Interpretation of mouth-bar and related lacustrine and fluvial sand bodies from the middle Green River Formation (Eocene), southern Uinta Basin, Utah

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ABSTRACT

The Uinta Basin of eastern Utah is an intermontane basin that contains an ~ 2-km-thick succession of mostly carbonate-rich mudrock assigned to the Eocene Green River Formation. In the southwest of the basin, along Nine Mile Canyon and its tributary canyons, the middle member of the Green River Formation contains numerous interbedded sandbodies. Previous researchers have interpreted these sand bodies variably as lacustrine deltaic mouth bars, terminal fluvial distributary bars, and various types of fluvial (delta plain/floodplain/braid plain) bar.

Using some modern western U.S. lakes as partial analogues, and taking into account the overall lacustrine basin context of a widely fluctuating, wave-influenced, alkaline-lake shoreline, we again interpret many of the sand bodies to be fluvial in origin. Several sand bodies both truncate and are capped by brown to red-maroon and variegated weak to non-calcareous mudstone with root and desiccation structures, indicating terrestrial deposition well away from the lake shoreline. Others display steep cutbanks from which noncalcareous, inclined heterolithic stratification laterally accreted as fluvial side bars. In contrast, utilizing helicopter-based Light Detection and Ranging (LiDAR), we investigated additional sandbodies that may be better examples of deltaic mouth bars. In contrast to the more commonly documented highstand progradational mouthbars of marine and open lake settings, these sand bodies are interpreted to have originated as late-lowstand or transgressive system tract fluvial channels that were then flooded and modified by waves following lake transgression. The examples illustrate that any large-scale sandy bed form present in the general vicinity of a closed basin's fluctuating lake shore may be expected to have formed

under more than one set of environmental conditions. A revised set of guidelines is therefore presented to aid in the interpretation of lacustrine deltaic mouth bars.

INTRODUCTION

Lake-basin sediments are prominent components of the Cenozoic to recent geologic record in the North American Cordillera of the western United States (Fig. 1A). Numerous lacustrine and alluvial depositional environments are consistently present, for example, profundal to encircling shoreface, shoreline, and backshore, and outside of these, fluvial, floodplain, and alluvial environments. Laterally, different process-related environments also may coexist, e.g., the "lake shore" region may consist of high- and low- energy carbonate and clastic beaches, organic-rich marsh/swamp and, where rivers empty into the lakes, various "river mouth" environments (Fig. 2). In rock outcrop or core, the recognition of deposits from such settings is purely interpretive: Geologists cannot observe, for example, a delta and its mouth bar; they observe the lithologies, mineralogies, sedimentary structures, geometries, etc., of the outcrop to interpret a delta and mouth bar. Ancient deltaic deposits are often only implied based on the interpretation of certain facies included in a number of deltaic models, although few process- or environmental- facies are uniquely diagnostic of deltas (e.g., Galloway and Hobday, 1996; Bhattacharya and Giosan, 2003; Olariu and Bhattacharya, 2006). Also, in vertical successions considered to be of alternating or cyclically terrestrial and subaqueous, what constitutes a possible mouth bar, levee, or distributary channel bar is particularly problematic. Local observation must be placed within the larger basin context before a succession can be validly interpreted (MacEachern et al. 1998; Bhattacharya and Giosan, 2003).

This contribution documents observations and provides interpretations of a somewhat controversial interval of sand bodies within an ~ 200-m-thick succession of exposed Green

River Formation strata in the southeastern Uinta Basin of Utah, previously described as the "Sunnyside delta interval" (e.g., Remy, 1992; Fig. 1B), and presents some modern day western U.S. lakes as partial analogues. The contribution interprets those parts of the studied succession that are fluvio-floodplain or lacustrine deltaic, and those sand bodies that represent fluvial bars or mouth bars (Ryder et al., 1976; Fouch et al., 1994, Keighley et al., 2003a; Taylor and Ritts, 2004; Keighley and Flint 2008; Schomacker et al., 2010; Moore et al., 2012). This is important partly because, at depth, the same interval contains sand bodies that form petroleum reservoirs in the basin (Morgan et al., 2003; Bereskin et al., 2004). Carefully considered interpretations of sandbody geometry and reservoir quality, which differ between fluvial bars and mouth bars, are required for accurate reservoir modeling. More broadly, it is also important to update the guidelines that can be used to aid interpretation of different sedimentary bar forms in other outcrop and subsurface fluvio-lacustrine settings.

GEOLOGIC SETTING

Basement-involved compressional tectonics of the Laramide orogeny commenced during Late Cretaceous time, partitioning an existing ramp-style foreland basin into various uplifted fault-bounded blocks, domes, and swells, partly along ancient structural trends (e.g., Dickinson et al., 1988; Constenius, 1996; Bump, 2003; DeCelles, 2004; Mulch et al., 2007). The Uinta Basin was one of the largest intermontane Laramide basins that developed between the uplifts. Its erosional remnant in eastern Utah and westernmost Colorado now crops out as a broad syncline with a gently dipping southern limb and more steeply dipping north limb (Fig. 1C). Several kilometers of relative uplift are recorded for the Uinta Mountains, which formed the northern margin of the basin (Osmond, 1964; Bradley, 1995). The southern margin of the original basin has been eroded, but it was likely delimited by more gently uplifted areas such as the Uncompahgre Uplift and San Rafael Swell (Dickinson et al., 1986).

Strata infilling the Uinta Basin are assigned to several units (Fig. 1B, C). The basal succession consists of conglomerate and sandstone included in either the Colton Formation or Wasatch Formation. Interfingering and cyclically onlapping these coarse clastic rocks, there are variably fissile fine-grained rocks ("shale" sensu lato) of the Green River Formation. The shale is predominantly carbonate rich (calcite, dolomite), with arenaceous and, up section, evaporitic (Na-carbonates, gypsum), tuffaceous, and organic-rich shale also locally abundant (e.g., Bradley, 1948; Milton and Eugster, 1959; Milton and Fahey, 1960; Williamson and Picard, 1974; Desborough, 1978; Brownfield et al., 2010; Keighley, 2015). Coarser-grained clastics of the Uinta Formation similarly interfinger, and gradually overlie westward, the Green River Formation in the north and east parts of the basin. Subdivisions of the Green River Formation have differed over the years (Ruble and Philp, 1998) and are rarely applicable basin-wide. The interval studied herein comprises part of the informal middle member of the Green River Formation ("middle Green River Formation") of Morgan et al. (2003) in the southwest of the basin, equivalent to the lower Douglas Creek Member in the east part of the Uinta Basin and southwest Piceance Creek Basin (e.g. Tänavsuu-Milkeviciene et al., 2017). The member is defined at its base by the top of the "carbonate marker unit" of Ryder et al. (1976), and it consists of greenish gray, gray, and black shale, but with several thick tongues of sandstone, thin limestone, and oil shale. The studied interval was termed the "delta facies" (Bradley, 1931) and subsequently the "Sunnyside delta interval" (Remy, 1992). However, these terms should be abandoned in the stratigraphic literature because "delta" is an interpretive term, and only one of several depositional environments interpreted as present within the succession. Within the middle Green River Formation and ~ 150 m above its base, a distinct and laterally continuous ostracodal grainstone and oncoidal rudstone interval has variously been identified as the D (Jacob, 1969; Remy, 1992), MGR3 (Morgan et al., 2003), and M1 marker (Keighley et al., 2002; 2003a, 2003b; Keighley and Flint, 2008). The contact with the base of the upper member is identified by the Mahogany oil shale zone, a basinwide interval that can be correlated across to the northeast part of the basin where it also forms the R7 interval (Cashion and Donnell, 1972), the richest of eight oil-shale dominated intervals in the succession.

The presence of basinwide, typically dolomitic, kerogenous oil shale is interpreted to represent predominantly meromictic, alkaline, anoxic, deep-water deposition that occurred when lakes in both the Uinta and adjacent Piceance Creek Basins expanded and merged as "Lake Uinta", culminating in deposition of the Mahogany oil shale zone (e.g., Cashion, 1967; Desborough, 1978; Johnson, 1981; Ruble et al., 2001; Keighley et al., 2003b; Keighley and Flint, 2008; Smith et al., 2008; Birgenheier and Vanden Berg, 2011). Carbonate grainstone intervals that interfinger with red beds are considered to partly represent open-lake highstands where spillovers were into adjacent closed (evaporitic) lakes (Keighley et al., 2003a; Smith et al., 2008). The red beds themselves, which have lower carbonate contents, indicate prevalent subaerial conditions and pedogenesis (Walker, 1967; Ryder et al., 1976;

Keighley et al., 2003a; Pearce et al., 2008), with the basin lake shoreline having shifted basinward to the vicinity of the basin's depocenter (Fig. 1C).

