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Pedogenic and lacustrine features of the Brushy Basin Member of the Upper Jurassic Morrison Formation in western Colorado: Reassessing the paleoclimatic interpretations

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Key words: Aridisol, Inceptisol, calcrete, palustrine, lacustrine, pedogenesis.

Abstract. Study of the pedogenic features of the Upper Jurassic Morrison Formation in western Colorado, USA, shows a clear difference in the types of paleosols between the strata of the lower and upper Brushy Basin Member. Lower Brushy Basin paleosols are mostly calcareous Aridisols with Stage I through Stage III calcrete Bk horizons, abundant root traces, occasional vertic features, but only rarely with ochric epipedons. Upper Brushy Basin paleosols are mainly thicker and commonly display ochric epipedons and well-developed Bt and Bw horizons. We assign these paleosols to the order Inceptisol. Limestones occur in the Brushy Basin Member and include both uniformly micritic limestones and limestones with strongly brecciated textures. The former contain sparse body fossils and charophyte debris, while the latter are characterized by clotted-peloidal fabrics with circumgranular cracking and silica replacement. We interpret these limestones as the deposits of carbonate in small water bodies on a low-gradient flood plain, with the textures resulting from pedogenic reworking of the carbonate sediment. We find no evidence for the presence of extensive lacustrine or wetlands (Lake T'oo'dichi') deposits in the study area. The paleoclimate suggested by all of these features is strongly seasonal, but subject to variations on orbital (precessional and higher) timescales causing intervals of semi-aridity during weaker monsoons, to alternate with sub-humid periods during stronger monsoons. The apparent long-term change in climate during Brushy Basin deposition potentially resulted from northward drift of North America.

INTRODUCTION

The collective continental sedimentary deposits of the Morrison Formation constitute one of the best known sedimentary units of Late Jurassic age in the world, mainly due to the continental fossil assemblage contained therein (*e.g.*, Dodson *et al*., 1980; Foster, 2003, 2007). The faunal diversity of the formation is justly famous; Chure *et al*. (2006), in their update of their own earlier biodiversity compilation (Chure *et al*., 1998), list over 30 genera of Morrison Formation dinosaurs alone. Indeed, it was this formation that fueled the "bone wars" of the Great Dinosaur Rush of the late nineteenth century (Lucas, 2007). Such obvious faunal diversity has naturally stimulated interest in understanding ancient ecosystems that were dominated by giant herbivores, as they were during the Late Jurassic (Foster, 2003, 2007).

Paleoecological study both requires and informs understanding of the concomitant paleoclimate, but the paleoclimatic interpretation for Morrison deposition has proved rather contentious. Studies of the Morrison paleoflora document a diverse and abundant flora of algae, bryophytes, ferns, gingkophytes, horsetails, cycadophytes, bennetites and

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conifers (Tidwell, 1990; Ash, Tidwell, 1998), suggesting moist soil conditions and a humid climate. This interpretation is consistent with local lacustrine faunal assemblages, including ostracodes, conchostracans, pulmonate gastropods, aquatic insects, many fish species and frogs and aquatic turtles, which indicate locally abundant supplies of fresh surface water (Gorman *et al*., 2008).

The interpretation of abundant moisture based on floral and faunal evidence is at apparent odds with those developed from sedimentary evidence, however. Initially, White (1886) envisioned Morrison Formation deposition as occurring mainly in lacustrine environments, but this inference was based almost entirely on the antiquated assumption that large sauropods led a mostly aquatic lifestyle due to their enormous bulk. Hatcher (1901) pictured the environment of the Morrison depositional basin as much like the modern Amazon basin, wet and lushly vegetated, a view also maintained by Mook (1915, 1916). Mook (1916) went so far as to develop a sedimentologic model for the Morrison, envisioning a series of low-gradient alluvial fans coalescing to form a broad alluvial plain, with lakes occupying abandoned valleys between the major streams. Stokes (1944) was the first to suggest that the marshy alluvial plain described by earlier authors was in fact drier than interpreted, that ephemeral, possibly seasonal lakes existed in a semi-arid climate regime, and also noted the presence of eolian sandstones in the lower Morrison (though this has been disputed, *e.g.*, Anderson, Lucas, 1998).

Craig *et al*. (1955) more-or-less followed Mook (1915, 1916) in interpreting the extensively mudstone-dominated sequences of the Brushy Basin Member as distal fluvial and overbank deposits. Dodson *et al*. (1980), in their still very useful study, specifically cited a lack of sedimentary evidence for the presence of large or deep lakes in the Morrison Formation. Nevertheless, Turner-Peterson *et al*. (1980) reinterpreted the same sediments as recording predominantly lacustrine deposition. Significantly, zeolites were recognized in the interpreted lacustrine sequence of the Brushy Basin Member (Bell, 1983; Turner-Peterson, 1985, 1987), leading to the development of the Lake T'oo'dichi' sedimentary model by Turner and Fishman (1991), who presented the occurrence of these minerals as evidence for a large, closed-basin alkaline lake that covered the eastern half of the Colorado Plateau during an interval of arid climate while the Brushy Basin sediments were deposited (Demko, Parrish, 1998). Turner and Peterson (2004) subsequently retreated from the original claim of a single, closed-basin lake of enormous size, but maintained the role of an arid climate in explaining the presence of the diagenetic minerals in an extensive complex of lakes and wetlands fed mainly by groundwater.

Nevertheless, deposition of the Brushy Basin sediments under a highly arid climate regime, whether in wetlands or lakes and streams, is difficult to reconcile with the iconic dinosaurian megafauna. Foster (2003), in his paleoecological analysis of the Morrison, noted that the formation is characterized by both great abundance and diversity of the megafauna, and that the structure of the paleocommunities remained relatively consistent throughout the history of deposition. Turner and Peterson (2004) compared the Morrison ecosystem to the modern savannah, with most forest limited to riparian zones, but relatively few perennial streams. But, this interpretation of the Morrison environment envisions a drier climate than that pictured by Moberly (1960), who compared the Morrison environment to the plains of modern-day northern Argentina, with abundant but very seasonal precipitation across a well-vegetated, partially forested plain dissected by streams, lakes and wetlands. In this interpretation, seasonal conditions alternated between dry and very wet with extensive flooding. Stevens and Parrish (1999), in attempting to understand the biomechanics of sauropods, concluded that the various genera of diplodocids that coexisted during Morrison deposition, *e.g.*, *Diplodocus* and *Apatosaurus*, fed on the same types of vegetation, which therefore was necessarily abundant. However, the issue is far from settled, as Englemann *et al*. (2004) maintain that this fauna was hosted primarily by riparian and wetland refugia in an otherwise inhospitable environment.

Here, we describe pedogenic and lacustrine features observed in three sections of the Brushy Basin Member of the Morrison Formation measured in western Colorado (Figs. 1–2) and discuss paleoclimatic implications of these features, particularly with regard to the context of earlier studies. Our goal is to use these new data to reconcile apparent disagreements between geological and paleontological evidence over the paleoclimate during Morrison Formation deposition. This is a companion study to the research of Galli (2014 – this volume), which focuses on the fluvial architecture of the channel sandstones of the Brushy Basin Member.

