Antecedent aeolian dune topographic control on carbonate and evaporite facies: Middle Jurassic Todilto Member, Wanakah Formation, Ghost Ranch, New Mexico, USA

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ABSTRACT
The Middle Jurassic Todilto Member of the Wanakah Formation is a carbonate and gypsum unit inset into the underlying aeolian Entrada Sandstone in the San Juan Basin. Field and thin section study of the uppermost Entrada and Todilto at Ghost Ranch, New Mexico, identified Todilto facies and their relationship to remnant Entrada dune topography. Results support the previous interpretation that the Entrada dunes, housed in a basin below sea level, were rapidly flooded by marine waters. Mass wasting of the dunes gave rise to sediment-gravity flows that largely buried remnant dune topography, leaving ca 12 m of relief that defined the antecedent condition for Todilto deposition. Previously interpreted as seasonal varves deposited in a stratified water body, the Todilto is reinterpreted as a microbial biolaminate. Most diagnostic are organic-rich laminae with structures characteristic of filamentous microbes and containing trapped aeolian silt, and clotted-texture laminae with a fabric associated with calcification of extracellular polymeric substances. The spatial arrangement of Todilto facies is controlled by the dune palaeotopography. A continuous basal laminated mudstone thickens over the dune crest, reflecting the optimum conditions for microbial mat development, and is interpreted to have been deposited when marine waters submerged the topography. Subsequent drying caused emergence of the crestral area, and formation of tepee structures and a dissolution breccia. Gypsiferous mudflats and periodic ponds occupied the dune flanks and interdune area, with gypsum concentrated within the interdune area. Entrada sands remained unstable during Todilto deposition with common injection structures into the Todilto, and a remnant slope caused the downslope movement and folding of Todilto strata on the upper lee face. Although some expansion of the gypsum occurred in the subsurface, facies architecture fostered development of a dissolution front adjacent to the interdune gypsum body with section collapse of gypsiferous limestone on the dune flanks.

Keywords Dunes, Entrada, gypsum, limestone, microbial mat, Todilto.

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INTRODUCTION

The Middle Jurassic (Callovian) Todilto Member of the Wanakah Formation is an enigmatic limestone and gypsum unit within a terrigenous and predominately aeolian province. The Todilto occurs inset into the underlying Entrada Sandstone within the San Juan Basin of the Western Interior of the United States. The Entrada Sandstone represents a major aeolian sand sea and inland sabkha system. Coupled, the stratigraphic horizon defined by the uppermost Entrada and the overlying Todilto exhibits sedimentary attributes that have generated controversial interpretations. In the subsurface of the San Juan Basin (Vincelette & Chittum, 1981) and in outcrop (Benan & Kocurek, 2000), the Entrada has been interpreted as showing preserved dune topographic relief, which is relatively rare in the stratigraphic record (e.g. Glennie & Buller, 1983; Eschner & Kocurek, 1988; Story, 1998; Jerram et al., 2000; Scherer, 2002). The Todilto infills remnant relief on this surface and drapes the Entrada.

The origin of the Todilto water body has been attributed to catastrophic marine flooding of the Entrada dune field, which was housed within a topographic basin below sea level (Benan & Kocurek, 2000). The flooding event resulted in partial mass wasting of the dunes, which generated sediment-gravity flows that buried much of the remnant dune topography. If initiated by a catastrophic flooding event, the Todilto water body persisted and continued marine connections are inferred based upon the thickness and areal extent of the deposits. Kirkland et al. (1995) report Todilto limestone and gypsum thicknesses up to 12 m and 30 m, respectively, and the projected zero-isopach for the Todilto (Kirkland et al., 1995, fig. 17) defines an area of ca 105 000 km² (larger than Lake Superior at 82 000 km², smaller than the Adriatic Sea at 136 000 km²). The Todilto water body has been interpreted as a saline lake (Anderson & Kirkland, 1960; Tanner, 1970), a restricted marine embayment (Harshbarger et al., 1957; Ridgley, 1989) and a salina (Lucas et al., 1985; Anderson & Lucas, 1994; Kirkland et al., 1995; Lucas & Anderson, 2000). The water body has been envisioned as highly stratified, and laminations in the limestone have been interpreted as seasonal varves (Anderson & Kirkland, 1960; Kirkland et al., 1995).

This work is a case study of the Entrada–Todilto stratigraphic horizon at Ghost Ranch in north/central New Mexico, USA. It expands the study area of preserved Entrada dune topography and blanketing sediment-gravity flow deposits addressed by Benan & Kocurek (2000), but emphasis is upon the Todilto. The Todilto is reinterpreted as representing a biolaminite deposited within an extreme environment largely devoid of marine fossils except for microbial mats. It is argued herein that the spatial distribution of facies within the Todilto is controlled by the antecedent Entrada dune topography, and that the facies architecture influenced fluid flow within the subsurface, resulting in collapse of much of the original depositional framework.

STRATIGRAPHIC CONTEXT

At the regional scale (Fig. 1), Middle Jurassic units were deposited as an eastward-tapering wedge within a retroarc foreland basin that extended from orogenic terrain in the west to the craton in the east marked by remnant elements of the Late Palaeozoic Ancestral Rockies (Bjerrum & Dorsey, 1995; DeCelles & Currie, 1996; Allen et al., 2000). The Entrada Sandstone occurs in the eastern portion of the basin, and consists of a complex assemblage of aeolian and inland sabkha strata (Kocurek & Dott, 1983; Blakey et al., 1988; Peterson, 1994). Eastern facies of the Entrada are aeolian-dominated, but these yield progressively westward to sabkha-dominated facies which are transitional into the Preuss Formation (Peterson, 1994, fig. 17; Jennings, 2014; Valenza, 2016). The Preuss has been interpreted as representing terminal fluvial systems that prograded eastward from the western uplands onto the inland sabkha (Hileman, 1973; Cook, 2016). In contrast to this western source, the aeolian quartz-rich sands of the Entrada were ultimately derived from transcontinental fluvial systems originating in the Appalachian Mountains (Dickinson & Gehrels, 2009, 2010), and more immediately from the sandy coast to the north (Sundance Formation) and transported southward by the wind (Peterson, 1988a). Later during the Callovian, the Preuss and Entrada systems were transgressed by marine systems from the north and north-west, as represented by the overlying Curtis Formation and equivalent units, and subsequently buried by basin-wide progradation of the continental Upper Jurassic Morrison Formation with sediment derived from the western uplands (Blakey et al., 1988, fig. 2; Peterson, 1994, figs 20/21).
The occurrence of the Todilto coincides with the San Juan Basin situated between the Late Palaeozoic Uncompahgre and Defiance uplifts (Fig. 1; Blakey et al., 1988, fig. 20; Blakey, 1988, fig. 15). Although differing in configuration from the later Laramide San Juan Basin, the Jurassic San Juan Basin was a site of active subsidence (Santos & Turner-Peterson, 1986; Blakey et al., 1988). The Todilto pinches out along the upper Entrada surface westward towards the Defiance Uplift, and is overlain by the Beclabito Member of the Wanakah Formation in the north, and by the aeolian Cow Springs Sandstone in the south where the Beclabito grades into the Cow Springs (Condon & Huffman, 1988, 1994; Condon & Peterson, 1986; Condon, 1989a,b). The poorly studied Beclabito consists of packages of sandstone and mudstone, and has been broadly interpreted as marginal marine in origin (Condon & Peterson, 1986; Condon, 1989a). Northward in Colorado, the Todilto-equivalent Pony Express Limestone again pinches out along the upper Entrada surface towards the Uncompahgre Uplift, and is overlain by the Bilk Creek Sandstone Member of the Wanakah Formation (O’Sullivan, 1986, 1992) which occupies the same stratigraphic position as the Beclabito Member (Condon & Huffman, 1988).

Although deposition of the Todilto marks the end-event of the Entrada sand sea within the San Juan Basin, the stratigraphic position of the Todilto within the overall Entrada system, especially with respect to the Curtis marine transgression to the north-west, is more difficult to define. The stratigraphic relationship of the Todilto to the Curtis is critical because Todilto marine waters have been inferred to have been derived via some connection to the Curtis sea (McKee et al., 1956; Harshbarger et al., 1957; Ridgley, 1989; Kirkland et al., 1995).