Collectively, lithofacies associations for the middle Green River Formation have been interpreted as representing both a "balanced-fill" and "overfilled" lake-basin succession (Carroll and Bohacs, 1999; Bohacs et al., 2000; Keighley et al., 2003a; Smith et al., 2008). At higher resolution, however, a marked cyclicity in the high-lake-level carbonate and oil shale and low-lake-level, calcite-poor fluvial sandstone and floodplain mudstone can still be interpreted from outcrop of both the middle and upper Green River Formation (Ryder et al., 1976; Fouch et al., 1994; Keighley et al., 2003a; Pearce et al., 2008; Keighley and Flint, 2008; Keighley 2015). Accordingly, there are some similarities to the "underfilled", and possibly contemporaneous, Wilkins Peak Member of the Green River Formation in Wyoming, salt-pan facies excepted (Smith et al., 2015). Geochronologic measurements, coupled with stable isotope and paleontological studies in adjacent basins, also correlate the middle Green River Formation of the Uinta Basin with the peak of, and initial decline from, the Early Eocene climatic optimum, and a warm, seasonally dry setting (e.g., Doebbart et al., 2010; Smith and Carroll, 2015). Paleogeographic reconstructions indicate that the Laramide basins have moved only slightly in latitude over the subsequent ~50 m.y. (Thrasher and Sloan, 2009; Hyland and Sheldon, 2013).

Notwithstanding the differing tectonic setting, and the global change from Eocene greenhouse to Quaternary icehouse climates, the current lakes of the Great Basin provide good local analogues for Laramide lithofacies successions (e.g., Bohacs et al., 2000; Keighley et al., 2003a; Pietras et al., 2003). Both were and are responding to monsoonal conditions in continental midlatitudes, and switching between closed and open, and separate to merged, lakes. Lateral and vertical facies associations in the Laramide lakes are considered comparable to the late Pleistocene, merged, Lake Bonneville and to the current spectrum of open, fresh-water lakes (e.g., Utah Lake), to closed, hypersaline but perennial lakes (e.g., Great Salt Lake) to salt-pans (e.g., Bonneville Flats).

STUDY AREA

This contribution focuses on the often-studied middle Green River Formation along Nine Mile Canyon and its side canyons in the southwest of the Uinta Basin (Fig. 1D). Here, an ~ 200-m-thick succession, dipping gently to the northeast, consists of numerous sandstone beds interbedded with red and gray shale, limestone (including the D, MGR3, or M1 marker) and rare oil shale (M2 of Keighley et al., 2002, 2003a) exposed across many kilometers of variable-quality outcrop. Earlier studies (Fig. 1D) have been focused around the junction of Nine Mile and Argyle Canyons, where the lower part of the succession is best exposed near the base of the canyon walls. This contribution expands the study area to include parts of Nine Mile Canyon further east where the lower part of the succession has entered the subcrop, leaving beds from higher in the succession well exposed lower in the canyon walls. Data availability in the field area is variable. Steeper topography equates to better exposure, as does a south-facing (= less vegetated) aspect. Outcrop is also better exposed toward the base of the canyons hence, with the gentle dip of the beds, different parts of the succession confer greater information at different locations along the canyon. In general, the presence of thick, blocky weathering sandstone promotes cliff development, steeper slopes, and better exposure, including exposure of intervening fine-grained units.

METHODOLOGY

This paper incorporates earlier work that focused on over 25 km^2 of outcrop midway along Nine Mile Canyon, where lithology, thickness, paleoflows, and lateral extent were determined directly from outcrop by cm-scale sedimentological logging (12 detailed logs), measuring stick, global positioning system (GPS), and pacing (or where inaccessible, by estimation) techniques and plotted, together with the markers, on 1:24,000 scale maps and photomontages (Keighley et al. 2002, 2003a,b; Keighley and Flint, 2008; Fig. 1D). The current study also incorporates observations from further east along Nine Mile Canyon (e.g., Spinnangr, 2014). Technological advances mean that, for these more eastern outcrops only, thicknesses and lateral extents have been more accurately measured and correlations improved by employing Light Detection and Ranging (LiDAR) methods. The LiDAR device consists of a laser that measures the distance and angles from the scanner to a reflective surface. When mounted onto a helicopter, the position of the scanner is geo-referenced by a global navigation satellite system antenna also attached to the helicopter, together with an internal measurement unit that records the helicopter orientation, and a digital camera (Rittersbacher et al., 2013). This method provides advantages over ground based LiDAR by being more mobile, allowing for easy avoidance of scan shadows caused by ledges and back-stepping benches, and by the rapid scanning of extensive outcrops in a short time frame. The collected raw point cloud and photographic data are then segmented into modules ~ 2 km long; the density of the point cloud was reduced and smoothed; a computerized 3D surface was created by triangulating between the points in the point cloud using the software PolyWorks (version 11); and the digital photographs were overlain on this surface. Interpretations used the software LIME version 0.5.3.

Field observations and measurements have resulted in the recognition of over 30 different, hierarchical lithofacies associations (Table 1). Four mudstone samples were analyzed by semi-quantitative whole rock X-ray diffraction (XRD), with an additional three samples analyzed quantitatively and on the < 2μm fraction to identify clay mineralogy. Analyses were undertaken by Activation Laboratories, Ancaster, Ontario, using standard techniques (see www.actlabs.com for details). Carbonate lithofacies were first identified by HCl-testing in the field, and subsequently confirmed by thin section and bulk inorganic geochemistry analysis (Pearce et al., 2008). Distinctive carbonate successions (the marker beds M1 to M11 discussed below) and > 3m-thick siliciclastic sandbodies were plotted onto either 1:24,000 maps and photomontages (western study area), or the LiDAR images (eastern study area). Regardless of plotting method, measurements of sand-body widths remain estimates where one or both margins of a sand body were not clearly exposed, due to either deteriorating outcrop, the end of an outcrop section, or truncation by another sand body.

OBSERVATIONS

The ~200-m-thick succession under investigation contains several distinctive, laterally persistent carbonate units, which are identified as markers M1 to M11 (Table 2; note that these markers are purely observational constructs identified without any preconceptions, and as such they do not always correspond to the subsequent interpretations of environmental facies or sequence stratigraphic intervals or boundaries; Keighley et al., 2002, 2003a). Originally identified between Sheep and Gate Canyons (Keighley et al., 2002), the markers are now recognized further east to beyond Daddy Canyon (Figs 3 and 4), except where the beds have dipped into the subcrop. Despite the expanded field area, there are still only minor variations in the thicknesses and lithologies of the marker beds and intervening strata.

The ~200-m-thick succession also can be divided into alternating intervals containing dominantly gray or red mudstone. Intervals containing mostly gray, calcareous mudstone, laterally persistent but thin, tabular, variably calcareous very-fine to fine-grained sandstone, and limestone (including most of the marker beds) are on average ~10 m thick (e.g., M1—M2, Table 3). Intervals of variably red and gray, sparsely to noncalcareous mudstone with laterally persistent, thin, and tabular very-fine-grained sandstone range from ~ 10 to 30 m thick (e.g., M2—M3). Downcutting fine to medium-grained sandbodies are present in all intervals but are uncommon in the gray calcareous intervals where they comprise less than 15% net:gross of an interval's cross-sectional area in the west (Keighley and Flint, 2008). Sandstone also may be a rare component (5% net:gross) in some of the red, sparsely calcareous mudstone intervals, where they form only isolated lenticular sand bodies (e.g., M4—M5). In other red mudstone intervals, lateral and vertical truncation of sand bodies by other sand bodies can result in extensive sheet sandstone preservation and ~ 35% net:gross (e.g., M2—M3).

Looking at some of the sand-body occurrences in more detail, the gray calcareous interval M1—M2, which is best exposed around the junction of Nine Mile and Argyle Canyons (Fig. 5A - D), contains only one major sand body that attains a thickness of up to 6 m and a measured width of > 275 m (Fig. 6 - see also Plate DR1). This noncalcareous to sparsely calcareous, fine to medium grained sandstone (Figs. 5B, 5C, 5E) sharply and steeply cuts down through fine-grained strata to locally sit on a stromatolitic limestone that forms part of the D Marker (M1). At the western truncating margin of the sand body (Fig. 5B), inclined heterolithic stratification and coset boundaries (Thomas et al., 1987; McKee and Weir, 1953) dip E-NE, but paleoflow from asymmetric (unidirectional) cross laminations and shallow trough cross-beds is to the W-NW. In contrast, southeast of the junction of the two canyons, the central part of the sand body has consistent orientation of set boundaries and paleoflow from cross-strata, both of which are toward the W-NW (Fig. 5F, see also Keighley et al., 2003a fig. 7D, E). Where the canyon turns east and northeast ("Argyle Point", Fig. 6), less-well-exposed inclined heterolithic stratification dips east and may be onlapped by flatter bedded, mud-dominated heterolithic strata. Symmetrical cross-lamination (bundled or opposing cross-lamination) is present only in isolated cosets. The few other scattered minor sand bodies in this interval include, in the southwest, lenticules of shallow inclined heterolithic stratification of very-fine to fine sand with silty drapes. Of particular note, the truncated strata include tabular very-fine grained calcareous mudstone and calcareous sandstone containing symmetrical and, or, asymmetrical cross laminations. Rarely, non-calcareous, mottled, reddish brown, or purplish gray mudstone displays evidence of root structures (Figs. 5G, 5H, and 5I).

