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The Moenave Formation: Sedimentologic and stratigraphic context of the Triassic–Jurassic boundary in the Four Corners area, southwestern U.S.A.

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Abstract

The Moenave Formation was deposited during latest Triassic to earliest Jurassic time in a mosaic of fluvial, lacustrine, and eolian subenvironments. Ephemeral streams that flowed north-northwest (relative to modern geographic position) deposited single- and multi-storeyed trough cross-bedded sands on an open floodplain. Sheet flow deposited mainly silt across broad interchannel flats. Perennial lakes, in which mud, silt and carbonate were deposited, formed on the terminal floodplain; these deposits experienced episodic desiccation. Winds that blew dominantly east to south-southeast formed migrating dunes and sand sheets that were covered by low-amplitude ripples.

The facies distribution varies greatly across the outcrop belt. The lacustrine facies of the terminal floodplain are limited to the northern part of the study area. In a southward direction along the outcrop belt (along the Echo Cliffs and Ward Terrace in Arizona), dominantly fluvial–lacustrine and subordinate eolian facies grade mainly to eolian dune and interdune facies. This transition records the encroachment of the Wingate erg. Moenave outcrops expose a north–south lithofacies gradient from distal, (erg margin) to proximal (erg interior).

The presence of ephemeral stream and lake deposits, abundant burrowing and vegetative activity, and the general lack of strongly developed aridisols or evaporites suggest a climate that was seasonally arid both before and during deposition of the Moenave and the laterally equivalent Wingate Sandstone. We interpret growth of the erg as a consequence of marine regression during the latest Triassic through earliest Jurassic that exposed sediments on the coast of the back-arc sea to eolian reworking. Tectonic processes that created accommodation space enhanced preservation of the erg.

Keywords: Moenave; Wingate; Erg; Eolian; Ephemeral stream

1. Introduction

The Late Triassic through Early Jurassic represents one of the most significant intervals of geologic time in the evolution of terrestrial tetrapods, an interval marked by the appearance of all crown-group archosaur clades (pterosaurs, dinosaurs, phytosaurs, crocodilomorphs, rauisuchids, and aetosaurs) and the disappearance of several reptile and amphibian groups. These evolutionary events, in concert with marine events that included multiple Late Triassic extinction pulses and Early Jurassic radiations, took place against an extraordinarily

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dynamic backdrop of tectonic, climatic, and environmental changes that included Pangaean rifting, eustatic sea-level fluctuation, and the voluminous eruptions of the Central Atlantic Magmatic Province, the last of which may have triggered severe atmospheric and climatic disruptions (Tanner et al., 2004). Evaluating the potential role of these varied processes in guiding the narrative of terrestrial evolution requires careful examination of the record of sedimentary processes at numerous locations globally.

The continental Upper Triassic and Lower Jurassic of North America are well-represented by two great sedimentary successions, the Newark Supergroup of eastern North America, and the Chinle-Glen Canyon Group of the Colorado Plateau. While the former is thicker and ostensibly forms the more continuous (i.e., more conformable) sedimentary record, the latter is much more extensively exposed and contains a significant fossil record. Many details of the sedimentology of the Glen Canyon Group, the Moenave Formation and Wingate Sandstone, have been reported previously (e.g., Clemmensen and Blakey, 1989; Clemmensen et al., 1989; Olsen, 1989; Blakey, 1994). In this paper, we report additional data pertaining to the sedimentology of the Moenave and associated formations that provide a palaeoclimatic and palaeogeographic context for examining the fossil record of the Late Triassic through Early Jurassic in the Four Corners area (see Lucas and Tanner, this volume).

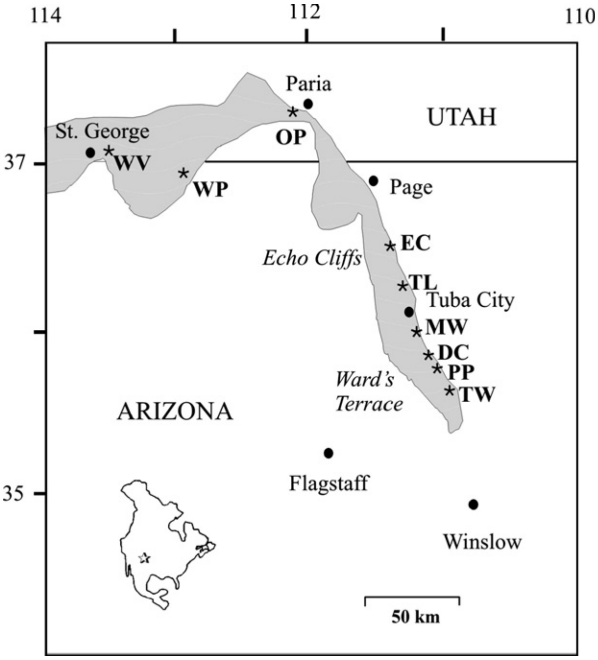

Fig. 1. Location map of the Moenave Formation outcrop belt. Locations referred to in the text at which sections were measured are DC = Dinosaur Canyon; EC = Echo Cliffs; MW = Moenkopi Wash; OP = Old Paria; PP = Paiute Point; TL = type location (measured near village of Moenave, north of Tuba City); TW = Tohachi Wash; WP = Whitmore Point; WV = Warner Valley.

2. Setting

2.1. Stratigraphy

In northern Arizona and southern Utah, the Moenave Formation comprises a succession of continental redbeds, averaging 100 m in thickness, that includes sandstone, siltstone, and mudstone deposited by fluvial, lacustrine, and eolian processes (Harshbarger et al., 1957; Wilson, 1967; Clemmensen et al., 1989; Olsen, 1989; Irby, 1996a; Lucas and Heckert, 2001; Tanner et al., 2002). These strata are exposed as cliffs, buttes, and hoodoos in northern Arizona along the Echo Cliffs and on Ward Terrace, and in Utah along the Vermillion Cliffs from Lee's Ferry to Zion National Park, and as far west as St. George, Utah (Fig. 1). Stratigraphic nomenclature currently in use assigns the Moenave to the Glen Canyon Group and subdivides the formation into the Dinosaur Canyon, Whitmore Point, and Springdale Sandstone members, in ascending order (Fig. 2). Over most of its outcrop belt, the Moenave Formation overlies the Owl Rock Member of the Chinle Formation (the Chinle Group of Lucas, 1993; Lucas et al., 1997) with pronounced unconformity and is overlain disconformably, in turn, by the Kayenta Formation (Glen Canyon Group). The basal Moenave-Owl Rock contact has been termed the J-0 unconformity and was long considered to coincide with the Triassic–Jurassic boundary (e.g., Pipiringos and O'Sullivan, 1978), but recent work suggests that basal Moenave strata are Late Triassic in age (Molina-Garza et al., 2003). East of the study area, the unconformity between the Owl Rock and Rock Point members of the Chinle Formation is termed the Tr-5 unconformity (Pipiringos and O'Sullivan, 1978).