The most comprehensive stratigraphic work in the region is the large assemblage of measured sections and correlations by the US Geological Survey, which were used in constructing Fig. 2. The lower portion of marine strata of the Curtis Formation pinches out between the Slick Rock...
Member and Moab Tongue of the Entrada Sandstone. Although the Curtis transgressive surface (J-3 unconformity of Pipiringos & O’Sullivan, 1978) is traceable for some distance, definition of the Moab Tongue is lost within the overall Entrada sand body in south-eastern Utah. Southward, stratigraphic relationships (and nomenclature) become complex because the main body of the Entrada splays into three informal, predominantly aeolian tongues (lower sandy, salmon and bed at Butler Wash, the latter of which is assigned to the Wanakah Formation). These tongues interfinger with sabkha-dominated units of the Entrada (Rehoboth Member) and Wanakah Formation (lower and middle members). Along the western edge of the San Juan Basin, the Todilto is underlain by the upper sandstone of the Entrada, and overlain by the Beclabito Member of the Wanakah and the Cow Springs Sandstone. Overall the cross-section shows the thinning of the section onto the craton from the retroarc foreland basin to the west, with the section again thickening into the San Juan Basin in the south-east. Note the extreme vertical exaggeration; in nature these strata are essentially flat-lying.

Fig. 2. Stratigraphic relationships for Middle Jurassic strata between the J-2 and J-5 unconformities of Pipiringos & O’Sullivan (1978) from north-west of Moab, Utah, to Todilto Park, New Mexico, as traced in insert in the lower right. Datum is the J-5 surface. Based upon 93 measured sections (tick marks along the J-5 surface) and correlations by O’Sullivan (1980, 1981, 1996, 1997), O’Sullivan & Pierce (1983), Condon & Peterson (1986), Condon (1989b) and Robertson & O’Sullivan (2001), and using revised terminology of Condon (1989a) and Robertson & O’Sullivan (2001). Predominately aeolian units are highlighted in yellow. Greatest complexity occurs where the Entrada Sandstone and Wanakah Formation intertongue. Note the stratigraphic position of the Todilto Member of the Wanakah Formation, underlain by the upper sandstone of the Entrada, and overlain by the Beclabito Member of the Wanakah and the Cow Springs Sandstone. Overall the cross-section shows the thinning of the section onto the craton from the retroarc foreland basin to the west, with the section again thickening into the San Juan Basin in the south-east. Note the extreme vertical exaggeration; in nature these strata are essentially flat-lying.
Entrada nor the marine-transgressive Curtis Formation.

In a sequence-stratigraphic analysis of the Entrada in south-eastern Utah, Carr-Crabaugh & Kocurek (1998) defined progradational parasequences capped by polygonally fractured surfaces. The Todilto pinches out in the basal sabkha deposits of the lower Wanakah and rests upon the polygonally fractured surface bounding the salmon (John, 2000; Makechnie, 2010). This stratigraphic position is at the base of the progradational parasequence defined by the Slick Rock and lower Wanakah, and which is bounded above by the surface correlative to the J-3 surface at the base of the Curtis transgression (Carr-Crabaugh & Kocurek, 1998, fig. 3). By this analysis, the Todilto pre-dates the Curtis transgression. This interpretation is in agreement with that of Peterson (1994, fig. 3), who assigned the Todilto to an earlier (i.e. pre-Curtis) transgression, which potentially manifested as transgressive facies within the Preuss Formation and the Hulett Member of the Sundance Formation, and is correlative to the sabkha-dominated middle member of the Entrada in south-central Utah (Peterson, 1988b). Combined, these regional stratigraphic relationships suggest that the Todilto flooding event occurred over widespread sabkha environments and pre-dates the Curtis transgression over the aeolian-dominated Upper Entrada (south-central Utah) and Slick Rock Member (east-central Utah).

PREVIOUS INTERPRETATION OF THE TODILTO

The most comprehensive treatment of the Todilto is by Anderson & Kirkland (1960) and Kirkland et al. (1995). Anderson & Kirkland (1960) interpreted a repetitive cycle of seasonal varves within the limestone, identifying three types of laminae: (i) microcrystalline carbonate laminae precipitated during the summer with increased temperature and evaporation; (ii) organic-rich laminae derived from the settling of planktonic material during the autumn and early winter; and (iii) laminae of clastic silt, primarily quartz, washed or blown in during the winter and spring. A fourth lamina type, gypsum, was identified as the limestone transitions into gypsum, and these laminae represented maximum evaporation during the summer and occurred after the carbonate in the cycle. Based upon the number of annual varve cycles per unit thickness, the limestone was interpreted to have been deposited in ca 14 kyr, with 6 kyr estimated for subsequent deposition of the gypsum. Kirkland et al. (1995) revised the depositional time to emplace the Todilto to 30 to 100 kyr. The Todilto water body was estimated as initially <90 m in depth with subsequent shallowing to gypsumiferous pools. The Todilto water body was envisioned as salinity stratified with anoxic bottom waters that prohibited a benthic fauna. Bioturbation is absent within the Todilto, and the unit is unfossiliferous with the exception of ostracods, and rare fish and aquatic insect fossils (Lucas et al., 1985, 2000; Kirkland et al., 1995; Lucas & Anderson, 2000). Most of the fossils appear to be from the basal section, for which initial normal marine waters are interpreted (see below), or along the margins of the basin. Bulk sample isotopic analysis (C, O and Sr) led Kirkland et al. (1995) to conclude that Todilto waters did not match Callovian seawater, but rather showed an additional fluvial influx. However, Guhl (2004) found significant C, O and Sr isotopic variations within single facies and even along single laminae, and argued that the isotopic signature did not represent the original waters, but rather diagenetic alteration of original marine sediments.

STUDY AREA AND METHODS

This study was conducted at Ghost Ranch, New Mexico, USA, utilizing the same outcrops as in Benan & Kocurek (2000), but extended laterally (Fig. 3). The Entrada Sandstone overlies the red beds of the Triassic Chinle Formation (Fig. 3A), and rests upon the regional J-2 unconformity of Pipiringos & O’Sullivan (1978). The Todilto occurs as a bench capping the Entrada cliffs, and is overlain by recessive mudstones and sandstones of the Beclabito (Fig. 3B). The general trend of the outcrop is 066° (Fig. 3A), which is approximately normal to the reconstructed dune-crestline orientation of 335° from Benan & Kocurek (2000).

The Entrada/Todilto contact and the top surface of the Todilto were surveyed using a total station (Fig. 3B). A best-fit line through the surveyed elevation data was used to remove decimetre-scale fluctuations in the surveyed points offset laterally along the sloping Entrada/Todilto contact. A gigapan photomosaic was made of the north-eastern portion of the outcrop. Twenty-one sections (Sections A to U, Fig. 3B)
were measured in the Todilto. Where the outcrop was inaccessible because of slope (primarily over the gypsum), thicknesses were estimated using a tape and the gigapan photomosaic; the greatest measurement error occurs with the gypsum outcrops. A cross-section of

Fig. 3. (A) The study area at Ghost Ranch as indicated by the rectangle (see regional position in Fig. 1). The Entrada Sandstone forms the white cliffs above the red slopes of the Triassic Chinle Formation. The solid line is the general outcrop trend of 066°. The dashed line is positioned over the dune crestal area and oriented in the reconstructed crestline trend of 335° from Benan & Kocurek (2000). The outcrop trend is in the general dune migration direction. Black arrows indicate the gypsum mound in the studied outcrop and a gypsum mound on a pedestal to the south-east. Given the dune crestline orientation, these now disjoined gypsum outcrops were deposited within the same Entrada interdune area. (B) Surveyed locations of Todilto measured Sections A to U (red dots). Small black dots show surveyed control points. Note the down-dropped fault block between Sections G and E, and covered interval. Location shown in (A).
the total outcrop was constructed from the survey data, photomosaic and measured sections. In the cross-section, measured sections were projected onto a vertical plane trending 066° or in the overall outcrop trend. Between Sections G and E there is a down-dropped fault block and covered interval (Fig. 3B). In the cross-section, this block was restored to be continuous with the sections on either side using a simple linear interpolation. Samples were collected at the measured sections for thin section analysis.

Thin sections were analyzed in plane-polarized and cross-polarized light using a Zeiss M2m petrographic microscope (Carl Zeiss Microscopy GmbH, Jena, Germany). In addition, several samples of the basal laminated mudstone (facies T1) were dissolved in a bath of ca 5% acetic acid (diluted household vinegar), following Anderson & Kirkland (1960). Samples of the residue, including organic material, were filtered, dried and mounted on stubs to be analyzed in a scanning electron microscope (JEOL JSM-6490LV SEM; JEOL Limited, Tokyo, Japan). Secondary electron (SE) images of the organic material were taken at low vacuum with uncoated specimens.