GEOLOGIC SETTING AND AGE

The Morrison Formation comprises mostly continental sedimentary deposits of Late Jurassic age deposited across a vast area of the Western Interior of North America, defining a depositional basin that extends from central New Mexico to southern Canada, and from central Utah to Nebraska and Kansas (Fig. 1). Sediment was transported to the basin from uplifted regions in the back-arc region to the south and west (Dickinson, 2001). Most of the Colorado Plateau is encompassed within this depositional basin, and Morrison sed-

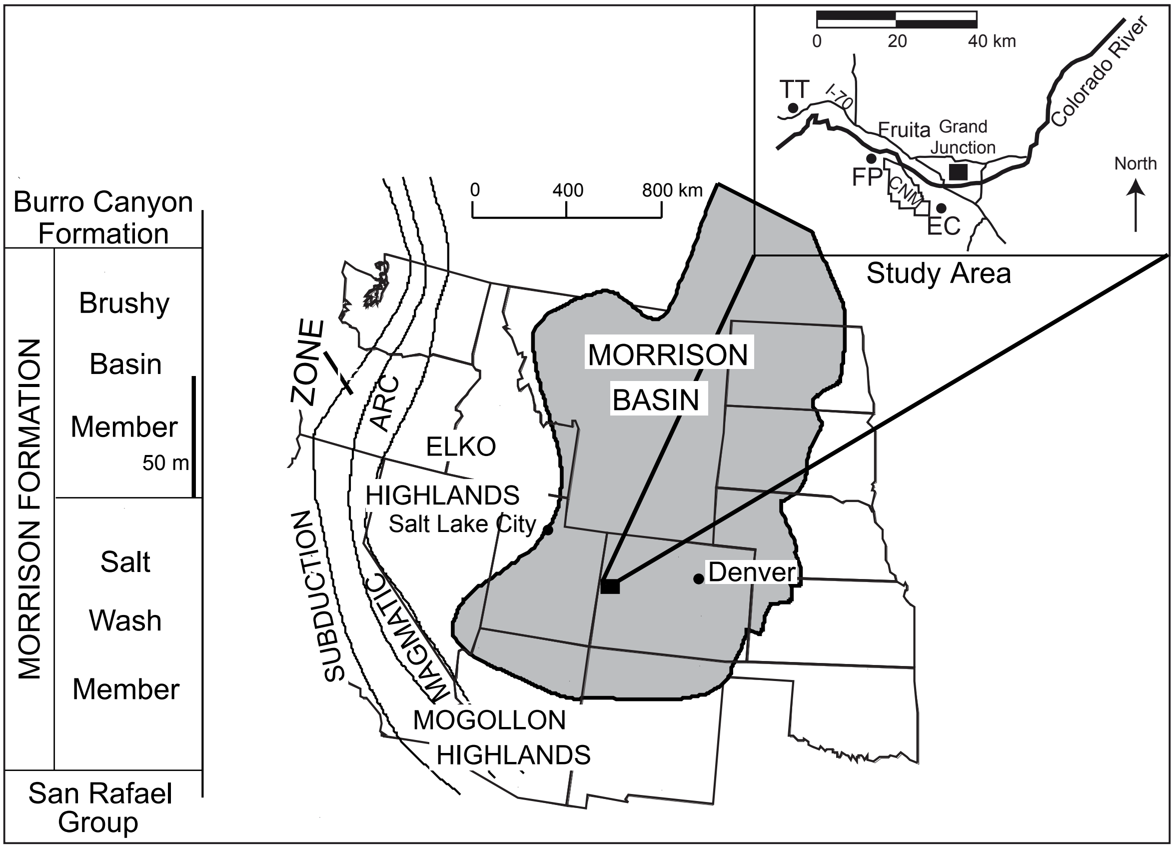

Fig. 1. Map depicting limits of the extent of Morrison Formation deposition, location of the area of this study relative to the Morrison depositional basin, and stratigraphy of the study area modified from Galli (2014 – this volume)

iments there are entirely continental. The age of the formation is fairly well-constrained by palynology to mainly Kimmeridgian, although the youngest strata are early Tithonian in age (Litwin *et al*., 1998). Numerical ages are provided by $^{40}Ar/^{39}Ar$ dating of sanidine from volcanic tuffs in the formation, indicating that deposition took place over an interval of about seven million years of the Late Jurassic (Kowallis *et al*., 1998).

Given the immensity of the depositional area of the Morrison Formation, the stratigraphic relationships between units both within and vertically adjacent to the formation are varied. In the study area, the basal Morrison Formation strata rest unconformably on strata of the Middle Jurassic San Rafael Group assigned variously to the Wanakah Formation (O'Sullivan, 1992) or the Summerville Formation (includes strata termed Tidwell Member of Morrison Formation by some workers; Lucas *et al*., 2006). In western Colorado, the formation is overlain unconformably by the Lower Cretaceous Burro Canyon Formation. Morrison strata are here subdivided into a basal Salt Wash Member that grades upward into the Brushy Basin Member, the latter constituting the bulk of the formation thickness (O'Sullivan, 1992). Within the Brushy Basin Member, a subdivision into lower and upper units has been advocated based on a perceived synchronous change in clay mineralogy, from illite-dominated reddish-brown mudstones in the lower Brushy Basin, to mainly smectitic, gray-variegated mudstones in the upper Brushy Basin (Turner, Peterson, 1999). Trujillo (2006), however, examined the clay mineralogy of the Brushy Basin outcrops across a broad area of the depositional basin and reviewed previous studies, and found no basis for the use of clay mineralogy within the Brushy Basin Member for stratigraphic correlation. The present study is based on measurement and sampling of three complete sections of the Morrison Formation in western Colorado, an area occupying the northeastern corner of the Colorado Plateau (Fig. 1).

METHODOLOGY

The Brushy Basin Member of the Morrison Formation was studied at three locations: 1) the Trail Through Time locality is several kilometers from the Colorado-Utah border, near the Mygatt-Moore Dinosaur Quarry, in the Rabbit Valley Research Natural Area; 2) the Fruita Paleontological Resource Area locality is north of the west entrance to Colorado National Monument, just over 19 km east of the Colorado-Utah border; 3) Echo Canyon is near the eastern boundary of Colorado National Monument, 42 km from the Colorado-Utah border. For additional details of these locations, see Galli (2014 – this volume).

Paleoclimate-sensitive lithologies, pedogenic features and carbonates in particular, were examined in the field and sampled for laboratory analysis. Standard-thickness petrographic sections were examined to determine microfabrics and fossil content. The mineralogy of cements was determined by X-ray diffraction analysis of bulk powder samples with a Bruker Phaser D2 X-ray diffractometer using Cu-kα radiation.

PALEOCLIMATE-SENSITIVE FEATURES

OUTCROP SECTIONS

In western Colorado, the Brushy Basin Member consists mainly of mudstone, with some thick sandstone intervals, primarily in the lower Brushy Basin Member, and thinner sandstones, limestones and volcanic tuffs throughout (see fig. 19 in Galli, 2014 – this volume). On outcrop, it is quite evident that the ledge-forming sandstones occur mainly lower in the Bushy Basin section, which is a zone transitional between the Salt Wash and Brushy Basin members, and the mudstone-rich upper Brushy Basin forms rounded hills. Although the use of clay mineralogy to distinguish between the lower and upper Brushy Basin does not pass rigorous examination, as noted above, it is true that in the study area the mudstones of the lower Brushy Basin are for the most part reddened compared to the gray to greenish variegated mudstones seen higher in the section (Fig. 2A). The change from lower to upper Brushy Basin is generally transitional, with locally abrupt contacts between the reddish lower strata and overlying gray strata, but commonly additional red mudstones occur higher in the section above the first appearance of the gray mudstones (Fig. 2B).