The Dinosaur Canyon Member, as described at the type area east of Cameron, comprises reddish-orange to light-brown siltstones and sandstones of mainly fluvial and eolian origin, in proportions that vary across the outcrop belt (Harshbarger et al., 1957). At the type location of the Moenave Formation, near Tuba City, Arizona, the formation consists entirely of the Dinosaur Canyon Member (Fig. 3A). The Whitmore Point Member, comprising lacustrine shales, mudstones, siltstones, and limestones (Fig. 3B), conformably overlies Dinosaur Canyon strata only in the northernmost part of the outcrop belt (in northernmost Arizona and

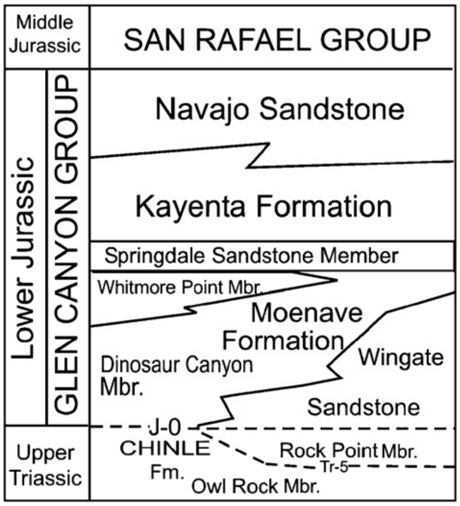

Fig. 2. Stratigraphy of upper Chinle Group–Glen Canyon Group currently in use by most workers. The Moenave and Wingate formations are assumed to be entirely Jurassic in age. The Rock Point Member is separated from the Wingate by a presumed unconformity that is considered to correspond to the Triassic–Jurassic boundary.

southern Utah). Dinosaur Canyon and (where present) Whitmore Point strata are overlain by the Springdale Sandstone. The Springdale Sandstone is a prominent cliff-former comprising rippled to cross-bedded ledges of coarse- to medium-grained sandstone, commonly conglomeratic at its base. Olsen (1989) noted, as did Marzolf (1993, 1994), that the lower contact of the Springdale is generally sharp, with erosional relief, and that the lithology of the unit generally bears greater similarity to sandstones in the overlying Kayenta Formation. Therefore, we follow earlier recommendations and consider the Springdale as a basal member of the Kayenta Formation (Marzolf, 1994; Lucas and Heckert, 2001; Tanner et al., 2002).

The Wingate Sandstone, which consists largely of eolian sandstones, is exposed to the east and south of the Moenave Formation outcrop belt and is considered a partial lateral correlative of the Dinosaur Canyon Member as the two units intertongue to some degree (Harshbarger et al., 1957; Clemmensen et al., 1989; Marzolf, 1993; Irby, 1996a). Earlier usage (e.g., Harshbarger et al., 1957) considered the Wingate Sandstone to consist of two units, the older Rock Point Member and the conformably overlying Luakachukai Sandstone Member. Later workers (e.g., Pipiringos and O'Sullivan, 1978; Clemmensen and Blakey, 1989) posited an unconformable relationship between these strata. Consequently, the Rock Point was assigned member status in the Chinle Formation, and the Lukachukai strata became synonymous with the Wingate Sandstone.

2.2. Age

Fossils of the crocodylomorph *Protosuchus* and tracks of the dinosaur ichnotaxon *Eubrontes* (Irby, 1996b) from the middle to upper Dinosaur Canyon Member (Fig. 4A) indicate that this part of the formation is likely of Hettangian age (Lucas and Heckert, 2001). The lower Dinosaur Canyon strata host *Grallator*-dominated trackways that lack *Eubrontes*, suggesting, although not proving, a latest Triassic (Rhaetian) age for the base of the formation (Morales, 1996; Lucas et al., 1997). However, palaeomagnetic evidence (Molina-Garza et al., 2003) and the lateral relationship with the Wingate Sandstone, which contains a demonstrably Upper Triassic fauna and ichnofauna in its lower part and Jurassic ichnofauna near its top (Lockley et al., 1992, 2004; Lucas et al., 1997; Lucas and Tanner, this volume), indicate that the Triassic–Jurassic boundary lies within both the Dinosaur Canyon Member and the Wingate Sandstone. The Whitmore Point Member was named by Wilson (1967) to designate lithologically distinct lacustrine strata, comprising gray-purple to reddish siltstones and shales stratigraphically high in the Moenave in the Vermillion Cliffs. These strata have produced fossils of semionotid fish and a palynoflora that is dominated by *Corollina* spp., indicating an Early Jurassic, probably Hettangian age (Peterson and Pipiringos, 1979).

The lithologies of the units measured in this study suggest that the uppermost strata of the Rock Point Member of the Chinle Formation are present in part of the study area. The Rock Point Member, where present, rests unconformably on the Owl Rock Member (the Tr-5 unconformity) and consists mainly of reddish-brown to red-orange siltstone and laminated to cross-bedded fine-grained sandstones deposited by sheetflood, eolian, and to a lesser extent, fluvial processes. It is generally non-fossiliferous, with the notable exception of the Whitaker (Ghost Ranch) *Coelophysis* quarry in northern New Mexico and numerous tracksites in Utah and Colorado (Lockley and Hunt, 1995; Lockley et al., 2004). Age-diagnostic palynomorphs and palaeomagnetic data suggest that it is mostly of Norian age (Litwin, 1986; Litwin et al., 1991; Molina-Garza et al., 2003; Lucas and Tanner, this volume). Current convention places an unconformity at the base of the Wingate Sandstone, truncating the Rock Point. This presumably disconformable surface has been correlated with the J-0 unconformity at the base of the Moenave, resulting in the assumption that the overlying Wingate Sandstone is entirely earliest Jurassic (Pipiringos