REVIEW OF PRESERVED DUNE AND REMOBILIZED SANDS, WITH NEW OBSERVATIONS

In agreement with Benan & Kocurek (2000), relief defined by the top of the cross-strata within the Entrada Sandstone is interpreted to reflect a partially preserved dune; this palaeotopography is buried beneath remobilized Entrada sands that are interpreted to have been deposited by subaqueous sediment-gravity flows (Fig. 4). As reconstructed from the compound cross-strata, the dune was a sinuous crescentic ridge with a crestline trending NNW (335°), migrating towards the WSW, and with along-slope migration of the sinuosity and/or superimposed lee dunes (Benan & Kocurek, 2000). In the reconstructed cross-section, interdune area, stoss slope, crestal area and lee slope can be identified (Fig. 4). The portion of the lee slope preserved is largely the upper lee slope, and the outcrop is truncated by erosion to the southwest. The eroded crestal area of the preserved dune is ca 25 m above the lowest point in the interdune area (Fig. 4).

Decimetre to metre-scale beds of remobilized Entrada sand progressively onlap and bury the stoss slope (Fig. 4; Benan & Kocurek, 2000, fig. 4). Deposition of these onlapping beds appears to have been relatively non-erosional. Although centimetre-scale scour into the stoss slope is evident, the surface of the stoss slope is retained, and small sets of cross-strata overlying the truncated large compound cross-strata occur, and these sets are interpreted as representing small dunes superimposed on the stoss slope. Overlying the stoss slope, the remobilized sand beds typically appear structureless or faintly laminated, but in the best exposures parallel-laminated sand at the bed base yields upward to contorted bedding and flame structures, thence to structureless sand (Fig. 5D). This succession of structures is interpreted as deposited by sediment-gravity flows with initial deposition in upper plane bed (i.e. parallel-laminated sand), followed by rapid deposition that gave rise to contorted bedding and fluidization upward (i.e. structureless sand) with water expulsion (cf. Lowe, 1979, 1982). In contrast to the stoss slope, the crestal area and lee slope show a metre-scale scoured profile (Fig. 4) and beds of remobilized sand fill scoured channels with over-steepened walls (Fig. 5A and B). On the lee slope, a possible coherent block of slumped cross-stratified sand occurs within the remobilized sand (Fig. 5C), but the nature of the outcrop prevented determination of whether this feature is fully enclosed within the remobilized sand and not a window into an irregularly scoured remnant of Entrada. The thickness of the unit of remobilized sand ranges from ca 20 m in the interdune area to <2 m over the crestal area (Fig. 4).

Using Fig. 4 as a two-dimensional proxy for sand volume, the sediment-gravity flow deposits have a cross-sectional area of 8100 m² whereas the cross-stratified dune has an area of 12 400 m². Considering that only the crestal area and lee slope show significant erosion, and therefore contributed sediment to the sediment-gravity flows, a significant imbalance exists between local dune erosion and emplaced remobilized sand. This imbalance implies that the sediment-gravity flows transported sand into the field site along interdune corridors. Moreover, it is likely that the sediment-gravity flows in the crestal area and on the upper lee slope, including those containing the possible coherent block, were locally derived from slumping and scouring of the preserved dune, whereas the beds onlapping the stoss slope represent remobilized sand transported along the interdune corridor.
The preservation of the remnant dune buried by sediment-gravity flows led Benan & Kocurek (2000) to conclude that a catastrophic flooding event submerged the Entrada dune field and initiated the Todilto water body. Ideally, outcrops of the lower lee slope and downwind interdune area would show whether dune migration continued during a slower-paced flooding (i.e. a steep angle of dune climb with rapid interdune accretion), or if flooding was instantaneous in terms of rates of dune migration. Given the absence of these outcrops, it is important to note that the interdune area on the stoss side does not show interdune deposits typical of the Entrada (for example, sabkha deposits), but rather is filled by sediment-gravity flows that even preserve dune stoss-slope sets. If initial flooding was sufficient to fully submerge the dune, a minimum water depth of ca 35 m is required (Fig. 4). Whereas events such as breaching of coastal areas during a storm surge can be envisioned as initiating the flooding, interpreted water depth and the continuation of the Todilto water body require that the dune field housed within the San Juan Basin was below sea level (Benan & Kocurek, 2000). Following final deposition of the sediment-gravity flows, ca 12 m of relief existed on the surface from the crestal area to the interdune area (Fig. 4). This palaeotopography formed the antecedent condition for subsequent Todilto deposition.

TODILTO FACIES

Seven facies (T1 to T7) were identified in the Todilto at Ghost Ranch and these were mapped spatially over the palaeotopography, using the measured sections as control points (Fig. 6). The facies were initially numbered by elevation on the palaeotopography (T1 to T5), with the gypsum (T6 and T7) identified lastly; this facies nomenclature corresponds to the archived samples. However, following interpretation of the facies, it was clear that elevation does not strictly track depositional (i.e. stratigraphic) order. Thus, facies retain their original nomenclature (to correspond with archived samples) but these are discussed below in interpreted stratigraphic order.

Facies T1 (basal laminated mudstone)

Description
The basal Todilto facies T1 consists of a laminated limestone (mudstone) with millimetre-scale dark and light laminae (Fig. 7A). Although well-laminated, the laminae are wavy to crinkly, non-parallel and difficult to trace laterally at the

Fig. 4. Cross-section constructed from survey data, measured sections and photomosaic, and projected onto a plane in the general outcrop trend 066°. Note the locations of measured Sections A to U. Interpreted positions on the preserved dune palaeotopography are given. Vertical exaggeration of 6× allows depiction of the thinnest Todilto, but distorts compound cross-strata in the Entrada. For simplicity, small sets along the upper surface of the stoss slope are not depicted.
metre-scale. In a few areas, the laminae define centimetre-scale stromatolitic domes (Fig. 7F). Facies T1 is the only facies that is continuous over the crestal area (36 to 42 cm at Sections D to F) and upper lee slope (37 to 50 cm at Sections A to C), thins irregularly down the stoss slope (23 to 38 cm at Sections G to Q) and is thinnest in the interdune area (11 to 17 cm at Sections R to U). Uncommon gypsum nodules (<1 cm) and centimetre-scale gypsum lenses occur, especially in the dune crestal area.

Sand injections into facies T1 from the underlying Entrada are relatively common over the entire outcrop (Fig. 6), and occur as: (i) incipient dykes that create antiforms in the overlying laminated beds; (ii) dykes that penetrate the laminated beds and create folds above (Fig. 7B); and (iii) dykes that splay into sills along bedding planes (Fig. 7C).

Additional folds occur in facies T1 that could not be linked to sand injections, and these are confined to the dune crestal area (Fig. 6). These folds in the crestal area consist of both continuous antiforms, and fractured antiforms with upturned edges and rotated blocks (Fig. 7D). Some fractured antiforms show gypsum interlaminated with the limestone, in some cases with displacive growth that arches overlying limestone laminae (Fig. 7E). The cores of some fractures show a cemented breccia of limestone clasts and gypsum (Fig. 7E).

In thin sections three laminae types are evident in facies T1: (i) black, organic-rich laminae; (ii) clotted-texture laminae; and (iii) micritic laminae (Fig. 8). Laminae are distinctly irregular
with wavy contacts and are discontinuous; in particular, the organic-rich laminae range in thickness from \(<0.1\) to \(3.0\) mm in a single thin section.

The organic-rich laminae contain silt-sized grains, primarily quartz. The silt is typically concentrated along the upper horizons of the laminae, where some grains rest beyond the angle of repose, but silt is also distributed throughout the organic-rich laminae (Figs 8 and 9A to E). Opaque minerals occur, including pyrite (Fig. 9E and J to L) and probable uranium minerals, as reported by Anderson & Kirkland (1960). Structures within the organic-rich laminae include millimetre-scale arches and stromatolitic domes with lateral pinch-out features, tufts and poorly developed tented to cuspatel features (Fig. 9A to E), using the terminology of Sumner (1997). In some of these cases, the tufts appear filamentous (ca 30 to 35 μm wide; Fig. 9A to D) but these are not preserved well enough for microfossils to be observed in thin section.