PEDOGENIC FEATURES

Lower Brushy Basin

As stated above, mudstones in the lower Brushy Basin Member mostly display reddish-brown color (5YR 4/4 to 5YR 4/6) and contain abundant and varied pedogenic features, including varying stages of horizonation, drab root traces, various pedogenic fabrics and carbonate nodules. Individual paleosol profiles in the lower Brushy Basin are generally thin (<1 m) and commonly truncated. Most paleosols in the lower Brushy Basin Member are calcareous to some degree, although most do not form prominent ledges on

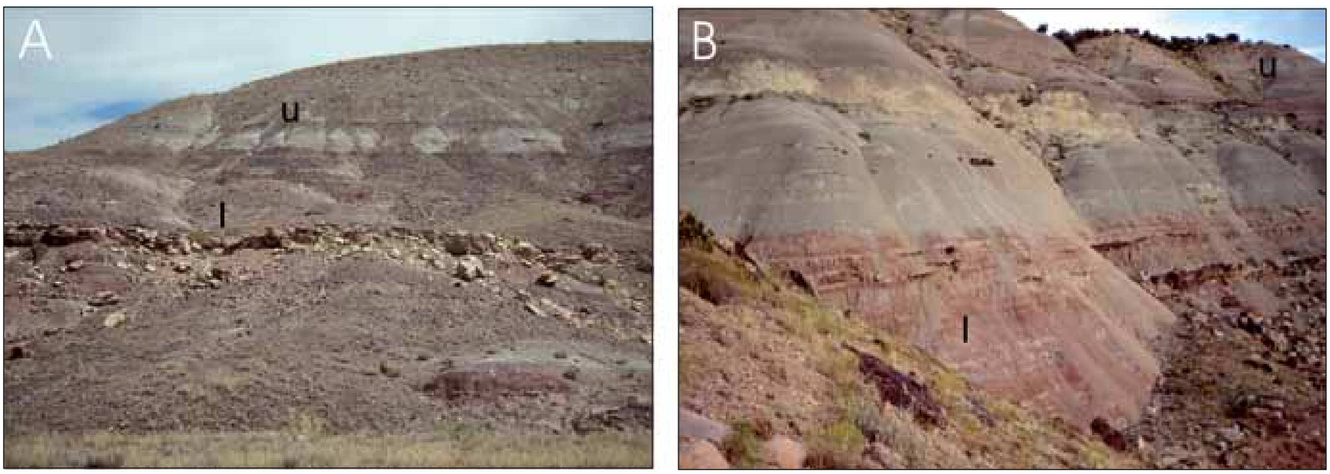

Fig. 2. Differences between lower (l) and upper (u) Brushy Basin Member lithologies

A. The mudstones of the lower Brushy Basin are reddened compared to the gray to greenish variegated mudstones seen higher in the section (Fruita Paleontological Resource Area). **B.** Locally the change from lower to upper Brushy Basin is transitional, with alternating reddish and gray strata in the contact zone (Echo Canyon section) weathering, as they lack sufficient accumulations of calcite in the B horizon to constitute K horizons; *i.e.*, Stage IV or higher calcrete horizons (Gile *et al*., 1966; Machette, 1985) are generally absent. Nodular carbonate accumulations of soil carbonate, which are considered here distinctly separate from pedogenically altered lacustrine or wetland, *i.e.*, palustrine carbonate, are more common in the lower Brushy Basin than the upper. This distribution may be related to the thinner soil profiles of the lower Brushy Basin, which suggests lower depositional rates during this time, contrary to the interpretation of Demko *et al*. (2004).

Calcrete maturity in the Brushy Basin Member varies from Stage I carbonate accumulation in rooted, burrowed mudstone to Stage III horizons of coalescing nodules. Mature (Stage III) calcretes have abrupt upper contacts and lower boundaries that grade downward through scattered nodular zones to calcareous mudstone. Nodular horizons tend to have only limited lateral continuity, up to several hundred meters at most, and are in some instances laterally equivalent to channel-fill sandstone bodies. Most lower Brushy Basin calcretes are Stage I and II, but, as noted by Galli (2003), the thickness and maturity of these calcretes increases upward in the lower Brushy Basin section. Most lower Brushy Basin paleosols lack a discernable A horizon, although there are exceptions, as seen at Echo Canyon (Fig. 3A). Although this profile is erosionally truncated, there is a pale A horizon about 20 cm thick directly below the ledge-forming carbonate body. Drab root traces (gray to greenish, 5Y 8/1 to 5G 8/1) extend downward from the profile top. The underlying B horizon is reddened, clay-enriched (Bt horizon) and displays a crumb fabric with argillans (clay skins) on the peds. The drab band below marks a hydromorphic zone with a crude prismatic fabric.

Most of the paleosol profiles in the Brushy Basin Member are simple profiles, representing prolonged intervals of pedogenesis during nondeposition, but composite profiles do occur, recording more frequent (but still very intermittent) depositional events. An example is near the top of the lower Brushy Basin as at Echo Canyon (Fig. 3B), where a composite profile occurs near the top of the lower Brushy Basin below a channel sandstone. At this location, Stage III calcrete (Bk) horizons are each 30 to 40 cm thick and separated by ~40 cm, and have abrupt upper contacts and gradational lower contacts.

In lower Brushy Basin outcrops, the carbonate nodules themselves have distinct to indistinct boundaries, but are recognizable by their nodular weathering. The nodules may host crystallaria. These horizons typically are associated with drab reduction haloes that surround root traces and, less commonly, burrows, contrasting with the red mudstone matrix (Fig. 3C). Prepared specimen slabs demonstrate the association of plant roots, which in some instances are calcified (rhizocretions), and invertebrate burrows with