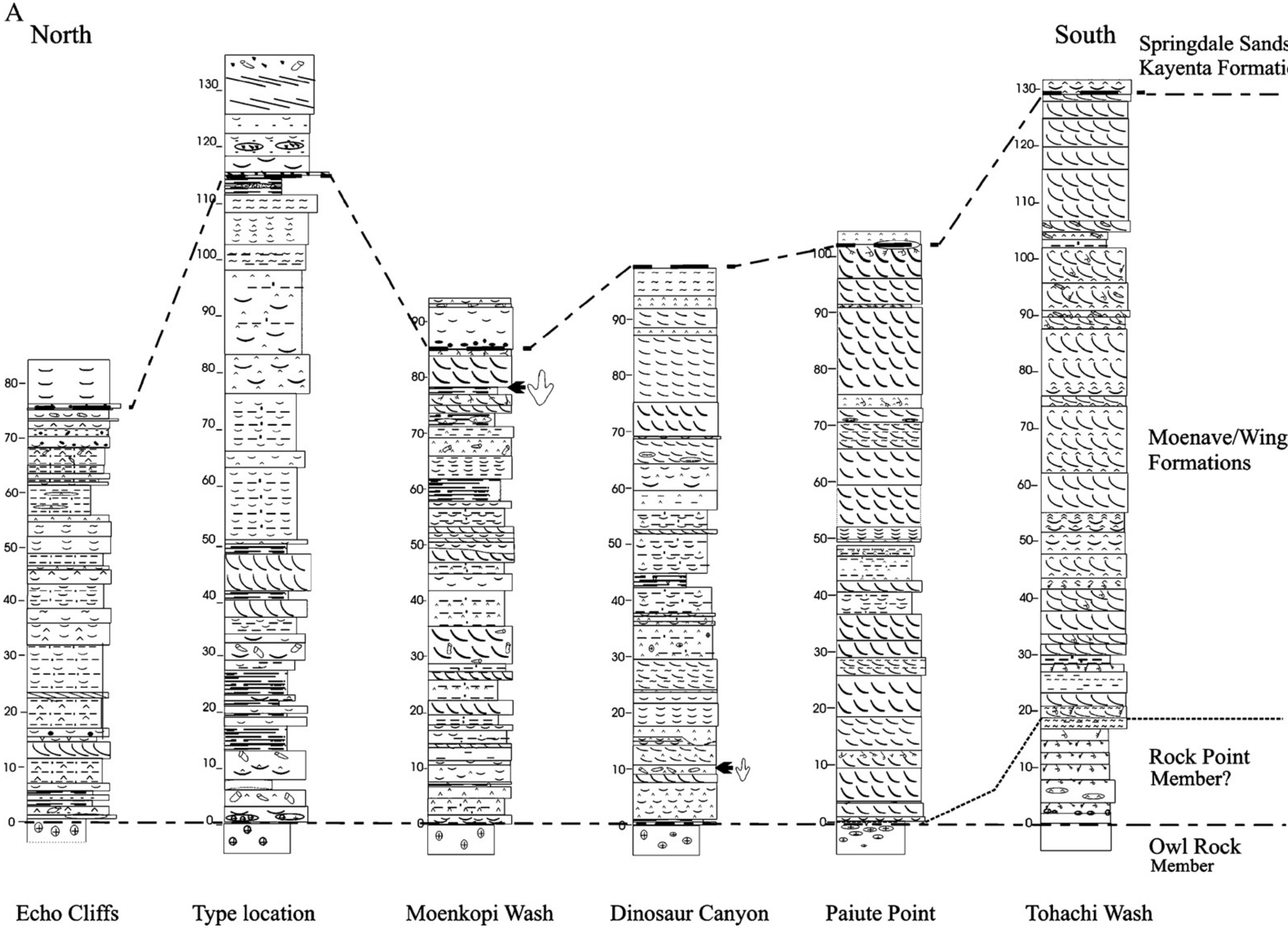

Fig. 3. Cross-sections illustrating lateral relationships of lithofacies for strata between the Owl Rock and Springdale contacts. (A) North–south section from the Echo Cliffs to the south end of Ward Terrace. Key to lithologic symbols in (B). (B) East–west cross-section across northern part of study area from Warner Valley (near St. George) to Old Paria. Base of the section at Whitmore Point may be in Petrified Forest strata, rather than Owl Rock. No horizontal scale is implied (see Fig. 1 for section locations).

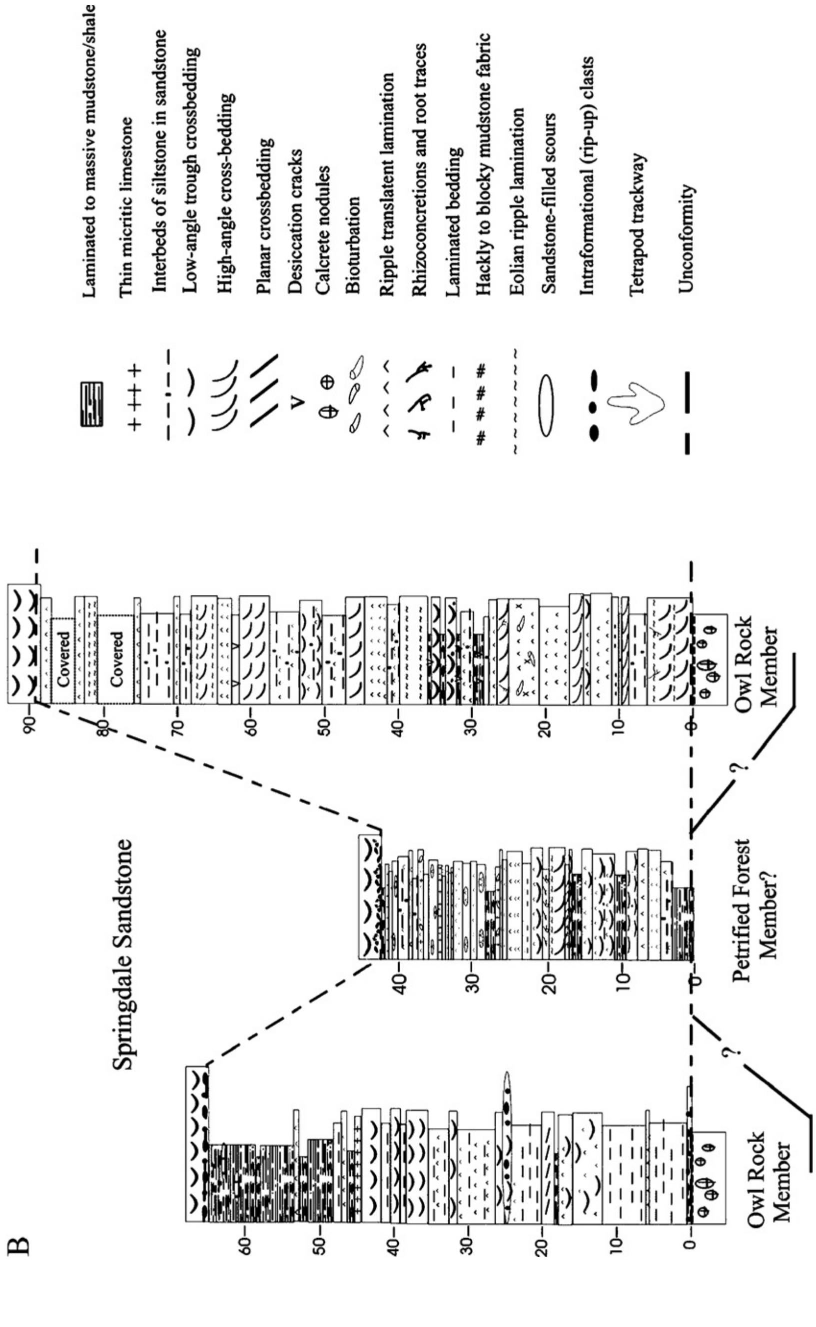

Fig. 3 (*continued*).