The clotted-texture laminae exhibit peloids, microclots and mesoclots (sensu Shapiro, 2000), and show lateral variability in thickness (Fig. 9F and G). Within the clotted fabrics are areas of deformation and some mesoclots appear to be rolled-up structures (Fig. 9F).

The micritic laminae show a range of textures that include: (i) an interlocking mosaic of micrite, commonly with patches of microspar (Figs 8A, 9B and 9E); (ii) a peloidal texture (Figs 8A and 9E), in some cases transitional to clotted-texture laminae (Figs 8B and 9G); and (iii) a darker micrite with apparent fragments of clotted-texture laminae (Fig. 8B).

Paralleling the thinning of facies T1 from the dune crestal area and the upper lee face into the interdune area is a change in the frequency of laminae types and their thicknesses. Over the crestal area, the organic-rich laminae and clotted-texture laminae are best developed (Fig. 8B). The organic-rich laminae can reach 5 mm in thickness, forming millimetre-scale stromatolitic
domes, tents and cuspsate or filamentous features. Further down the dune stoss slope and into the interdune area, organic-rich laminae become thinner; there are fewer instances of stromatolitic features and clotted-texture laminae, and the micritic laminae become thicker and more abundant (Fig. 8A).

Residue from dissolution of samples of facies T1 consisted of fine-grained sediment (clay and silt) but some small (millimetre-scale), cohesive

Fig. 7. Outcrop-scale structures of facies T1. (A) Laminated mudstone in outcrop. Centimetre-scale ruler for scale. (B) Sandstone dyke (arrow) from underlying Entrada ('E') folds and cuts across layers of facies T1 before terminating beneath an antiform. Pencil (14 cm) for scale. (C) Sandstone sill (arrow) between layers of facies T1. The sill originates from a vertical injection from the Entrada ('E') to the right of the image. (D) Folds in facies T1 overlying the Entrada ('E') and not associated with an injection. These folds (ca 40 cm thick) are interpreted as tepee structures. (E) Tepee structure with upturned edges and core of cemented breccia of limestone and gypsum. Note the gypsum lens (arrow) with arched limestone laminae above. Boulder from facies T1 at Section E, centimetre-scale bar on the left. (F) Small stromatolite (arrow) between Sections L and M.
organic films were recovered (Fig. 9H and I). No filamentous microstructure was observed in SEM analyzes of these organic films (Fig. 9J to M); however, clays (Fig. 9M) and small (ca 10 \( \mu m \)) rare pyrite framboids (Fig. 9J to L) were observed. In addition, some of the laminae surfaces have a structure that is reminiscent of a microbial mat with trapped grains (Fig. 9J).

**Interpretation**

Based upon the composition, fabric and arrangement of the laminae in thin sections,
facies T1 is interpreted as a microbial-laminated bindstone. Larger manifestations of stromatolites (for example, Fig. 7F) occur, but are not common, indicating that microbial development within the Todilto was primarily as a surface mat.

The black, organic-rich laminae, originally identified by Anderson & Kirkland (1960) and interpreted as settled planktonic material, are reinterpreted as microbial mats. Observed structures, including millimetre-scale arches and stromatolitic domes, tufts and tented to cuspat e structures, have been described from modern mats and used as diagnostic structures in the identification of ancient microbialites (Walter et al., 1976; Mayall & Wright, 1981; Sumner, 1997; Flannery & Walter, 2012; Martindale et al., 2015).

Although dissolution residues did not yield cohesive layers [‘sapropel layers’ as reported by Anderson & Kirkland (1960)] and filamentous microstructures were not observed, the recovered millimetre-scale organic films and ghost structures reminiscent of microbial mats are consistent with a microbial mat interpretation. Degradation of the original microbial mat is accredited to the age of the unit and diagenesis that it has undergone. Clay minerals and pyrite framboids may also be related to the mat as biogeochemical byproducts of the microbial community. Pyrite framboids (authigenic) have been found associated with modern and ancient microbial mats (e.g. Popa et al., 2004; Perri et al., 2018). Pyrite is the ultimate stable mineral from sulphate-reducing bacteria within microbial mats (Gerdes, 2007), suggesting that these microbes were a component of the microbial community.

Concentrations of clastic silt along the upper surfaces of the organic-rich laminae, identified as a separate laminae type by Anderson & Kirkland (1960), as well as silt dispersed within the laminae, are interpreted as integral to the organic-rich laminae, and represent grains trapped by filamentous micro-organisms. The trapping and binding of detrital grains, especially at angles greater than the angle of repose, is a diagnostic feature of well-developed microbial mats (Riding, 2000; Dupraz & Visscher, 2005; Gerdes, 2007; Basso et al., 2013; Frantz et al., 2015). While filamentous structures were observed, it was not possible to determine whether the filaments were cyanobacterial or algal in origin, given their current preservation. The silt-sized grains are compatible with aeolian suspension-transport to the water body.

Equally inherent to the interpretation of a microbialite are the laminae that exhibit a clotted texture. This texture is commonly associated with the calcification of extracellular polymeric substances (EPS) which are secreted in copious amounts in microbial mats (Riding, 2000; Dupraz et al., 2004). Deformation and possible rollup structures within these laminae suggest cohesive microbial binding (Gerdes, 2007); rollup structures have been described from microbial mats dominated by filamentous cyanobacteria (Beraldi-Campesi & Garcia-Pichel, 2011). For both the organic-rich and clotted-texture laminae, the change in laminae thickness laterally, as well as their crinkly and non-parallel structure, are consistent with a microbial mat interpretation, but not compatible with sediment settling from suspension through the water column.

Micritic laminae, identified by Anderson & Kirkland (1960) and interpreted as an abiotic precipitant, are the most problematic. Determination of the origin of micrite within ancient microbialites is typically not straightforward (e.g. Petryshyn et al., 2016). Not all microbial mats are calcified, and the environmental conditions and microbial processes that promote calcification are not fully understood (e.g. Reid et al., 2000; Riding, 2000; Dupraz & Visscher, 2005). Calcification of microbial mats initially focused upon the incorporation and cementation of trapped and bound sediment, typically carbonate, within the accreting biofilm (e.g. Ginsburg & Lowenstam, 1958; Logan, 1961). More recent work, however, has identified a variety of geomicrobiological processes occurring within the microbial mat that biologically mediate carbonate precipitation (e.g. Visscher et al., 1998; Grotzinger & Knoll, 1999; Dupraz et al., 2004, 2009, 2013; Decho et al., 2005; Glunk et al., 2011; Bouton et al., 2016; Pace et al., 2018). The calcification of EPS into micrite (Riding, 2000; Dupraz & Visscher, 2005; Glunk et al., 2011), with subsequent recrystallization of micrite to micropar (Dupraz et al., 2004, 2009), are also recognized processes in modern microbial systems.

Micritic laminae within the Todilto that show a peloidal texture, or are associated with clotted-texture laminae, or contain fragments of clotted fabrics may be reasonably interpreted as having arisen through the calcification of EPS. Other micritic laminae appear integral to the microbial mat structure and alternate with organic-rich laminae that show well-developed

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microbial mat structures (for example, Fig. 9A to E). These micritic laminae are also reasonably interpreted as carbonates precipitated within the microbial mat. Most problematic are mosaic micritic laminae, most of which contain microspar patches. These micritic laminae occur most typically where the microbial mat structure is poorly developed, such as facies T1 within the interdune area. The precipitation of these laminae may have been microbially influenced but, as described for facies T2 where this fabric is much better developed, the laminae more likely represent abiotic carbonate precipitation or the calcitization of gypsum.

The environmental context for microbial mat development in facies T1 can be partly reconstructed from the geomorphic context. Given the interpretation of a flooding event that completely submerged the dunes, an increase in water depth from the interdune area to the crestal area of ca 12 m occurred, and water depth over the crestal area is not known. Presuming that flood waters were marine, initial Todilto waters were probably near normal marine in salinity. Microbial mat processes and structure are known to strongly reflect external forcing, as well as the composition of the microbial community (Dupraz et al., 2004; Gerdes, 2007; Petryshyn et al., 2012; Basso et al., 2013; Bouton et al., 2016). Changes observed in the microbial mat from the crestal area to the interdune area can reasonably be interpreted as depth-related, with the most obvious factor being a reduction in light intensity. Although submerged, the absence of current-formed structures within facies T1 is attributed to surface cohesion created by the microbial mat.