At the same locations and laterally eastward, the overlying interval M2—M3 contains many laterally truncating and stacked, sharp-based, mostly non-calcareous sandstone bodies (Fig. 5A, C, D). Although widespread (~35% net:gross), these amalgameted sand bodies only very locally truncate the M2 oil shale and incise into underlying sandbodies of M1—M2 (Figs. 5D, 6B; Keighley and Flint, 2008). Sets and cosets of asymmetric cross-lamination and low-angle cross-beds within the sand bodies may be over two meters thick. Amalgamated sand bodies are more common in the lower half of the interval; lenticular sandstone with inclined heterolithic stratification is more common up section. South of the junction of Argyle and Nine Mile Canyons, one broadly lenticular sand body truncates red mudstone and has a convex upward geometry that, on the northwest side, is onlapped by dark gray brown non-calcareous mudstone (Fig. 5J, Table 4). Laterally, red-maroon, commonly non-calcareous mudstone with sporadic reduced horizons, subvertical downward tapering structures, branching root-like structures, microconcretions, and rare polygonal cracks, is incised by lenticular sand bodies (Figs. 5K, 5L, 5M also Keighley et al., 2003a, fig. 6F). Paleocurrent data in this interval are predominantly toward the NW but some inclined heterolithic sandstone shows eastern flow directions. The widespread occurrence and amalgamation of sand bodies makes it difficult to correlate and map individual sand bodies between outcrops (Fig. 6B).

The gray calcareous interval M6—M7 and overlying red-gray sparsely calcareous interval M7—M8 are locally well exposed in the west (Plates DR1 and DR2 [see footnote 1]; Keighley et 281 al., 2003a, their figures 6c and 6d), but are best studied east of Blind Canyon where they crop out near the foot of the canyon. Near the base of this outcrop, noncalcareous and sparsely calcareous, very-fine to fine-grained gray sandstone contains abundant low-angle cross-strata and asymmetric cross-lamination. Up section, sporadic concretions, cracks, and root-like structures are found in variegated gray-green and red mudstone (Figs 4 and 7A) that is interbedded with more calcareous very-fine grained sandstone (Fig. 7B; unit 1 in Fig. 8A, B). These grade up into interbedded thin gray mudstone and calcareous sandstone containing uncommon symmetrical cross-lamination, and small-scale cross-strata including swales and hummocks. One gray sandstone grades up into a carbonate grainstone with abundant shelly and oolitic particles and is capped by a dark gray, locally fossiliferous micrite (M6 marker, unit 2). There is then a return to interbedded red and gray mudstone and very fine-grained sandstone (unit 3 in Fig. 8C) that is shallowly truncated by similarly fine-grained lenses containing inclined heterolithic stratification (unit #4A). This unit is mostly inaccessible, but to the northwest, the uppermost sandstone lens is capped by a more tabular, highly calcareous bed (unit 4B). These latter three units and the next thin gray, locally red, mudstone (unit 5) are locally truncated by an otherwise sheet-like, variably calcareous, fine-grained sandstone that can be over 6 m thick (unit 6). Although this unit is widely soft-sediment deformed, scour-and-fill features with northeast paleoflow, and sets of asymmetric cross-laminations (locally soft-sediment deformed to a vertical orientation) are present. The sandstone is at a stratigraphic position where, further west, other sandstone units truncate a 1 m thick carbonate grainstone (M7 marker in Fig. 3). Overlying this unit is another succession of interbedded red and gray mudstone and very-fine grained sandstone (unit 7) that is locally truncated by laterally amalgamated lenticular sandstone with inclined heterolithic stratification (unit #8): the M7—M8 interval.

The M8—M9 interval is internally more complex than the underlying units. The interbedded gray calcareous mudstone and oolite- and ostracode-bearing micritic carbonate succession (M8 marker) is ~ 5 m thick at Argyle Point, but its equivalent NE of Trail Canyon is an ~ 10-m-thick calcareous siliciclastic succession (Keighley et al. 2003a, their fig. 9c, sections 2 and 11). No marker bed caps this gray calcareous succession and, southwestward along Sheep Canyon, M8 is progressively truncated by sandbodies with a broad sheet-like geometry locally. The upper part of M8—M9 is another red-gray sparsely calcareous succession where the predominantly gray, uncommonly red, mudstone is truncated by only a few lenticular sand bodies.

Between Dry and Daddy canyons, the complexity of the lower M8—M9 interval is well exposed (Figs. 8D, E). At the base of the canyon wall, there is a similar succession of strata to that described above for M6—M7 and M7—M8. The latter interval is capped with interbedded red and gray mudstone and very-fine grained sandstone (Fig. 7C; unit A in Fig. 8E), locally truncated by a 2 – 3-m-thick, noncalcareous to sparsely calcareous sandstone containing shallow-dipping inclined heterolithic stratification (unit B). This sandstone is itself shallowly truncated by very fine-grained calcareous sandstone and mudstone of similar thickness and containing shallow dipping inclined heterolithic sandstone with an overall fining upward character (unit C1). Laterally, the tops of the inclined heterolithic stratification are overlain by tabular calcareous sandstone beds up to 1 m thick (unit C2). Up to three C1 units were observed successively truncating each other laterally, but with the paired C2 units sitting atop older units (Figs. 8F, 8G). Rarely, shell fragments, ooids, indistinct cross lamination, and vertical trace fossils (cf. *Skolithos*, fugichnia) are present in these paired units. A capping gray micrite locally coarsens into an oolitic or ostracodal grainstone (unit C3). These calcareous beds collectively attain a thickness of 8 - 10 m, and are considered to be equivalent of the M8 marker. They are very similar to the succession previously logged east of Trail Canyon (section S11, Keighley et al., 2002, 2003a). Once again, the succeeding strata of the M8—M9 interval consist of interbedded red and gray mudstone and very fine-grained sandstone (Unit D), locally truncated either by noncalcareous to sparsely calcareous, lenticular, very fine-grained sandstone (unit #1) or variably calcareous very-fine to fine-grained sandstone (unit E2). The latter can truncate over 10 m of section to vertically amalgamate with ~ 5 m-thick sandstone of the underlying M7—M8 interval (Figs. 8D and 8E).

INTERPRETATIONS AND DISCUSSION

Most sandstone of the middle Green River Formation in Nine Mile Canyon is interpreted to have formed in various fluvio-deltaic subenvironments present in highly seasonal (monsoonal) climatic settings (e.g., Bradley, 1931; Ryder et al., 1976; Remy, 1992; Fouch et al., 1994; Keighley et al., 2002, 2003a, 2003b; Morgan et al., 2003; Keighley and Flint, 2008; Pearce et al., 2008; Plink-Björklund, 2015). Keighley et al. (2002, table 1 in Keighley et al., 2003a) also interpreted the presence of minor deltaic mouth-bar lithofacies associations, but more widespread mouth-bar deposits have since been interpreted to be present by Taylor and Ritts (2004), Schomacker et al. (2010), Moore et al. (2012), and also, east of Nine Mile Canyon, by Rosenberg et al. (2015). The interpretation of fluvial and deltaic mouth-bar sand bodies is the focus of the following discussion.

Fluvio-lacustrine delta settings

A fluvial system can be divided into a tributary and downstream trunk system, where channels tend to truncate progressively wetter, flatter, floodplain deposits; a distributary system may extend further downflow from a distributary apex. The largest channel of a fluvial system, the trunk channel (Fig. 2; Olariu and Bhattacharya, 2006), may be recognizable all the way to the river terminus at a marine or lake shoreline. In marine settings, a delta is defined by a convex-basinward deviation (or bulge, Bhattacharya and Giosan, 2003) of the shoreline in the region of the active distributary outlets. In lake basins, the term has been more loosely applied to cover any river terminus that periodically has active distributaries. This is reasonable given that many lakes sit within elongate, fault- or bedrock-bounded constrictions and their axial rivers will empty out at locations along an estuary-like concave shoreline. Also, where lakes are endorheic and experience rapid changes in water level, frequent fluvial adjustments to the changing base level can cause temporary loss of a bulge or cause outlet patterns to change from distributary to single mouth. For example, the Walker River outlet at Walker Lake has been termed a delta (e.g., Blair and McPherson, 1994). Satellite images from 1994 and 2014 indicate no bulge and at most two active distributaries, but at intervening times a bulge and many small distributaries were present (Fig. 9A).

