusively planktic individuals as might be expected, one of which is populated by exclusively planktic individuals (*“Streptochilus”*), and three clades of which are populated by both benthic and planktic individuals. *Bolivina* and *Streptochilus* are the same foraminifer that can express two different test morphologies (e.g., Smart and Thomas, 2007; Kucera et al., 2017), and thus, *Streptochilus* is a junior synonym of *Bolivina* (Darling et al., 2009). *Bolivina variabilis* might be able to grow to adulthood and express both *Bolivina* and *Streptochilus* test morphologies in both planktic and benthic environments (Darling et al., 2009). *Bolivina variabilis* has been reported living in Lake Qarun in Egypt (Abu-Zied et al., 2007), but to our knowledge there are no records of *Streptochilus* living in lakes. A detailed assessment of *Bolivina variabilis* test morphologies in Lake Qarun might be an informative endeavor. The isotopic data presented here provide compelling evidence that the SBF contains the first benthic (e.g., Miranda-Martinez et al., 2017) and lacustrine occurrence of *Streptochilus*.

A benthic *Streptochilus* interpretation carries important consequences for the SBF debate. The presence of planktic foraminifers has long been offered as strong evidence for a marine environment (Smith, 1970; McDougall, 2008; McDougall and Miranda Martinez, 2014). *Streptochilus* spp. comprise ~97% of the rare planktic foraminifers identified from the SBF (McDougall and Miranda Martinez, 2014; Bright, 2017). If the Bouse *Streptochilus* species are simply morphological variants of benthic *Bolivina*, regardless of their test morphology, then their ecological constraints implying open-marine conditions would be lost. Numerous *Bolivina* species have been reported from extant and paleolakes (Arnal, 1958; Resig, 1974; Patterson, 1987; Abu-Zied et al., 2007) and from dilute coastal lakes (Kane, 1967; Rao et al., 2000).

Evidence for a Lacustrine Population of Amphibalanus subalbidus

The barnacle species found in the SBF is *Amphibalanus subalbidus* (Van Syoc, 1992; Pitombo, 2004). Although barnacles are typically considered marine organisms, the salinity tolerance of *A. subalbidus* extends below 3000 mg L⁻¹ (Poirier and Partridge, 1979; Dineen and Hines, 1994), which approaches the freshwater category as defined by Williams (1964). Several other barnacles in the genus *Amphibalanus*, such as *Amphibalanus improvisus* and *Amphibalanus amphitrite*, have also been found successfully living at freshwater or near-freshwater salinities (Fyhn, 1976; Kennedy and DiCosimo, 1983). Barnacles have been reported from unquestionably nonmarine environments, such as pumping stations (Shatoury, 1958), desert oases (Pint et al., 2017), and lakes (Plaziat, 1991; Geraci et al., 2008).

Well-preserved barnacle fragments were analyzed from multiple exposures of the basal carbonate unit in the Blythe basin, including four samples throughout a critical outcrop containing sigmoidal bedding featured in O’Connell et al. (2017), which is often, but not always, associated with tidal environments (Tinterri, 2011). Sigmoidal bedding has been described from both a littoral shoal or bar-form environment in the Green River Formation (Keighley et al., 2003), and from a littoral dune bed-form environment in Lake Bonneville (White, 1996), for example. None of the SBF barnacle fragments yielded δ18O-δ13C values that are consistent with calcification in normal seawater or with calcification in a Colorado River estuary (Fig. 12).