Additional processes and evolution of the Todilto water body can be inferred from the outcrop-scale sedimentary structures. Sand injections such as dykes and sills argue for unconsolidated Entrada sand well into and after deposition of facies T1. These injections may have arisen with liquefaction of Entrada sands because of water–sediment loading (Lowe, 1975, 1976), escape of air trapped within the rapidly flooded dunes (Glennie & Buller, 1983), wave loading with high-wind events (Lee et al., 2007) or earthquakes within this retroarc foreland setting.

The continuous and fractured antiforms not associated with injections and that occur in the crestal area are interpreted as tepee structures, which can show a range from incipient antiforms to fractured antiforms with upraised edges (Lokier & Steuber, 2009, fig. 8). Well-formed tepees are peritidal structures that are cross-sectional representations of three-dimensional polygonal fractures with buckled edges that result from expansion of the surface layer because of carbonate/evaporite crystallization, thermal expansion/contraction, wetting/drying, and other factors (Adams & Frenzel, 1950; Assereto & Kendall, 1971, 1977; Warren, 1982, 1983). Tepee structures are commonly associated with microbial mats (Warren, 2006) and some of the Todilto antiforms may be a combination of stromatolitic and tepee growth. Tepees developed within facies T1 generally consist of folded to fractured layers, but do not show the fenestral fabric typically associated with tepees that form contemporaneously with deposition (Assereto & Kendall, 1977; Warren, 1983). However, gypsum associated with the tepee structures appears syndepositional. For these reasons, the formation of the tepees is interpreted to have occurred after deposition of facies T1 when water level fell below the crestal area, and an increase in salinity initiated gypsum precipitation.

Facies T2 (recessive laminated mudstone)

Description
Facies T2 is identified as a recessive, crumbly, thin (13 to 26 cm) unit overlying facies T1 on the lower stoss slope and in the interdune area (Sections O to U). In the vicinity of Sections N and O, facies T1 and T2 interfinger (Fig. 6), and facies T2 is not recognized progressively higher on the stoss slope. Facies T2, therefore, represents the stratigraphic equivalent of the upper portion of facies T1 prior to the onset of the deposition of the lower gypsum (facies T6) within the interdune area.

In thin section, portions of facies T2 are similar to facies T1 on the lower stoss slope and in the interdune area, and consist of very thin, discontinuous, black, organic-rich laminae separated by much thicker micritic laminae (Fig. 10A). The organic-rich laminae are similarly associated with concentrations of quartz silt, but unlike in facies T1, the organic material is thickest in depressions, and no stromatolitic domes or other structures readily associated with microbial mats were identified. In addition, clotted-texture laminae are absent. Micritic laminae mostly show a mosaic of micrite with abundant patches of microspar, but weakly-developed peloids characterize some laminae.
The micritic laminae with patches of microspar commonly show an upper surface characterized by sub-millimetre protrusions (Fig. 10B). Considered with the organic-rich laminae, an upward trend is defined in which an organic-rich lamina is overlain by dense micrite, commonly with peloids, that yields upward to micrite with abundant microspar (Fig. 10B). In other thin sections of facies T2, the black, organic-rich laminae are scarce or absent, and quartz silt is abundant in a mudstone matrix.

**Interpretation**

Although no definitive microbial mat features (for example, stromatolitic domes, tufts and cuspatate structures) were identified in facies T2, the black, organic-rich laminae are interpreted as representing a poorly-developed microbial mat by analogy to the poorly-developed mats of facies T1 in the interdune area. Facies T2, therefore, shows a continuation of the crest to interdune trends observed in facies T1, but enhanced such that the microbial mat is yet more poorly developed than in underlying facies T1. Overall, facies T2 is interpreted as a laminited bindstone where there is very weak microbial mat development, and a quartz-silt bearing mudstone where microbial mat development is largely absent.

The lamina-scale upward trends in facies T2 (Fig. 10B) are thought to indicate the onset of gypsum precipitation within the Todilto water body. Whereas the lower portions of micritic laminae, showing a dense micrite with peloids, may be associated with the microbial mat, the upper portions of the laminae characterized by surface protrusions and abundant microspar are very similar to calcitized evaporitic fabric identified by Kendall (2001). Given that facies T2 is equivalent to the upper portion of facies T1, interpreted relict evaporitic fabric in facies T2 represents the onset of periodic gypsum precipitation in the Todilto water body, and probably coincides with crestal emergence and initiation of the tepee structures. Continued drawdown of the Todilto water body and brine concentration is manifested by subsequent deposition of the lower gypsum (facies T6) within the interdune area.

**Facies T5 (dissolution breccia)**

**Description**

Todilto facies T5 occurs only in the dune crestal area, forming a 10 to 15 cm thick rim overlying facies T1 (Fig. 6). On the dune lee side, facies T3 (described below) thins, onlaps and pinches

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**Fig. 10.** Facies T2 in thin section, plane-polarized light. (A) Black, organic-rich laminae (examples marked with red arrows) are very thin and commonly discontinuous, with thickening in depressions (black arrow). Micritic laminae with microspar (examples marked by white arrows) dominate, but some micritic laminae have a more peloidal structure (yellow arrows). Dashed box in upper left shows location of (B). Section U. (B) Upward trends (white triangles) defined by organic-rich laminae overlain by micrite with increasing microspar upward. Upper surfaces are marked by protrusions (yellow arrows). The microspar and surfaces are interpreted as calcitized evaporitic textures. The protrusions appear to form surfaces for subsequent microbial mat development.
out over facies T5 in the crestal area (Fig. 6). In outcrop, facies T5 is a breccia consisting of centimetre-scale mudstone intraclasts with up to centimetre-scale, spar-filled pores between intraclasts (Fig. 11A). In thin section, the angular intraclasts consist largely of clotted-texture laminae with microclots and mesoclots (Fig. 11B to D), as characterizes the crestal area of facies T1. Pores are filled by two generations of cements: (i) a rim of isopachous, bladed to equant spar crystals radiating from the intraclasts; and (ii) a mosaic of large, equant spar crystals filling the pores (Fig. 11D). In some cases, tepee structures in facies T1 lose definition upward into facies T5, but in other cases the tepees are apparent within facies T5.

**Interpretation**

Facies T5 is interpreted as a dissolution breccia, and represents the remnants of a thick microbialite similar to the well-developed microbial mat in the crestal area of facies T1. The intraclasts consisting of clotted-texture laminae are interpreted as derived from calcification of EPS, as discussed above for facies T1. Facies T5 and crestal portions of facies T1

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**Fig. 11.** Facies T5. (A) In outcrop facies T5 overlies facies T1 in the crestal area and consists of angular laminated intraclasts (examples marked by black arrows) separated by spar (yellow arrow). Note centimetre ruler for scale. Section E. (B) Thin section in plane-polarized light showing angular intraclasts with spar infilling. Intraclasts show a dark clotted texture. (C) Thin section in plane-polarized light showing intraclasts with clotted-texture laminae and spar infill. Note the possible fenestrae (white arrows). (D) Thin section in cross-polarized light showing two generations of cements: relatively isopachous bladed to equant cement rind and then a second, large, equant cement mosaic. All thin sections are from samples taken near Section E.
represent a well-developed microbial community that developed within the crestal area where light and other conditions (for example, oxygenation) would have been optimized. Largely absent in facies T5 are the black, organic-rich laminae which were probably less preservable. With brecciation, the organic-rich laminae probably served as parting planes (also observed in facies T4, as discussed below) and are thought to have been subsequently flushed from the system. The dissolution breccia is interpreted to have formed with subaerial exposure of the crestal area, and was developed within the exposed crestal portions of facies T1. Because some tepee structures are terminated upward by facies T5, the dissolution breccia is believed to have developed after formation of the tepees. Because facies T3 onlaps facies T5, initial development of the dissolution breccia is further bracketed as before deposition of facies T3. However, there is no direct evidence for subsequent submersion and burial of the crestal area during Todilto time (Fig. 6) and development of the dissolution breccia probably continued in exposed areas of the crestal area for the duration of Todilto time. Because the dissolution breccia represents a collapsed fabric, the original elevation of the crestal area was higher than its current elevation. Cementation of the dissolution breccia is thought to have occurred in the subsurface, as with facies T3 and T4 (discussed below).

