1   A chronostratigraphic framework for the Upper Jurassic
2   Morrison Formation, western USA

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ABSTRACT

The fluvial, overbank, and lacustrine deposits of the Upper Jurassic Morrison Formation of the Western Interior, U.S.A. have been intensively studied due to their diverse and well-preserved dinosaurian fauna, and the presence of economic quantities of uranium and vanadium ores. The formation crops out over 12 degrees of latitude and 1.2 million km², and is an excellent case study for the examination of paleoecology, community structure, and evolutionary dynamics at a time in Earth's history when the climate was significantly warmer than today. However, paleoecological studies have been hampered by lack of correlation across the formation. Assuming a primarily tectonic control on fluvial architecture, we propose the first chronostratigraphic framework of the formation, which is based on sequence stratigraphy, magnetostratigraphy, and radiometric dating. The formation can be divided into three sequences each represented by a period of degradation followed by aggradation. This chronostratigraphic framework indicates that the formation youngs to the north, and was deposited over about 7 million years during the late Kimmeridgian and Tithonian. This framework provides a foundation for future sedimentological, stratigraphic, and paleobiological studies of the iconic dinosaurian fauna known from the Morrison.

INTRODUCTION

The terrestrial deposits of the Upper Jurassic Morrison Formation have been intensively studied since the discovery of their diverse and well-preserved dinosaurian fauna in the latter part of the 19th century. Some of the most easily recognisable and iconic dinosaurs, such as *Stegosaurus*, *Diplodocus*, and *Brontosaurus*, are known from the formation. The formation also hosts the United States' most productive uranium-vanadium deposits (Turner-Peterson 1986; Chenoweth 1998), and was mined for uranium until 1990 (Chenoweth 1998). Outcropping from Montana in the north to New Mexico in the south, the fluvial, overbank, and lacustrine deposits of the formation cover about 1.2 million km² and a latitudinal range of around 12 degrees (Fig. 1). The age of the formation has been difficult to definitively constrain, but U/Pb (Trujillo et al. 2014) and ⁴⁰Ar/³⁹Ar dating (e.g., Kowallis et al. 1998; Kvale et al. 2001; Trujillo and Kowallis 2015), palynology (Litwin et al. 1998), and invertebrate biostratigraphy (Schudack et al. 1998; Lucas and Kirkland 1998) suggest a Kimmeridgian to Tithonian age.

Due to excellent exposure in the Western Interior and the long record of paleontological collecting and sedimentological research in the region, the Morrison is an ideal candidate for a case study of ecology and community structure at a time in Earth's history when the climate was significantly warmer than today (Dodson et al. 1980; Turner and Peterson 2004 and references therein; Sellwood and Valdes 2008). However, attempts to examine the paleoecology of terrestrial vertebrate faunas of the Morrison (e.g., Dodson et al. 1980; Turner and Peterson 1999; Foster 2000, 2003, Whitlock et al. 2018) have been severely hampered by a lack of long-range correlation: temporal relationships between outcrops in Montana, Wyoming, and on the Colorado Plateau remain entirely unknown, and thus the relative ages of dinosaur quarries and their faunas cannot be explored. Herein, we propose the first chronostratigraphic framework for the entire Morrison Formation. The framework is formulated based on the principles of continental sequence stratigraphy, and is tested and supplemented with magnetostratigraphy and radiometric dating. It is

55 based on direct sedimentological observations supplemented with the vast literature on Morrison
56 Formation sedimentology, and we propose it as a hypothesis, which can be directly tested by new
57 observations, and upon which future work can build. The framework lays the foundations for future
58 paleobiological research in the formation, allowing dinosaur quarries to be located temporally
59 relative to each other for the first time.

60

61 **BACKGROUND**

62 *Geological Setting*

63 The Morrison Formation was deposited in the back bulge depozone of a retroarc foreland
64 basin (DeCelles and Burden 1992; Royse 1993; DeCelles and Currie 1996; Currie 1998; DeCelles 2004;
65 Roca and Nadon 2007; Fuentes et al. 2009) generated during the early stages of the Sevier orogeny
66 (DeCelles and Burden 1992; DeCelles and Currie 1996; Cooley and Schmidt 1998; Currie 1998;
67 DeCelles 2004; Fuentes et al. 2009; Christiansen et al. 2015). The formation overlies Middle Jurassic
68 marine deposits of the Sundance Sea to the north, and sabkha and aeolian sediments deposited on
69 the sea's arid shores farther south (Peterson 1994; Anderson and Lucas 1997, 1998; Turner and
70 Peterson 1999). To the south, the earliest deposits of the formation are braided and anastomosing
71 fluvial sandstones (Robinson and McCabe 1997; Kjemperud et al. 2008) of a distributive fluvial
72 system (Tyler and Ethridge 1983; Owen et al. 2015a) that extended from west to east. Farther east
73 and to the north, these fluvial deposits are not recognized, presumably due to downstream fining
74 and perhaps because the base of the formation is not everywhere contemporaneous. Overlying the
75 distributive fluvial sands in the south and representing the whole of the formation elsewhere,
76 variegated overbank silty mudstones, in many places containing abundant volcanic ash (Christiansen
77 et al. 2015), lacustrine silts and limestones, and fluvial channel sandstones predominate. The
78 formation is overlain by Early Cretaceous terrestrial sediments deposited in the foreland basin of the
79 Sevier orogenic belt (DeCelles and Burden 1992; Currie 1997, 1998; Aubrey 1998), which are known

80 by a variety of names in different geographic areas (Cedar Mountain Formation in Utah; Burro
81 Canyon Formation in Colorado; Cloverly Formation in Wyoming; Kootenai Formation in Montana,
82 Lakota Formation in South Dakota) but generally comprise purple-gray mudstones with coarse
83 conglomerates at their bases.

84 A variety of sedimentological and paleontological evidence suggests that climate varied
85 across the geographic extent of the Morrison Formation, and this is supported by general circulation
86 models (GCMs). In the southwest, clay mineralogy suggests the presence of an alkaline-saline lake
87 (Turner and Fishman 1991) or groundwater system (Dunagan and Turner 2004), and this is
88 supported by nodular gypsum observed in overbank deposits (SCRM personal observation 2014),
89 and GCMs, which indicate that the area was desert and summertime temperatures may have
90 exceeded 40°C (Sellwood and Valdes 2008). In the southeast, however, lacustrine limestones
91 indicate a freshwater lacustrine or wetland system (Lockley et al. 1986; Dunagan 1999; Dunagan and
92 Turner 2004; Schumacher and Lockley 2014). To the north, abundant ferns (Parrish et al. 2004) and
93 coals (Cooley and Schmidt 1998) at the top of the formation in Montana indicate a much wetter,
94 cooler climate, and GCMs indicate a mixed forest biome in this region (Sellwood and Valdes 2008).

95 *Morrison Lithostratigraphy*

96 A proliferation of “Members” in the Morrison Formation have been named, and
97 consequently the stratigraphy and nomenclature has become cumbersome. Most of the members
98 are recognized only locally. However, across Utah, Colorado, and New Mexico, the formation is
99 readily divided into two members, a lower, river-dominated Salt Wash Member, and an upper,
100 overbank-dominated Brushy Basin Member. These members are not recognized in outcrops off the
101 Colorado Plateau, where most of the formation resembles the Brushy Basin Member.

102 Piringos and O’Sullivan (1978) recognized several major unconformities present across the
103 Western Interior. The “J5” unconformity was thought to separate Middle Jurassic and Upper Jurassic
104 strata, while the “K1” was located at the base of the Cretaceous succession. Thus, the base of the

Morrison Formation is usually considered to be the J5 unconformity, while the top of the formation is the K1. Unfortunately, throughout most of the Western Interior, Jurassic and Cretaceous rocks are paraconformable, and without very precise dating, unconformities can be difficult to recognise in many field localities. Furthermore, it is now recognized that the J5 unconformity is not as regionally extensive as previously thought, and in many places across the Western Interior, Middle Jurassic and Upper Jurassic rocks are conformable. The use of unconformities, difficult to recognise in the field and inherently time-transgressive in their nature, has led to a great deal of debate about what constitutes the base of the Morrison Formation.

Underlying the formation to the south, in Utah, western Colorado, and New Mexico are strata of the Middle Jurassic San Rafael Group, which commonly comprise aeolianites, playa-lake, and sabkha deposits indicative of arid environments (Condon and Peterson 1986; Peterson 1994; Lucas and Anderson 1997; Anderson and Lucas 1998). The early nomenclatural history of the Morrison Formation in the southwest has been reviewed in detail by Anderson and Lucas (1997; 1998). Briefly, Gregory (1938) described aeolian and sabkha deposits underlying fluvial sandstones in Utah as the Recapture and Bluff Sandstone Members of the Morrison Formation. Although Gregory (1938) suggested that fluvial sands overlying the Recapture and Bluff Members might be equivalent to the Salt Wash member, he named them the Westwater Canyon Member. Stokes (1944) then correlated the Recapture and Bluff Sandstone Members with the Salt Wash Member, meaning that the Westwater Canyon Member sandstones must be at a higher stratigraphic level. This interpretation was followed by subsequent workers, and caused a variety of curious correlations across the region (e.g., see correlation charts in Condon and Peterson 1986; Turner-Peterson 1986; Peterson 1984; Godin 1991; Turner and Fishman 1991; Peterson 1994; Turner and Peterson 1999; Dunagan and Turner 2004). Anderson and Lucas (1997, 1998) argued that the Recapture and Bluff Sandstone Members were genetically more closely related to San Rafael Group strata. This allowed recognition that the Westwater Canyon Member was synonymous with the Salt Wash Member, and

they placed the base of the Morrison in this area at the base of the Salt Wash (Lucas et al. 1995; Lucas and Anderson 1997; Anderson and Lucas 1997, 1998).