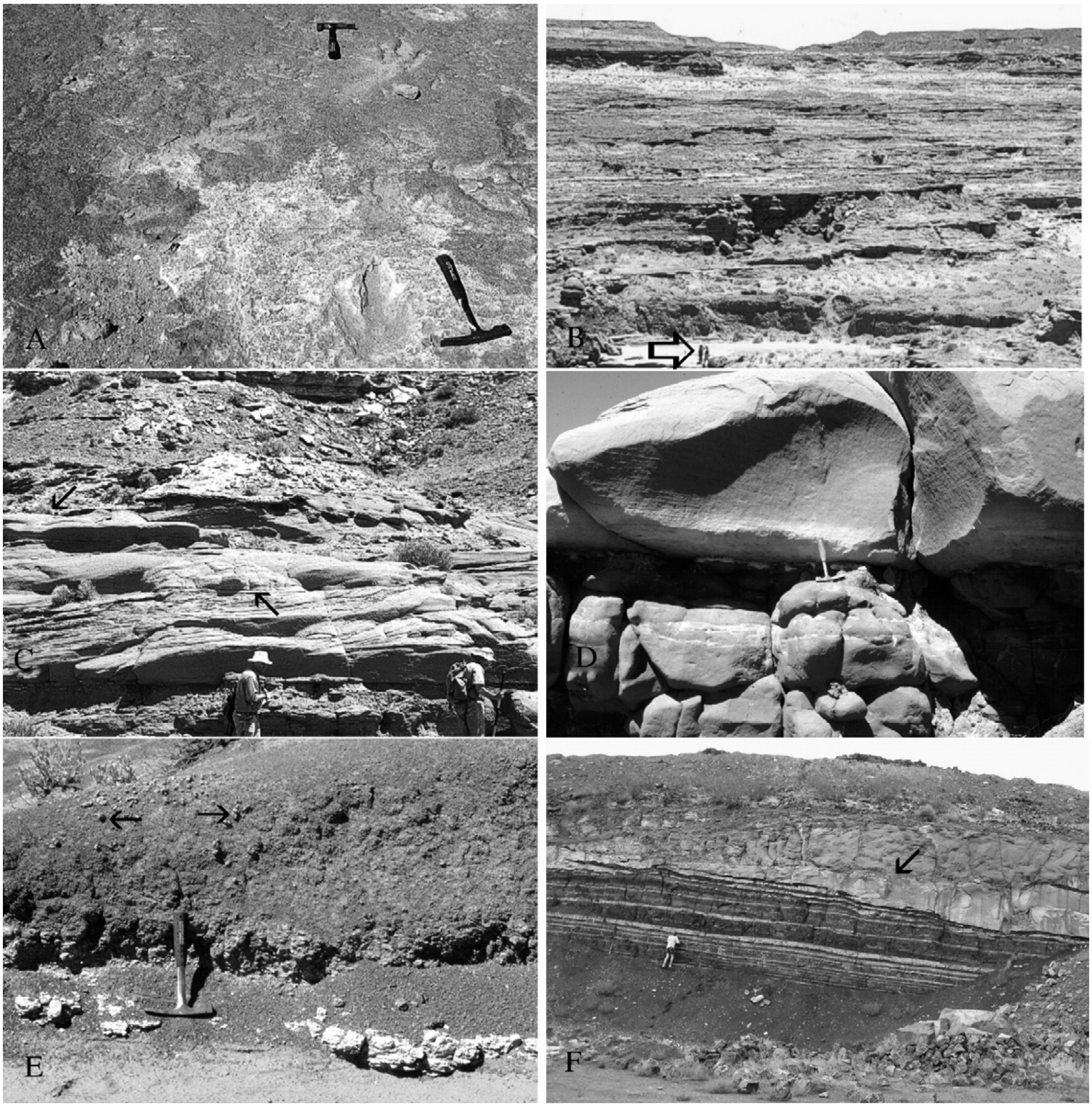

Fig. 4. Lithofacies and other sedimentologic features of the Moenave strata. (A) Theropod tracks (*Eubrontes*) occur in ripple (translatent) laminated sandstone in upper portion of section at Moenkopi Wash. (B) Overview of the lower portion of the section at Moenkopi Wash. Two figures at base are at approximate contact with Owl Rock Formation. The section in this view consists mainly of sheet-like and tabular bodies of sandstone facies LF-1 and LF-2 interbedded with LF-6. (C) Steepening-upward cross-bed sets (LF-3) in stacked sets (at Moenkopi Wash). Individual set thicknesses are less than 1 m, but a 1.5-m co-set is visible, delineated by arrows. (D) (At Tohachi Wash) Upper unit (above hammer) is a thick bed of sandstone with crudely parallel lamination, alternating fine and coarse grain (LF-4). The lower unit consists of laminated to massive siltstone with drab mottled horizon (LF-7). (E) Mudstone with blocky fabric and centimeter-scale micritic nodules (arrows), near the base of section at Tohachi Wash. (F) Section in the uppermost Moenave Formation at St. George, Utah (correlative with uppermost section at Warner Valley). Laminated grayish-purple mudstones alternate with gray and red rippled to laminated siltstones and pink limestones. Arrows mark contact with Springdale Sandstone.

and O'Sullivan, 1978). However, tetrapod fossils (the phytosaur *Redondasaurus* and the crurotarsan track *Brachychirotherium*: Lucas et al., 1997; Lockley et al., 2004) and palaeomagnetic data (Molina-Garza et al., 2003) indicate a Late Triassic age for at least the lower part of the Wingate.

3. Lithofacies assemblages

Published examinations of the sedimentology of the Moenave Formation largely have been limited to study of the Dinosaur Canyon Member in southeastern Utah (i.e., between Paria and Kanab; Clemmensen and Blakey, 1989; Clemmensen et al., 1989; Olsen, 1989). Unpublished work (Edwards, 1985) also has extended the available data on Moenave and Wingate sedimentology (see review in Blakey, 1994). We extend these earlier works by examining sections at numerous localities along the Moenave outcrop belt in southeastern Utah and northeastern Arizona. We measured complete stratigraphic sections of the entire thickness of the Dinosaur Canyon and Whitmore Point members, or equivalent Wingate strata, from the basal unconformity on Chinle Formation strata to the overlying Springdale Sandstone (or the contact with other Kayenta Formation strata where the Springdale is not present). We describe these strata as comprising the following lithofacies, in proportions that vary greatly across the outcrop belt.

3.1. Lithofacies descriptions

LF-1) *Trough-cross-bedded sandstone:* Beds up to 3 m thick of fine to coarse sandstone displaying small- to medium-scale (0.1–0.5 m) trough cross-bedding, commonly with mudstone or calcrete clast lag deposits. Beds consist of single to multi-storey sand bodies, and have an erosive base. Horizontally laminated sandstone is present locally.

LF-2) *Fining-upward sandstone:* Tabular to sheetlike beds of medium to fine-grained sandstone form laterally extensive sheets up to 4 m thick (Fig. 4B). Sand bodies are single-storey and display horizontal lamination, low-angle cross-bedding, and/or small-scale (0.1 to 0.3 m) trough-cross-bedding that commonly grades upward to ripple translatent strata or climbing ripples. Indistinct to distinct (e.g., *Skolithos*) traces are common and some bedding plane exposures contain tetrapod trackways. Beds have nonerosive to erosive bases and commonly form repetitive sequences up to 12 m thick.

LF-3) *Sandstone with low- to high-angle cross-beds:* Beds of fine- to medium-grained sandstone display trough and planar cross-bed sets 0.2 to 2.2 m thick (Fig. 4C). Cross-bed foresets dips are variable, but typically over 20° and locally up to 35°. Sets are arranged in tabular to wedge-shaped compound sets up to 7 m thick. Foresets dip consistently east to south-southeast. Bedding is locally obliterated by indistinct burrows and rhizoliths, particularly at the tops of beds marking major set breaks.

LF-4) *Laminated sandstone sheets:* Sheetlike beds of fine to coarse sandstone are up to 5 m thick and commonly display crudely parallel horizontal lamination, typically with small-scale alternation of coarse and fine laminae (Fig. 4D).

LF-5) *Sandstone with alternating small-scale cross-beds and horizontal lamination:* Sandstone units occur in bodies up to 15 m thick and consist of tabular beds up to 0.7 m thick that display small-scale trough cross-beds. These are interbedded with medium- to coarse-grained sandstone in beds up to 3 m thick that display crude horizontal lamination with alternating coarse and fine laminae. The lower contact is nonerosive.