**Facies T3 (laminated mudstone with gypsum nodules) and facies T6 (lower gypsum)**

**Description**

Todilto facies T3 is a well-laminated limestone (mudstone) with spatially-varying degrees of development of gypsum nodules (Fig. 12A). Facies T3 overlies facies T1 except on the lower stoss slope where it overlies facies T2 and at the crestal area where it pinches out over facies T5 (Fig. 6). On the upper lee face (Sections A and B) the unit is 95 cm thick, then rapidly thins towards the crestal area (56 cm at Section C; 21 cm at Section D), and pinches out over facies T5 between Sections D and E (Fig. 6). The stratigraphic relationship between facies T3 and T5 is obscured by the covered area on the upper stoss slope, but facies T3 is ca 60 to 70 cm thick on the exposed stoss slope (Sections G to P), and thickens to 1.1 m at Section Q (Fig. 6). Between Sections Q and R, facies T3 abruptly transitions into gypsum (facies T6) over ca 15 m laterally (Fig. 13). In spite of the change from limestone to gypsum, the upper contact of facies T3 can be traced across the gypsum, thereby separating lower (facies T6) and upper (facies T7) portions of the gypsum (Fig. 6). Between Sections T and U, the gypsum abruptly transitions back to facies T3 limestone with abundant gypsum nodules within a few metres laterally (Fig. 6).

Detailed study of the gypsum was not possible owing to steep slopes and surface weathering,
but it is generally described as having a chicken-wire fabric of coalesced gypsum nodules with thin, brecciated limestone horizons (see also Kirkland et al., 1995). In addition, the lower gypsum (facies T6) has a more blocky and nodular appearance than the upper gypsum (facies T7). The coalescing of gypsum nodules to form the chicken-wire fabric in the interdune area, although abrupt, follows a trend of increasing nodule size, abundance and degree of coalescing down the stoss slope. Nodules in facies T3 are centimetre-scale and abundant on the lower stoss slope, where they commonly coalesce into wavy beds with the limestone laminations deformed around the nodules (Fig. 12A). Nodules decrease in size and abundance and become isolated up the stoss slope such that these are sparse and <0.5 cm in diameter by Sections I and J, and are largely absent crestward. Similarly, centimetre-scale nodules occur at the most leeward sections (Sections A and B), but decrease in size and abundance crestward.

Facies T3 on the upper lee slope shows two types of folding. As seen in Fig. 14, strata of facies T3 are arched into an antiform over a
large sand dyke that penetrates through facies T1. In addition, downslope from the antiform, folding along bedding planes occurs in facies T3, and some beds thin and toe-out downslope (Fig. 14). Contained within the folds are brecciated blocks with rotation (Fig. 15A).

In thin section facies T3 shows very thin black, organic-rich laminae with scattered silt-sized quartz and thicker micritic laminae with patches of microspar (Fig. 12B). Laminae have been cohesively deformed around common millimetre-scale lenses filled by a mosaic of equant spar crystals. Within the organic-rich laminae there are no obvious microbial structures such as stromatolitic domes, tufts or cuspatte structures. Clotted-texture laminae were also not observed.

**Interpretation**

Although lacking in definitive microbial mat structures, by analogy to facies T1, the silt-bearing, black, organic-rich laminae are interpreted as representing poorly-developed microbial mats. Given the absence of associated clotted-texture laminae and the complexity of the laminae as they deform around the gypsum nodules and calcite-filled vugs, it is difficult to determine whether the micritic laminae were precipitated within the microbial mats or represent abiotic micrite deposited over and incorporated into the mats. The latter possibility argues for a periodic alternation of microbial mat and micrite laminae. In either case, the microbial mat structure (i.e. organic-rich laminae) is pervasive and probably provided cohesion, thus facies T3 is interpreted as a poorly-developed microbial mat-laminated bindstone.

Cavities filled with spar are interpreted as pseudomorphs of small gypsum nodules that have been dissolved and subsequently filled by calcite during a later stage of cementation. Given the presence of the larger gypsum nodules in the facies, the small gypsum nodules are thought to have been more susceptible to dissolution, while the larger nodules remained coherent. Because the laminations cohesively deform around the gypsum nodules and the small pseudomorphs, the nodules are taken as having grown displacively beneath the surface. Gypsum nodules are characteristic of the vadose to upper phreatic zones of sabkhas (Kendall, 1978; Warren & Kendall, 1985; Warren, 2006). The increase in nodule size and abundance down the stoss slope and into the chicken-wire gypsum of facies T6 is interpreted as representing an increase in brine saturation, such that the upper portions of facies T3 were deposited on a sabkha surface that graded downslope into progressively wetter conditions, with probable periodic ponding in the interdune area. Because facies T3 pinches out over the flanks of facies T5, the crestal area is interpreted as having been emergent during deposition of facies T3.

The large sand dyke that caused folding of facies T3 strata on the upper lee slope demonstrates that sand injections from the Entrada continued even after deposition of facies T3. The folding downslope of the antiform, however, represents yet another fold type and one that appears to be confined to the lee face [see
Lucas et al. (2014) for an overview of fold types in the Todilto regionally. These folds, in combination with the rotated blocks and the thinning and downslope-toeing of some layers, suggest a downslope movement of layers, along glide planes where toeing occurs. These features, as well as the late-stage sandstone dyke, probably resulted from instability associated with the remnant slope on the upper lee face, which exceeded 2.5° as estimated from Fig. 4. In contrast, the remnant slope on the reconstructed stoss slope (see below) was ca 1°.

The thickness of the lower portion of the gypsum (facies T6) is estimated as 2.5 to 3.5 m (Fig. 6), and sections of adjacent carbonate facies T3 are significantly thinner (1-1 m at Section Q; 1.4 m at Section U). In outcrop the gypsum (facies T6) clearly rises higher than facies T3 strata (Fig. 6). Although displacive gypsum growth may raise a depositional surface (Warren, 2006), during facies T3 time, the surface probably extended from below the exposed crestal area as a low slope into the interdune area where gypsum was concentrated. Thickness differences between gypsum facies T6 and adjacent carbonate facies T3 imply that in the subsurface section collapse of facies T3 (ca 60%) has occurred and/or expansion of the gypsum body has occurred. Gypsum strata may expand vertically with additional crystallization, contract with conversion to anhydrite or collapse with dissolution and development of a class of karst topography (Warren, 1999, 2006; Johnson, 2005). Unlike facies T4 (discussed below), facies T3 does not show a fabric suggesting significant section shortening. Carbonate laminae are commonly deformed, but this appears to be largely around gypsum nodules and their pseudomorphs. It is probable, however, that gypsum nodules, especially smaller nodules, and gypsum laminae were dissolved, allowing for section collapse. Expansion of the gypsum (facies T6), however, cannot be ruled out. A dynamic model for selective dissolution and collapse of the section adjacent to the gypsum body is discussed in the interpretation of facies T4, where significant section collapse is manifested.

Regardless of the origin of the elevation differences, the highest point of deposition of facies T3 (between Sections D and E in Fig. 6), and the upper surface of facies T6 at Section S and the upper surface of facies T3 at Section Q is ca 9 m and 11 m, respectively. Crestal facies T5 rises ca 1 m yet higher. These measurements provide an estimate of remnant relief at the end of facies T3 deposition and the beginning of deposition of facies T4.

**Facies T4 (brecciated laminated limestone) and facies T7 (upper gypsum)**

**Description**

Todilto facies T4 is distinctively laminated, vuggy limestone (mudstone) in which the millimetre-scale laminae are brecciated into centimetre-scale, elongated intraclasts (Fig. 16A). The intraclasts are oriented generally sub-horizontal, but rotated and on-end orientations are common. The overall fabric of the rock is a semi-chaotic stacking of elongated intraclasts with irregular, commonly calcite-filled vugs between intraclasts. The unit consists of three decimetre-scale beds separated by bedding planes (Fig. 6), which are marked by a recessed, centimetre-thick gypsum powder. On the upper lee face, facies T4 rapidly thins from 3 m at Section A to 0.8 m at Section C, and it then pinches out immediately before Section D. All three beds thin up the lee face, but bedding distinction is lost before the pinch-out. Downslope from the covered interval on the stoss slope, facies T4 regularly thickens from 0.5 m at Section G to 1.8 m at Section L, and varies from 1.2 to 1.8 m on the lower stoss slope (Sections M to Q). Between Sections Q and R, the two lower beds of facies T4 (T4a and T4b) abruptly grade into the upper gypsum (facies T7) over several metres laterally, and in the same area of the interdune depression as where facies T3 grades into the lower gypsum (facies T6; Fig. 13). The upper bed of facies T4 (T4c) continues as a 0.5 m thick limestone bed that grades into gypsum near Section R. The estimated maximum thickness of the upper gypsum (facies T7) is ca 8 m. Bedding planes in facies T4 (i.e. contacts between T4a, T4b and T4c) can be traced through the upper gypsum (Fig. 6). Between Sections T and U, the gypsum abruptly transitions back into facies T4 limestone with three distinct beds. As with facies T3, facies T4 shows folds on the upper lee face, and within the folds are pronounced roll-over structures with shear planes (Fig. 15B).

