Climate impact on fluvial-lake system evolution, Eocene Green River Formation, Uinta Basin, Utah, USA

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ABSTRACT

In light of a modern understanding of early Eocene greenhouse climate fluctuations and new highly seasonal fluvial system faces models, the role of climate in the evolution of one classically-cited continental, terminal lake system is re-examined. Detailed stratigraphic description and elemental abundance data from fifteen cores and seven outcrop regions of the Green River Formation were used to construct a ~150 km cross section across the Uinta Basin, Utah, USA. Lake Uinta in the Uinta Basin is divided into five lake phases: (1) post-Paleocene Eocene Thermal Maximum, (2) peak Eocene hyperthermal, (3) waning hyperthermal, (4) Eocene Climatic Optimum (EEOC), and (5) post-EECO regimes, based primarily on climatically driven changes in fluvial style in combination with sedimentary indicators of lacustrine carbonate deposition, organic matter preservation, salinity, and lake depth. Basinwide siliciclastic dominated intervals were deposited by highly seasonal fluvial systems and record negative organic carbon isotope excursions associated with early Eocene abrupt, transient global warming (hyperthermal) events. Carbonate dominated or organic rich intervals record stable, less seasonal climate periods between hyperthermals, with lower siliciclastic sediment supply allowing the development of carbonate and organic matter preservation. The stratigraphic progression from alternating organic rich and lean zones to the overlying organic rich Mahogany and R8 zones represents the global transition out of the pulsed early Eocene hyperthermal climate regime to a time of sediment starvation and lake stratification, sequestering sedimentary organic carbon. This study provides a novel approach to terrestrial paleoclimate reconstruction that relies largely on unique sedimentary indicators and secondarily on isotopic proxy records within the context of a large basin-wide sedimentologic and stratigraphic data set, thus setting the stage for future detailed geochemical terrestrial paleoclimate proxy development.

INTRODUCTION

The stratigraphic and vertical heterogeneity and variability of deposition rates preserved in terrestrial successions, as compared to their deep marine counterparts, makes developing geochemical (e.g., stable isotopic) or paleontological terrestrial paleoclimate proxy records challenging. Typical terrestrial paleoclimate reconstruction studies rely heavily on developing stable isotopic or paleontological proxy records on relatively thin successions from one or two localities (e.g., Abels et al., 2012), which can ultimately lack a firm regional sedimentary context. Commonly, the understanding of sedimentologic and stratigraphic framework of terrestrial successions is incomplete as compared to their marine counterparts, due in part to inherent preservation potential challenges as well as limited age constraints. This study uniquely provides a terrestrial paleoclimate proxy record based on a robust, thick, basin-wide sedimentary, geochemical and stratigraphic data set.

The Eocene Green River Formation is the most well-cited example of an ancient lacustrine system, and it was deposited during a period of critical global climate transition. In light of recent advancements in understanding early Eocene global climate shifts (Bowen et al., 2006; Nicolo et al., 2007; Sexton et al., 2011; Abels et al., 2012; Abels et al., 2016), the role of climate in lacustrine deposition needs to be re-evaluated (Fig. 1A, 1B). The record of the Green River Formation from the Uinta Basin, Utah, USA offers an opportunity for re-examination in this context.