In this exercise, a marine analogy is delineated by the δ18O-δ13C values of late Pleistocene planktic foraminifers from the central Gulf of California, from fully marine barnacles from various sites around the globe, and from modern barnacles from the intertidal zone of Isla Montague (Fig. 1). Modern barnacles from the saline lacustrine Salton Sea (Fig. 1) are included for context. An estuarine analogy is delineated by the δ18O-δ13C values in individual growth bands of pre-dam-era *Mulinia* clams from the modern northern Gulf of California (Fig. 1). The lowest *Mulinia* δ18O-δ13C values (Fig. 12) are from Isla Montague (Fig. 1) and are thought to represent calcification during peak pre-dam-era Colorado River runoff, when voluminous spring snowmelt discharge on the river lowered the δ18O-δ13C values in the northern Gulf of California (Rodriguez et al., 1901).
Nearly all of the SBF barnacles analyzed thus far yielded $\delta^{18}$O values lower than even the most river-influenced estuary analog (Fig. 10). Accounting for a roughly $+1\%$ offset between the *Mulinia* shells (aragonite) and the barnacle plates (calcite) does not affect the outcome of this comparison in any appreciable manner. Several SBF barnacle samples have isotopic compositions more indicative of calcification in freshwater (Fig. 12). Thus, the SBF barnacles likely calcified in a lacustrine rather than a marine or estuarine environment.

**A Brackish Lake–Freshwater Plume Model for the Southern Bouse Formation**

A brackish lake and freshwater plume model (Fig. 15) can explain the isotopic complexity preserved in SBF carbonates (Figs. 13A and 13B), as well as explaining the slight offset in $^{87}$Sr/$^{86}$Sr ratios measured in micrite and biologic material at Hart Mine Wash (Fig. 5). In this model, the high $\delta^{18}$O$_{ost}$ values and the higher $^{87}$Sr/$^{86}$Sr ratios in the fish bone and *Cyprideis* calcite are interpreted as reflecting the average open-water $\delta^{18}$O value and $^{87}$Sr/$^{86}$Sr ratio of paleolake Blythe (Fig. 15). The much lower $\delta^{18}$O$_{sed}$ values and lower $^{87}$Sr/$^{86}$Sr ratios in the micrite represent more seasonally limited epilimnic calcium carbonate production in paleolake Blythe during spring/summer floods on the early Colorado River (Fig. 15).

This model is analogous to the freshwater plumes that are observed on Lake Turkana, Kenya (Yuretich and Cerling, 1983), and on Lake Van, Turkey (Reimer et al., 2009). At Lake Turkana, reduced surface-water total dissolved solids (TDS) values are seen as far as 50 km from the mouth of the Omo River during flood season (Yuretich and Cerling, 1983; Halfman et al., 1989), indicating that density and salinity contrasts between the flood plume and the open lake water can be maintained for some distance and for a period of time. A similar, but smaller, offset between $\delta^{18}$O$_{ost}$ and $\delta^{18}$O$_{sed}$ values also occurs in...
Lake Turkana (Halfman et al., 1989). Circulation patterns in Lake Turkana also vary seasonally, such that Omo River water interacts with different parts of the lake depending on the time of year (Yuretich and Cerling, 1983). At Lake Van, carbonate production occurs during the summer, when inflowing rivers provide additional calcium to the lake (Reimer et al., 2009). The resulting carbonate is then distributed within the basin by eddies and currents (e.g., Stockcheck et al., 2012). Similar interannual (or longer-scale) variability in the distribution of the proposed Colorado River plume might have promoted considerable complexity in the distribution of isotopic values in SBF carbonates (Figs. 14B and 14D).