LF-6) *Siltstone with interbedded sandstone:* Meter-scale beds of siltstone with fine horizontal to ripple lamination are interlayered with thin sheet to lenticular sandstones that display small-scale (0.1–0.2 m) trough cross-beds and and/or ripple-translatent strata. Individual bed thicknesses range up to 0.5 m, and unit thicknesses range up to 13 m. Locally, this facies is extensively burrowed and hosts tetrapod trackways.

LF-7) *Siltstone/mudstone with discontinuous sandstone laminae:* Beds of finely laminated to massive siltstone to mudstone, typically reddish brown, contain discontinuous thin (0.5 cm) layers of fine- to coarse-grained sandstone that locally displays parallel lamination or low-amplitude ripple forms. Bed thickness is up to 4 m. Beds commonly display laterally extensive horizons of drab mottling (Fig. 4D).

LF-8) *Laminated brown mudstone:* Homogeneous laminated mudstone commonly contains sandstone-filled desiccation cracks. This facies is truncated laterally by facies LF-1 or LF-2.

LF-9) *Massive mudstone with calcrete:* Beds of mudstone weathered to a blocky–hackly fabric contain scattered centimeter-scale micritic nodules and drab root traces up to 15 cm long (Fig. 4E).

LF-10) *Variegated siltstone and shale with interbedded limestone and sandstone:* Red, purple, and gray mudstone and siltstone occurs in beds 0.1–0.3 m thick. The beds are finely laminated and commonly contain desiccation cracks (Fig. 4F). These are cyclically interbedded with thin (0.1 m) mottled micritic limestones and ripple laminated sandstones, 0.1–0.3 m thick.

3.2. Lithofacies interpretation

The lithofacies described above represent deposition of sediments by a spectrum of process involving traction flow, both channelized and unconfined, suspension settling, and eolian transport.

3.2.1. Fluvial processes (LF-1, LF-2, LF-6)

In the type location of the Moenave Formation, as well as along the northern half of Ward Terrace and in the Echo Cliffs, the contact between the Dinosaur Canyon Member and the Owl Rock Formation is overlain by (LF-1) trough cross-bedded sandstone that typically contains lags of mudstone rip-ups and calcrete clasts clearly derived from the underlying strata. This facies records deposition in bedload-dominated streams locally incised into the underlying unconformity surface. The preponderance of small- to medium-scale trough cross-bedding indicates sediment was transported primarily by migrating dunes in shallow channels, although the contribution of horizontally laminated sandstone demonstrates at least some depositional events in the upper flow regime; the scarceness of planar cross-bedding and absence of lateral accretion surfaces suggests a general lack of mid-channel or side-attached bar development in the channels that formed during initial deposition of the formation. Similarly, Tooth (2000a) has noted that the channels of many dryland rivers consist of a single thread with a shallow planar bed. LF-1, where it occurs at the base of the formation, typically is overlain by an abrupt transition to finer grained facies, primarily LF-2 or LF-6. Stratigraphically higher in the formation, planar cross-bedding, low-angle cross-bedding, and horizontally laminated sandstone occur in greater proportions in LF-1, although still subordinate to trough cross-bedding. Outcrops of LF-1 at the formation base may exhibit either single- or multi-storey geometry, but higher in the section, this facies almost always consists of single-storey sand bodies.

The fining-upward sandstone bodies of LF-2 are more common throughout much of the Dinosaur Canyon section along Ward Terrace. Abrupt contacts may superimpose LF-2 and essentially all other lithofacies. This facies most likely records deposition in broad, ephemeral sandy streams in which sand sheet deposits, which comprise low-relief bedforms of horizontally laminated and/or low-angle cross-bedded sandstone (Miall, 1996; Reid and Frostick, 1997; Tooth, 2000a), are laterally equivalent to weakly defined channels containing small-scale dunes. The type of deposition suggested by the vertical succession to ripples and climbing ripples is one in which sheetflood processes dominate, with rapid deposition occurring during waning flow, as is typical of many sandy ephemeral rivers on distal braid plains (Miall, 1996; Reid and Frostick, 1997; Tooth, 2000a). This style of sheetflood over low-relief barforms has been described in modern desert streams in the Sinai (Sneh, 1983; Miall, 1996). Recurring flood cycles during basin aggradation produced thicker sequences of repetitive fining-upward sequences.

Ephemeral sheetflood is also interpreted for LF-6, which constitutes a significant proportion of the Dinosaur Canyon Member at the type location and other localities on the northern end of Ward Terrace. Beds of siltstone, massive to horizontally laminated, were deposited by sheetflow over non-channelized, or interbar areas during flooding events; the lack of appreciable finer grained (i.e., clay-size) sediment in the overbank environment is typical of sandy ephemeral streams, which top their banks easily during floods. The modern desert stream systems described by Sneh (1983) remain an appropriate analog; in the downstream direction, these modern systems evolve from bank-confined channels to open floodplains hundreds of metres to kilometres wide, covered with barforms having amplitudes of just a few tens of centimetres, with broad muddy areas between the bars. Tooth (2000b) similarly described the downstream decrease in the size and number of interchannel bars as streams approach low-gradient surfaces, or floodouts, where bedload transport terminates. Tunbridge (1984) invoked such a model of deposition to describe sheet sandstones and siltstones in a Devonian ephemeral stream system. On the Moenave floodplain, episodic flood events of significantly greater strength caused deposition of coarser bedload material in laterally extensive sheets of trough cross-bedded and ripple laminated sandstone. Relatively constant floodplain aggradation is suggested by the general (but not total) lack of palaeosol development.

Olsen (1989) analyzed in considerable detail the fluvial facies in the Dinosaur Canyon Member in the northern part of the outcrop belt (Paria to Kanab). From his study of the geometry and architectural elements of the sandstone bodies, Olsen concluded that the sandstone sheets were deposited mainly by sheetflow. He also described most single-storey channel sandstones as narrow bodies displaying lateral accretion surfaces (a feature we did not observe to the south on Ward Terrace) and therefore deposited by sinuous ephemeral streams. The mean palaeoflow direction for the ephemeral stream facies studied was north-northwest. Multi-storey channel sandstones are sheetlike bodies with multiple internal scour surfaces, deposited under upper-flow regime conditions.

3.2.2. Lacustrine facies (LF-8, LF-9, LF-10)

Fine-grained (mud or clay-size) sedimentary rocks are generally insignificant in the Dinosaur Canyon Member exposures to the south on Ward Terrace. Reddish-brown laminated mudstone beds (LF-8), tens of centimetres thick, occur with increasing frequency towards the northern end of the outcrop belt (in the Echo and Vermillion cliffs). These beds commonly contain desiccation features and may be laterally discontinuous due to lateral truncation by superimposed fluvial facies (LF-1 or LF-2). The laminated mudstones represent conditions of deposition from suspension in bodies of standing water, but evaporation and desiccation of these water bodies is evident. The typically limited lateral extent and relationship with fluvial facies indicate that these water bodies were not laterally extensive and probably represent deposition in ponded water filling topographic lows on the floodplain, possibly incised by migrating channels (Miall, 1996). Similar facies were described in the model of Devonian ephemeral stream and lakes sedimentation of Tunbridge (1984).