As traced from the photomosaic in Fig. 4 (see also Benan & Kocurek, 2000, fig. 4), beds of the Beclabito Member overlying the gypsum mound drape downward over the north-eastern flank of the mound. The south-western flank of the gypsum mound is obscured by cover, but ravine outcrops show a chaotic mosaic of facies T4.
fragments and Beclabito sandstone. Underlying beds of facies T4 show fractured layers thrusted over one another. Coarser sandstone veins, most likely to be derived from the Beclabito, were found as low in the section as facies T3.

In thin section, facies T4 is remarkably constant across the outcrop as an intraclast breccia in which the intraclasts are laminated mudstone (Fig. 16B). In agreement with outcrop observations, the fabric of the rock shows intraclasts with an overall sub-horizontal orientation, but rotated and on-edge intraclasts are common. Some intraclasts are oversized with respect to laminae thickness and may have been deformed. Pores filled by spar occur between the irregularly stacked intraclasts. Laminae appear to have brecciated along very thin dark, organic-rich horizons with concentrations of coarse silt, as evident along the upper surface of some intraclasts. In some thicker intraclasts, these organic-rich laminae with silt are evident within the intraclasts. The predominant composition of the laminae, however, is an aphanitic brown micrite.

**Interpretation**

Facies T4 is interpreted as a collapsed set of beds in which a once laminated mudstone was brecciated into intraclasts as significant amounts of material were removed. The top of the equivalent gypsum (facies T7) is ca 1.5 m below the crestal area, but is ca 0.5 m above the point at which facies T4 pinches out on the upper lee face (Fig. 6). As with facies T3, it is assumed that the depositional surface of facies T4 originally extended as a low slope from below the emergent crestal area to the top of the equivalent gypsum (facies T7). If no gypsum expansion is assumed, the section collapse in facies T4 is ca 75%, using 8 m as the gypsum (facies T7) thickness and 1.8 m as the thickest adjacent facies T4 (Section P). The observation that the upper surface of facies T7 is ca 0.5 m above the point where facies T4 pinches out, argues that some gypsum expansion in the subsurface has also occurred.

However, most of the thickness differences between the upper gypsum (facies T7) and the adjacent carbonate (facies T4) must have occurred with collapse of facies T4. The most probable mechanism is the development of a dissolution front adjacent to the gypsum body. Within the subsurface, bodies of gypsum can lose effective porosity and permeability and act as aquitards or aquicludes (Warren, 1999, 2006; Sarg, 2001). Fluids are channelled around the gypsum bodies and dissolution fronts are set up such that the greatest dissolution occurs along the edge of the gypsum body (e.g. Ezersky & Frumkin, 2013). In this dynamic process, dissolution would occur within facies T3 and T4 while the gypsum body remained intact. This process is similar to the sag model of Gutierrez & Guerrero (2008), in which interstratal dissolution of evaporites results in the downward movement of the section. Down-draping of Beclabito beds on the north-eastern flank of the gypsum mound is taken to have occurred with
collapse of the section below, and the chaotic bedding on the south-western flank is interpreted as karst topography developed with dissolution. Stapor (1972) reached the same conclusion at Ghost Ranch and attributed the sagging beds adjacent to the gypsum mound to have resulted from evaporite dissolution within the 'brecciated' portion of the Todilto, and having occurred after deposition of the overlying beds. Elsewhere in the San Juan Basin, Hunter et al. (1992) report down-dropped pipes of Beclabito and overlying Horse Mesa and Cow Springs strata as the result of evaporite dissolution and strata collapse within the Todilto. This observation indicates that evaporite dissolution occurred after emplacement of the Beclabito and Cow Springs, and most likely during subsequent Upper Jurassic Morrison time.

Given the absence of evidence for gypsum nodules, conjecture as to what material was removed from facies T4 includes gypsum laminae that were susceptible to dissolution. In this hypothesis the original structure of facies T4 would have consisted of a triplet of organic-rich laminae with silt concentrations, micritic laminae and gypsum laminae. As with other facies, the organic-rich laminae are interpreted to represent poorly developed microbial mats, which served as parting planes during brecciation and have been largely flushed from the section. The aphanitic micritic laminae show the least affinity to microbial mat structure of any micritic laminae within the Todilto, and may represent abiatic precipitation of micrite. An upward progression of laminae from microbial mat to micrite to gypsum could represent a repetitive saline concentration within interdune ponds. Highest salinity would have been in the centre of the ponds, giving rise to the gypsum (facies T7). Because regional field evidence suggests that brecciation did not occur until deposition of the Upper Jurassic Morrison, emplacement of the spar cement between intraclasts did not occur until the Late Jurassic or later. This cementation event may include the late-stage spar cement evident in the other Todilto facies.

Given conjecture that facies T4 laminae and facies T7 gypsum originated within interdune ponds, and that facies T3 is interpreted to represent a largely emergent sabkha that graded downslope to periodic gypsum pools (facies T6), facies T4 represents renewed flooding of the remnant interdune areas. The presence of three beds within facies T4 implies that the flooding was repeated three times, the culmination of each being more widespread precipitation of gypsum, as represented by the powdery gypsum along bedding planes.

DISCUSSION

Figure 17 is a schematic summary of the interpreted stages of development for the uppermost Entrada and Todilto stratigraphic horizon at Ghost Ranch. Stage 1 is the initial configuration of a dune field housed within a basin below sea level. During Stage 2 a rapid marine flooding of the dune field occurred, accompanied by dune mass wasting, which gave rise to sediment-gravity flows that buried much of the remnant dune topography. A microbial community developed on the submerged, remnant palaeotopography (facies T1, T2 and T5) during Stage 3. The best microbialite development was in the crestal area (facies T1 and T5). Although initial waters were normal marine, falling water level gave rise to brines and initial precipitation of gypsum in the interdune depression (facies T2). By Stage 4 the crestal area was emergent and tepees formed (facies T1). Continued exposure of the crest gave rise to development of a dissolution breccia (facies T5). Downslope of the emergent crestal area, microbial mats formed on sabkha mudflats with gypsum nodules (facies T3) and gypsum (facies T6) was deposited in the ponded interdune area. During Stage 5 ponding events below the emergent crest occurred, with a hypothesis of repetitive deposition of microbial mat, carbonate and gypsum laminations (facies T4). Gypsum (facies T7) was again concentrated within the interdune area. Stage 6 depicts subsurface collapse of facies T3 and, to a much greater extent, facies T4 along a dissolution front developed adjacent to the gypsum body.

Although depositional environments rarely develop from a 'blank slate', but rather originate and evolve from the antecedent geomorphic surface, the entire depositional sequence of the Todilto facies is arguably controlled by the antecedent Entrada dune topography. In turn, subsurface development of a dissolution front with accompanying section collapse was dictated by the facies architecture.

Subsidence of the San Juan Basin during the Jurassic is poorly constrained, but the Entrada is regionally a wet aeolian system in which the water table was near the surface and the desert floor was just above sea level (Crabaugh & Kocurek, 1993). Sabkha deposits increase in
abundance towards and into the San Juan Basin (Fig. 2). Accommodation space for the marine flood waters may have formed with subsidence of the basin to below sea level while coastal dune barriers existed to the north and west. After the initial flooding and development of microbialites in postulated normal marine waters, the area evolved into an extreme environment. The initial flooding and development of microbialites in postulated normal marine waters, the area evolved into an extreme environment.