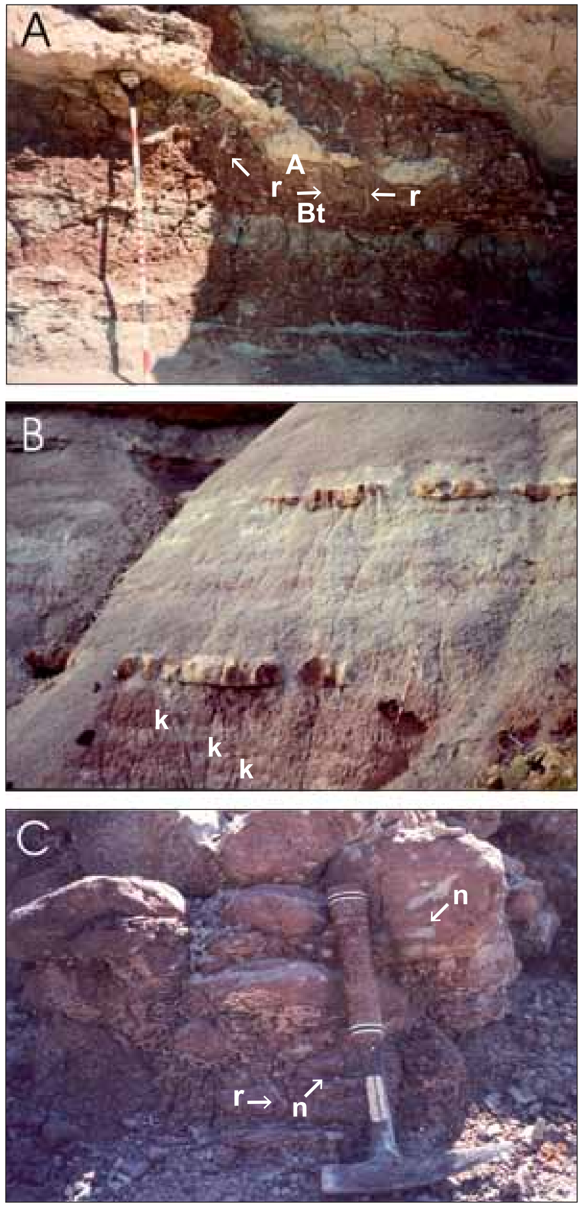

Fig. 3. Features of lower Brushy Basin Member paleosols

A. The profile immediately below the carbonate body (above the staff) is unusual in containing a partially preserved edipedon (A) directly below the ledge, underlain by a clay-rich (Bt) horizon. Root traces (r) are present near the profile top (Echo Canyon section). **B.** A composite profile occurs below the sandstone ledge adjacent to the staff. Stage III calcrete horizons (k) have sharp tops and gradational bases (Echo Canyon section). A similar composite profile may be present above the sandstone ledge next to the staff. **C.** Weakly developed calcareous Aridisol with fine, branching root traces (r) and indistinct carbonate nodules (n)

calcareous nodules (Figs. 4A–B). In thin section, the carbonate nodules show mostly distinct boundaries, but generally only partial displacive calcite cementation.

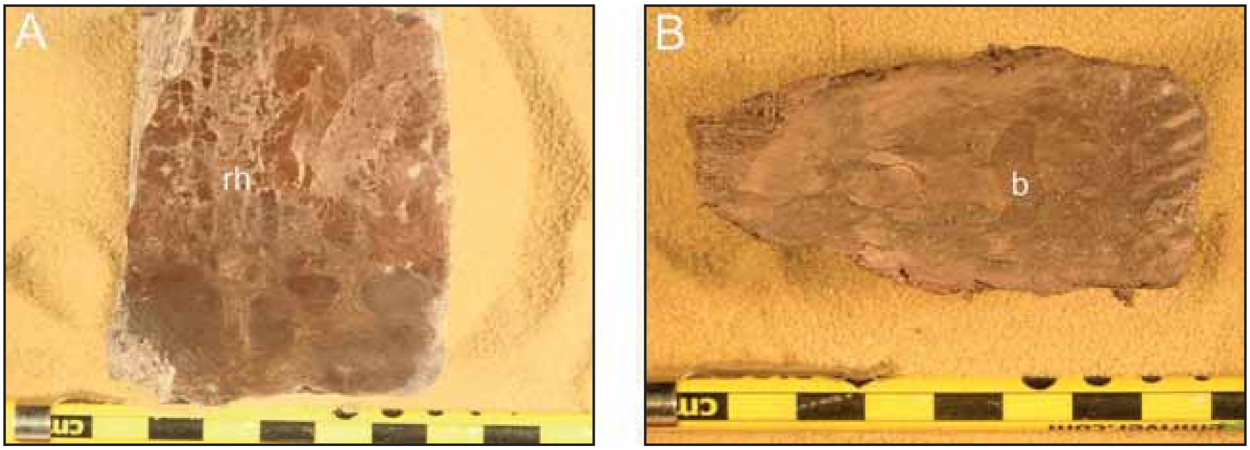

Fig. 4. Features of lower Brushy Basin calcrete

A. This prepared slab displays downward tapering and branching rhizocretions (rh) and a brecciated mudstone fabric. The base of the specimen is a coalescing nodule layer (Trail Through Time section). **B.** Nodules in this specimen vary from distinct (sharp, discrete boundaries) to indistinct. Cross-cutting of nodules by the invertebrate burrow (b) indicates a low degree of induration of the nodules soon after formation (Trail Through Time section)

Noncalcareous paleosols also occur in the lower Brushy Basin, although infrequently. In Figure 5, for example, from the lower Brushy Basin section at Echo Canyon, sandy brown (2.5YR 3/3) mudstone exhibits a semi-platey ped fabric with prominent vertic features and pedogenic slickensides.

Fig. 5. Clay-rich paleosol horizon displaying pedogenic slickenside (ps) surfaces (lower Brushy Basin Member at Echo Canyon)

Upper Brushy Basin

In the upper Brushy Basin Member, paleosol profiles tend to be thicker (several meters), with pronounced lighter-toned A horizons, gradually darkening downward, coincident with increasing clay or cation content in Bt or Bw horizons. This is well-exemplified at Echo Canyon (Fig. 6A). In this section, the “transition” between the lower Brushy Basin (red mudstone) and upper Brushy Basin (gray mudstone) at Echo Canyon is marked by a well-exposed paleosol profile that is 1.6 m thick and relatively complete with A and Bw horizons, developed over a thin channel sandstone body. Possible vertic features are exposed in the fresh exposure of the outcrop to the left of the staff. This profile is overlain by a darker, grayish purple (10R 5/2) B horizon of the profile above. This pattern of repeated thicker, more complete profiles observed in the Fruita Paleontological Resource Area appears typical of the upper Brushy Basin Member in the study area (Fig. 6A).

Interpretation

Strata of the lower Brushy Basin Member host a variety of pedogenic features, but generally lack complete or well-developed paleosol profiles. The mudstones hosting these features are well-oxidized, and the observed features include vertic fractures (*i.e.*, pedogenic slickensides), platey to crumb ped fabrics, root traces and Stage I to Stage III calcretes. Although profiles including epipedons are rare, most

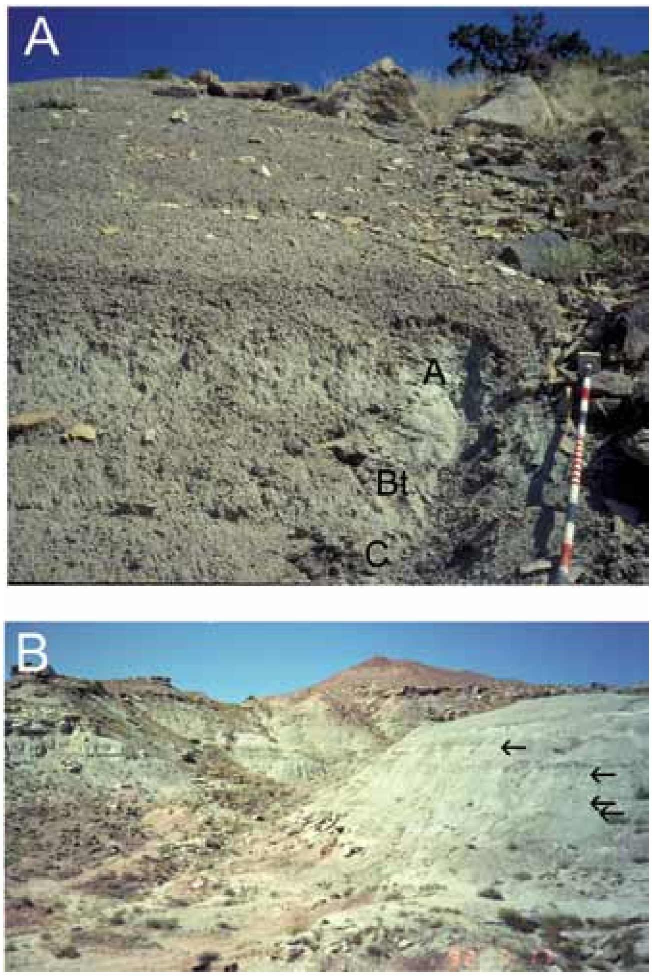

Fig. 6. Upper Brushy basin paleosols

A. Profile to the left of the staff shows a downward darkening from the A horizon (A), through the clay-enriched Bt horizon (Bt) to a lower C horizon (Echo Canyon section). **B.** Multiple A/B pairs (arrowed) are visible (Fruita Paleontological Reserve)

of the paleosols in the lower Brushy Basin appear to be assignable to calcic Aridisols and possibly Entisols, as well as Vertisols. All of these paleosols appear to have developed on alluvial sediments, *i.e.*, on the floodplains of the Morrison Formation streams. Pedogenic carbonate accumulates during the dry season in regions that experience significant seasonal fluctuations in climate, but this encompasses a broad range of climates, from semi-arid to sub-humid (250 to >750 mm precipitation per year). Seasonality of precipitation is also a requirement for the development of vertic features in soils.