The late Paleocene through early Eocene was a greenhouse period in which global temperatures rose, culminating in the Early Eocene Climatic Optimum (EECO), followed by the transition to icehouse conditions and development of Antarctic ice sheets during the late Eocene (Zachos et al., 2001). Superimposed on the warming phase of the late Paleocene to early Eocene greenhouse period were transient hyperthermals: brief periods of widespread, extreme climatic warming, driven by catastrophic release of isotopically light carbon (e.g., methane) to the atmosphere (Thomas and Zachos, 2000; Bowen et al., 2006; Kroon et al., 2007). The best documented hyperthermal, the Paleocene/Eocene Thermal Maximum (PETM), was followed by numerous, at least eight, early Eocene hyperthermal events that occurred during a warming phase before and during the EECO (Fig. 1A, 1B; Zachos et al., 2000; Sexton et al., 2006). The first documented early Eocene hyperthermal, the H1 event, or ETM-2, (Cramer et al., 2003), is associated with surface and deepwater ocean warming of 2–3 °C (ELMO clay horizon; Lourens et al., 2005). Early Eocene hyperthermals, likely driven by orbital forcing of insolation, are estimated to have lasted 40 k.y. with 100 k.y. to 400 k.y. spacing between events (Sexton et al., 2011). Specifically, it has been proposed that transient warming events are associated with eccentricity minima (Smith et al., 2014) or maxima (Cramer et al., 2003). Hyperthermals caused elevated chemical weathering rates and changes in precipitation patterns and intensity (Robert and Kennett, 1994; Kump et al., 2000). Additional studies from the Rocky...

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Lake Uinta Phases

Phase 1: post-PETM
Fresh lake, carbonate dominated, low siliciclastic sediment supply

Phase 2: Peak hyperthermal regime
Shallow lake, alternating siliciclastic (hyperthermals) & thin carbonate (inter-hyperthermal) dominated intervals

Phase 3: Waning hyperthermal regime
Rising lake, alternating carbonate and siliciclastic intervals

Phase 4: Post hyperthermal regime
High, stratified lake organic rich

Phase 5: post-EECO
Hypersaline, closing lake, migration of evaporites east to west

Regional Stratigraphy

Central & Western Uinta Basin

Eastern Uinta Basin

Willow Creek/ Indian Canyon

Nine Mile/ Gate Canyon

e.g., Evacuation/ Texas Creek

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Figure 1. Stratigraphy of Green River Formation (GRF), Uinta Basin, Utah, USA using known age constraints, as compared to the global climate-proxy record. (A) Δδ18O, paleotemperature, and δ13C data from Zachos et al. (2001) updated to Gradstein et al. (2012) global timescale (Vandenbergh et al., 2012). (B) δ13C data from Zachos et al. (2010) adjusted to Gradstein et al. (2012) global timescale (Vandenbergh et al., 2012). Paleocene/Eocene Thermal Maximum (PETM)-ETM1 (orange), known early Eocene hyperthermal events (solid dark orange), include H1 (ELMO/ETM-2), H2, H1, 12, J, K/a (X/ETM-3); lower magnitude, less well-studied Eocene hyperthermals b and c (dotted dark pink lines) and Early Eocene Climatic Optimum (EECO) (light pink) are plotted on Gradstein et al. (2012) global timescale (Vandenberghe et al., 2012) using noted age constraints. *marks dated tuff; ¥ marks 54.0 Ma Remy (1992) date at base of Carbonate Marker Unit, with approximate error estimate marked in brackets. Error estimate comes from the duration of Wasatchian adjusted to Ogg et al. (2016) timescale. References: 1Zachos et al. (2010), 2Sexton et al. (2006), 3Remy (1992), 4Smith and Carroll (2015); 5Smith et al. (2010); 6Westerhold et al. (2018); 7Smith et al. (2014). Ages of hyperthermals plotted on Green River Formation come from Zachos et al. (2010) with ages adjusted Gradstein et al. (2012) global timescale (Vandenberghe et al., 2012). Age of H1/ELMO/ETM-2 and H2 plotted on Green River Formation reflect revised published age of ETM-2 in Westerhold et al. (2018). (C) Uinta Basin, Utah, USA stratigraphy and interpreted lake phases, based on basin-wide detailed correlation of outcrop, core, and log data sets. GR—Green River Formation; Ls—Limestone; EECO—Early Eocene Climatic Optimum; PETM—Paleocene/Eocene Thermal Maximum; R—rich; L—lean.
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Mountain region during the PETM and early Eocene suggest that the highly seasonal precipitation patterns (i.e., monsoonal) climate had significant impact on the stratigraphy, driving increased physical erosion and chemical weathering rates (Smith et al., 2008a; Hyland and Sheldon, 2013), expansion of alluvial facies (Abels et al., 2012; Smith et al., 2014; Battle et al., 2016), and in some locations, alternations between alluvial and lacustrine sedimentation on the 100 k.y. short-eccentricity timescale (Aswasereelert et al., 2013).