Peterson (1984) described the Tidwell Member at the base of the Morrison Formation in Utah and New Mexico. The Tidwell comprises red to gray mudstones with abundant gypsum, and has been interpreted as being deposited in an arid coastal-plain sabkha environment (Lucas and Anderson 1997; Anderson and Lucas 1998). Peterson (1984) identified it as a member of the Morrison Formation because he identified the J5 unconformity at its base. Anderson and Lucas (1998) argued that the Tidwell was probably correlative with the Recapture Member, which they had assigned to the Summerville Formation, and thus should not be part of the lithostratigraphically defined Morrison Formation. Furthermore, they argued that the J5 unconformity was present above Tidwell and Recapture strata. Subsequent workers (e.g., Owen et al. 2015a, 2015b), have recognized the Tidwell Member as the distalmost deposits of the Salt Wash distributive fluvial system, and have recognized the J5 underlying the Tidwell Member on the west side of the San Rafael Swell based on a slight angular unconformity (J. Foster, personal communication 2016). Thus, the Tidwell Member is generally considered part of the Morrison Formation system.

In northern Utah and Wyoming, the Windy Hill Member has been described as part of the Morrison Formation (e.g., Currie 1997, 1998; Aubrey 1998; Turner and Peterson 1999). Originally considered part of the underlying, exclusively marine, Sundance Formation (Brenner and Davies 1973; McMullen et al. 2014), the Windy Hill comprises shallow marine bioclastic limestones, glauconitic sandy limestones and calcareous mudstones and, at least in Wyoming, represents a conformable sequence of sediments indicating a shallowing-upward trend as the Sundance Sea regressed during the early part of the Late Jurassic (Brenner and Davies 1973; McMullen et al. 2014; SCRM personal observation 2014). In the Uinta mountains of northern Utah, the Tidwell Member has been identified as overlying the Windy Hill (Currie 1997, 1998; Turner and Peterson 1999); in this

154 area the Tidwell is considered shallow marine (Currie 1998). The J5 unconformity has been identified
155 underlying the Windy Hill (McMullen et al. 2014).

156 The Sundance Sea, which occupied the Western Interior region in the Middle and early Late
157 Jurassic, withdrew to the north, and thus it is to be expected that the base of the lithostratigraphic
158 Morrison Formation is time-transgressive, younging to the north. It would not, therefore, be
159 surprising if the oldest Morrison deposits in Utah and New Mexico are contemporaneous with
160 shallow marine deposits farther north in Wyoming and Montana.

161 *Morrison Chronostratigraphy*

162 **The “clay change”**.---Peterson and Turner (1998) and Turner and Peterson (1999) suggested
163 that in the Brushy Basin Member a consistent change from illitic clays below to smectitic clays above
164 could be observed, that this represented an isochronous horizon because it represented a time
165 when input of volcanic ash increased, and that it was readily identifiable in the field thanks to the
166 propensity of smectitic clays to swell on contact with water. Virtually all later work requiring
167 correlation of the Morrison Formation has used this so-called “clay change” as an isochronous
168 horizon (e.g., Evanoff et al. 1998; Hasiotis and Demko 1998; Kowallis et al. 1998; Litwin et al. 1998;
169 Schudack et al. 1998; Dunagan and Turner 2004; Demko et al. 2004; Kirkland 2006; Galli 2014). Two
170 subsequent studies (Jennings and Hasiotis 2006; Trujillo 2006) have demonstrated that there is no
171 consistent change from illite to smectite through the formation, and that illite and smectite occur in
172 equal abundances throughout. In the field, in certain areas, there are clearly zones which appear to
173 contain more swelling clays than others (SCRM personal observation 2014–2015; Kirkland 2006; Galli
174 2014); however, the studies of Jennings and Hasiotis (2006) and Trujillo (2006) demonstrate that the
175 clay change is not an isochronous horizon, and changes in the content of swelling clays likely vary
176 from place to place and may represent local variations in the input of volcanic ash, degree of
177 weathering, and local environmental conditions (Jennings and Hasiotis 2006). The “clay change”
178 cannot, therefore, be used for correlation in the Morrison basin.

179   **Paleosols.**---Demko et al. (2004) identified several well-developed paleosols in the Morrison
180 Formation that were interpreted as sequence-bounding unconformities. However, not all of these
181 paleosols were present in every section, confounding attempts to use them for regional correlation.
182 Furthermore, there are many paleosols in the Morrison Formation, and in the field it is difficult to
183 identify which paleosols are those considered of stratigraphic significance by Demko et al. (2004)
184 and which represent less long-lived exposure of the ground surface (SCRM personal observation
185 2014-2015). The reliance of the scheme on the clay change as an isochronous horizon has meant
186 that it has not been employed subsequently.

187   **Biostratigraphy.**---A number of studies have used palynomorphs, invertebrate,s and
188 vertebrates in attempts to both date and correlate the Morrison Formation. The distribution of
189 charophytes, ostracods (Schudack et al. 1998), palynomorphs (Litwin et al. 1998), conchostrachans
190 (Lucas and Kirkland 1998), and molluscs (Evanoff et al. 1998) have been investigated to attempt to
191 derive stratigraphic schemes. Most of these, however, relied on the clay change as an *a priori*
192 lithostratigraphic marker, and many of the organisms are temporally wide-ranging, making them
193 sub-optimal for biostratigraphic purposes. Dinosaur biostratigraphy has also been attempted, both
194 locally and regionally (Bakker et al. 1990; Carpenter 1998; Turner and Peterson 1999; Ikejiri 2006).
195 Local studies suffered from a lack of specimens, although quarries can be well correlated (Bakker et
196 al. 1990; Carpenter 1998), while regional studies were reliant on the clay change (Turner and
197 Peterson 1999; Ikejiri 2006), and thus these results are questionable.

198   **Radiometric Dating.**---A number of radiometric dates, using a variety of methods, have
199 attempted to date the Morrison Formation or various quarries within it. Older dates tended to have
200 very large error bars, and constrained deposition only to the Late Jurassic (e.g., Bilbey, 1998; Kvale et
201 al. 2001). Kowallis et al. (1998) provided the first multi-sample study of Morrison ash deposits using
202 Ar/Ar dating of sanadines, and achieved error bars of < 1 Ma. These dates have recently been
203 recalibrated to account for re-dating of the Fish Canyon Tuff, the standard for this method (Trujillo

and Kowallis 2015). Recalibration of the dates suggests that the Tidwell Member of the Morrison Formation is about 156–157 Ma, while other samples studied fell between about 152–150 Ma. Trujillo et al. (2014) recently published a U/Pb date for the Mygatt-Moore quarry in western Colorado, dating it to ~152 Ma, while Galli et al. (2018) studied sections in the nearby Fruita Paleontological Area as well as the Trail Through Time at Mygatt-Moore and established younger dates of 150–149 Ma, also using U/Pb dating. While radiometric dating can constrain the ages of specific quarries with error bars of ~500,000 years, it can be carried out only where ash layers or ash-rich mudstones occur. These occur predominantly in the lower part of the Brushy Basin Member in Colorado and Utah, and have not been identified farther north. There are currently no reliable radiometric dates from the Morrison of the Bighorn Basin of Wyoming or from Montana, which means that radiometric dating alone cannot be used to correlate sections across the whole of the Morrison depositional basin.

Magnetostratigraphy and Sequence Stratigraphy.---Both magnetostratigraphy (Steiner et al. 1994; Swierc and Johnson 1996; Steiner 1998; Maidment et al. 2017) and sequence stratigraphy (Currie 1997, 1998) have been used to correlate the Morrison Formation locally, and are favorable because they do not rely on a lithostratigraphic marker as a starting point. Some success has been achieved in both cases, and both methods warrant broader, regional studies to assess their utility for regional correlation. Herein, a chronostratigraphic framework is proposed based on sequence stratigraphy and supplemented with magnetostratigraphy.

Continental Sequence Stratigraphy

Nonmarine sediment architecture results from a complex array of autocyclic and allocyclic controls that are not mutually exclusive. Accommodation space in the terrestrial realm may be controlled by the combined effects eustatic sea-level change, tectonic subsidence and uplift, and climate, such that changes in precipitation may change local base levels. Sediment supply reflects a combination of tectonic uplift and climate-controlled erosion of the hinterland. Autocyclic behaviors,

such as channel avulsion, act at various scales and can be superimposed upon allocyclic controls, or be the predominant mechanism controlling sediment architecture (Shanley and McCabe 1994; Richards 1996), particularly in the case of internally drained basins (Fisher and Nichols 2013).