The plume model hypothesizes that the open water of paleolake Blythe was mildly brackish with a high, evaporation-driven $\delta^{18}O$ value and a high, possibly more regional groundwater-derived $^{87}$Sr/$^{86}$Sr ratio (Fig. 15; e.g., Crossey et al., 2015). The early Colorado River flood plume would have been composed of very dilute and $^{18}O$-enriched snowmelt sourced in the Rocky Mountains. Predam-era *Anodonta* shells from the modern Colorado River yield $\delta^{18}O$ values approaching $-18\%$ (Dettman et al., 2004), for example. The $^{87}$Sr/$^{86}$Sr ratio of the early Colorado River is not precisely known. However, the $^{87}$Sr/$^{86}$Sr ratio in rivers with large watersheds that incorporate a variety of bedrock lithologies can fluctuate on seasonal timescales (Rai and Singh, 2007; Santos et al., 2015). The $^{87}$Sr/$^{86}$Sr ratio of the modern Colorado River (0.70945 ± 0.00082, n = 9) is slightly lower than the $^{87}$Sr/$^{86}$Sr ratio of modern regional groundwater (0.71120 ± 0.00492, n = 44), and the $^{87}$Sr/$^{86}$Sr ratio of modern Colorado River tributaries is even lower still (0.70874 ± 0.00074, n = 7; Crossey et al., 2015). The high $^{87}$Sr/$^{86}$Sr ratios found in the 12–6 Ma Hualapai Limestone (0.7114–0.7190) provide context for the $^{87}$Sr/$^{86}$Sr ratio of regional groundwater in the western Grand Canyon area just prior to the deposition of the Bouse Formation (Crossey et al., 2015). Perhaps spring floods (Fig. 12) on the early Colorado River had a low $^{87}$Sr/$^{86}$Sr ratio derived more from surface water in its watershed, while baseflow conditions had a higher $^{87}$Sr/$^{86}$Sr ratio derived from an increase in the relative proportion of groundwater in the river. Incomplete mixing of this seasonal flood plume in paleolake Blythe would have led to episodic carbonate having a much lower $\delta^{18}O_{\text{calc}}$ value and a lower $^{87}$Sr/$^{86}$Sr ratio than carbonate formed deeper in the lake (e.g., Halfman et al., 1989; Sun et al., 2011). The low episodic $\delta^{18}O_{\text{calc}}$ values could have been enhanced by high summer temperatures and additional contributions from kinetic disequilibrium effects that are associated with rapid calcification (e.g., Gabitov et al., 2012; Watkins et al., 2014).

Previously published bulk sediment samples and biologic carbonates from multiple locations around the Blythe basin yielded highly variable $\delta^{18}O$–$\delta^{13}C$ values (Figs. 14B and 14D) and can display clear location- and material-dependent variability (Fig. 14D). Some of the heterogeneity in the bulk carbonate results (Fig. 14B) might be explained by variability in the strength, mixing, and spatial distribution of the early Colorado River plume (e.g., Yuretich and Cerling, 1983; Fig. 15), although some of the variability appears to be related to the material analyzed (Fig. 14D). Note that previously published $\delta^{18}O$–$\delta^{13}C$ values from bivalves and from shells of the intertidal marine gastropod *Batillaria* at Milipitas Wash (Fig. 1) overlap with the *Candona* $\delta^{18}O$–$\delta^{13}C$ values from the lacustrine NBF in the Chemehuevi basin (Figs. 13B and 14D), over 100 km to the north. The offset in $\delta^{18}O$ values between the aragonitic bivalve and snail shells (roughly +1‰) and the calcitic ostracode valves is on par with common *Candona* vital effects (+1‰ to +2‰; e.g., von Grafenstein et al., 1999). At the scale of this data set, these materials can be directly compared. The southern Blythe bivalves and snails calcified in a water mass that was isotopically similar to the water in paleolake Chemehuevi; a marine or estuarine origin for these shells is not required.