Landward of a delta's shoreline is the delta-top environment, potentially with its own interdistributary lake deposits or sand bodies associated with adjacent crevasses, levees, active and abandoned river channels. Where the shoreline transects a river mouth, a convex-upward sandbody located closely downflow can be termed a "mouth bar" (Fig. 2). Wright (1977) reviewed how different mouth-bar geometries relate to differing combinations of active forces. On steep slopes, inertia may dominate producing lunate bars with classic Gilbert delta foresets. On shallow shoreface slopes with common high outflow velocities and bed-load transport, bed friction may form midground bars enclosed by bifurcating subaqueous channels. Where there are strong density contrasts, high depth-width ratios to the river mouths, and fine-grained sediment loads, buoyancy effects (including salt-wedge intrusion into the river channel and hypopycnal flow of the river water) support the development of straight, digitate subaqueous levees and lunate mouth bars. Buoyancy effects can be precluded by wave activity in the standing water body (ocean or lake), which results in narrow crescentic mouth bars and broad, shoal-like subaqueous levees with swash bars. In wave-influenced deltas, the mouth bars may be accompanied by, or transition into, barrier bars, barrier islands, or spits sheltering their own lagoons and smaller-scale bayhead delta (Fig. 2; Bhattacharya and Giosan, 2003). Although these bars generally form nonerosively based, coarsening upward sand bodies, erosively based sands that fine upward or show no discernible vertical trend may exist on low-gradient floors where flood discharge first scours the bed on which the sediment then accumulates (e.g., Fielding et al., 2005).

In lakes, wave action would only dominate where a large fetch exists, the river outlets are located on the down-fetch side, and the shoreface slope is not too flat and broad to dissipate wave energy. In midlatitudes, where westerly winds predominate, wave action could be important on unsheltered northeast to southeast lake shores. Otherwise fluvial processes should prevail. In planform, deltas dominated by fluvial processes can be modelled as a multiple distributary lobate-type feature building out into shallow water, versus an elongate-, or birds-foot-, type feature with few distributaries building out into deep water (Olariu and Bhattacharya, 2006). Along shores that are wave influenced, there will be a complete spectrum from river-dominated deltas with minor reworking to primarily a strand-plain system with minor delta promontories. Across this spectrum, deltas may symmetric, with low net sediment drift at the outlet, or asymmetric with greater longshore movement (Bhattacharya and Giosan, 2003).

Identifying a delta, or a mouth bar or associated subaqueous levee, can be problematic when assessing Holocene and older deposits. Without high-resolution 3-D seismic, the mapping of a shoreline bulge, and hence distinction of a delta from, for instance, a strand plain, is difficult. Few processes and environments are unique to deltas (Galloway and Hobday, 1996), and so there are few unique sedimentologic characteristics. Mouth bars may grade, or be variably reworked, into beach and shoreface deposits in wave-influenced deltas, and a significant proportion of longshore-drift-derived sediment can be incorporated. These factors also make the distinction between wave-formed shorefaces and wave-dominated deltas difficult (Bhattacharya and Giosan, 2003). Bar-form sand bodies within, or on the channel margin of rivers at base level (sea- or lake-level) may exhibit some mouth-bar characteristics such as wave ripples, or mineral precipitates that reflect the chemistry of the standing water body (lake-water or sea-water). At any point in time, tongues of the standing water (e.g., a salt wedge intrusion) may extend considerable distances upstream dependent on fluvial discharge and channel gradient. The maximum upstream limit of the standing water's influence would be the elevation at which the base of a fluvial channel intersects the base level of the lake, but the actual limit will vary dependent on the annual discharge regime: Flashy discharge may indeed permit a standing water influence to invade upstream for most of the year when fluvial flow is minimal.

Varying fluvial and groundwater discharge usually also results in a change of lake level. For closed lakes, a changing lake level is the standard situation, and, where contained on a gently sloped basin floor, minor changes in lake level accompany extensive regression and transgression (e.g., Kroonenberg et al., 1997). For example, the floor of the present-day Great Salt Lake has in places an average gradient of ~ 0.4 m/km (Keighley et al., 2003b). Historic low-lake elevations (AD 1963, 2015) of ~1277 m and historic highs (AD 1873, 1987) of ~1284 m translate to localized shoreline migrations of over 17 km. If shorelines regress, mouth-bar deposits may become stranded onshore, and over time, they may be gradually reworked or even removed by eolian or pedogenic processes, or by the lateral migration, lengthening, or downcutting and terracing of river channels (e.g., Walker Lake and Pyramid Lake deltas, Fig. 9A; Blair and McPherson, 1994; Adams, 2012). In contrast, extended periods of increased discharge and higher lake level will cause existing fluvial channels to be flooded by transgression, leaving in-channel bars submerged lakeward of the shoreline, potentially in a mouth-bar location. However, only if the deposits are modified over time by lake-related processes might the sandbody evolve to contain structures indicative of mouth bars although, as previously noted, they may further evolve into other shoreface deposits. Accordingly, the preservation potential of mouth bars can be quite low, requiring plenty of accommodation space and frequent channel avulsion (Hornung and Hinderer, 2011; Li et al., 2011). Preservation may be achieved where increasing discharge and base-level rise lead to accelerated vertical growth and onlapping of sedimentary bodies (Kroonenberg et al., 1997). Thus, in balanced-fill lake basins mouth bars are associated with the formation and preservation of thick transgressive systems tracts (Keighley et al., 2003a; Hornung and Hinderer, 2011) rather than with highstand and falling stage tracts common in marine settings. However, we are not aware of modern analogues detailing the evolution and internal geometry of modern mouth bars in both dominantly transgressive and wave-influenced settings (although exclusively wave-influenced settings have been studied by e.g., Bhattacharya and Giosan, 2003; Smith et al., 2005; Evans and Clark, 2011; see also discussion in Anthony, 2015). Finally, as indicated in the discussion above, the grouping of a series of sand bodies in outcrop simply as a delta succession (e.g., the Sunnyside delta) may be an oversimplification. There may be spatial and temporal transitions between delta and strand plain, and between different types of delta.

Depositional cyclicity in Nine Mile Canyon

General consensus among workers suggests that the succession of middle Green River Formation strata along Nine Mile Canyon represents cycles of lacustrine and terrestrial deposition (Ryder et al., 1976; Remy, 1992; Fouch et al., 1994; Keighley et al., 2002, 2003a, b; Morgan et al., 2003; Keighley and Flint, 2008; Pearce et al., 2008). Gray calcareous mud and sandstone, and a variety of carbonate strata including those comprising the marker beds (Table 2), recur in intervals ~ 10 m thick, such as M1—M2, M3—M4, M6—M7, and the lower parts of the succeeding intervals. These intervals indicate the prevalence at this location of the significant carbonate factory that was Lake Uinta or its component smaller basin alkaline lakes (Bradley, 1948; Milton and Eugster, 1959; Milton and Fahey, 1960; Williamson and Picard, 1974). Symmetric cross laminated (wave ripple) calcareous sandstone and carbonate grainstone lithofacies, also incorporating coated grains (ooids and oncoids), may imply wave-processes were of importance on what would have been the southeast shore of the lake (various aqueous high energy sandstone (SA) and limestone (LA) beach, bar and shoal lithofacies associations, Table 1). We envision a setting intermediate between river- and wave-dominated (similar to the marine Brazos Delta of Rodriguez et al., 2000 or Burdekin delta of Fielding et al., 2005). Algal stromatolites may further indicate shallow water, photic conditions (Bradley, 1948; Ryder et al., 1976). Dark gray fossiliferous micrite and dolomitic oil shale would have been deposited as quieter water lacustrine strata (the LM lithofacies associations, Table 1): variation in organic geochemistry indicating shallow, lagoonal, and deep, profundal settings respectively (Keighley et al., 2003a). However, the presence of fine-grained terrestrial (FT) lithofacies associations (Table 1) defined by rare red mudstone and gray mudstone with root-like structures, and polygonal cracks from desiccation, indicate periods of floodplain emergence within these lacustrine-dominated intervals.

Gray to red, sparsely to noncalcareous mudstone and laterally persistent, thin and tabular very-fine-grained sandstone recur in intervals ~ 10 to 30 m thick (such as M2—M3, M4—M6, M7—M8, and the upper parts of the succeeding intervals). Here, the common and typically noncalcareous red or variegated mudstone, associated polygonal cracks, root-like, blocky ped-like, and concretionary structures indicate that the intervals are from a well drained, floodplain dominated setting away from the (less extensive) alkaline lake (Keighley et al., 2003a, Pearce et al., 2008, and cf. Pietras and Carroll, 2006 in the Green River Formation of Wyoming). However, some beds of micritic, oolitic or fossiliferous limestone, such as the M5 marker, indicate the occasional southward expansion of the lake within these intervals. The presence of subaqueous deposits also indicates that when the shoreline was located around the Nine Mile Canyon area the lake was endorheic: Because sublacustrine deposits are noted, transgression could continue and lake level could rise further before reaching its spill point. Keighley et al. (2003a) suggested the base of the carbonate shoaling successions (parasequences) could represent times when the lake was at a spill point, with dolomitic oil shale deposited when the lake was even deeper after merger with adjacent lakes (Fig. 10).