Massive mudstones with pedogenic fabrics (LF-9; Fig. 4E) are uncommon in the Dinosaur Canyon Member, but are present in strata possibly equivalent to the Rock Point Formation (see below). At Tohachi Wash, for example, dark brown mudstone displays a blocky fabric and contains micritic nodules up to 3 cm in diameter and drab root traces up to 10 cm long. We interpret the mudstone as an ephemeral lake modified by pedogenesis, producing an immature (Stage II) calcrete (Machette, 1985).

More fully developed lacustrine facies in the Moenave Formation are limited to the Whitmore Point Member, as described by Wilson (1967). Sections at widely separated locations at Warner Valley (near St. George, Utah) and at Whitmore Point display similar facies in the upper part of the formation (below the Springdale Sandstone). Variegated beds comprise red, blue-gray, purple, and green siltstone and shale, parallel laminated to ripple laminated, and interbedded with thin (5–20 cm) beds of micritic to sandy limestone that is pink to mottled-green (LF-10; Fig. 4F). Interbedded sandstones and siltstones display translatent ripple and climbing ripple stratification; desiccation cracks, fish scales, thin layers of mud-chip conglomerate, and *Skolithos* burrows occur locally, and the limestones display stromatolitic laminations in isolated instances.

Deposition of these fine-grained sediments took place in lakes that generally were perennial and of greater extent than described above for LF-8. However, the depth was insufficient to cause thermal stratification of the water column, as indicated by the lack of organic-rich sediments. Episodic sheetfloods on the floodplain brought incursions of bedload sediment into the lakes. We view the deposition of lacustrine carbonate in an otherwise siliciclastic-dominated depositional system as analogous to that in the Shuttle Meadow Formation, a similar-aged formation of the Newark Supergroup in which deposition in an ephemeral system was dominantly by sheetflood (Gierlowski-Kordesch, 1998). In this case, the source of the carbonate is interpreted as primarily from surface run-off, with precipitation occurring when the availability of clay-size sediment is limited. Thus, the Whitmore Point strata represent deposition on the terminal floodplain of the Moenave alluvial system (cf. Sneh, 1983), where a perennially high water table formed a base level above the topographic surface. Episodes of lowered lake level may have been triggered by climatic fluctuations that allowed desiccation; however, the lack of evaporite minerals, as either crystals or molds, indicates that the climate was not strongly arid.

3.2.3. Eolian dune and interdunal facies (LF-3, LF-4, LF-5, LF-7, LF-9)

Strata of the Dinosaur Canyon Member display abundant features indicating deposition by eolian processes at the margin of the Wingate erg, as interpreted by Clemmensen et al. (1989); the proportion of these facies in the Dinosaur Canyon Member increases from north to south along Ward Terrace. The high-angle cross-bedding present in LF-3 records deposition by migrating dunes, but the varied scale and type of cross-bedding suggests multiple dune types. Langford and Chan (1993), for example, suggested that trough cross-beds record migration of crescentic or barchanoid dune forms, while planar cross-beds result from two-dimensional dune forms that migrate in a direction transverse to the mean wind direction. Clemmensen and Blakey (1989), in studying the Wingate eolian system, found dune cosets up to 30 m thick, considerably greater than the cosets at the erg margin, and interpreted these thick trough-cross-bedded sandstones as the deposits mainly of oblique dunes.

Exposures of LF-3 in the Echo Cliffs and at the type location are limited to isolated cosets up to 7 m thick. Along Ward Terrace, multiple compound sets are superimposed in places. At Moenkopi Wash, for example, the top of one compound set is marked by pronounced pedogenesis, including bleaching and formation of an extensive rhizolith horizon (Fig. 5A and B). These features suggest that this horizon marks the end of an accumulation event and is a potential supersurface (Havholm et al., 1993). Correlation of this surface to the erg interior is necessary to confirm this hypothesis, however. Nation (1998) interpreted seven major erg sequences, separated by bounding surfaces, in the Wingate sequence. More intense pedogenesis is noticeable immediately below the Springdale erosional surface, as at Paiute Point; here, the uppermost 2 m of eolian cross-bedded sandstone displays

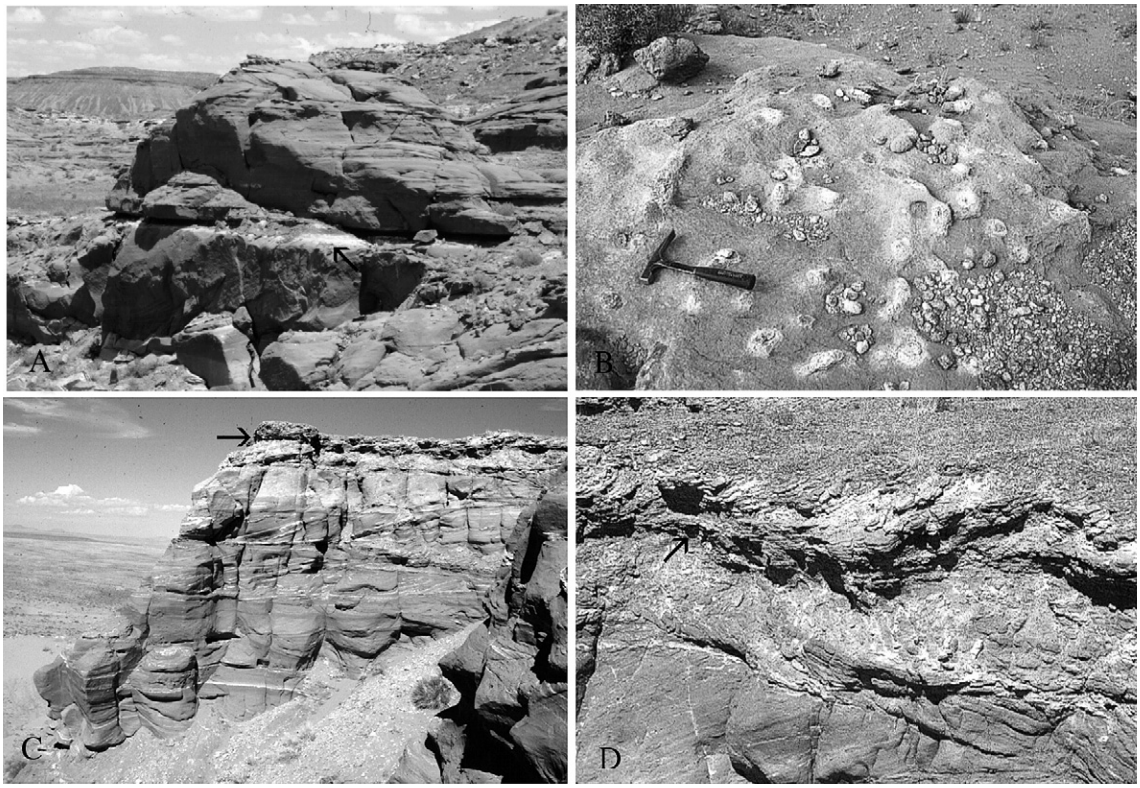

Fig. 5. Features of eolian deposition. (A) Bleached horizon (arrow) at top of eolian cross-bed co-set overlain by eolian sheet sandstone constitutes a supersurface (at Moenkopi Wash). (B) Same horizon as indicated in (A) is packed locally with rhizoliths containing a calcite core and bleached halo. (C) Coarse sandstone and intrabasinal conglomerate of Springdale Sandstone overlies (arrow at contact) eolian strata at the top of section at Paiute Point. Uppermost eolian bed sets are mottled gray to purple. (D) Detail of (C) illustrating pedogenic obliteration of eolian fabric.

a pronounced downward colour gradation from gray to purple in an otherwise reddish orange formation (Fig. 5C). The uppermost 0.7 m is also intensely disrupted by drab-coloured rhizoliths (Fig. 5D). Significantly, the proportion of dune-bedded sandstone increases southward along Ward Terrace, as does the thickness of the cosets. These observations support the conclusion of Clemmensen et al. (1989) of a greater depositional influence of the Wingate erg.