The low maturity of the calcretes in the Morrison Aridisols could reflect either greater moisture or lower seasonality. Interestingly, Myers *et al*. (2014) used measurement of the chemical index of alteration and the calcium-magnesium weathering indices of samples from the B horizons of vertic paleosols to determine that precipitation during Morrison deposition was highly variable, but averaged nearly 700 mm per year. This is considerably more humid than predicted by computer modeling; the GCM of Moore *et al*. (1992a) predicted a hot and dry climate (<500 mm precipitation per year) for the Western Interior during the Late Jurassic.

Paleosols in the upper Brushy Basin display thicker, more complete profiles, and typically display eluvial horizons near the top, *i.e.*, they contain ochric epipedons as the uppermost preserved horizons. Although carbonate bodies occur interbedded in these strata, these carbonates are lacustrine or pedogenically modified wetland deposits (see below), and mature pedogenic calcretes (Stage III or higher) are largely absent. The lower portions of these profiles are darkened primarily by illuviated clays and cations. The B horizons are typically calcareous, but generally lack discrete calcareous nodules. Vertic features occur locally. By comparison with modern soils, these profiles most closely resemble Inceptisols to vertic Inceptisols.

The differences between lower and upper Brushy Basin Member paleosols, including increased thickness and decreased calcite accumulation, suggest a more humid climate or higher sedimentation rate, or both, during deposition of the upper Brushy Basin Member sediments. Demko *et al*. (2004) remarked on the likely pattern of a decreasing rate of sedimentation through the earlier interval of Brushy Basin deposition, followed by increasing sediment accumulation rates later, although they project the inflection point of minimum sediment accumulation much later, with sedimentation continuing to decline at least half-way into deposition of the upper Brushy Basin Member. Notably, Demko *et al*. (2004) also inferred an increase in the humidity of the Late Jurassic paleoclimate during upper Brushy Basin deposition.

LIMESTONES

Lacustrine and palustrine limestones

Limestone beds in the Brushy Basin Member are thin, laterally continuous (at outcrop scale) tabular bodies to lensoidal beds ranging in thickness from less than 10 cm to 6 cm. The beds have textures that vary from nearly uniform to extensively brecciated, and most weather to a nodular upper surface on outcrop (Figs 7A, B). Beds typically have sharp upper and lower contacts, and colors vary from gray to reddish brown to reddish-green mottled. The beds consist of low-magnesium calcite, but many contain masses of silica as chert or chalcedony.

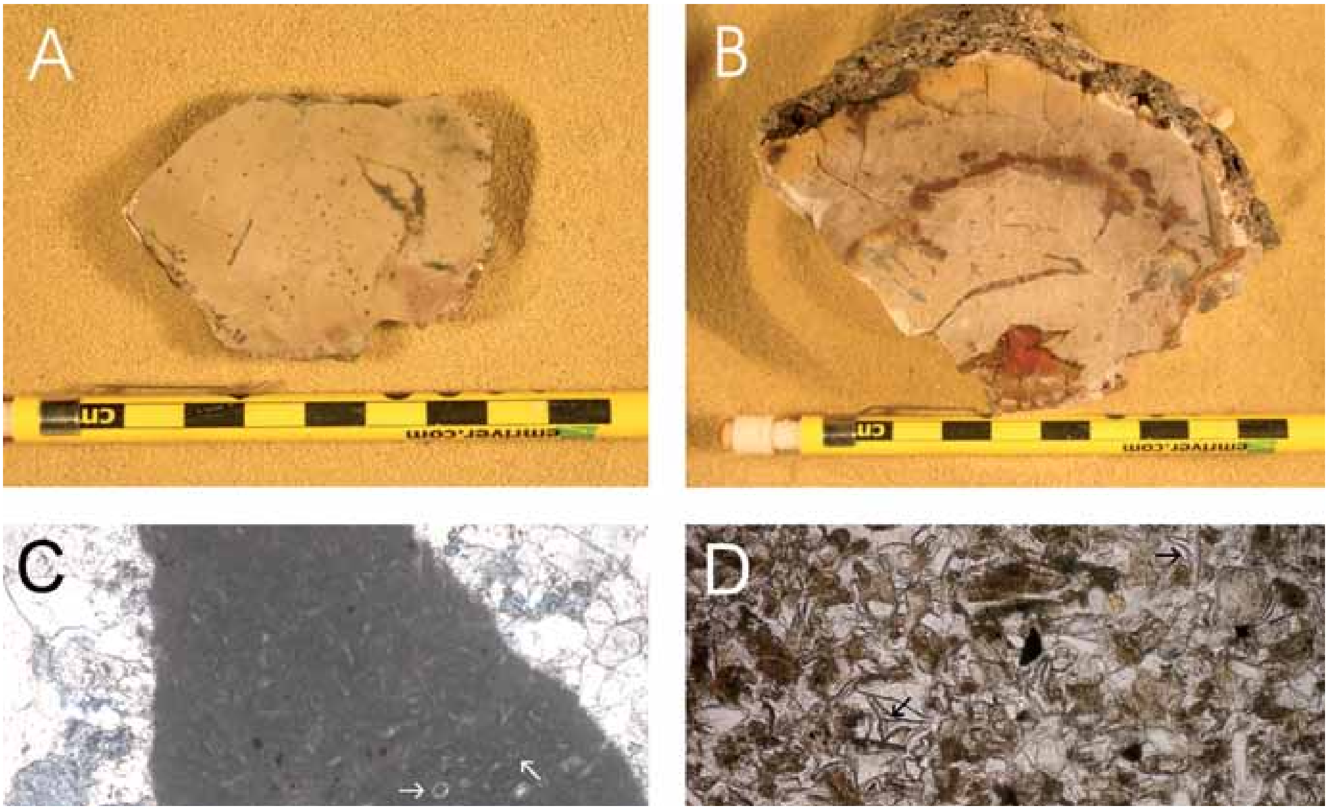

Fig. 7. Features of Brushy Basin limestones

A. Prepared slab of uniform micrite (upper Brushy Basin at Echo Canyon). **B.** Slab from limestone with nodular-appearing upper surface shows brecciated fabric and replacement by chalcedony (upper Brushy Basin at Fruita Paleontological Resource Area). **C.** Charophyte debris (arrows) is abundant in this brecciated limestone with calcite spar-filled veins (upper Brushy Basin at Fruita Paleontological Resource Area, 40×, nicols uncrossed). **D.** Vitric shards (arrows) are abundant in this limestone (lower Brushy Basin at Trail of Time section, 100×, nicols uncrossed)