Local observation and interpretation of the thicker lenticular and sharp-based sheet-like sand bodies must be made within this larger basin context (MacEachern et al. 1998; Bhattacharya and Giosan, 2003). Initially, where the sandbodies both truncate and are capped by FT lithofacies, particularly red mudstones with soil features indicative of a well-drained floodplain, Occam's Razor is applied and the default hypothesis is that the sandbody is entirely fluvial in origin: delta or braid-plain or alluvial floodplain (Keighley et al., 2003a) or upper mud flat (Ryder et al., 1976). Predominant paleoflow directions in the sand bodies further indicate the fluvial system drained northward (or NW, NE), and any lake shore lay in that direction. Where the sand bodies both truncate and are capped by the various SA, LX, or LM lithofacies associations that are most diagnostic of subaqueous deposition, it is equally reasonable for the default to be that the sandbodies are lacustrine influenced, for example the mouth bar, subaqueous levee, and fluvial distributary environments summarized above.

More intuitively, where the sand bodies truncate lacustrine lithofacies but are capped by terrestrial lithofacies, shoreline regression is inferred. In this situation and away from the margin of the lake-basin floor, preservation of delta deposits is unlikely through reworking and erosion (Figs. 9A and B). The default hypothesis would be fluvial, with a regressive surface best placed at the channel base. Mouth bars might be found at the furthermost (northernmost) progradation of the fluvial system into the lake (fall to rise turnaround of Hornung and Hinderer, 2011). This would approximate to the time-stratigraphic position of where lenticular, inclined heterolithic stratification-dominated sandbodies become prominant up section in the floodplain-dominated intervals along Nine Mile Canyon. Subsequent to this turnaround, fluvial deposits may be subject to flooding, and reworking by lacustrine processes as the shoreline transgresses; sand bodies are truncating terrestrial lithofacies but are being capped by lacustrine lithofacies. The flooding surface may drape or be within the sand body. The surface is unlikely to be at the very base of the sand body since, once the channel is submerged beneath the lake, any sediment already deposited is likely to be preserved because the fluvial flow is retarded and downcutting erosive potential is lost (Wright, 1977). Preservation of delta deposits is enhanced. In all cases, further detailed lithofacies investigation of the individual sandbodies is required to confirm or modify an initial hypothesis.

M1—M2 sandbodies: delta distributary and trunk channels

The M1 marker has been interpreted as a set of up to three lacustrine carbonate parasequences sitting on a regional flooding surface (Keighley et al., 2003a). Parasequences commonly culminate in a coarse-grained carbonate bed indicating a down-fetch storm-wave dominated upper shoreface to beach deposit (Ryder et al., 1976; Keighley et al., 2003a). The M2 marker is considered a regional maximum flooding surface (MFS) that followed further lake deepening and merger (Keighley et al., 2003a). A spillpoint highstand for a single nested basin is equivalent to a stillstand when nested basins merge. Between these markers the gray fine-grained clastics and tabular, coarsening upward, calcareous siltstone to wave-rippled fine-grained sandstone suggest prograding lacustrine shoreface deposits. Current-rippled crevasse-splay deposits may also be present. Rare root and ped-like structures indicate only local exposure in what remained a high watertable, paludal setting.

The main convex-up sandbody of this interval (and many of the sandbodies of the succeeding intervals) contains numerous sedimentary structures that are consistent with fluvial deposits from monsoonal or highly seasonal settings (Plink-Björklund, 2015), although those structures indicating waning flow equally could be considered to have formed due to flow deceleration at the river mouth. Furthermore, southeast of the junction of Nine-Mile and Argyle canyons, the main sandbody contains coset boundaries and IHS that dip in a similar NW direction to the cross strata. From this outcrop alone (Fig. 5E), the sandbody could represent a downflow-accreting mid-channel fluvial bar or mid-ground mouth bar. However, viewing the outcrop more broadly, the bar likely forms only part of the fill of a steeply incising, ~ 5 m deep channel that in map view shows very low sinuosity for almost 3 km (Figs. 5, 6A). On the western margin of the channel, steep IHS indicate lateral accretion (Fig. 5B), but the mapped straightness of the channel suggests a side bar rather than a point bar. Petrographic analysis shows that the sandstone of the IHS close to the channel margin has minimal carbonate cement, only weathered, reworked calcite grains (Fig. 5D). The side bar is best interpreted as having initially formed while confined within a low-sinuosity fluvial trunk channel (Keighley et al., 2003a; Keighley and Flint 2008), with the lack of carbonate cement suggesting a position landward of the lacustrine carbonate factory and its saline wedge. The low-sinuosity fluvial interpretation is also consistent with progradation and lengthening of a fluvial system (Schumm, 1993; Keighley et al., 2003a; Keighley and Flint, 2008) that cut across the preceding M1 shoreface deposits.

Isolated lenticular sand bodies are small-scale features, typically <4 m thick and <100 m wide, consisting of inclined heterolithic stratification of shallow-dipping, (very) fine-grained sandstone to mudstone (Fig. 5G). In some adjacent lenses, the IHS have dips in opposing directions suggesting the outrop cuts across a meander loop (Fig. 6A). They have previously been interpreted as distributary channels of the main trunk (Keighley et al., 2003a; Keighley and Flint, 2008), and can be further explained as part of the network of shallow terminal channels that form during delta progradation during stillstand periods (Olariu and Bhattacharya, 2006).

Sandstone lenses comprising IHS that are immediately below a flooding surface could represent the increasing sinuosity of fluvial channels associated with early base level rise (Schumm, 1993). Similarly, IHS cosets near the top and downflow end of the main convex-up sandbody (Fig. 5) locally contain wave ripples and a carbonate-cement mineralogy, suggesting commencement of a lacustrine influence. Around Argyle Point the flat-lying, onlapping, mud-dominated heterolithic strata may have been deposited as the lake subsequently transgressed and flooded the trunk channel (see supplementary data: Plate 1, around UT18-19).

Schomacker et al. (2010: their S-15-1, 1000-1400 m) alternatively interpreted the onlapping part of the sandbody to show the transition between an erosively and depositionally based mouth bar. The rest of the main sandbody in this interval (their stage 1) was described simply as a symmetrical mouth bar, and they placed the shoreline somewhere south of Nine Mile Canyon. In concluding that the sandbody should be reinterpreted as a deltaic mouth bar rather than a fluvial point bar, their list of "diagnostic" criteria included the presence of a convex upward bedding pattern, basinward paleocurrent orientation parallel to the migration direction of the sandbodies, and an absence of cutbank terminations. However, these recorded criteria have long been documented on low sinuosity floodplain rivers (e.g., Coleman, 1969). Also, Schomacker et al. (2010) did not examine the sandbody in outcrop west of Argyle Canyon, and so did not identify the ~ 5 m deep cut bank with laterally accreting side bar, the presence of non-calcareous sandstone within the sandbody, nor the presence of reddened, rooted horizons (Fig. 5A, B, C, E, G, H, I).

M2—M3 sandbodies: floodplain (sandy braid plain) channels

The deep lake M2 oil shale is overlain by carbonate shoal deposits and subsequently by numerous, often well-developed, noncalcareous red-bed paleosols, thin tabular/sheetlike sandstone, and many extensive, laterally and vertically amalgamated sand bodies. The thin (< 1.5 m thick), sheet-like sandstone interbedded with the paleosols are interpreted to be sheetflood or crevasse splay deposits. The many lateral and basal truncations of amalgamated sandbodies impair their detailed interpretation but, where they truncate mudstone to form cut banks and are overlain by red-beds, they are interpreted to be fluvial deposits on a well-drained floodplain or braidplain. Following Keighley et al. (2003a), the lower part of the succession reflects a dramatic switch from a deep, more extensive lake (that likely had merged with lakes in adjacent basins) to a closed lake in the Uinta Basin. Closed basin conditions likely coincided with a drier climate, less annual run-off, lower lake level, and thus less accommodation space. Any mouth bars and terminal distributary channels associated with the shoaling deposits and the northward regression of the lakeshore across the study location had little preservation potential (negative shoreline trajectory of Li et al., 2011). Upflow, rivers would have been incising through unconsolidated basin margin lake-highstand terraces that were readily available for sediment reworking with redeposition on the now exposed former lake floor (Keighley et al., 2003a). The lack of any carbonate interbedding with the red-beds precludes interpretation of shoreline proximity. Also, there is no good evidence from which to interpret whether the preserved sandbodies were sufficiently downflow to be part of any delta system. As a possible analogue, before emptying into the Great Salt Lake via its delta, the Weber River crosses the tectonically active Wasatch Front and cuts through through its Quaternary Lake Bonneville highstand terraces and fan delta (Fig. 9B; Lemons et al., 1996). Alternatively, ephemeral rivers drain from the Agai Pah Hills (Gillis Range) and traverse across the axial trough of the Walker Basin and Lahontan highstand terraces toward Walker Lake, forming a braid plain (Figs. 1A, 9A).