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evaporitic environment of emergent dune crestal areas, and interdune mudflats and ponds with microbial mats. The persistence of the Todilto water body after the initial flooding event necessitates a limited connection with marine waters, which compares well to the salina model of Lucas et al. (1985) that envisions marine subsurface recharge and periodic surface flows. The regional bull’s eye pattern with gypsum confined to the interior while carbonate continued along the periphery (Fig. 1) is typical of many salinas (e.g. Warren, 2006). However, the Todilto body is much larger than modern salinas [for example, Lake MacLeod in western Australia, given by Warren (2006) as the largest salina, is 4800 km² compared to the 105,000 km² of the Todilto]. The Todilto is more comparable to embayment salinas where a coastal embayment loses its open marine connection and a salina environment develops (Warren, 2006). An example is the Late Pleistocene coastline of Egypt that was flooded during the last interglacial period and subsequently lost its connection to the Red Sea (Plaziat et al., 1995).

Although the geological circumstances that gave rise to the Todilto are unusual, partial analogues exist in the rock record of dune fields that occupied subsided basins and that were flooded by marine waters with dune topography partially preserved (e.g. Glennie & Buller, 1983; Eschner & Kocurek, 1988; Story, 1998). Because aeolian dunes are readily reworked by tidal and wave currents, examples of preserved dune topography are thought to require a transgression that is both rapid and quiescent (Eschner & Kocurek, 1988). Although scour and collapse of the crestal portions of Entrada dunes at Ghost Ranch occurred and the sediment-gravity flows were energetic (for example, upper plane bed), subsequent deposition within the Todilto water body was low energy. In contrast, the Curtis transgression of the large Entrada dunes in north-eastern Utah was characterized by tide-dominated marine conditions (Eschner & Kocurek, 1986). For this example, deep erosion into dune sand occurred and sediment-gravity flows shed from the dunes bury trains of tidal dunes.

Modern partial analogues include dune fields transgressed by the Holocene rise in sea level after the last glacial maximum lowstand, with development of interdune sabkhas further inland. Extensive sabkhas within interdune areas between large linear dunes of coastal Mauritania originated with the marine transgression of Sahara dunes (Kocurek et al., 1991) which extended onto the continental shelf during the last glacial maximum (Sarnthein, 1978). The microbial mats of Cayo Coco lagoon in Cuba formed over dune topography flooded with the Holocene transgression (Bouton et al., 2016). An extensive area of marine-flooded dune fields in which the relict dune topography strongly influenced subsequent sedimentation is along much of the southern Arabian Gulf (Warren, 2001, 2006; Kirkham, 1998). During the last glacial maximum, the Arabian Gulf was a desiccated basin with widespread dune fields. Aeolian deposits underlie Holocene marine carbonates as a result of the Holocene transgression. Thickening patterns of Holocene strata indicate initial, relatively passive marine encroachment of interdune depressions, and interdune sabkhas, many with microbial mats, persist inland where the surface is near sea level (Warren, 2006).

Although microbial communities that form mats are pervasive within modern hypersaline environments (e.g. Dupraz et al., 2004; Glunk et al., 2011), discerning the biotic community that formed the Jurassic Todilto microbialites is extremely difficult because of the poor preservation of biotic microstructures (for example, Fig. 9). Although the composition of the Todilto microbial mat (for example, cyanobacteria, algae and sulphate-reducing bacteria) is unknown, the palaeoenvironmental and facies interpretations would not be largely different if, for example, these microbial mats were algal, cyanobacterial or a combination thereof. Overall, the microbial mats in the Todilto Formation may have been similar to hypersaline mat communities found in Qatar, which mediate the precipitation of carbonate and clay minerals as well as pyrite frambooids (Perri et al., 2018).

The reinterpretation of Todilto laminae as microbially influenced may, in part, explain the isotopic variations observed by Guhl (2004), who credited these variations to the degree of recrystallization of the micrite. The presence, composition or absence of a particular microbial community could account for isotopic variations within single facies and even along single laminae (for example, areas with a healthy algal mat or a poorly developed cyanobacterial mat). In contrast, if the laminae were varves, as posited by Anderson & Kirkland (1960), a consistent isotopic signature within a lamina would be expected. Further, the degradation and remineralization of these mat communities could alter the local
geochemistry early in diagenesis (e.g. Guy et al., 1993; Meyers, 1994; Schouten et al., 2001; Loyd et al., 2012).

The poorly developed microbial mat laminae in the deeper part of the interdune areas (for example, facies T2 and facies T1 at Sections O through to U) may be related to a stratified water column or lower oxygen levels in the depression; however, these could also be explained equally well by a photosynthetic microbial community that was healthier and better developed in the palaeobathymetric highs (i.e. the dune crest) than in the lows. Further geochemical work [for example, sulphur isotopes ($\delta^{34}$S) in sulphate from carbonate-associated sulphate precipitation; however, these could also be explained equally well by a photosynthetic microbial community] may elucidate the conditions involved in the microbial mat formation. From a taphonomic point of view, microbial mats may have also enhanced the fossil preservation observed in the Todilto Formation, as observed in other exceptional fossil deposits (e.g. Wilby et al., 1996).

It should be stressed that this study is focused upon a small portion of the overall upper Entrada and Todilto stratigraphic horizon, and other exposures of this stratigraphic horizon were not investigated. Given the nature of the basin structure, antecedent dune field and flooding event, the depositional environment and resulting stratigraphic record can be expected to be spatially heterogeneous. The Todilto facies development at Ghost Ranch is the result of the aeolian dunes being rapidly flooded and submerged. Similar facies development may be expected elsewhere within the deep parts of a topographic basin (e.g. Vincelette & Chittum, 1981). To the west, towards the Defiance Uplift, shoreline facies of the Todilto are recognized and dune palaeotopography is not apparent (John, 2000). The Todilto flooding event appears to have occurred during a time of extensive sabkha development within the Entrada (Fig. 2). The spatial extent of the dune field evident at Ghost Ranch is not known; where the dune field was not present, the Todilto may be represented by limestone and gypsum overlying Entrada sabkha strata. Finally, results of this study of the Todilto at Ghost Ranch differ from the analysis of Anderson & Kirkland (1960) and Kirkland et al. (1995), primarily in the interpretation of the origin and arrangement of the laminae, and the identification of facies and additional sedimentary structures. There is concurrence in envisioning a flooding event that submerged the dunes, followed by subsequent drying to interdune pools.

CONCLUSIONS

The primary conclusion of this work is the reinterpretation of the Todilto Formation at Ghost Ranch as a biolaminit or microbialite. This interpretation is based upon the nature of the laminae within the limestone. Black, organic-rich laminae, which form millimetre-scale stromatolites, tufts and other microbial structures, are interpreted as representing filamentous microbial mats with concentrations of trapped aeolian silt. Larger (centimetre-scale) stromatolites occur but are uncommon. Clotted-texture laminae show a fabric consistent with calcified extracellular polymeric substances secreted by well-developed mats. Most problematic is a spectrum of micritic laminae. Those with a peloidal texture, contain clotted-texture fragments or grade into clotted-texture laminae are the best candidates for biologically mediated precipitation within the microbial mat. Other micritic laminae arguably have origins as abiotic sediment or replacements of evaporites. Microbial mat development occurred initially over a marine-flooded, remnant dune topography, and persisted as the system evolved into an evaporitic environment of sabkha mudflats and interdune ponds.

Microbialite development and the spatial distribution of facies were strongly influenced by the antecedent dune topography. During the marine flooding, microbial mats, initially formed in normal marine waters, were best developed over the dune crestal area where light intensity was greatest. With system drying and emergence of the crestal area, gypsiferous mudflat and pond facies formed downslope and culminated in gypsum deposition in the interdune area. Tepee structures and a dissolution breccia developed in the emergent crestal area. Facies architecture in the subsurface controlled development of a dissolution front adjacent to the more impervious gypsum body. Section collapse, speculated to have occurred with dissolution of small gypsum nodules and gypsum laminae, as well as the downwarping of overlying strata and development of karst topography, occurred in association with the dissolution front.

New observations support the interpretation that the Todilto depositional environment began with a geologically instantaneous marine flooding of the Entrada dune field that was housed in a basin below sea level. Sediment-gravity flows that bury the remnant dune topography can be differentiated into those that originated with local slumping of the crestal area and lee slope,
and those that represent more distal transport along interdune corridors. Buried dune strata remained unconsolidated well into Todilto time with common upward sand injections. Remnant slope on the upper lee face promoted the downslope movement of Todilto strata.

Although far from resolved, the current best stratigraphic data indicate that the Todilto flooding event pre-dates the regional Curtis transgression and occurred during a time of extensive sabkha development within the Entrada sand sea. The overall Todilto environment may be best characterized as an embayment salina, in which the initial marine connection was lost and subsequent development occurred with marine subsurface and periodic surface flow.

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