**Reconciling Marine Sedimentology in a Lacustrine Environment**

The proposed tidal sedimentary structures recognized by O’Connell et al. (2017), such as potential tidalites and sigmoidal bedding, seemingly require a connection to the ocean, yet the broader isotopic and ecological constraints on the SBF presented here are incompatible with a marine or normal estuarine environment. Two possible explanations that could reconcile tidal features with a nonmarine environment include deposition in a freshwater estuary (i.e., slight modification of Crossey et al., 2015) or deposition in a tidally influenced Colorado River–fed lake situated well inland of the coast but at or near sea-level (Fig. 16). This tidally influenced lake would have possessed a long residence time that would have allowed for a mild increase in salinity and for the high $\delta^{18}O$ values measured in a variety of biologic carbonates, similar to what is observed at Lake Tanganyika (Dettman et al., 2005). Freshwater estuary environments with meter-scale tidal amplitudes are known to occur hundreds of kilometers inland from the ocean (e.g., Geyer and Chant, 2006; Gugliotta et al., 2017) and are populated by continental organisms (e.g., Sousa et al., 2005; Gugliotta et al., 2017). Examples of ancient freshwater rhythmite deposits entirely lacking marine fossils have also been reported in the literature (Kvale and Mastalerz, 1998). However, the freshwater portion of a normal estuary would be strongly influenced by river water (e.g., Ingram et al., 1996a) and might not produce the high, nearly marine-like *Candona* $\delta^{18}O$–$\delta^{13}C$ values measured in this study (Fig. 13B). Alternatively, the Lake Wahloha-Waipori complex (Schallenberg et al., 2003; Schallenberg and Burns, 2003) and Pitt Lake, British Columbia (Ashley, 1977; Ashley and Moritz, 1979), are restricted freshwater lakes situated 12 and 30 km inland, respectively, yet they experience tidal ranges of up to 0.5 m and up to 1–2 m, respectively (Hodgins and Quick, 1972; Ashley, 1977; Schallenberg et al., 2003; Schallenberg and Burns, 2003). To our knowledge, the isotope systematics of the Lake Wahloha-Waipori complex or Pitt Lake have not been studied. Nonetheless, there are numerous modern analogues where tidal influences can be observed in freshwater lacustrine environments, and nonanalogue environments may have existed in the past.

A freshwater estuary or tidally influenced lake model could potentially reconcile many of the peculiarities of the SBF. Southern Bouse carbonates would have been deposited at or near sea-level but a considerable distance inland and well away from seawater or a salt wedge, and thus, they would completely lack any marine geochemical indicators. In this model, the diverse foraminifer
assemblage that is often cited as evidence for marine conditions (McDougall and Miranda Martínez, 2014, 2016; Dorsey et al., 2018) would instead comprise a large proportion of autochthonous euryhaline benthic foraminifers, such as *Ammonia beccarii*, which were living in relatively dilute conditions (e.g., Otvos, 1978; Boudreau et al., 2001). The autochthonous benthic foraminifers are mixed with small numbers of exotic planktic foraminifers that were reworked inland by strong flood tidal currents or other processes (e.g., Arnal et al., 1980; Ghosh et al., 2009). Foraminifer species richness is highest at the southern end of the Blythe basin (McDougall and Miranda Martínez, 2014), which could lend support to this model. Perhaps small amounts of seawater periodically entered Paleolake Blythe during storm surges or during the highest tides (e.g., Crossey et al., 2015), as sometimes happens at the Lake Waihola-Waipori complex (Schallenberg and Burns, 2003). If so, then they would have been short-lived events, and subsequent dilution or flushing (e.g., Schallenberg and Burns, 2003) would have limited or even prevented the preservation of a marine geochemical signal in SBF carbonates.

Projecting a Gulf of California tidal influence nearly 60 km inland to the Blythe basin is not unreasonable. The historical tidal range on the modern Colorado River estuary is ~10 m, and tidal influences are observable 50 km upstream (Thompson, 1968; Nelson et al., 2013; Zamora et al., 2013). This distance is comparable to the distance between the southern Blythe basin and Yuma (Fig. 1). If the uppermost reaches of the early Gulf of California during the late Miocene was narrower and more constricted than the modern bathymetry, then the paleo–tidal range could have exceeded 10 m. A modern analogy might be the narrow upper reaches of the Bay of Fundy, New Brunswick, where maximum tidal ranges are on the order of 15–16 m (Desplanque and Mossman, 2001) and are currently one of the largest tidal ranges on Earth (Archer, 2013). Nonanalogue paleo–tidal ranges in excess of 15 m in the early Gulf of California could have conceivably existed in the past.