The uppermost sandbodies of this interval are lenticular with IHS indicating point-bar deposition of meandering streams developing during subsequent early base level rise (Schumm, 1993), which culminated with M3 carbonate deposition. South of the junction of Argyle and Nine Mile canyons, the onlapping brown mudstone on the downflow side of one convex-up sandbody (Fig. 5D) is interpreted to be a channel abandonment fill on the floodplain (ox-bow lake fill). Results from XRD show a lack of carbonate phases that would be expected of mud deposited in Lake Uinta (Table 4). Based on the earlier discussion it remains possible that some of the other sandbodies may have been modified to mouth bars and subaqeuous levees when flooded by rising lake levels, although our studies have not found any remarkable examples.

The interpretations of this study differ from those of Schomacker et al. (2010) who include the entire M1—M3 interval as one sequence (stage S-15), comprising three stages, with the shoreline persisting south of the study area. They identified an M2 limestone but no oil shale and overlying shoaling deposits. Certainly the facies shift at M2 is rapid, occurring over less than a meter (Keighley et al., 2003a, figs. 6b, 9a), and sandbody truncation or pedogenic modification limits the recognition of the shoaling succession, but there is no supposed "gap in the stratigraphic motif" (Schomacker et al., 2010, p. 1083). The M2—M3 interval is equivalent to their stages 15-2 and 15-3, but while their stage 15-2 delta development supposedly was terminated by a rise in lake level they acknowledge no limestone bed is present from such an event. Overlying the M2 marker around Argyle Point, these authors also interpreted a stack of lenticular sandbodies over 15 m thick to be a mouth bar and associated distributary channel sandstone succession. However, they did not recognize any red, variegated, or pedogenic mudstone that is truncated by and overlies the sandstone at this location, or elsewhere in their S-15 study area and beyond (but see Fig. 5F, L, M; and figures in Keighley et al., 2003a). They also consider that the depositional style and sand:mud ratios remained very much the same for all three of their S-15 component stages, whereas examination of a broader study area had already revealed higher sand:mud ratios in M2—M3 than in M1—M2 (Fig. 6; Keighley et al., 2002, 2003a; Keighley and Flint, 2008).

Schomacker et al. (2010) also presented several "diagnostic" criteria for deltaic mouth bar and terminal distributary channels. One useful criterion was the absence of cutbank terminations, but the presence of cutbanks in outcrop (Fig. 5L) contradict these authors' own interpretations. The absence of subaerial facies and structures and interdigitation of sandstone with lacustrine mudstone, were also listed as diagnostic criteria. Similarly, the widespread occurrence of pedogenic red and variegated mudstone both truncated by and overlying the sandbodies (Figs. 5F-I, L, M) disprove these authors' interpretation of the outcrops being lakeward of the shoreline. In contrast, we consider as non-diagnostic their criteria (*ibid*, p. 1086) of "(i) a convex-upward bedding pattern…" and "(ii) basinward palaeocurrent orientation…".

M6—M7 sandbodies: delta distributary channels (locally lake modified)

The M6 marker has been interpreted as littoral bars of carbonate grainstone and distributary channel abandonments filled with fossiliferous micrite (Keighley et al., 2003a). This possibly indicates a location on the updrift part of a wave-influenced delta (cf. Bhattacharya and Giosan, 2003). The succession up to M7 (more beds of carbonate grainstone indicating littoral bars) is interpreted to be lacustrine dominated, similar to M1—M2, but lacking microbial and oil-shale carbonates. The presence of rare red-bed mudstone signifies terrestrial conditions were rarely established and that progradation must have been active to bring the depositional surface into the terrestrial realm at this location. More commonly than in the underlying intervals previously described, sandbodies cut down through the carbonate markers and rework individual (i.e., at that time unlithified) coated grains and ostracodes. The microbial carbonates and oil shale of M1—M2 were more resistant to downcutting by later channels (Keighley and Flint, 2008). This reworking, of course, makes the sandbodies more calcareous and thus the distinction of lake influenced versus upriver sandstone is more problematic.

Two locations, northwest and northeast of Dry Canyon, record a fluvial system bringing siliciclastic material to the wave-influenced delta. Northwest of Dry Canyon (Fig. 8A—C) the shoreface M6 marker beds (#2) grade up into a thin succession of floodplain deposits associated with delta progradation (#3). The floodplain is shallowly truncated by a very-fine grained sandstone, with gently dipping cosets of IHS marking gradual lateral fill of a terminal distributary channel (#4A). The upper, flat coset passes laterally westward into a distinctly calcareous sandstone bed (#4B). This bed may be interpreted as formed during a slight rise in lake level, which flooded the distributary and partly modified the uppermost sand in a marginal lacustrine shoal setting.

Renewed progradation and, or, relative fall in base level returned this location to a terrestrial setting with red-bed deposition (#5). This red-bed unit is widely truncated by a fine to medium grained sandstone, which locally (#6, far right of Fig. 8B) includes rotated slump blocks of the same lithology and a loaded base. Rapid deposition across a waterlogged substrate is suggested. Also in unit #6, gently inclined IHS and internal coset boundaries show a NW progradation of a barform with later cut-and-fill (chute channel?) on the eastern side. Some of the later-deposited IHS include very calcareous lithologies. Unit #6 is near the stratigraphic level of the M7 marker (cf. Fig. 3). Because of imperfect exposure and often inaccessible outcrop, two interpretations are possible for #6. The channelized sandstone may have truncated M7, with associated carbonate material being eroded upstream of this locality to be reworked and deposited as the downdip IHS. In this interpretation, the deposits of at least one cycle of lake expansion and retreat have been removed from this outcrop location. Alternatively, the sandstone may represent the lateral equivalent of the M7 littoral bars, formed somewhere around the delta mouth as the lake expanded just far enough south to reach this outcrop location. In this scenario, the distributary channel bar may have been either subject to gradually greater lacustrine influence (incorporating longshore drifted carbonate material and alkaline pore waters), or after flood-scouring of the channel base, to have prograded out into the lake as a sharp-based mouth bar (Fielding et al., 2005).

A return to terrestrial conditions is shown by units #7 and #8. The unit #8 sandstone cuts down into #7 thick red mudstones and is also capped by red-beds. Most of #8 is inaccessible but LiDAR images indicate lateral amalgamation of sandbodies containing IHS with a westerly dip component. They are interpreted to be sinuous point bar deposits on a well-drained floodplain located on the upper delta plain or further upslope than the delta apex. A similar interpretation is appropriate for the same interval in outcrop between Dry and Daddy canyons (Fig. 8D, E). Unit #4 in this case has incised more deeply and incorporates several angular boulders of limestone, possibly the micrite of M6. Unit #6 is vertically stacked on #4 and similarly contains carbonate-cemented sandstone (Fig. 7B).

M8—M9 sandbodies: delta distributary channels with evolved mouth-bars or shoals

Northeast of Trail Canyon, the succession passing through the stratigraphic level of M8 was broadly interpreted to be a delta mouth bar and distributary channel succession that passes westward into a carbonate shoal, and that is incised to the southwest by an angular sequence boundary caused by basin tilting (Keighley et al., 2003a). Between Daddy and Dry Canyons (Fig. 8E-G), the M7—M8 interval concludes with dry floodplain-sheetflood deposits (unit #A) truncated by 2-3 m thick fluvial channel sandstone (#B). In turn these units are both truncated by the complex geometries of unit #C. At the SW end of this unit (Fig. 8F), the #C1(a) fining up cosets of IHS are not highly calcareous and may represent a delta distributary. The overlying highly calcareous (grains and cement) #C1(b) that laterally transitions into the similarly calcareous #C2 units, with poorly preserved possible hummocky and symmetrical cross lamination, can be viewed as lacustrine, and thus deposited in a location immediately lakeward of the river mouth. Such a position would be the classic location for a mouth bar deposit. If true, the gently incising to depositional basal contact of units #C1 and #C2 represents a flooding surface. Likely following another, or continuing, transgressive pulse, the succession was capped by a carbonate shoal (#C3). Additional #C1-#C2 pairings are present in the overlying succession before floodplain deposits again predominate.