Recent advances in our understanding of highly seasonal fluvial facies models (Fielding, 2006; Fielding et al., 2009; Allen et al., 2013; Plink-Björklund, 2015; Gall et al., 2017), along with documentation of highly seasonal fluvial deposits at specific localities in the Green River Formation (Gall et al., 2017), warrants a systematic examination of the nature and distribution of fluvial facies in the Green River Formation as a paleoclimatic proxy for highly seasonal climate conditions. Fluvial facies models are based largely on modern studies of temperate North American perennial river systems, which create an inherent bias in ancient interpretation (Bridge, 2006; Fielding et al., 2009). In fact there is a wide range of fluvial morphodynamics in modern river systems as a result of the annual fluvial discharge range or variability as compared to annual mean discharge (Plink-Björklund, 2015). Highly seasonal fluvial system deposits contain a spectrum of distinctive sedimentary structures and barform features that record rapid changes in flow strength associated with pronounced seasonal peak in precipitation and runoff (Fielding et al., 2009; Allen et al., 2013; Plink-Björklund, 2015).

Wyoming, USA, Green River Formation deposits are particularly well known as the basis for Carroll and Bohacs’ (1999; 2001) genetic lacustrine model, in which the long term tectonic and climatic controls on accommodation and sediment plus water fill are considered. The model emphasizes the impact of long-term tectonic drivers on the preserved lacustrine record through accommodation, including tectonic sills and stream capture that controlled lake water communication between basins (Carroll and Bohacs, 1999; 2001). The role of aridity and humidity and their control on lake expansion and contraction are also discussed (Pietras and Carroll, 2006), particularly in regards to short term cyclicity. Considering aridity and humidity alone as climatic controls on the lake system is an oversimplification of climatic processes and drivers. In this study, we instead take an integrated and long term view of climatic control on the lake system that also considers siliciclastic sediment delivery.

The main goal of this study is to explore the role of climate in fluvial-lacustrine deposition and lake evolution in light of a new understanding of early Eocene hyperthermal events using a newly constructed, detailed basin-wide sedimentologic and stratigraphic framework of the Green River Formation. Identifying facies changes in the stratigraphy allows for an examination of how climate sensitive, basin margin fluvial deposits change laterally (along paleodepositional strike and dip) and stratigraphically.

Geologic Background

The Eocene Green River Formation is the record of a continental interior, terminal lake basin system that covered a large area across north-eastern Utah, western Colorado (Uinta-Piceance Basin, respectively, Lake Uinta), and southwestern Wyoming (Greater Green River Basin, Lake Gosiute), USA (Fig. 2). A Laramide uplift, the Douglas Creek Arch marks the eastern edge of the Uinta Basin and acted as a topographic sill in Lake Uinta controlling the water body connection between the Uinta and Piceance basins, depending on lake level (Fig. 2). Fluvial systems from the Uinta Mountains to the north delivered sediment to ancient Lake Uinta on the steeply dipping northern margin. The more gently dipping southern margin of the basin also records fluvial-deltaic deposition into the lake. Provenance data suggest that the large fluvial-deltaic system of the underlying Wasatch/Colton and Green River Formation was part of a larger paleoriver system sourced from the Cordilleran magmatic arc in California, USA (paleo-California river), rather than sourced locally from Laramide uplifts to the south like the San Rafael Swell and Uncompaghre Uplift (Davis et al., 2010; Dickinson, 2012). A period of “freshwater” lacustrine carbonate deposition (Uteland Butte Limestone) was followed by a period of increased lake depth and salinity (Carbonate Marker Unit) (Fig. 1C). Overlying is a period of shallow lake deposition in which marginal fluvial deltaic deposits are interfingered with lacustrine carbonate deposits (Sunny Side Delta Interval, central Uinta Basin, and Douglas Creek Member (eastern Uinta Basin) that transition to organic rich carbonate mudstone and oil shale basinward (Fig. 1C).