Accommodation space and sediment supply can be combined to define the accommodation/sediment supply ratio (A/S) regardless of the mechanism by which either is generated. Where A/S is very low, repeated phases of channel cutting, channel avulsions, and lateral migrations across the floodplain may result in very widespread fluvial incision and sediment bypass. Such incision surfaces may have significant relief (e.g., Plint et al. 2001) or may be relatively smooth (e.g., Olsen et al. 1995; Holbrook 1996). If both accommodation and sediment supply are low, prolonged exposure of the land surface results in the formation of paleosols (McCarthy and Plint 1998; Demko et al. 2004). As A/S increases, incised valleys may be infilled with the amalgamated deposits of the incising river systems, resulting in widespread deposition of sheet sands (Shanley and McCabe 1994; Richards 1996; Holbrook 1996; Currie 1997; Fisher and Nichols 2013). Downstream from sources of sediment supply, paleosols and smaller, fixed channels may occur (Fisher and Nichols 2013; Owen et al. 2015a). As A/S increases, floodplain and channel aggradation occurs, resulting in a lower degree of amalgamation of channel sands, the preservation of floodplain fines, and fewer paleosol horizons (Fisher and Nichols 2013). Fluvial style may also change (Holbrook and Schumm 1999); for example, anastomosing river systems are often related to high aggradation rates (Nadon 1994; Cooley and Schmidt 1998; Holbrook and Schumm 1999). If A/S is high, lacustrine or wetland systems may develop.

A variety of sequence stratigraphic terminology has been applied to the above pattern to facilitate correlation. Shanley and McCabe (1994) promoted the use of marine sequence stratigraphic terminology in order to correlate with sea level changes observed in marine packages. Thus, the lowstand systems tract corresponds to multi-story sheet sandstones deposited in low A/S settings, the transgressive systems tract corresponds to decrease in channel amalgamation and the

increased preservation of floodplain fines, and the highstand systems tract corresponds to increased amalgamation of channel sands as A/S decreases (Shanley and McCabe 1994; Richards 1996). Sequence boundaries can be identified due to a basinward shift in facies; where a fluvial incision surface overlies marine mudstones, this is relatively uncontroversial. However, in basins in which there is no marine influence, sequence boundaries must be identified based on an abrupt change in facies, changes in grain size, petrographic changes indicating different source regions, different flow directions, changes in fluvial style, and degree of amalgamation (Shanley and McCabe 1994; Olsen et al. 1995; Catuneanu and Elango 2001; Lawton et al. 2003). The maximum-flooding surface in a nonmarine basin is characterized by the presence of poorly drained floodplains, swamps, or lakes (Richards 1996) in the absence of tidally influenced facies or marine fossils.

While correlation of marine systems tracts onshore is extremely useful, it is appropriate only for the lower reaches of fluvial systems in which sea level is likely to be the major autocyclic control (Shanley and McCabe 1994; Aitkin and Flint 1995; Holbrook 1996; Holbrook et al. 2006; Alqahtani et al. 2017). In the upper reaches of fluvial systems, A/S may be entirely independent of sea level, controlled instead by tectonics and climate (Shanley and McCabe 1994; Olsen et al. 1995; Holbrook and Schumm 1999; Catuneanu and Elango 2001; Holbrook et al. 2006; Owen et al. 2015b; Alqahtani et al. 2017). It is therefore misleading to extrapolate terminology for marine systems tracts to fluvial systems for which the coeval shoreline was many hundreds of kilometers away, or that are internally drained, and other terminology has been proposed. Currie (1997, 1998) used the terms “degradational systems tract” to correspond with low A/S conditions when sheet sandstones are deposited and paleosols developed, and ‘aggradational systems tract’ to correspond with higher A/S conditions, when the floodplain aggrades, channels are not amalgamated, and paleosols are not formed.

The Morrison Formation was deposited as the Sundance Sea retreated to the north. The Tidwell Member, at the base of the formation in Utah, is considered to have been deposited in

sabkha and marginal marine environments, and the lowest parts of the fully terrestrial Salt Wash Member in Utah, New Mexico, and Colorado may have been deposited while the coastline lay just a few hundred kilometers to the north in southern Wyoming (see below). The Sundance Sea must have quickly regressed to the north, however, because Kimmeridgian-Tithonian marginal marine deposits are known from the Morrisey and Mist Mountain Formations of the Kootenay Group in Alberta and British Columbia, Canada (Mastalerz and Bustin 1997; Raines et al. 2013), indicating that the shoreline was at least 700 km to the north of the northernmost Morrison outcrops in Montana during most of Morrison deposition (Raines et al. 2013; Owen et al. 2015a). To the south, Late Jurassic marine rocks are known from the Bisbee Group of southern New Mexico (Lucas et al. 2001), but these were separated from the Morrison depositional basin by the Mogollon highland, a rift shoulder related to Late Jurassic transtension in the region (Lucas et al. 2001; DeCelles 2004), and consequently it appears unlikely there was a marine connection to the south (Peterson 1994; Lucas et al. 2001). Although the very earliest deposits of the Morrison Formation may have been influenced by sea-level changes, the distance to the coeval shoreline indicates that throughout the majority of its deposition, A/S in the Morrison Formation would have been controlled by tectonics and climate rather than sea level change (Holbrook et al. 2006; Owen et al. 2015a). In order to avoid connotations with causative mechanisms, herein, sequence stratigraphic terminology is avoided and the systems tracts are described in terms of A/S conditions.

METHODS

Sequence Stratigraphy

The Morrison Formation was studied at 19 locations across the Western Interior region from southeastern Utah in the south to central Montana in the north (Fig. 1; see online supplementary

data for coordinates of measured sections). Locations were chosen because either little previous work had been published at the locality, or the locality was very well documented, allowing comparison of the results to previous studies. The Morrison Formation varied in thickness in the sections studied from ~ 20 m thick in central and western Montana to ~200 m thick in northeastern Utah. At each locality, the exposed section was logged at a resolution of 1 cm = 1 m, and measurements were taken with a tape measure. Sedimentological data was recorded and any available paleocurrent data (three-dimensionally preserved cross-beds, ripple-crest laminae, lateral-accretion surfaces) were measured using a compass-clinometer.

The Morrison Formation has been the subject of intensive sedimentological and paleoenvironmental research, and a large number of studies have documented the sedimentology at specific sites. Field data was therefore supplemented with 245 published sedimentological logs (online supplementary material) from all parts of the Formation, which were used to build correlations (Fig. 1).

Magnetostratigraphy

Magnetostratigraphic samples were taken at each locality where permission had been granted by the landowner or land-management agency. Magnetostratigraphic samples were cut using rock core drill with a water-cooled diamond drill bit. Azimuth and plunge of drilling direction were recorded using a sun and/or magnetic compass and an inclinometer. Cores were oriented by scoring the top surface with a brass wire.

We aimed to collect samples at 1 m intervals throughout the section; however, we were unable to drill cores in mudstone sections because the mudstone disaggregated on contact with drilling water. Furthermore, many sandstones proved difficult to core because they were weakly cemented, or the cores broke along cross-lamination surfaces. Consequently, we were able only to collect samples in sandstones, and we recovered only ~30% of cores we drilled. This resulted in very low sample coverage across most sections: in the future, magnetostratigraphic sampling the

Morrison Formation should be carried out by collecting hand specimens (see Maidment et al. 2017 for more details).

In total, 753 cores were collected across 20 sites (Fig. 1; see online supplementary information for stratigraphic locations of cores). Cores were cut into 1 inch specimens. Some cores disaggregated in the rock saw, while others were long enough to cut into two 1 inch specimens. Specimens were measured on a JR5A spinner magnetometer with a noise level of $\sim 5 \times 10^{-10} \text{ A m}^2$ at Imperial College London. Thermal demagnetization was carried out in controlled-field paleomagnetic ovens, because alternating-field demagnetization was ineffective in separating magnetic components and fully demagnetizing the specimens. Specimens were heated in helium between five and ten steps between 100°C and 690°C.

Thermal-demagnetization data were analyzed using Puffin Plot (Lurcock and Wilson 2012). Specimens were grouped by bed. Tectonic tilt and regional modern geomagnetic field were corrected for, and magnetization directions were identified by hand-picking points with well-defined directions using Zijderveld plots (Zijderveld 1967). Typically, three to four samples were used to obtain a mean direction, although in some cases, only two specimens were available. This was followed by a principal-components analysis to generate a best-fit line through a single component of the data (Kirschvink 1980; Tauxe 2010). Fisher statistics were used to calculate the mean direction and a confidence limit for each bed, giving either positive or negative inclinations (Butler 1992). Bed-by-bed data and results can be found in online supplementary data.