Non-dune eolian facies also are well-developed in the Dinosaur Canyon strata (LF-4, LF-5). Sheet sandstones consisting entirely (LF-4) or partially (LF-5) of laminated sand displaying subparallel horizontal laminae with alternating fine and coarse grain size were deposited by migrating eolian ripples. In LF-5, eolian–ripple laminated sandstone is interbedded with sets of small-scale dune cross-bedded sandstone (LF-5). Both LF-4 and LF-5 locally display burrows and root traces. Clemmensen et al. (1989) noted the presence of these facies, collectively referred to as eolian sand sheets, in their study area in Utah. Sand sheets, as defined by Lancaster (1993), are an interdunal facies consisting dominantly of parallel wind-ripple laminae and are formed in interdune areas during conditions of limited sand supply with sediment by-passing. The presence of abundant, small-scale cross-bedding in this facies, as noted in this study and the previous work of Clemmensen et al. (1989) confirms the importance of small dunes in the interdune sandflats; potentially, many of these small dunes formed in the lee of vegetation.

Wet interdune facies also occur in the Dinosaur Canyon strata. Laminated to massive siltstone and mudstone with interlayered coarse eolian ripple laminae (LF-7) may record deposition of fine-grained sediments by sheetflood on a mudflat, as interpreted for LF-6, with modification by migrating eolian sand ripples. Low-amplitude ripple forms may be isolated lenticles of sand trapped by adhesion on the wet surface. Although no evaporites were observed in the sections we studied, these strata resemble the facies described as “sabkha” by Clemmensen et al. (1989). Current usage of the term sabkha implies an association with a marine supratidal setting for which there is no evidence in the Moenave Formation. Therefore, we prefer to describe this facies as playas, or mudflats dominated by sheetflood and eolian deposition (Hubert and Hyde, 1982).

4. Palaeoenvironmental reconstruction

4.1. Palaeogeography

4.1.1. Lateral facies relationships

As described by Clemmensen et al. (1989), the Wingate erg covered 110,000 km^2 or more (see their Fig. 1). Across much of this area, the Wingate Sandstone is characterized by large-scale cross-stratified sandstone deposited by migrating dunes and draas. In their reconstruction, the Wingate and Moenave formations intertongue along the southwestern margin of this erg where deposits of the north-northwest-flowing Moenave streams (Olsen, 1989) were reworked by winds blowing to the southeast (Peterson, 1988) to form migrating oblique dunes. Clemmensen et al. (1989) recognized the following facies assemblages in the Dinosaur Canyon Member in the Paria-Kanab area: large-scale cross-bedded sandstones with high-angle foresets (to 35°), representing eolian dunes; sandstone sheets containing small- to medium-scale, high-angle cross-beds, low-angle cross-beds, and wind-rippled sandstones, collectively interpreted as eolian sand sheet deposits; interbedded cross-bedded sandstones and mudstones, deposited in dunes and adjacent muddy ephemeral streams; evaporitic or mottled sandstones with muddy interbeds or mud drapes, deposited in sabkhas; multi-storey sandstones, representing sandy ephemeral stream systems; and thinly interbedded mudstones and sandstones, deposited as muddy ephemeral stream sheet and single-storey ribbon sandstones. These facies are consistent with the model for erg-margin deposits developed by Langford and Chan (1993); erg margin deposits may be recognized by wide areas of interdune deposition, extensive bioturbation, abundant sheet sand deposits, common non-eolian facies, smaller and more varied dune types than in the erg center, and common erosional surfaces. Our recognition of these same or very similar facies in sections in the Echo Cliffs and Ward Terrace extends the interpretation of Moenave erg margin deposition south of the study area of Clemmensen et al. (1989).

Importantly, our sections illustrate significant encroachment of the erg in the southern end of Ward Terrace (Fig. 3A), consistent with earlier work (Edwards, 1985; Blakey, 1994). In the Echo Cliffs section, distinctly eolian facies (LF-3) constitute only 5% of the section thickness (from the Owl Rock Member contact to Springdale Sandstone contact). At the Moenave type location, all eolian facies constitute about 15% of the section thickness, and this proportion increases steadily along the outcrop belt. At Dinosaur Canyon, for example, eolian facies comprise roughly 50% of the section, and at Tohachi Wash, near the southern end of Ward Terrace, non-eolian facies constitute less than 5% of the section thickness. Consistent with this trend, we note also that individual dune cross-bed sets in the Echo Cliffs and type locality sections are typically less than 1 m thick, and cosets are generally no more than 3 m thick. South of Moenkopi Wash, individual set cross-bed set and coset thicknesses increase markedly. As modeled by Clemmensen et al. (1989), this transition records erg encroachment as the environment shifts from a fluvially dominated erg margin in the north, through an intermediate erg margin–erg setting, dominated by interdune environments, to a predominantly dune inner erg setting in the southern part of the study area.

4.1.2. Erg sediment supply

An interesting point arises in considering the source of sediment for the growth of the Wingate erg. Clemmensen et al. (1989) presumed that the north-northwest-flowing Moenave stream system (Olsen, 1989) supplied the sediment that was reworked inland and supplied the erg. Although the true size of the Moenave terminal flood basin cannot be known due to the erosional limits of the formation, reconstructions show a probable extent substantially smaller than that of the erg supposedly derived from it (Riggs and Blakey, 1993; Blakey, 1994). However, the Moenave streams ultimately drained into a back-arc sea whose eastern shoreline would have extended across eastern Nevada (Marzolf, 1994; Riggs et al., 1996), and deposits of this shore likely provided some of the upwind sediment load (Peterson, 1988). Zircon isotope geochronology indicates clearly that some of the eolian sediment supply was derived from the Laurentian shield sources to the north via southerly longshore drift along the coast of the back-arc sea (Dickinson and Gehrels, 2002).

Allen et al. (2000) and Blakey (2000) proposed that the preservation of sand in the great Jurassic ergs (including the Wingate erg) of the Western United States resulted from tectonically created accommodation space. The former authors described crustal flexure associated with shortening events in the continental margin arc, but these authors also noted that a positive sand budget was required for this net accumulation. As Chan and Kocurek (1988) and Swezey (2001) have noted, the supply of sediment to inland ergs may have a eustatic component of control; episodes of regression expose greater areas of sediment to eolian reworking. We suggest, therefore, that the sand supply to the

Wingate erg was controlled by sea level. We note, for example, that the initial growth of the Wingate erg, which we now understand took place during the Rhaetian stage, not the Hettangian as previously believed, coincided with an episode of falling sea level (Haq et al., 1987). Thus, we suggest that erg initiation may have been linked to the Rhaetian regression and the consequent exposure of an enlarged sediment source area on the coastal plain of the back-arc basin. This hypothesis implies that subsequent transgression would have limited the sediment supply and caused shrinkage and eventual abandonment of the erg. Potentially, this could explain the transition to the fluvial-dominated deposition of the overlying Kayenta Formation during the early Sinemurian, which was a time of rising sea level (Haq et al., 1987). In sum, we propose a combination of eustatic and tectonic processes as controls for erg formation and preservation.