A period of high lake level and stratification followed and is recorded by the regionally extensive organic oil shale rich (R) and lean (L) zones that alternate stratigraphically (Parachute Creek Member; Fig. 1C). The Green River Formation of the Uinta Basin, Utah, is one of the most prolific source rocks in the world, well known for its thermally immature oil shale resource of 1.32 trillion barrels (total in-place oil; Johnson et al., 2010). The most organic rich of the R zones is the Mahogany zone (R7; Fig. 1C). In the final period of deposition, Lake Uinta became hypersaline, consisting of a thick succession of moderately organic rich oil shale (R8) overlain by evaporite deposits as well as delta deposits (Horsebench Sandstone) and volcanoclastics that represent the closing of the lake (Fig. 1C; Vanden Berg and Birgenheier, 2017).

The Early Eocene Green River Formation records the period leading up to and during the Early Eocene hyperthermal events, the EECO, as well as the period following the EECO (Fig. 1B). The base of the Carbonate Marker Unit of the lower Green River Formation in the western Uinta Basin is dated to 54.0 Ma from paleontological evidence from Nine Mile Canyon and palynomorphs, ostracods, charophytes, and mollusks from Price Canyon. The fauna are Wasatchian (55.5–50.5 Ma, Ogg et al., 2016), which is broadly equivalent to the Ypresian, early Eocene (Fig. 1B). The Flagstaff/North Horn Formation, that underlies the Wasatch/Colton Formation, is Paleocene in age (Fig. 1B; Remy, 1992). The Skyline tuff from the R4 zone in the Uinta Basin has been dated to 49.58 ± 0.28 Ma (Smith and Carroll, 2015). The Curly and Wavy tuffs provide bracketing ages of the Mahogany zone (R7) of 49.32 ± 0.30 and 48.66 ± 0.23 Ma, respectively (Figs. 1B and C; Smith et al., 2008b; Smith et al., 2010). The Fat tuff, found within the Saline Facies stratigraphic interval of the western Uinta Basin, is dated to 46.62 ± 0.13 Ma (Figs. 1B and C; Smith et al., 2008b; Smith et al., 2010).

From the age constraints available, it is evident that fluvial-lacustrine deposition of the middle Green River Formation broadly coincides with the timing of the early Eocene hyperthermals and the subsequent Early Eocene Climatic Optimum (EECO), which occurred ca. 52.6–50.1 Ma (Fig. 1; Smith et al., 2014; Gall et al., 2017). This includes the Douglas Creek Member and Sunnyside Delta Interval of the middle Green River Formation as well as the overlying transition interval and basal Parachute Green Member (Fig. 1C). Age constraints do not allow for resolution of individual hyperthermal events within the Green River Formation stratigraphy.

**METHODS**

A systematic detailed sedimentologic, stratigraphic, and geochemical study was performed on fifteen cores ranging in thickness from 100 m to 500 m (Fig. 2; Appendix A, measured sections). Key features noted in each core include grain size, stratification, sedimentary structures, mineralogy, bioturbation, microbially-influenced features, and body and plant fossils. Visual inspection of core was supplemented with geochemical analyses and data sets including Fischer assay, a measure of organic richness, and X-ray fluorescence (XRF) analysis, which provided elemental abundance as a proxy for mineralogy (Appendix A, measured sections; Appendix B, XRF methods [see footnote 1]).

In addition to the core data, high quality 3-D exposures covering several square kilometers were chosen for study, allowing for depositional dip parallel, strike parallel, and oblique views of the ancient depositional system. In each of the seven outcrop regions studied, we measured several detailed stratigraphic sections as well as walked out key stratigraphic units to document stratigraphic architecture and interpret outcrop photomosaics (Fig. 2; Appendix A, measured sections). Detailed core and outcrop data sets were collected over a period of nine years and constructed iteratively into a basinwide detailed stratigraphic correlation using known age constraints in the upper Green River Formation and key proximal to distal facies relationships and stratigraphic stacking patterns, including correlation of aggradational, progradational, or retrogradationally stacked parasequence sets (Gall et al., 2017; Smith et al., 2010; Rosenberg, 2013; Toms, 2014; Rosenberg et al., 2015; Rosencrans, 2015).