Uniformly micritic limestones lack any sedimentary structures or significant contributions of bioclastic debris, although fossils are present, mainly ostracodes and charophyte debris (Fig. 7C). Relict shards of volcanic ash are present in some of these limestones in greatly varying concentrations (Fig. 7D). Beds with high concentrations of shards contain albite cement. Rarely, the limestone has a reddish brown color that reflects a significant, fine organic component. What appear on outcrop as carbonate bodies consisting of densely coalescing nodules display lenticular shapes and abrupt to semi-gradational lower boundaries, suggesting that rather than mature calcretes, these are palustrine carbonates that formed by accumulation in water bodies at the surface and were subsequently modified by pedogenesis, not in the subsurface in the B horizon of a soil. In the example shown in Figure 3A, the carbonate body has an obvious erosional lower boundary, suggesting carbonate deposition as a small, discrete water body, possibly an abandoned channel. By contrast, soil carbonate bodies (K horizons) generally have very gradational lower boundaries. The brecciated fabric of these Brushy Basin limestones consists of angular clast domains up to 10 cm in diameter, separated by millimeter- to centimeter-scale polygonal fractures filled by either calcite or silica (Fig. 8A). Fractures between clasts are straight to curved and have any orientation between vertical and horizontal. Clotted to peloidal fabrics with circumgranular cracking are common in this facies (Fig. 8B).

Interpretation

Although distinctly lacustrine sedimentary signatures, *i.e.*, body fossils or sedimentary structures, are generally scarce, there is little doubt that these limestones, both those with uniform and those with brecciated textures, have a lacustrine origin. The limited lateral extent and association of the carbonate bodies with alluvial facies suggests that carbonate deposition took place in small ponds on a low gradi-

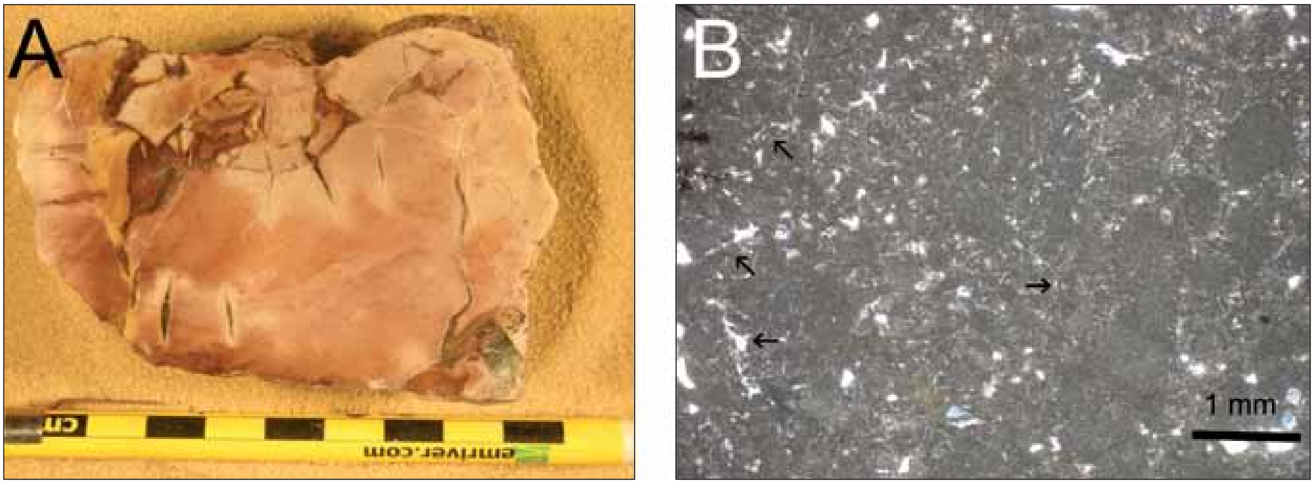

Fig. 8. Palustrine limestone features

A. Prepared slab of limestone with a fabric of large angular clasts and pockets of chert. The coloring is mottled due to ferrugination (upper Brushy Basin at Fruita Paleontological Resource Area). **B.** Limestone displaying peloidal microfabric with circumgranular cracking (arrows; upper Brushy Basin at Time section, 40×, nicols uncrossed)

ent floodout of the Brushy Basin stream system. The clotted peloidal fabric and circumgranular cracking indicate pedogenic modification of the original carbonate texture, a process that often destroys original lacustrine sedimentary fabrics and fossils (Platt, 1989, 1992). The circumgranular cracking resulted from shrinkage cracking during desiccation, potentially on a seasonal basis, and was followed by infilling of the fractures by phreatic cements (Armenteros *et al*., 1997). Color mottling and reddening (ferrugination) is a result of a fluctuating water table pH and/or Eh and vertical translocation of Fe oxides and hydroxides (Platt, 1989). Calcite was replaced locally by silica in meteoric waters, likely sourced from devitrification of volcanic ash. Microbrecciated and peloidal fabrics and mottling are typical features of palustrine carbonates (Platt, 1989, 1992; Platt, Wright, 1992; Armenteros *et al*., 1997; Tanner, 2000). Generally, the abundance of palustrine features in the carbonate suggests semiarid and likely seasonal climate conditions.

DISCUSSION

It is important to note that the observations and interpretations presented above represent the characteristics of the Brushy Basin Member in the study area of western Colorado, and are not necessarily applicable over the broader area of the Morrison Formation depositional basin. Sedimentary indicators of Morrison paleoclimate demonstrate profound regional differences, such as the presence of eolian sandstones to the south and coals in the far northern reaches of the depositional basin (Demko, Parrish, 1998; Demko *et al*., 2004). Previous workers have noted regional differences in Brushy Basin paleosols, with calcic Aridisols dominating to the south, Histisols and Inceptisols more common to the north (Jennings *et al*., 2011), and a west-to-east gradient of vertic Aridisols to argillic Aridisols (Demko *et al*., 2004; Myers *et al*., 2014). Hence, the conclusions of this study should not be considered applicable in their entirety to the entire Morrison depositional basin.

LAKE T'OO'DICHI'

Careful examination of the sedimentary characteristics of the Brushy Basin Member strata in the study area of western Colorado provides data that directly contradict the original Lake T'oo'dichi' supposition, as well as its later iteration as an extensive lacustrine/wetlands complex. First and foremost is the paucity of distinctly lacustrine wetland facies in the study area, which lies within the mapped outlines of Lake T'oo'dichi' (Turner, Fishman, 1991; Dunagan, Turner, 2004). Dodson *et al*. (1980), Anderson and Lucas (1997) and Galli (2003) have noted specifically the lack of sedimentary facies that would be predicted by the existence of a lake occupying the area of Lake T'oo'dichi', even one episodically evaporated to dryness. In particular, these authors, in unrelated studies, described depositional facies, such as crevasse and levee deposits, and channel-fill deposits of high-sinuosity channels, that suggest a depositional setting broadly dominated by alluvial processes. Additionally, these authors commented on the lack of facies typical of large lake bodies, *e.g.*, shoreline features and deltas, or even those features typical of ephemeral lakes. Galli (2003) remarked on the inconsistency of the fluvial paleocurrents measured in the Brushy Basin Member channels with the presence of a large lacustrine body located centrally in the Morrison depositional basin. Furthermore, taphonomic strudies of vertebrate fossil assemblages within the hypothesized confines of Lake T'oo'dichi' (*e.g.*, Morris *et al*., 1996; Richmond, Morris, 1996; Kirkland, 2006) show fluvial transport as a substantial concentrating mechanism of dinosaur carcasses and bones, inconsistent with a lacustrine setting.