This study does not directly address questions about the original elevation of the SBF. The late Miocene (and younger) tectonic history of the Blythe basin is a topic of active research (Beard et al., 2016; Karlstrom et al., 2017). Additional information is clearly needed to test the viability of the tidally influenced lake model and to determine if the SBF can be reconciled with formation near sea level. This study does suggest that the early Colorado River was present during deposition of the SBF, similar to interpretations by Crossey et al. (2015). Thus, the SBF must be younger than 5.6 Ma and is likely younger than 5.3 Ma, based on upstream chronological constraints for the arrival of the early Colorado River in the Cottonwood basin (Fig. 1; House et al., 2008; Pearthree and House, 2014; Crow et al., 2017).

The large and abrupt decreases in δ¹⁸O values in both the NBF and SBF (Fig. 9) are more consistent with the overspilling chain-of-lakes model for
the integration of the lower Colorado River corridor (e.g., House et al., 2008; Peartree and House, 2014; Bright et al., 2016). A tidally influenced lake model (Fig. 16) for the SBF is supported by multiple lines of evidence (O’Connell et al., 2017) and can also comprehensively integrate many of the seemingly irreconcilable data sets generated thus far from the SBF. However, the abrupt decrease in the δ¹⁸O values across the distinctive clay layer in Hart Mine Wash and Milpitas Wash (Fig. 9) seems better explained by a geologically instantaneous event associated with lake overspilling (e.g., Bright et al., 2016), which would not be consistent with direct tidal influence on the paleolake. The distinctive clay layer may be a xenocoformity (e.g., Carroll, 2017) and perhaps deserves additional study. Nonetheless, the detailed comparison with the lacustrine NBF in Chemehuevi basin provides compelling, reproducible, and internally consistent evidence that the basal carbonate unit of the SBF was deposited in a mildly brackish lacustrine environment. A marine environment (e.g., Dorsey et al., 2018) is unable to account for the ecological complexities of the mixed microfossil assemblage or the isotopic constraints provided by the various types of carbonates (inorganic and biologic) detailed in this study. No evidence for an estuarine environment was found either. The northern and southern Bouse Formations were deposited in mildly brackish (~5000 to <10,000 mg L⁻¹) lakes with high water δ¹⁸O values and compositions that mimicked dilute seawater. Both lakes were populated by an interesting but calculable mix of continental and opportunistic marginal marine organisms (i.e., euryhaline benthic foraminifers) before abruptly overspilling and transitioning to through-flowing freshwater environments (e.g., Anadón et al., 2008; Bright et al., 2016).

### CONCLUSIONS

We analyzed one outcrop of lacustrine NBF sediments and four outcrops of controversial SBF sediments to further understand the depositional environment of the SBF. Ecological constraints implied by the respective ostracode faunas, combined with the isotopic composition of the actual ostracode valves themselves, were used to test competing marine, estuarine, and lacustrine models for the origin of the SBF.

The NBF and SBF share a surprisingly large number of similarities in their microfossil assemblages and their stable isotope characteristics. Such similarities would be unexpected if the NBF was a lacustrine deposit and the SBF was a marine or estuarine deposit. Similarities would be expected if the NBF and SBF were both deposited in similar, early Colorado River–fed, lacustrine environments. Collectively, the results from this study strongly suggest that the SBF was deposited in a mildly brackish lake, similar to that proposed by Spencer et al. (2008, 2013), before finally draining. The results of this study are strongly incompatible with a marine (e.g., Dorsey et al., 2018) or estuarine depositional environment for the SBF. The combination of a freshwater Colorado River plume mechanism into a terminal-to-open-basin lacustrine transition can comprehensively explain the broader faunal and isotopic discrepancies that have hindered the development of an internally consistent and cohesive interpretation for the origin of the SBF. An alternative model invoking a freshwater plume into a tidally influenced brackish lake that was located at or near sea level should also be considered. Regardless of its original elevation, the SBF was deposited in the last of a chain of mildly brackish to freshwater lakes that record the southward advance of the terminus of the early Colorado River as it migrated toward the early Gulf of California.

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