Radiometric Dating

Chemical abrasion, thermal ionization, mass spectrometric U-Pb dating (CA-TIMS) of zircons from two mudstone samples collected at GU-DC (Fig. 1) was carried out at the University of Wyoming. The samples were mechanically crushed and zircons were concentrated by a combination of ultrasonic deflocculation and Wilfley shaker table separations. Zircons were then purified by magnetic separation and heavy-liquid flotation of lighter minerals. Euhedral zircons yielding

characteristics typical of ash-fall deposition (elongate tips, longitudinal bubble tracks) were chosen for dissolution and U-Pb analysis. Zircons were annealed at 850°C for 50 hours, then dissolved in two steps in a method modified from Mattinson (2005). Pb and UO₂ isotopic compositions were determined on a Micromass Sector 54 mass spectrometer. Ages were calculated using PbMacDat and ISOPLOT/EX after Ludwig (1991, 1998).

In addition, recent radiometric dates from the literature (Trujillo et al. 2014; Trujillo and Kowallis 2015; Galli et al. 2018; Deblieux et al. 2018) were projected onto the nearest logs from the literature (see online supplementary data) and/or sampled localities. These radiometric dates were used to relate magnetostratigraphic data to the Geomagnetic Polarity Timescale of Malinverno et al. (2012), and calculate absolute ages for the systems tracts.

RESULTS

Facies Associations

The following builds on numerous previous studies of Morrison sedimentology and facies, and is thus relatively brief. More detailed accounts of facies and facies associations can be found in Dodson et al. (1980), Godin (1991), Dunagan and Turner (2004), Kirkland (2006), Kjemperud et al. (2008), Jennings et al. (2011), Galli (2014), and Owen et al. (2015a), and in further references given in each facies association below.

Amalgamated Channel Belt Facies Association (Fig. 2A).----Defined by gravel to fine-grained, trough and planar cross-bedded and current-rippled sands. Occasional red to greenish gray silty to muddy interbeds. Sands are laterally continuous, forming sheets, and multi-story, with many phases of channel cut and fill, generally bedded on a 0.5 m to 1 m scale (Owen et al. 2015a). In the sections studied, this facies association ranged from around 5 m to over 25 m thick. Although sometimes interpreted as the deposits of a braided-river system (Peterson 1984; Robinson and McCabe 1997;

Kjemperud et al. 2008), many authors have identified bedforms suggestive of low sinuosity but laterally stable channels in this facies association (Tyler and Ethridge 1983; Miall and Turner-Peterson 1989; Kjemperud et al. 2008). Owen et al. (2015a) interpreted it as the deposits of a mixed meandering to braided system with flashy discharge, formed in conditions of generation of flow accommodation space but high sediment supply.

Isolated Channel Fill Facies Association (Fig. 2B).---Laterally restricted gravel to fine-grained sandstones; sometimes lens-shaped in cross-sectional geometry; single story. May be planar and trough cross-bedded, contain lateral-accretion surfaces where flow direction can be determined, current-rippled and climbing rippled, sometimes bioturbated. Rarely more than 1.5 – 2 m thick (Owen et al. 2015a). Interpreted as fixed, anastomosing or meandering fluvial channels (Kirkland 2006; Cooley and Schmidt 2008; Kjemperud et al. 2008; Galli 2014; Owen et al. 2015a) with flashy discharge (Kirkland 2006; Galli 2014). Formed when accommodation-space generation and sediment supply are approximately equal.

Well-Drained Floodplain Facies Association (Fig. 2C).---Red, reddish-brown and reddish-purple silty mudstone to claystone. May contain calcareous nodules, green or purple mottles, burrows, rootlets, soil slickensides and other evidence of soil-forming processes. Interpreted as variably developed paleosols formed during exposure of the land surface above the water table (Dodson et al. 1980; Demko et al. 2004; Kirkland 2006; Jennings et al. 2011; Owen et al. 2015a), formed under conditions of low accommodation space and sediment supply.

Poorly Drained Floodplain Facies Association (Fig. 2D).---Green to purplish-green silts and mudstones. Laminated on a mm to cm scale, or structureless. May contain purple mottles, rare, small calcareous nodules, and sparse evidence for root traces (Kirkland 2006; Jennings et al. 2011). Deposited on a poorly drained floodplain where the water table was near the surface (Dodson et al. 1980; Jennings et al. 2011). Closely associated with the carbonate-dominated wetland or lacustrine

facies association (Dunagan and Turner 2004; Jennings et al. 2011). Formed on interfluvies when accommodation-space generation and sediment supply are approximately equal.

Carbonate-Dominated Wetland or Lacustrine Facies Association (Fig. 2E).---Greenish gray silty mudstone, laminated on a mm to cm scale or structureless, with abundant large calcareous nodules distributed throughout. Massive, nodular, hard, gray limestone beds, sometimes laterally discontinuous. May contain freshwater sponge spicules (Dunagan 1999), other invertebrate fossils (Dunagan and Turner 2004; Kirkland 2006), and plant material (Kirkland 2006). Closely associated with the poorly drained floodplain facies association (Dunagan and Turner 2004). Deposited in a perennial to ephemeral lacustrine environment (Dunagan and Turner 2004; Jennings et al. 2011) when accommodation-space generation exceeded sediment supply, and there was little clastic input.

Clastic-Dominated Wetland or Lacustrine Facies Association (Fig. 2F).---Gray, homogeneous, structureless silty mudstone. May weather to a distinctive popcorn texture, and contain rare gypsum nodules, particularly in the southern part of the study area. Occasional carbonate nodules, where present often occurring in bands. Thin (< 50 cm), hard, laterally discontinuous, bright green welded tuffs commonly distributed throughout; lens-like sand bodies in this facies are also commonly bright green and massive (Dodson et al. 1980). May contain unioid bivalves and gastropods. Deposited in perennial to ephemeral shallow lakes that in the south of the area have been interpreted as alkaline-saline (Turner and Fishman 1991; Dunagan and Turner 2004). Deposited when accommodation-space generation exceeded sediment supply, in an environment where there was some clastic input.

The six facies associations above were identified on sedimentological logs measured in the field and, where enough sedimentological detail had been provided, on the 245 published logs from the literature (online supplementary material). Repeated patterns of facies associations were observed across the logs, and were used to build the sequence stratigraphic framework outlined

below. Magnetostratigraphic data and radiometric dates were then plotted on the sequence stratigraphic framework to (1) test its robustness and (2) provide absolute dates for each systems tract.

Magnetostratigraphy

The majority of specimens were magnetically weak with low intensities ($< 1 \times 10^{-3}$ A/m), which resulted in noisy data with high mean angular deviations and high α_{95} (95% confidence limit) values for the bed mean directions, in common with other magnetostratigraphic studies of the Morrison Formation (Steiner et al. 1994; Steiner 1998). During the demagnetization process, some specimens reached intensities lower than the magnetometer's sensitivity levels, and measurement was terminated. We excluded sections where we did not observe any changes in north-south inclination, because we are unable to exclude diagenetic overprinting of the magnetic signal at a later date. Based on a reconstruction of virtual geomagnetic poles relative to polar-wander curves, Maidment et al. (2017) suggested that magnetic overprinting had occurred at DNM-US40 (Fig. 1), and this is supported by observations of a thin section of a sandstone at this locality, which contained epidote, which probably formed during a postdepositional phase of fluid flow through the area, and may have overprinted the depositional magnetism. Beds that exhibited reliable paleomagnetic data are shown in Fig. 3, and a summary of the paleomagnetic data, tied to the Geomagnetic Polarity Timescale, is shown in Fig. 4.

Sequence Stratigraphy

The upstream controls of tectonics and climate were the predominant causative mechanisms of A/S in the Morrison basin (Roca and Nadon 2007; Owen et al. 2015b). Numerous previous works have examined the climate of the Morrison Formation (e.g., Dodson et al. 1980; Demko and Parrish 1998; Ash and Tidwell 1998; Tidwell et al. 1998; Hasiotis and Demko 1998; Demko et al. 2004; Dunagan and Turner 2004; Engelmann et al. 2004; Good 2004; Turner and Peterson 2004; Jennings and Hasiotis 2006; Jennings et al. 2011; Myers et al. 2014; Tanner et al.

2014), and generally suggest a seasonally arid climate in the south, with perhaps more humid conditions to the north. There is also some evidence that humidity increased over the time represented by the formation. However, no large-scale climatic changes during Morrison deposition have been identified.

Although the tectonic setting of the Morrison depositional basin has been debated (Currie 1998; DeCelles 2004; Christiansen et al. 2015; Owen et al. 2015a), a large amount of evidence now suggests that Sevier orogenic loading began in the Late Jurassic (Royse 1993; Decelles 2004; Fuentes et al. 2009), and that the Morrison basin was located in the back-bulge depozone of the Sevier foreland basin (DeCelles and Burden 1992, DeCelles and Currie 1996; Currie 1997, 1998; Decelles 2004). Subsidence in the basin was likely enhanced by dynamic processes relating to the subduction of the Farallon plate under the North American craton (Currie 1998; DeCelles 2004; Christiansen et al. 2015). Changes in A/S in the Morrison basin were therefore likely driven by regional tectonics, a conclusion also reached by Robinson and McCabe (1997), Roca and Nadon (2007), Jennings et al. (2011), and Owen et al. (2015b). An assumption herein is that tectonic events affected accommodation space across the Morrison basin nearly synchronously, allowing its effects on fluvial architecture to be temporally correlated using sequence stratigraphy. While we acknowledge that this is an assumption, it does not appear unreasonable in the light of the tectonic history of the basin, the numerous previous studies that have considered regional tectonics to be a primary driver of A/S in the Morrison basin (Robinson and McCabe 1997; Roca and Nadon 2007; Jennings et al. 2011; Owen et al. 2015b), and general support for the sequence stratigraphic framework by magnetostratigraphy and radiometric dating, which we detail below.