The Dry-Daddy outcrop is comparable in cross section (Figs 8D, 11G) to that illustrated for the rapid evolution of terminal distributary channel and mouth bar systems in modern river-dominated deltas (Olariu and Bhattacharya, 2006; see Fig. 11H). However, the modern mouth bars are documented to have prograded onto subaqueous delta plain, delta front and prodelta deposits, with a classical shoaling up profile in vertical logged section. In the Dry-Daddy outcrop, the *overall* succession (#A to #C3) is interpreted as due to the deepening and areal expansion of a wave-influenced lake. With only a few detailed facies distributions reported for modern mouth-bar locations in transgressive wave-influenced deltas (e.g., Penland et al., 1988, Roberts 1997), there is no well-developed model explaining exactly how distributaries and interfluves might be flooded and evolve into mouth bars or related sandbodies. Based on our outcrop interpretations, we suggest that as lake shorelines transgress (Fig. 11 A-C), a few scenarios are possible that allow for mouth bars to then develop and aggrade within an overall landward-stepping facies succession. For example, in a homopycnal situation, any erosion and most sediment transport and deposition might be expected to occur along the path of the now-flooded channel, whereas the flooded interfluve may become a wave-modified/reworked barform (Fig. 11D). Once deposited, sediment may also be reworked onto the developing mouth bar (old interfluve). In a hypopycnal situation, the fluvial jet and buoyant plume may spread out extensively over the flooded channel and more elevated interfluve, draping sediment over a wide area (Fig. 11E). As transgression continues, a more classical progradational or aggradational mid-ground bar, located between bifurcating jets, may develop over the old interfluve (Fig. 11F).

Taylor and Ritts (2004) and Moore et al. (2012) have also interpreted outcrop of the M6 to M9 interval in Argyle and Parley canyons to include mouth bars, although there are issues with how these interpretations were determined. For example, the authors tabulate the existence of red mudstone but do not distinguish such in their figures, where they interpret all the fines as offshore or marginal lacustrine. They also consider some laterally extensive sandstone to be gradationally, not sharply, based. Accordingly, these authors interpret the succession as a series of parasequences with mouth bars overlying marginal lacustrine deposits and truncated by distributary channel sandstone. In contrast to Moore et al.'s (2012, fig. 6) measured section 5, Keighley et al. (2002, and see supplementary data, plate 2) at the same location but commonly under a thin veneer of reddened soil, had recorded significant thicknesses of red, occasionally distinctly hematitic, mudstone (sample X31, Table 2). These red-beds are from the same intervals between M6 and M9, and thus laterally equivalent to those illustrated in Figures 7A, 7C and 8. In addition, laterally extensive sandbodies are predominantly sharp-based, often truncating underlying deposits such as gray mudstone or limestone. The red-beds in Argyle and Parley canyons therefore should be interpreted as subaerial, dry floodplain, possibly including paleosols. They underlie tabular limestone (M8) with a flat base that marks a lacustrine flooding surface, except where the overlying, downward truncating, sandstone, interpreted as fluvial (potentially distributary channels), has removed the carbonate.

CONCLUSIONS

The middle Green River Formation (GRF) in Nine Mile Canyon and its tributaries contains ~ 200 m of cyclic lacustrine-terrestrial sediments. Included are eleven carbonate marker beds, M1 to M11, which are most distinctive of the lacustrine part of the cycle and can be correlated across many km of outcrop. Red-bed mudstone with features of subaerial exposure and accompanying pedogenesis are the most distinctive terrestrial components. Calcareous sandstone suggests an alkaline-lake influence, and when coupled with symmetric, hummocky or swaley cross stratification, leads to a shoreface barform interpretation. Non-calcareous sandstone containing asymmetric cross-strata, laterally accreting bed sets or cosets of inclined heterolithic stratification (IHS), and steeply dipping lateral truncations of precursor strata by such sandstone leads to interpretation of an in-channel fluvial barform (e.g., sandbodies in the M2—M3 interval of this contribution). Some interbedded sandstone units are more difficult to assign to a particular environment because they lack, or contain a mixture of many of these features.

Based on the above characteristics and the position of the sandbody within the cyclic statigraphy, we documented a few problems regarding reinterpretations of sandbodies as simply deltaic mouth bar and terminal distributary channel deposits, and so modify the "diagnostic" features derived from these reinterpretations (Schomacker et al., 2010). First, although terminal distributaries may pass laterally into mouth bars, lumping together their diagnostic criteria, negates recognizing that these are composite bodies: one is sublacustrine (or submarine), the other terrestrial, with the shoreline (potential flooding surface) in between. Secondly, while some of their criteria may indeed distinguish a deltaic mouth bar/distributary channel from a fluvial point bar, other interpretations—such as a low sinuosity bar—, can equally be possible. Thirdly, few process- or environmental- facies are diagnostic of deltas (Galloway and Hobday, 1996; Bhattacharya and Giosan, 2003; Olariu and Bhattacharya, 2006); the same is also true for their component subenvironments.

We suggest that a new set of guidelines for the interpretation of mouth bars is required.

A) Any succession is first placed in its correct basin setting and sequence stratigraphic context by identification of the more easily resolved lithofacies (well drained floodplains, carbonate shoals, etc). For predominantly low-gradient closed-lake settings such as is interpreted for the current study area, the subsequent guidelines for mouth-bar interpretation could then include the following.

B) The sand body should contain a combination of basinward-dipping bed sets /cosets /foresets, sedimentary structures indicating basinward paleoflow, and/or, structures indicating rapid flow deceleration. Basal surfaces should be depositional or, if sharp and associated with channel bifurcation, should not form cut banks of more than a few degrees slope. Top surfaces should be convex-up or flat. Ideally, the sandbody should interfinger within a succession interpreted to be lacustrine (subaerial facies and structures absent).

C) Where a sandbody is both underlain and overlain by red-beds there must be evidence of a flooding event at the same stratigraphic level. The sand body must (i) pass laterally into a bed interpreted to contain a lacustrine shoreface, basinward into shoreface or delta front, (e.g., unit #6 in the M6—M7 interval) or (ii) laterally truncate such a bed as a sharp-based mouth bar.

D) Where the sandbody is at the approximate stratigraphic position where a flooding surface (e.g., red-bed below, carbonate above) or regressive surface (e.g., carbonate below, red-bed above) must be placed, detailed study is required. A considered interpretation should assess, where possible, detrital and early diagenetic mineralogy as well as sedimentary structures, paleoflow data, bedding contacts, etc., across the entire sandbody, and in any laterally transitional beds. Changes in mineralogy and sedimentary structure between any early and late-formed IHS may be particularly informative.

When either of the latter two guidelines are used, the dynamic nature of closed-lake basins must be remembered. As noted for Great Salt Lake, shorelines and adjacent supralittoral and littoral environments can even relocate hundreds of meters on an annual to decadal basis, with sediment deposited one year in one environment then subjected to a different environment the next year. As a result, any large scale sandy bedform located in the general vicinity of the lake shore does not necessarily have to have formed only under one set of environmental conditions. During transgression, near-surface parts of a former fluvial bar may be modified to, or overlain and backstepped by, mouth- and shallow lacustrine bars (e.g., unit #C in the M8—M9 interval). Such an interpretation of a lacustrine succession containing mouth bars provides a very different succession to that of delta fronts in a marine setting, which typically show a coarsening up profile associated with prograding shorefaces. Mouth bar preservation would not be favored during regression across a low gradient basin floor due to reworking and erosion by the lengthening fluvial systems.

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Table 1: Lithofacies associations and their interpreted depositional setting (modified from Keighley et al., 2003a).

Table 2: Summary of Marker Horizons, M1 to M11 (modified from Keighley et al., 2002)

Table 3: Summary of the siliciclastic succession in central Nine Mile Canyon (updated from Keighley et al., 2002). For carbonate succession, see Table 2.

Table 4: Semi-quantitative, and quantitative*, whole-rock XRD of mudstone from the middle GRF of Nine Mile Canyon (§and Parley Canyon; †Muscovite includes illite).

Figure 1 (A) Location of selected modern, Quaternary, and Cenozoic lakes of the western US. (B) Stratigraphy of the Uinta Basin (modified from Keighley, 2015; revised dates from Smith and Carroll, 2015). (C) Geological map of the Uinta Basin. Approx. depositional axis, which migrated through time, from Cashion (1995). Synclinal axis & Laramide thrust faults from Silliphant et al. (2002). (D) Location of studied outcrops along central Nine Mile Canyon.

Figure 2 Lake terminology. For the deltaic inset (modified from Bhattacharya and Giosan, 2003), general terms: A = Apex, B = River-influenced delta, C = Wave-influenced delta. Subaerial delta plain: α = fluvial in-channel bar/island, β = subaerial levee, γ = swamp-marsh (palludal), δ = beach/barrier sands. Subaqueous delta plain: *a* = distributary channel, *b* =

terminal distributary channel, *c* = abandoned channels, *d* = crevasse splay, *e* = trunk channel, *f* = delta-top lake. Delta front (lacustrine): 1 = mid-ground mouth bar, 2 = lunate mouth bar, 3 = subaqueous levee, 4 = interdistributary bay, 5 = shoal, 6 = lagoon.

Figure 3. LiDAR processed images of outcrops east of Gate Canyon. (A) Image from Blind to Daddy canyons, no vertical exaggeration (v.e.). (B) Same image with lithologic overlay, x3 v.e. (C) Lithologic overlay only, x3 v.e.

Figure 4. Summary sedimentological log of the succession exposed east of Blind Canyon. See table 1 for lithofacies association codes.