Carbonate bodies in the Brushy Basin Member are of limited lateral extent and most exhibit some pedogenic modification, some extensively so. Anderson and Lucas (1997) noted that the size and distribution of Brushy Basin carbonates was more consistent with their deposition in small, floodplain ponds than a vast, aerially extensive lake. We continue to support this assertion. Notably, Dodson *et al*. (1980) recorded a meters-thick partly silicified carbonate unit in the upper Brushy Basin at Dinosaur National Monument, which they interpreted as calcrete, but also noted that a similar lithofacies was lacking in sections at other locations they had studied. A calcrete unit of such thickness would require landscape stability for such an extended interval of time, on the order of 10^5 years (Machette, 1985), that calcrete should be quite common at the same stratigraphic level over a broad region. As this is not the case, it is quite likely that the unit they examined was in fact a pedogenically modified lacustrine carbonate.

Dunagan and Turner (2004) retreated from the original Lake T'oo'dichi' construct, describing the upper Brushy Basin depositional environment as a wetlands-lacustrine complex, which received only a minor proportion of its water from surface flow. They explained the presence of the authigenic minerals through invocation of a groundwater divide created by the Uncompahgre uplift, resulting in a hydrologically closed groundwater basin. Surface waters were alkaline in the arid to semi-arid climate, causing the volcanic ash to undergo syndepositional alteration. However, this hypothesis suffers from several weaknesses, prominent among which is the need for a barrier to groundwater flow (but not surface flow) to explain the presence of high-alkalinity water bodies in the up-gradient direction, while the distal basin contained freshwater bodies. While the Uncompahge uplift had positive topographic expression during the Late Jurassic, the primary regional paleoflow was from highland areas to the west and south towards the north-northeast (Blakey, 2011). There is no sedimentary evidence to suggest that the Uncompahgre Uplift created a barrier to either surface or subsurface water flow and created a hydrologically-closed basin during the Late Jurassic.

Furthermore, there are no carbonate lithofacies in the upper Brushy Basin Member that can be attributed inarguably to spring activity, although spring carbonates tend to have a very limited aerial distribution. Dunagan and Turner (2004) present isotopic data for the Brushy Basin carbonates, but noted the lack of covariance of the oxygen and carbon isotopes. At face value, these data indicate a hydrologically open regime, as isotopic covariance is a feature typical of carbonates in hydrologically closed lakes, and both carbon and oxygen isotopes become evaporatively enriched (Talbot, 1990; Talbot, Kelts, 1990). The authors explain the lack of covariance in the Morrison Formation as a function of continued replacement of evaporated waters by groundwater flow. However, this premise fails to explain the lack of evaporative enrichment of the wetland waters that would result regardless of groundwater influx without some mechanism for removing the accumulated heavier isotopes. Hence, the isotopic data should be accepted at face value as consistent with carbonate deposition in hydrologically open water bodies.

Initially, Lake T'oo'dichi' was a hypothetical construct to explain the perceived pattern of zeolite and authigenic feldspar distribution in strata of the upper Brushy Basin Member. Specifically, Turner-Peterson (1985, 1987) and Turner and Fishman (1991) described lateral changes in the diagenetic facies of the mudstones that could be interpreted as concentric zonation, and cited similar concentric patterns in modern alkaline lakes in arid environments. This became the basis for interpreting much of upper Brushy Basin deposition as occurring within an extensive ($>1 \times 10^5$ km^2) lacustrine basin. Indeed, Turner and Fishman (1991) stated specifically that the perceived pattern of zonation was "the most compelling reason to infer a syndepostional origin" for the authigenic minerals. However, as Anderson and Lucas (1997) noted, Turner and Fishman (1991) presented too few data to document a convincing concentric zonation of zeolites. Their two cross-sections only suggest smectite-dominated rocks to the north (Grand Junction area in Colorado) and south (near Toadlena in New Mexico) with either clinoptilolite, analcime or K-spar dominated mudrocks at some points in between. Thus, the data presented by Turner and Fishman (1991, figs. 1, 4–5) are inadequate to define the concentric pattern of zeolite facies (their figure 12) presented therein. Furthermore, we observe that our data do not match the pattern presented by Turner and Fishman (1991), as we find albite cement in ash-rich beds near Grand Junction, Colorado, where only smectite should be present in the ash beds, based on their zonation pattern.

Moreover, we find the hypothesis of a syngenetic origin for the diagenetic minerals flawed due to the assumption that they could not originate by burial diagenesis. They cite the interpreted maximum burial temperature of 75°C (based on a maximum burial depth of ~1800 m) as insufficient for albite formation, citing work by Pittman (1988) that set a minimum temperature for albitization at 85°C. However, other studies have shown zeolites to form from the alteration of volcanic glass at much lower temperatures. For example, Ogihara (2000) found the conversion of volcanic glass to clinoptilolite at temperatures of 45° to 51°C, and Nähr *et al*. (1998) documented clinoptilolite formation at temperatures below 20°C. Albite is the highest-temperature diagenetic mineral recorded in the Brushy Basin strata, but Aargaard *et al*. (1990) documented diagenetic albitization at temperatures as low as 60°C, and van de Kamp and Leake (1996) found that authigenic albite formed in Newark Basin strata, where burial depths were between one and two kilometers, at temperatures of only 50°C or possibly lower. Hence, the requirement of high alkalinity for mineral authigenesis is countered by the burial history of the strata, which provided sufficiently elevated temperatures. Thus, there are multiple lines of evidence, based on sedimentary facies, vertebrate taphonomy and zeolite distribution and chemistry, for rejecting the hypotheses of Lake T'oo'dichi' or similarly, of an extensive wetland covering much of the Colorado Plateau during Brushy Basin Member deposition.

PALEOCLIMATE INDICATORS

Previous workers have noted that the fluvial channel deposits in the Brushy Basin Member exhibit architectural elements indicative of both low and high sinuosity stream systems, suggesting low depositional gradients and discharge characteristics varying from ephemeral to perennial (Kirkland, 2006; Galli, 2003, 2014 – this volume). Contrary to the interpretation of Dunagan and Turner (2004), surface flow delivery of water to the western Colorado portion of the depositional basin was significant, suggesting a climate more humid than described by these authors. The work of Galli (2003, 2014 – this volume) describes the Brushy Basin stream systems as mainly anastomosing, a pattern typical of streams that are largely perennial and carrying a high sediment load.

As described by Tidwell (1990) and Ash and Tidwell (1998), the Morrison Formation hosted an abundant and diverse flora, but taphonomic issues could interject biases toward preservation of those plants growing in riparian and wetland environments. Hence, Turner and Peterson (2004) likened the Brushy Basin Member ecosystem to that of the savannah, with a warm, highly evaporative climate, and vegetation abundant in riparian and wetland zones, but sparsely vegetated in intervening arid regions. We note, however, that this model is at odds with the pedogenic evidence presented above for a sub-humid to semi-arid climate and well-vegetated inter-channel floodplains.