Sequence A, Systems Tract 1.---The lowest systems tract in the Morrison comprises the amalgamated channel belt facies association in the west, while in the east, at CC-DV, it comprises the isolated channel fill facies association (Figs. 5, 6). Where the base of the systems tract is visible (e.g., MO-CL; DNM-US40; CC-DV), the change to fluvial facies is abrupt, and channels incise into

underlying sabkha and shallow marine deposits, often with meter-scale relief (Owen et al. 2015b), so the base is interpreted as a sequence boundary. The systems tract is part of distributive fluvial systems identified extending from west to east across the region (Owen et al. 2015a, 2015b). Systems tract A1 is interpreted as being deposited under low A/S conditions, with A/S increasing slightly across the fluvial system from west to east, probably representing a decrease in sediment supply in the distal parts of the fluvial system. The systems tract is recognized in Utah, western Colorado, New Mexico, and CC-DV in the southern part of the Colorado Front Range. Farther east and north of this region, the systems tract was either not deposited or is correlative with shallow-marine strata.

Magnetostratigraphic results at TO-TSC and DMN-DQ indicate that the base and top of the A1 systems tract is normal polarity, with a reversal in the middle (Fig. 3). A radiometric date obtained close to the top of the overlying A2 systems tract indicates an age of 152.29 +/- 0.27 (see below; Fig. 4; Trujillo and Kowallis 2015) in a reversed interval. The reversal is therefore probably CM25An.2r (152.38–152.24 Ma; Malinverno et al. 2012). Since the A2 is conformable with the A1, and assuming that no reversals are missing at the base of the A2, the normal-polarity interval at the top of the A1 is probably CM25An.3n (152.58–152.38 Ma), the reversal in the middle of the A1 is CM25Ar.1r (152.73–152.58 Ma), and normal polarity at the base of the A1 is CM25Ar.1n (152.86–152.73 Ma). The maximum age of the A1 is therefore 152.86 Ma and the minimum age is 152.38 Ma, giving maximum duration of 0.48 million years for its deposition, although it was probably somewhat shorter, because the top of CM25An.3n extends into the A2 and the bottom of CM25Ar.1n is unsampled.

This interpretation is consistent with Steiner (1998), who also interpreted the base of the Salt Wash to represent CM25. Radiometric dates from the underlying Tidwell Member obtained close to DNM-DQ and TO-TSC indicate ages of 156.84 +/- 0.59 Ma and 156.77 +/- 0.55 Ma respectively (Trujillo and Kowallis 2015), and indicate that Tidwell deposition occurred before CM30

(Malinverno et al. 2012). This suggests that the sequence bounding unconformity at the base of systems tract A1 may have been as long as 4 million years in duration.

Sequence A, Systems Tract 2.---Systems tract A2 comprises the isolated channel fill facies association, and it represents an increase in A/S relative to systems tract A1 (Figs. 5, 6). It is identified across Utah, New Mexico and Colorado, although not at FCO-HT2, and it appears not to have been deposited there. A radiometric date close to the top of the systems tract at CC-DV indicates an age of 152.29 +/- 0.27 (Fig. 4; Trujillo and Kowallis 2015; correlated onto our logged section using data from Kowallis et al. 1998 and the published logs in Carpenter 1998).

Magnetostratigraphic data from TO-TSC and MO-CL (Fig. 3) indicate that the middle part of the A2 systems tract was normal polarity, while the upper part is reversed. The radiometric date near the top of the A2 at CC-DV correlates the reversal to CM25An.2r, and the underlying normally polarized section is therefore CM25An.3n (see above; Fig. 4). The maximum age of the systems tract is therefore 152.58 Ma and the minimum age is 152.24 Ma. The maximum duration of the systems tract was 0.34 million years, although CM25An.3n extends downwards into the A1, so it is likely that it was somewhat shorter.

Sequence B, Systems Tract 3.---This systems tract comprises the amalgamated channel belt facies association in some areas (e.g., DNM-DQ; BL-SH; CC-DV; BR-MDQ; Fig. 7A), while at others, it comprises the amalgamated channel belt facies association at the base, but the isolated channel fill and well-drained floodplain facies associations towards the top (e.g., DEN-I70; SH-RG). The systems tract was deposited under low A/S conditions, and the base is interpreted as a sequence boundary due to an abrupt change in fluvial amalgamation (Fig. 7A); the amount of relief on this surface is difficult to determine from individual outcrops (Figs. 5, 6, 8). In proximal areas, primarily in the south west of the depositional basin, A/S was very low early on during the systems tract, and increased slightly upwards, perhaps representing a decrease in sediment supply. To the east and north, A/S was higher throughout the systems tract, indicated by the presence of isolated channel fill and well-

526 drained flood plain facies associations throughout, and this probably represents conditions of lower
527 sediment supply distal to the main distributive fluvial systems (e.g., Fig. 9, SH-RG). The systems tract
528 is present across Utah, New Mexico, Colorado, and in the eastern and northern parts of the Bighorn
529 Basin of Wyoming and southern Montana.

530 Trujillo and Kowallis (2015) reported a date from the base of this systems tract at Little
531 Cedar Mountain (correlated using data from Currie 1997, 1998) of 152.14 ± 0.51 . New single-
532 crystal U-Pb radiometric dates obtained from zircons in mudstones above and below an ~1.5-m-thick
533 sand at GU-DC near the top of this systems tract recorded ages of 151.1 ± 0.58 Ma (2 sigma) and
534 151.43 ± 0.36 Ma (2 sigma; Fig. 4). Unfortunately, GU-DC appears to have been magnetically
535 overprinted subsequent to deposition, but magnetostratigraphic data from TO-TSC, MO-CL, DMN-
536 DQ, and DEN-I70 consistently indicate reversed polarity at the base and top of the interval and
537 normal polarity in the middle. The radiometric date at the base of the systems tract suggests that
538 the reversal there is a continuation of CM25An.1r (152.07–151.81 Ma; although given the error on
539 the date, this reversal could also be CM25An.2r; Malinverno et al. 2012). The radiometric dates at
540 the top of the systems tract suggest the reversed interval there is CM25r (151.56–151.36 Ma;
541 although given the error bars on the dates, the reversal could also be CM24Ar; Malinverno et al.
542 2012). The normally polarized interval in the middle of the systems tract therefore lies in
543 CM25An.1n. Steiner (1998) interpreted the top of the Salt Wash as being deposited in CM24n at
544 sites in New Mexico and western Colorado, but we find it to be slightly older due to refinements in
545 dating of the GPTS (Malinverno et al. 2012). The B3 systems tract therefore extended from a
546 maximum of 152.07 Ma to a minimum of 151.36 Ma, giving a maximum duration of 0.71 million
547 years. The sequence-bounding unconformity at the base of the systems tract was probably just a few
548 hundred thousand years long (Fig. 4). The Kimmeridgian-Tithonian boundary, currently dated at
549 152.1 ± 0.9 Ma (Cohen et al. 2013), probably lies at the base of this systems tract.

550   **Sequence B, Systems Tract 4.**---In the southwest of the study area, in Utah, western
551 Colorado, and New Mexico, this systems tract predominantly comprises the clastic-dominated
552 wetland or lacustrine facies association, sometimes interbedded with the poorly drained floodplain
553 facies association (Figs 5, 6, 7B). In the Colorado Front Range it comprises the carbonate-dominated
554 wetland or lacustrine facies association interbedded with the poorly drained floodplain facies
555 association (Fig. 8), while in the Bighorn Basin, poorly-drained floodplain deposits are interbedded
556 with clastic-dominated wetland or lacustrine facies (Figs 6, 7C, 8, 9). The isolated channel fill facies
557 association is sometimes observed surrounded by the poorly-drained floodplain facies association
558 or, more rarely, the clastic-dominated wetland or lacustrine facies association. This systems tract is
559 interpreted as representing very high A/S conditions, leading to widespread wetland or lacustrine
560 facies across almost the entire depositional basin. Anastomosing-river deposits, which are often
561 associated with rapid aggradation (Holbrook and Schumm 1999), have been observed in this systems
562 tract (Cooley and Schmidt 1998) at BZ-SCQ (Fig. 9). The B4 systems tract is the most widespread, and
563 is present across the whole depositional basin.