Figure 5. Outcrop photographs from adjacent to the junction of Nine Mile and Argyle canyons. Locations of thin section (TS) and XRD (X) samples arrowed. (A) General outcrop photo of Nine Mile Canyon illustrating the sandbodies of the intervals M1—M2 and M2—M3: by contrast, the overlying interval (M3—M4) is mostly of poorly exposed fine-grained strata. (B) UAV photomontage of the western truncating margin of the major sandbody in M1—M2. Red arrows mark the truncating surface, blue arrows mark example IHS (Inclined Heterolithic Strata) that dip to the east-northeast, indicating the presence of a fluvial side bar. (C) Outcrop photomontage showing more detail of the western truncating margin of the sandbody. A flooding surface (FS) is interpreted close to the top of the sandbody. Red arrows mark the same locations as in the previous image. (D) General outcrop photo of Nine Mile Canyon at the junction with Argyle Canyon showing rare truncation of M2. Red chevron arrow indicates the location of red/maroon mudstone

interpreted as floodplain. (E) Cross-polarized photomicrographs of thin section TS21, taken from the western truncating margin of the major sandbody in M1—M2: sandstone containing porosity, p, and reworked carbonate grains (x) that exhibit marginal dissolution and iron oxide coatings. Carbonate cement (z) is very rare. (F) Detail of macroform from M1—M2, interpreted as a mid-channel bar; white arrow marks a downflow-accreting set boundary. Two beds of red/maroon mudstone interpreted as floodplain (chevron arrows) are approximately laterally equivalent to the brown mudstone of X02 in Fig. 5J.

(G) Outcrop photo from the M1—M2 interval taken ~400 m west of Fig. 5C. Note the mottled red-purple-gray color of the weathered mudstone beneath an interpreted flooding surface, FS. Hammer for scale. (H) Possible root structures (red chevrons) in mottled brown-gray mudstone east of Fig. 5H. One-cent coin for scale. The white arrow locates the concretion split open in Fig. 5I. (I) Detailed image of concretionary mudstone in H. Red chevrons point to interpreted ferruginous oxidized plant roots. (J) A sparsely calcareous, convex-up sandbody is onlapped by non-calcareous brown mudstone, interpreted to be the fill of an abandoned fluvial channel. Hammer, circled, for scale. (K) TS01, top of sandbody in M2—M3, northwest of Argyle Point: porous, p, sandstone is again lacking in carbonate cement. (L) Gently inclined, very-fine grained IHS, white arrowed, form the sandbody in M1—M2 (see also Keighley et al., 2003a, Fig. 6g, h for detail). Stacked sandbodies in M2—M3 truncate (red chevron arrows) and form cut banks in red/maroon mudstone interpreted as floodplain. For scale, thickness from base M1—M2 is ~ 10 m. (M) Red to variegated red-gray (mottled) silty mudstone, truncated by medium-fine sandstone, M2—M3 interval, Argyle Point. The mudstone is interpreted to show variable pedogenic modification. The lowermost visible horizon is mostly structureless, whereas the overlying slightly siltier, microconcretionary horizon retains original lamination. Beneath the truncating sandstone is a reduced zone. Meter stick for scale.

Figure 6. Maps of sandbody distribution in intervals M1—M2 and M2—M3 (modified from Keighley and Flint, 2008). These intervals crop out only in the western part of the study area.

Figure 7. (A) Red to variegated red-gray (mottled) silty mudstone from below M6, at the measured section northwest of Dry Canyon, with structures interpreted as the desiccation cracks and roots of a poorly developed paleosol. ~~thin-section, cross-polarized photomicrographs. (A) TS21, western truncating margin of the major sandbody in M1—M2: sandstone containing porosity, p, and reworked carbonate grains, x, that exhibit marginal dissolution and iron oxide coatings. Carbonate cement absent. (B)~~ (B) TS11, associated with M7, near Daddy Canyon: sandstone with pore space occluded by extensive calcite cement (x). (C) Red to variegated red-gray (mottled) interbedded mudstone and tabular very-fine-grained sandstone, M7—M8 interval, between Dry and Daddy canyons. Again evidence exists for variable pedogenic modification. Of note are the large circular 'reduction' spots (center left), and bifurcating reduction patterns (roots?) and vertical sand filled cracks (above the 33 cm-long hammer for scale).

Figure 8. (A) LiDAR-processed image of outcrop between Blind and Dry canyons. (B) Interpretation of (A), from below M6 to above the bed correlated with M8. Note the lateral

amalgamation of the lenticular sandbody (#8) that contains IHS interpreted as a fluvial point bar. (C) Detailed outcrop photo of the interval above M6. (D) LiDAR processed image of outcrop between Dry and Daddy canyons. (E) Interpretation of (D), from below M6 to M9. (F), (G) Detailed outcrop photos of the sandbodies around the stratigraphic level of M8. Whereas the LiDAR and interpretive images are vertically exaggerated for clarity, outcrop photos have no vertical exaggeration. See table 1 for lithofacies codes.

Figure 9. Selected 'river mouth' maps from western US. (A) Walker River and Walker Lake. Recent deltas from Blair and McPherson (1994). The Sehoo highstand terrace at 1332 m (~13 ka) is little incised whereas the Eetza highstand terraces at 1350-1360 m (~140 ka) are highly degraded. (B) Weber River and Great Salt Lake. The Bonneville highstand terrace is at ~1550 m (12 ka) whereas the Provo highstand terrace is at ~1450 m (10 ka). Quaternary lake levels from Lemons et al., (1996-GSL); Adams and Wesnousky (1999).

Figure 10. Suggested paleogeographic models for the Uinta Basin constructed for various stages of basin fill. A), B) Initially, an increased hydrologic input:output ratio causes the lakes to transgress extensively over the low-gradient basin floor, even with limited relative rise of the base level. Rivers increase in sinuosity. The Uinta lake reaches threshold and starts to spill into the lower elevation Piceance Creek lake. C) Stable base level promotes progradation in Lake Uinta, and the shoaling upward carbonate parasequences. Progradation also straightens the fluvial planform near the river mouths, leading to low sinuosity trunk channels. This is the situation inferred for the succession below and through M1 (and M7, M8). D) If the two lakes merge and inputs still exceed output, the merged lake can deepen,

potentially up to the elevation of the next threshold (here speculated to be west of the Uinta lake). Clastic material no longer reaches much of the lake, which, under deep water, accumulates organic-rich sediment (which lithifies to an oil shale). E) A shift to output exceeding input results in a forced regression, exposing recently deposited, likely unconsolidated basin-margin material well above base level. F) The basin-margin material will have formed steep basin-margin terraces, with pronounced breaks in gradient. Fluvial systems actively incise through these unconsolidated terraces, reworking them out onto the basin floor as lowstand fluvio-deltaics. This situation is considered akin to the succession through and above M2

Figure 11. Modelled interpretations of lacustrine delta 'mouth-bar' development in the middle GRF. Labels #A, #B, #C1, #C2, and #C3 refer to the units in Figure 8. The break in each block diagram represents the present day outcrop in the canyon wall near the junction of Nine Mile and Daddy canyons. (A) The lake shoreline is well to the north. Fluvial channels truncate redbed soils and, locally, earlier carbonate beds. Channel sands are typically noncalcareous due to freshwater flow, or sparsely calcareous where there has been reworking. At the river outlet sediment in the fluvial outflow jet may be traction-dominated if the fluvial and lake waters have similar density (homopycnal) or a buoyant plume may spread out into the lake if the lake water is denser (hypopycnal) - hyperpycnal state not considered. (B) As in (A), but due to the low gradient basin floor a saline tongue may extend upriver during low fluvial flow, increasing alkalinity and salinity in sediment pore waters and potentially allowing for carbonate precipitation. In the lake, the influence of oblique to shore waves results in down (longshore-) drift reworking of any mouth bars and

incorporation of updrift strandplain carbonate particles. (C) As the lake transgresses, channel fill is increasingly influenced by lakewater composition and longshore transport of sediment. Eventually channels (terminal distributary or trunk) are submerged and the intervening interfluve deposits are now located lakeward of the river mouth, i.e., in a mouth bar location. In a homopycnal situation, (D), the old interfluve may now be wave modified while most sediment transport and deposition still occurs along the path of the now-flooded channel. Once deposited, sediment may also be reworked onto the developing mouth bar (old interfluve). In a hypopycnal situation, (E), the fluvial jet and buoyant plume may spread out extensively over the flooded channel and interfluve, depositing sediment over a wide area. As transgression continues (F), a more classical progradational or aggradational mid-ground bar, located between bifurcating jets, may develop over the old interfluve. Eventually (G) the entire area may form a lower shoreface, replete with carbonate sand shoals. Similar cross-sectional profiles (H) have been illustrated for river-dominated terminal distributary channels in the Atchafalaya delta (redrawn from Olariu and Bhattacharya, 2006).

Supplementary data (Plates 1 & 2)

Photomontages and interpretive panels of the outcrops around Argyle and Nine Mile Canyons (see Fig.8 for key). Modified from Keighley et al., (2002, 2003a).