4.2. Palaeoclimate

At the end of the Triassic period, the Four Corners area was located at a low palaeolatitude, probably no farther north than 15° (Scotese, 1994; Molina-Garza et al., 1995; Kent and Olsen, 1997). Latest Triassic through earliest Jurassic time has been interpreted as an interval during which climate was strongly seasonal, a consequence of a very pronounced monsoonal effect (Kutzbach and Gallimore, 1989; Parrish, 1993). Climate-sensitive facies (evaporites and carbonates), combined with the restriction of coal formation to high latitudes, provide abundant evidence of overall warm and dry conditions across much of Pangaea during the Late Triassic (Frakes et al., 1992; Lucas, 1999). The succession of lithologies and palaeosols in the Chinle Formation and Glen Canyon Group, in particular, has been suggested as indicating a gradual drying trend during the Late Triassic on the Colorado Plateau (Blakey and Gubitosa, 1983; Lucas, 1999; Tanner, 2000a, 2003).

Seasonally arid conditions, denoted by the pedogenic fabrics of the Owl Rock Member, appear to have continued or possibly intensified during subsequent deposition of Rock Point strata, as suggested by the abundance of eolian facies. Notably, however, Rock Point eolian facies are wet, consisting primarily of interdune sand sheets and mudflats (Lucas et al., 1997; Tanner, 2000b), similar to the Moenave erg margin environment. Moreover, burrowing organisms and vegetation were abundant during deposition of these strata, and evaporites and mature arid palaeosols (aridosols) are generally lacking. Widespread development of these facies in eolian systems is a clear indication of a high water table (Kocurek, 1996). Therefore, we conclude that there is no evidence of increased aridification during growth of the Wingate erg. As we suggested above, the late Rhaetian–Hettangian regression may have provided abundant sediment for eolian reworking to inland regions, resulting in erg growth, and tectonic events created accommodation space for its preservation.

Despite the lack of evidence of an overall trend of aridification during Rock Point-Wingate deposition, shorter term climatic variation is suggested by intermittent expansions of the erg into the marginal areas, as noted by Clemmensen et al. (1989). We note, for example, that adjacent sections measured along Ward Terrace commonly display eolian facies interbedded with the non-eolian facies in approximately correlative positions. Furthermore, the formation of supersurfaces in the erg interior (Clemmensen and Blakey, 1989; Nation, 1998) suggests episodes during which the water table rose significantly, a potential signal of climate change (Lancaster, 1993; Kocurek, 1996).

5. Formation boundaries

5.1. Moenave/Wingate

The gradational nature of the transition from erg margin to erg interior demonstrated along Ward Terrace raises the question of how best to draw boundaries between the Moenave and Wingate formations. Clearly, previous workers (e.g., Harshbarger et al., 1957) have associated the Moenave Formation with mainly non-eolian (fluvial and lacustrine) facies. The measured section from the type location therefore best exemplifies the lithologies of the Moenave Formation, while the strata at the southern end of the Ward Terrace outcrop belt, in which eolian facies dominate at the expense of non-eolian facies, have a clear Wingate affinity. This suggests that the boundary between the two formations might reasonably be placed south of Dinosaur Canyon. More precise definition of the formation boundary might require consideration of the relative proportions of dune and interdune lithofacies and comparisons of dune and sediment grain size between the erg margin and the erg interior.

5.2. Basal unconformity: J-0?

At most locations in the Moenave outcrop belt, fluvial facies of the Moenave rest unconformably on the Owl Rock Member of the Chinle Formation, which is typically recognizable as grayish pink mudstone with abundant calcrete nodules. At the southern end of Ward

Terrace, however, Owl Rock strata are overlain unconformably by sheetflood and eolian sand sheet facies that resemble the Rock Point Member at its type location (Lucas et al., 1997). We suggest that these strata may be correlative with the Rock Point Formation. Importantly, we find no evidence of an unconformity between these "Rock Point" strata and the overlying Moenave/Wingate strata. In our area of study, therefore, our findings are consistent with the original conclusion of Harshbarger et al. (1957). Tentatively, we accept the conclusion of Marzolf (1994) that all strata bounded by the unconformities between the Owl Rock Member (or the older strata where the Owl Rock has been truncated) and Springdale Sandstone constitute a single tectonosequence bounded below by the Tr-5 unconformity and above by the sub-Springdale/Kayenta unconformity.

6. Conclusions

Deposition of Moenave Formation sediments took place during the latest Triassic to earliest Jurassic in a mosaic of fluvial, lacustrine, and eolian subenvironments. Channelized flow in ephemeral streams deposited single- and multi-storeyed trough cross-bedded sandstones; laminated siltstones were deposited by unconfined flow (sheetwash) in interchannel areas. Sheetwash also deposited mudstone and siltstone on mudflats and in ephemeral lakes (playas). Deposition of shales, mudstones, and micritic limestones took place in perennial lakes that were subject to episodic desiccation. Eolian processes deposited sandstones with high-angle cross-beds in large and small dunes that migrated east to south-southeast, and as sheets of laminated sandstone formed by the migration of low-amplitude eolian ripples.

The distribution of these facies varies greatly across the outcrop belt. The northern part of the study area (in Utah) was a terminal floodplain dotted by broad, shallow lakes that was located marginal to the Wingate erg. In a southward direction along the outcrop belt (southward along the Echo Cliffs and Ward Terrace), dominantly fluvial-lacustrine lithofacies are replaced by mainly eolian dune and interdune deposits, recording encroachment of the Wingate erg. Thus, Moenave outcrops expose a north–south lithofacies gradient from distal (erg margin) to proximal (erg interior).

The prevalence of ephemeral stream and eolian environments during deposition of these strata indicates a seasonally arid climate during the latest Triassic to earliest Jurassic. We see no sedimentologic evidence for significant climate change at the Triassic–Jurassic boundary, or at any time encompassed by this sedimentary succession. Indeed, the growth and incursion of the Wingate erg into the Moenave fluvial system may have been driven by the availability of sediment in the upwind source area, the coastal plain and coastline to which the Moenave streams delivered sediment. We interpret a eustatic signal as responsible for formation of this erg as long-term regression during the Rhaetian and continued Hettangian lowstand exposed a broader area of shallow-marine sediments to eolian reworking.

The north–south lithofacies transition from fluvial–lacustrine erg margin to eolian dune–interdune indicates that the contact between the Moenave and Wingate formations is gradational, as indicated by earlier workers. The presence of Rock Point-like lithofacies at the base of stratigraphic sections at the south end of the Moenave outcrop belt suggests that the Rock Point Formation underlies the Moenave Formation in this part of the study area. The contact between these formations appears conformable.

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