564   Sparse magnetostratigraphic data obtained in this systems tract at DEN-I70 and DNM-DQ
565 indicate normal polarity (Fig. 3), although a higher sampling density would be preferable. Thanks to
566 the bentonitic nature of the mudstones in this systems tract in the southwest, several radiometric
567 dates have been measured from horizons in it (Fig. 4). Trujillo and Kowallis (2015; correlated using
568 logs in Turner and Peterson 1999) reported four radiometric dates at Notom, Montezuma Creek, and
569 Dinosaur National Monument, Utah; Trujillo et al. (2014; correlated using Kirkland and Carpenter
570 1994) reported a date at the Mygatt-Moore dinosaur quarry, Colorado; Galli et al. (2018; correlated
571 using Galli [2014] and Kirkland and Carpenter [1994]) reported dates at Mygatt-Moore and the
572 Fruita Paleontological Area, Colorado; and Deblieux et al. (2018; directly correlated to our logs BL-
573 SM and BL-ZH) reported a date from a location close to Blanding, Utah. Trujillo et al. (2014) and Galli
574 et al. (2018) reported conflicting dates for the Mygatt-Moore dinosaur quarry. Trujillo et al. (2014)
575 found a date of 152.18 +/- 0.29 Ma for the quarry itself, while Galli et al. (2018) resolved a date of

150.218 +/- 0.2 Ma for a horizon 4 m below quarry level. The discrepancy led Galli et al. (2018) to suggest that the Trujillo et al. (2014) date may have been based on zircons reworked from a lower level, and thus we disregard the older date here. Ignoring the Mygatt-Moore date of Trujillo et al. (2014), dates from the B4 systems tract range from 151.88 Ma to 149.1 Ma incorporating errors. The young age of 149.1 Ma was obtained from a sample at Montezuma Creek (Trujillo and Kowallis 2015) which has a large error (149.74 +/- 0.64 Ma); the other dates range from 151.88 Ma to 150.02 Ma. This age range overlaps with that of systems tract B3 (see above), with which systems tract B4 is conformable. Given the conformable nature of B3 and B4, normal polarity in B4 probably represents CM25n or a normal interval in CM24r (Fig. 4). Magnetic reversals in B4 have very probably been missed because sampling was restricted to sparse sandstones.

Sequence C, Systems Tract 5.---Systems tract C5 generally comprises the isolated channel fill facies association interbedded with well-drained floodplain deposits (Fig. 7B, C); in some areas, notably CC-DV, in central parts of New Mexico, and in much of the Bighorn Basin, the amalgamated channel belt facies association is also observed (Figs. 8, 9). Systems tract C5 is interpreted as being deposited in low A/S conditions, when both accommodation space and sediment supply were low, leading to the widespread formation of paleosols. In areas where sediment supply was slightly higher, stacked channels and sheet sandstones are observed. Its base is a sequence boundary, identified by the sharp change from lacustrine deposits to stacked paleosols (Fig. 7B, C). Present over most of the depositional basin, systems tract C5 seems to be absent in the southwesternmost exposures in Utah (Fig. 7D), where it was either never deposited, perhaps because the Sevier forebulge had begun to migrate northeastwards (Currie 1998) or was subsequently eroded during forebulge migration.

A radiometric date obtained near the top of the C5 systems tract by Galli et al. (2018; correlated based on Galli 2014 onto our MO-CL section) has an age of 149.451 +/- 0.19 Ma. Magnetostratigraphic results at DNM-DQ, DEN-I70, and FCO-HT1 indicate reversed polarity at the

base and top of the systems tract and normal polarity in the middle (Fig. 3). The radiometric date at the top of the systems tract lies within CM23r.2r (149.78–149.21 Ma; Malinverno et al. 2012). The normal-polarity interval in the middle of the systems tract is therefore CM24n (150.24–149.78 Ma; Fig. 4), and the reversal at the base is CM24r.1r (150.44–150.24 Ma). This indicates a maximum age of 150.44 Ma and a minimum age of 149.21 Ma, and a total maximum duration of 1.23 million years for deposition of the C5; the sequence-bounding unconformity at the base of the C5 was probably very short lived. Maidment et al. (2017) suggested that the reversal at the base of the systems tract observed at DNM-DQ was CM22Ar, but this was based on an older version of the GPTS (Tauxe 2010). Steiner (1998) also identified a reversed interval near the top of Morrison sections in the Four Corners region, and interpreted it as CM22r, slightly younger than is found herein, again due to refined dating of the GPTS (Malinverno et al. 2012).

Sequence C, Systems Tract 6.---Systems tract C6 comprises the clastic-dominated wetland or lacustrine facies association and the poorly drained floodplain facies association (Figs. 8, 9). Systems tract C6 is was deposited under high A/S conditions in areas where sediment supply was low. C6 appears to be primarily restricted to the Bighorn Basin of Wyoming; its presence in Montana was difficult to constrain due to lack of exposure, but it appears to be present in the logged section of Turner and Peterson (1999) at Toston, Montana. The absence of the systems tract in the south of the depositional basin is probably because it was never deposited there due to northeastwards migration of the Sevier forebulge across the region (Currie 1998). Radiometric dates are not known from this systems tract, making age constraints difficult, and no reliable magnetostratigraphic data were obtained, but the systems tract comprises the youngest deposits in the Morrison basin. Facies associations for each systems tract for each sampled locality and for each of the 245 published logs from the literature (online supplementary material) were plotted geographically using ArcGIS (ESRI 2011). These were used to build paleogeographic maps for each systems tract, and are shown in Figs. 10-12.

626

627

DISCUSSION

628 *Autocyclic vs. Allocyclic Controls on Fluvial Architecture*

629   The application of sequence stratigraphy to terrestrial strata is controversial because of the
630 numerous factors that can control fluvial architecture. Autocyclic controls, such as river avulsion, can
631 produce fluvial architecture strongly reminiscent of changes produced by allocyclic controls like
632 tectonics and climate. Although autocyclic processes undoubtedly had an influence on local
633 sediment architecture in the Morrison basin (Owen et al. 2015a, 2015b), especially in terms of
634 downstream amalgamation of channel deposits, three factors suggest that allocyclic controls were
635 the dominant mechanism controlling A/S: (1) Consistency of magnetostratigraphic data. Where
636 magnetostratigraphic data was obtained, polarity data were found to generally support the
637 sequence stratigraphic framework proposed (Fig. 5). (2) Support by radiometric dating and previous
638 magnetostratigraphic studies. The framework is supported by previously published radiometric
639 dates, which in general have relatively small error bars (< 1 million years), and using these
640 radiometric dates to tie new magnetostratigraphic data to the global Geomagnetic Polarity
641 Timescale results in conclusions very similar to those of Steiner et al. (1994) and Steiner (1998), the
642 previous magnetostratigraphic studies of the formation (Fig. 6). (3) Good correlation with the results
643 of previous studies. As demonstrated below, several authors have previously attempted to correlate
644 the Morrison Formation across small parts of the basin. Their conclusions are generally in very good
645 agreement with those presented herein, suggesting general support for the framework.

646   Nonetheless, further detailed magnetostratigraphic studies and radiometric dating are
647 required to test the hypothesis that fluvial architecture was predominantly allocyclically controlled in
648 the Morrison basin, and robustly test the proposed framework.

649 *Comparison with Previous Studies (Fig. 13)*

650 Currie (1997) examined the Morrison Formation of northern Utah in an explicitly sequence
651 stratigraphic context. Currie (1997) considered the Windy Hill, Tidwell, and Salt Wash as
652 representing a conformable highstand systems tract which prograded into a marine basin. The lower
653 part of the Salt Wash represented a late-stage aggradational systems tract, where generation of
654 accommodation space slowed and sands began to amalgamate. This corresponds with the A
655 sequence identified here, and differs in that we consider the lower part of the Salt Wash to be a
656 degradational systems tract in the terminology of Currie, separated from the underlying Tidwell by a
657 sequence boundary. Currie (1997) considered there to be a second sequence in the Salt Wash, with
658 a sequence boundary at the base, which he referred to as the UJ-2 sequence. Currie's UJ-2 sequence
659 corresponds with the B sequence herein. The scheme presented by Currie (1997) and that presented
660 herein are therefore entirely complementary in the Uinta mountains of Utah, with the only clear
661 difference being whether the lower part of the Salt Wash represents a late-stage aggradational
662 systems tract or a degradational systems tract.

663 Currie (1997) did not study the Morrison in the Bighorn Basin, but correlated the Uinta
664 mountains outcrops with those in the Bighorn Basin based on the work of DeCelles and Burden
665 (1992). Currie (1997) correlated the Pryor Conglomerate (also known as the Lakota Conglomerate) at
666 the base of the overlying Cloverly Formation with the top of the Salt Wash, a situation that results in
667 the entirety of the Brushy Basin in Utah correlating with the Cloverly Formation of the Bighorn Basin.
668 This scenario is unsupported herein based on several lines of evidence. Firstly, radiometric dating
669 and biostratigraphy clearly demonstrate an Early Cretaceous age for the Cloverly (e.g., Ostrom 1970;
670 Oreska et al. 2013; Cifelli and Davis 2015), but a Late Jurassic age for the Brushy Basin (Trujillo and
671 Kowallis 2015). The dinosaurian fauna of the Brushy Basin in Utah is much more similar to the fauna
672 of the Morrison Formation in the Bighorn Basin than it is to the fauna of the Cloverly Formation (e.g.,
673 Ostrom 1970; Foster 2000). Widespread coarse conglomerate (the Buckhorn Conglomerate of the
674 Cedar Mountain Formation), lithologically similar to the Pryor Conglomerate, is found at the top of
675 the Brushy Basin in Utah (e.g., Aubrey 1998; Roca and Nadon 2007), in the same stratigraphic

676 position occupied by the Pryor (= Lakota) Conglomerate in the Bighorn Basin. Therefore, in line with
677 most other workers, herein the Buckhorn Conglomerate and Pryor (= Lakota) Conglomerate are
678 considered roughly time-correlative, and correlation of the Pryor Conglomerate to the Salt Wash is
679 rejected.