Accumulations of dinosaur bones in the Morrison Formation are typically found in channel deposits where the remains generally have a very low degree of articulation (*e.g.*, Dodson *et al*., 1980); the general assumption is that the animals died on the floodplain and their decayed, disarticulated skeletons were swept into the channels during large floods. However, Dodson *et al*. (1980) noted that bone accumulations also occurred in levee and oxbow environments, in addition to the characteristic channel environment. This would indicate sufficiently frequent sedimentary events on the floodplain to ensure burial in some instances. Furthermore, these authors noted differences in taphonomic distribution between the sauropod dinosaurs, particularly *Diplodocus* and *Camarasaurus* (Fig. 9), which were gregarious and appeared to favor the riparian environments, and *Stegosaurus*, which may have favored more open environments, implying sufficient food sources in the inter-channel environment to support the food requirements of megaherbivores (see below).

PALEOECOLOGICAL CONSIDERATIONS

Published estimates of dinosaur food requirements suggest that a 20 ton sauropod dinosaur consumed 50–60 kg of vegetation per day (Béland, Russell, 1979; Russell *et al*., 1980; Alexander, 1989). This implies the presence of a tremendous amount of vegetation in the Morrison Fm. depositional system to support what were apparently large populations of sauropods (Fig. 9). Such reasoning, in part, underpinned early depiction of the Morrison depositional system as one of rain forest/jungle analogous to the modern Amazon Basin. Nevertheless, Englemann *et al*. (2004) accepted the idea of a seasonal, semi-arid Morrison climate and explained away the apparent food needs of abundant sauropods by speculating that sauropods were unusually efficient herbivores capable of enduring long periods of starvation, thus reducing their food needs, and also able to migrate over large distances, thus reducing the need for dense vegetative cover. They also argued that sauropods are preferentially preserved in the Morrison depositional system because of the durability of their very large bones. Thus, taphonomy was used to undermine the notion of sauropod abundance in the Morrison depositional basin. Aside from this taphonomic contention, however, the arguments of Engelmann *et al*. (2004) that reduce the amounts of vegetation in the Morrison depositional system are purely speculative. The physiology, periodic starvation and migration of sauropods they suggest are plausible, but constitute a complex set of *ad hoc* reasons designed to reconcile the presence of abundant megaherbivores with a climate/landscape with limited (or locally

Fig. 9. Scene during Brushy Basin Member deposition showing the sauropod dinosaur *Diplodocus* ("Seismosaurus") being attacked by the theropod dinosaur *Saurophaganax* on a well vegetated landscape. Artwork by Mary Sundstrom and Matt Celeskey, courtesy of New Mexico Museum of Natural History and Science

clumped) vegetation. The simpler interpretation, that there was sufficient vegetation available in the Morrison basin to feed the many sauropods, is consistent with Morrison paleobotany, with apparent Morrison sauropod abundance and with the climatic/paleohydrological inferences made here based on Morrison paleosols.

PALEOENVIRONMENTAL INTERPRETATION

We favor a paleoenvironmental interpretation similar to that of Moberly (1960) and as amended by Dodson *et al*. (1980); Brushy Basin sedimentation took place on a broad, low gradient alluvial plain crossed by many streams. Many of these streams were perennial, at least for extended periods, as evidenced by the presence of aquatic invertebrates and small vertebrates. That many, but not necessarily all, of these streams were perennial does not negate that the climate was undoubtedly seasonal, as indicated by the pedogenic features, and supported by models that indicate the continued operation of monsoonal climate systems during the early stages of Pangean rifting (Moore *et al*., 1992b; Hallam *et al*., 1993). Precipitation was sufficient to fill numerous lakes and ponds on the broad alluvial plain, many of which were likely formed by channel avulsion and abandonment. Although many of these water bodies were perennial, the water levels fluctuated substantially, as demonstrated by seasonal growth banding in gastropods (Good, 2004). Thus, the broad Brushy Basin alluvial plan was crossed and dotted by streams and lakes, some perennial, some not, which provided abundant sources of water for vegetation, but the intervening plains were also vegetated, perhaps not heavily, but clearly they were not dry, barren expanses.

Contrary to some previous interpretations (*cf.* Demko *et al*., 2004), a mountainous arc to the west of the Morrison depositional basin would not have produced an orographic desert due to the summer monsoonal flow from the incipient Gulf of Mexico (Moore *et al.*, 1992b). Thus, the strength of monsoonal flow would not have been constant as orbital forcing of solar insolation would have caused variations in the strength of the monsoons over periods of tens to hundreds of thousands of years (Crowley *et al*., 1989; Kutzbach, Gallimore, 1989; Kutzbach, 1994). Consequently, the Morrison depositional basin environment would have experienced long-term variations in precipitation, potentially significant in scale, as are experienced in northern Africa (Rossignol-Strick, 1983, 1985), where major latitudinal shifts occur in the distribution of grasslands and desert ecosystems due to orbital forcing (Foley *et al*., 2003). Similar shifts were likely in the Morrison ecosystems, and they can account for the often apparently conflicting evidence sedimentary and fossil proxy evidence.

Thus, the paleoclimate during Brushy Basin Member deposition was not constant, but very likely alternated between conditions best characterized as sub-humid and semiarid over the time scales of precessional and obliquity forcing (tens of thousands of years). During these changes, the distribution of Late Jurassic biota likely migrated latitudinally in response to climate change. During intervals of stronger northern hemisphere monsoons, vegetation was more lush and the landscape supported larger populations of megafauna. In the intervening drier periods, much of the megafauna may have migrated northward as lakes and streams dried, and more arid-climate soils formed. Finally, the gradual northward drifting of North America during the Late Jurassic brought portions of the depositional basin of the Morrison Formation out of the center of the subtropical climate zone controlled by Hadley circulation. This latitudinal drift of the basin permitted establishment of a more consistently subhumid climate allowing for wetter conditions during later Brushy Basin deposition, as noted by other authors (*e.g.*, Demko *et al*., 2004).

CONCLUSIONS

Study of the pedogenic and lacustrine features of the Morrison Formation in western Colorado allows a more nuanced interpretation of paleoclimate than has been produced to date. Lower Brushy Basin Member paleosols are mostly calcareous Aridisols with Stage I through Stage III calcrete Bk horizons, abundant root traces, occasional vertic features, but only rarely with preserved epipedons, whereas Upper Brushy Basin Member paleosols are mainly thicker and commonly display ochric epipedons and well-developed Bt and Bw horizons that we assign to the order Inceptisol. Limestones in the Brushy Basin Member include both uniformly micritic limestones with sparse body fossils and charophyte debris and limestones with strongly brecciated textures displaying clotted-peloidal fabrics with circumgranular cracking and silica replacement. We interpret all of these limestones as the deposits of carbonate in small water bodies on a low gradient flood plain, with the brecciated textures resulting from pedogenic reworking of the carbonate sediment.

The overall climate during Brushy Basin deposition suggested by of the features in the study area is strongly seasonal (monsoonal), with precipitation amounts ranging between semi-arid and subhumid. Longer-term (tens of thousands of years) variations in climate conditions were likely controlled by orbital forcing and account for potential conflicts between sedimentary evidence (paleosols and palustrine limestones) and the fossils, particularly the dinosaurian megafauna and floral evidence for moist conditions. A long term change in climate from lower Brushy Basin to upper Brushy Basin deposition, specifically an increase in humidity, is suggested by the differences in paleosols and may be accounted for by the northward drift of the continent during the Late Jurassic, bringing the study area to a slightly higher, more humid latitude. Although these conclusions are based on a very localized study, certain aspects of our paleoclimatic interpretation, such as strong seasonality and potential orbital forcing, likely are applicable to a large portion of the Morrison depositional basin.

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