680 Kjemperud et al. (2008) studied the Tidwell and Salt Wash in south-central Utah. They
681 identified four cycles that they attributed to changes in accommodation space and sediment supply.
682 Cycle 1 corresponds to the Tidwell. They then recognized three cycles of degradation followed by
683 aggradation: three sequences. Herein, the Salt Wash is considered to comprise two sequences, but it
684 may be that an additional sequence is present at the base of the Salt Wash in the most proximal
685 parts of the depositional basin, and this is especially likely since the Salt Wash has been described as
686 a progradational distributive fluvial system (Owen et al. 2015a, 2015b). Further
687 magnetostratigraphic work in the region would allow correlation of the sequences of Kjemperud et
688 al. (2008) to those identified here.

689 Demko et al. (2004) recognized depositional sequences in the Morrison Formation based on
690 paleosols. They recognized basal Morrison, mid-Morrison, and top-Morrison sequence-bounding
691 unconformity paleosols; however, not all unconformity paleosols could be recognized in every
692 section, and it is difficult to distinguish between “unconformity” paleosols and other paleosols in
693 individual sections. Thus it is not possible to compare the framework presented herein with that
694 presented by Demko et al. (2004).

695 No other authors have explicitly recognized sequences in Morrison Formation. However,
696 several workers have correlated locally based on lithology, and these schemes can be compared to
697 that presented here. Turner-Peterson (1986), Peterson (1984), Godin (1991), and Robinson and
698 McCabe (1997) all studied the Salt Wash in various locations on the Colorado Plateau. Turner-
699 Peterson (1986), Peterson (1984), and Godin (1991) recognized three “cycles” in some outcrops of

the Salt Wash that appear to broadly correlate with the A1, A2, and B3 systems tracts proposed here.

Owen et al. (2015b) examined vertical trends in the Salt Wash in Utah. They interpreted the Salt Wash as a progradational distributive fluvial system and considered that local changes in channel amalgamation vertically were likely related to autocyclic controls such as local avulsion events. However, Owen et al. (2015b) noted a large degree of variability in vertical trends, with some sites showing evidence for retrogradation of the system. The conclusion of progradation by Owen et al. (2015b) is complementary to the conclusion reached here: the distributive fluvial system could have been overall progradational, with a degree of aggradation being represented by systems tract A2. This aggradational systems tract might not be recognized in the most proximal parts of the basin, where the degradational deposits of B3 may have removed flood plain deposits by incision and erosion. The magnetostratigraphic data presented herein suggest that the change from degradation/progradation to aggradation between systems tracts A1 and A2 is time-correlative. Further magnetostratigraphic studies would help to test this hypothesis.

Currie (1998) and Galli (2014) examined the Brushy Basin Member in the Uinta Basin and the Grand Valley near Grand Junction, Colorado, respectively. Both identified a lower section comprising reddish mudstones and an upper section comprising gray-purple ash-rich mudstones topped with red, color-banded siltstones. The lower section represents the top of systems tract B3, the grayish section is B4, and the reddish, color-banded section corresponds to C5. Both authors found good local correlation using these color-based distinctions, and they accord well with the scheme suggested herein.

Peterson and Turner (1998) described the stratigraphy along the Front Range near Denver. They divided the Morrison into a lower member, corresponding to systems tracts A2 and B3, a middle member, which is B4, and an upper member, which is C5. The division of their "members" in the Front Range corresponds exactly with those proposed here. In the only detailed examination of

Morrison stratigraphy in the Bighorn Basin, Ostrom (1970) divided the formation into three units, I-III. Ostrom's units appear to correspond exactly with systems tracts B4, C5, and C6.

It is clear that other workers have identified very similar divisions of the Morrison Formation when examining local stratigraphy, and this lends strong support to the framework proposed.

Morrison Chronostratigraphy vs Lithostratigraphy

The J5 unconformity, originally envisaged as separating Middle Jurassic from Upper Jurassic deposits in the Western Interior (Pipringos and O'Sullivan 1978), may be represented by the sequence boundary at the base of the A1, a proposal put forward by Anderson and Lucas (1997, 1998). Others (Peterson 1994, Currie 1998, Owen et al. 2015a, 2015b), however, have envisaged the J5 unconformity lying below the Tidwell Member. Currie (1998) and Owen et al. (2015a, 2015b) considered the Tidwell deposits to be genetically related to the overlying fluvial deposits of the Salt Wash Member; in particular, Owen et al. (2015b) described the Tidwell as distal deposits of a distributive fluvial system. Radiometric dating of the Tidwell at Notom and Dinosaur National Monument constrain it as no younger than 156.22 million years old (Trujillo and Kowallis 2015), while the combination of radiometric dating and magnetostratigraphy presented here suggests that the base of the A1 systems tract may be around 153 million years. The temporal offset between the Tidwell and base of the Morrison is supportive of a sequence boundary at the base of the A1 systems tract that represents an unconformity of ~ 3 million years in duration, and thus it is possible that the J5 of Pipringos and O'Sullivan (1978) occurs at the base of the A1. Further radiometric dating, detailed sedimentological analyses, and field mapping are required to test this proposal.

In the Colorado Plateau, the Morrison can be divided lithostratigraphically into the lower sand-dominated Salt Wash Member and the upper mud-dominated Brushy Basin Member. The boundary between the two is usually placed at the last stacked sand, although in some locations (e.g., MO-CL) it is difficult to place, and the members interfinger. The concepts of continental sequence stratigraphy allow the recognition that during conditions of low A/S, where stacked

channel sand sheets may form in areas of high sediment supply, paleosols may form in areas of lower sediment supply. The amalgamated channel belt facies association, which characterizes the Salt Wash, and the well-drained floodplain facies association, which characterizes part of the Brushy Basin, can therefore be part of the same systems tract. At the base of the Brushy Basin in many sites (TO-TSC; MO-CL; DNM-US40, DEN-I70) the well-drained floodplain facies association is identified and is interbedded with either amalgamated channel belt sands or isolated channel fill sands. At other sites (BL-ZH, DNM-DQ, CC-DV), amalgamated channel belt sands are directly overlain by the clastic or carbonate-dominated wetland or lacustrine facies association. The well-drained floodplain facies association, which would lithostratigraphically be considered part of the Brushy Basin Member, is actually genetically and temporally related to the amalgamated channel belt facies association, and is part of the same systems tract (B3). The base of the Brushy Basin Member is therefore not a contemporaneous horizon; sometimes it lies within the B3 systems tract, while at other locations it is at the base of the B4 systems tract.

Future Directions

The proposed stratigraphic framework is the first attempt to divide the ~ 9 million years of time represented by the Morrison Formation into packages that are more biologically and ecologically meaningful in the context of evolution: it is chronostratigraphic rather than lithostratigraphic. Detailed accounts of dinosaur quarry stratigraphy provided by many publications will allow dinosaur occurrences to be assigned to a systems tract based on this framework. These data will allow faunal change through time, patterns of biodiversity, the evidence for cladogenesis vs. anagenesis, stratigraphic congruence of existing phylogenetic trees, ecological partitioning, and community structure among Morrison dinosaur faunas to be examined.

The framework is a hypothesis, however, and requires farther testing. Several areas were not included in this study, and published data for them are lacking. The stratigraphy of the Morrison in the Black Hills of South Dakota and Wyoming requires farther documentation in the light of the

framework proposed here. Likewise, the Morrison in Arizona is largely undocumented, and studies there will shed light on the most proximal parts of the depositional system. Magnetostratigraphy appears to be a promising method of correlation, and could be used in these areas to test the framework. Much of the Morrison in Montana is poorly exposed, and there is little outcrop on public land, meaning that access proved difficult during the course of this study. Radiometric dating could prove important in linking outcrops in Montana to those farther south.

CONCLUSION

The Morrison Formation was deposited over a time period of around 9 million years, from about 156 to 147 million years ago, in the Late Jurassic. The primary driving mechanism behind the balance of accommodation space and sediment supply was probably tectonics: subsidence in the basin was driven by a combination of flexural subsidence associated with early stages of the Sevier orogeny and dynamic subsidence due to subduction of the Farallon plate under the North American craton. Assuming that subsidence occurred more or less contemporaneously across the basin, continental sequence stratigraphy can be used to correlate regionally. The formation can be divided into three depositional sequences and six systems tracts each representing a period of low ratio of accommodation space to sediment supply followed by high value of this ratio, and this division is supported by available radiometric and existing and new magnetostratigraphic data. The formation youngs to the north; in southern Utah, the base of the A1 systems tract is around 153 million years old while the top of the formation is around 150 million years old, but in the Bighorn Basin of Wyoming the base of the formation may be as young as 151 million years, while the top could be substantially younger than 148 million years. This chronostratigraphic framework provides a foundation for studies of paleoecology, community structure, and evolutionary dynamics in the iconic Morrison dinosaur fauna.

800

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1133

1134 **FIGURE CAPTIONS**

1135 **Figure 1:** The study area. Studied localities are as follows: **1**, SD-ZA: Zeigler anticline, near Sheridan,
 1136 Montana; **2**, BZ-SCQ: O’Hair quarry, near Bozeman, Montana; **3**, LT-FHR, roadcut along Fish Hatchery
 1137 Road, Lewiston, Montana; **4**, BR-MDQ, Mother’s Day quarry near Bridger, Montana; **5**, BR-SQ,
 1138 *Suuwassee* quarry near Bridger, Montana; **6**, SH-SQ, Red Canyon Ranch quarry near Shell, Wyoming;
 1139 **7**, SH-RG, Red Gulch, near Shell, Wyoming; **8**, SH-DQ, Dana quarry, near Hyattville, Wyoming; **9**, CO-
 1140 PI, exposures south of Cody, Wyoming; **10**, RA-SB, Sunshine Beach, Seminoe State Park, near
 1141 Rawlins, Wyoming; **11**, FCO-HT1, Horsetooth Reservoir section 1, near Fort Collins, Colorado; **12**,
 1142 FCO-HT2, Horsetooth Reservoir section 2, near Fort Collins, Colorado; **13**, DEN-I70, I-70 road cut,
 1143 near Denver, Colorado; **14**, CC-DV, Death Valley, Garden Park Paleontological Area, near Canyon
 1144 City, Colorado; **15**, GU-DC, Dinosaur Cove, Curecanti National Recreation area, near Gunnison,
 1145 Colorado; **16**, DNM-US40, exposures south of US-40, near Vernal, Utah; **17**, DNM-DQ, section near
 1146 the visitor center, Dinosaur National Monument, near Vernal, Utah; **18**, MO-CL, Cisco Landing, near
 1147 Moab, Utah; **19**, BL-SM, exposures near Shumway mines, near Blanding, Utah; **20**, BL-ZH, Zeke’s
 1148 Hole, near Blanding, Utah; **21**, TO-TSC, exposures south of highway 24 between Torrey and

1149 Hanksville, Utah. Coordinates for measured sections are given in Online Supplementary Materials;
1150 coordinates for dinosaur quarries are not given to protect paleontological resources. Gray squares
1151 represent the locations of previously published sedimentological logs of the Morrison Formation
1152 that were used to supplement data collected in this study.

1153 **Figure 2:** Facies associations. **A)** stacked fluvial channels characteristic of the amalgamated channel
1154 belt facies association at TO-TSC; **B)** cross-bedded single story fluvial sand characteristic of the
1155 isolated channel fill facies association at CO-PI; **C)** red mudstones characteristic of the well-drained
1156 floodplain facies association at MO-CL; **D)** greenish-gray laminated silts characteristic of the poorly
1157 drained floodplain facies association at RA-SB; **E)** nodular carbonate in gray siltstone characteristic of
1158 the carbonate dominated wetland or lacustrine facies association at DEN-I70; **F,** gray, bentonitic
1159 mudstone characteristic of the clastic-dominated wetland or lacustrine facies association at MO-CL.

1160 **Figure 3:** Correlation showing sedimentological logs and magnetostratigraphic results. All sites at
1161 which magnetostratigraphic reversals were obtained are shown. For site names and locations, see
1162 Figure 1. For sedimentological symbols see Figs 5, 6, 8 and 9. **1,** location of a radiometric date
1163 obtained by Trujillo and Kowallis (2015), of 150.91 +/- 0.43 Ma.

1164 **Figure 4:** Summary of results showing radiometric dates, magnetostratigraphy (**MS**), magnetochrons
1165 (**MC**), simplified sedimentology (**Sed**), systems tracts (**ST**), facies associations, and maximum and
1166 minimum ages of systems tracts. See text for sources of radiometric dates. **ACF**, amalgamated
1167 channel fill; **ICF**, isolated channel fill; **PDFFP**, poorly drained floodplain; **WDFP**, well-drained
1168 floodplain; **WL**, carbonate or clastic dominated wetland or lacustrine; **Ma**, millions of years before
1169 present; **mst**, mudstone; **sst**, sandstone.

1170 **Figure 5:** Correlation of systems tracts from west to east across southern Utah and Colorado,
1171 flattened on the base of the B4 systems tract, which is present across the whole of the Morrison
1172 depositional basin. Note variation in scale of sedimentary logs.

1173 **Figure 6:** Correlation of systems tracts from south to north across eastern Utah, Wyoming, and
 1174 central Montana, flattened on the B4 systems tract, which is present across the whole of the
 1175 Morrison depositional basin. Note variation in scale of sedimentary logs.

1176 **Figure 7. A)** transition from systems tract A2 to B3 at TO-TSC; **B)** transition from systems tract B4 to
 1177 C5 at MO-CL; **C)** transition from systems tract B4 to C5 at CO-PI; **D)** transition from systems tract B3
 1178 to B4 at TO-TSC.

1179 **Figure 8:** Correlation of systems tracts from south to north across the Colorado Front Ranges and the
 1180 eastern side of the Bighorn Basin, Wyoming, flattened on the B4 systems tract, which is present
 1181 across the whole of the Morrison depositional basin. Note variation in scale of sedimentary logs.

1182 **Figure 9:** Correlation of systems tracts from west to east across southern Montana and northern
 1183 Wyoming, flattened on the B4 systems tract, which is present across the whole of the Morrison
 1184 depositional basin. Note variation in scale of sedimentary logs.

1185 **Figure 10:** Paleogeography of the A sequence based on 21 measured sections and supplemented
 1186 with data from published logs, the location of which are shown in Fig. 1. Location of Sevier
 1187 highlands, Mogollon highlands, Salt Wash DFS, and Vernal DFS follows Owen et al. (2015a). Location
 1188 of ancestral Rockies follows Peterson (1994). **A)** A1 systems tract showing maximum progradation of
 1189 DFS; **B)** A2 systems tract showing retrogradation of DFS and the formation of anastomosing channels
 1190 and ephemeral lakes. **DFS**, distributive fluvial system.

1191 **Figure 11:** Paleogeography of the B sequence based on 21 measured sections and supplemented
 1192 with data from published logs, the location of which are shown in Fig. 1. Location of Sevier
 1193 highlands, Mogollon highlands, Salt Wash DFS, and Vernal DFS follows Owen et al. (2015a). **A)** B3
 1194 systems tract showing maximum progradation of DFS beyond the extent of the A1 DFS, and the
 1195 development of a new DFS in the north; **B)** B4 systems tract showing lacustrine highstand. The name
 1196 “Lake T’oo’dichi” was proposed by Turner and Fishman (1991) for a lake occupying a location around

the four corners. In southeastern Colorado, lacustrine facies have been termed “Dinosaur Lake” (Lockley et al. 1986). Clastic-dominated lacustrine facies dominate to the west of the lake, while carbonate-dominated lacustrine facies dominate to the east, as would be expected given source areas to the west. The Bighorn DFS comprises anastomosing streams at this time. **DFS**, distributive fluvial system.

Figure 12: Paleogeography of the C sequence based on 21 measured sections and supplemented with data from published logs, the location of which are shown in Fig. 1. Location of Sevier highlands, Mogollon highlands, Salt Wash DFS, and Vernal DFS follows Owen et al. (2015a). **A)** C5 systems tract showing progradation of the Bighorn DFS and the migration of the Sevier forebulge over the southwestern portion of the study area; **B)** C6 systems tract showing poorly drained floodplain and carbonate-dominated wetlands, and the migration of the Sevier forebulge over the majority of the former depositional basin. **DFS**, distributive fluvial system.

Figure 13: A comparison of the sequence stratigraphic scheme presented in this study with units identified in previous works. Turner-Peterson (1986) and others: see text for a full list of references. **Lithostrat**, lithostratigraphy; **Seq**, sequence; **Tid**, Tidwell Member.

- Morrison outcrop
- Studied locality
- Published log

Radiometric dates	MS	MC	Sed mst sst	ST	Facies Assoc	Age (Ma)	
149.45+/-0.19 Ma					C6	WL/PDFP	
150.91+/-0.43 Ma		23r.2r			C5	ICF/WDFP	150.44 -
150.67+/-0.32 Ma		24n					149.21
151.23+/-0.54 Ma		24r.1r					
151.34+/-0.54 Ma		25n?			B4	WL/PDFP	151.88 -
150.22+/-0.20 Ma							150.02
149.74+/-0.64 Ma		25r			B3	ACF/ICF/ WDFP	152.07 -
151.10+/-0.58 Ma		25An.1n					151.36
151.43+/-0.36 Ma		25An.1r					
152.14+/-0.51 Ma		25An.2r			A2	ICF/WDFP	152.58 -
152.29+/-0.27 Ma		25An.3n					152.24
		25An.3n			A1	ACF/ICF	152.86 -
		25Ar.1r					152.38
		25Ar.1n					

Normal polarity

Reversed polarity

No data

Sandstone

Red mudstone

Grey mudstone

Green mudstone

— Sequence boundary   - - - - - Systems tract boundary   ..... Undefined boundary