In concert with detailed sedimentologic and stratigraphic descriptive, mudstone facies were sampled from two outcrop composite measured sections (10; Nine Mile Canyon, 16; Hay Canyon) for stratigraphic bulk organic carbon isotope analysis, as well as organic carbon and nitrogen content. Fist-size samples of fluvial floodplain derived, pedogenic siltstones, and lacustrine mudstones were collected by trenching until fresh rock was exposed. Samples were dried and powdered. Inorganic carbon was removed from samples using an acid slurry method. Samples were immersed in 1M HCl and reacted until the pH of the solution remained at <2. The acid solution was removed via vacuum and filter and the sample placed in a drying oven overnight. Samples were weighed into tin foil cups and analyzed using a Carlo Erba 1108 elemental analyzer connected to a Thermo Finnigan Delta XL stable-isotope gas ratio mass spectrometer at the University of Utah, Salt Lake City, Stable Isotope Ratio Facility for Environmental Research laboratory.

The %Corg and δ13Corg analyses, along with average, minimum, maximum, and standard deviation and analytical precision (standard error, 1σ) are reported in Appendix C, organic carbon isotope data (see footnote 1). To isolate for organic matter source, only results from fluvioglacial floodplain derived siltstones with varying degrees of pedogenic modification are reported here, similar to the approach of Magioncalda et al. (2004). Lacustrine mudstones were not utilized. Carbon isotopic compositions of the bulk organic matter fractions are reported in per mil (%ε) relative to the Vienna Pee Dee belemnite standard. Analytical error is around 0.1% and 0.2%ε, and replicate analyses are ± 2% and ± 0.16%ε of average values at given stratigraphic height for Corg and δ13Corg, respectively. δ13Corg stratigraphic records from Nine Mile Canyon and Hay Canyon are shown in Figures 3 and 4.

**RESULTS**

Full detailed measured sections and relevant elemental abundance (XRF) data from outcrop and core at the 21 study localities are reported in Appendix A, measured sections. These data were used to construct the west to east cross section shown in Figures 3 and 4. We define 21 facies that are grouped into seven facies associations (FA) including: (FA1) fluvial deposits; (FA2) deltaic deposits; (FA3) siliciclastic lake flat deposits; (FA4) low siliciclastic sediment supply, littoral to sublittoral carbonate deposits; (FA5) siliciclastic sediment starved, progradational carbonates deposits; (FA6) evaporite deposits; and (FA7) volcanic deposits (Table 1; Appendix A, measured sections). We present a description of the facies and facies association scheme below. Detailed facies descriptions from outcrop and core localities can be found in (Gall et al., 2017; Birgenheier and Vanden Berg, 2011; Rosenberg, 2013; Toms, 2014; Rosenberg et al., 2015; Rosencrans, 2015).

**Facies**

**FA1 Fluvial deposits**

**Description.** Facies Association 1 (FA1) consists of very fine to medium sandstone channel bodies (F1.1 and F2.2) with adjacent and interbedded pedogenically modified siltstone and gray to green to red or purple paleosols (F1.3; Fig. 5).
Figure 3. Basin-wide cross section of the Green River Formation, Uinta Basin, Utah, USA displaying genetic stratigraphic packages and five lake evolution phases linked to climatic controls. Established stratigraphic units within the Green River Formation are marked, along with interpreted early Eocene hyperthermal events (orange and yellow intervals). Numbered data sets shown in cross section correspond to locality numbers and color and circle symbology shown and used in Figure 2. Master datum: Curly tuff (49.3 Ma, Smith et al., 2010); secondary datums: Fat tuff (46.6 Ma, Smith et al., 2010) and C Marker (Remy, 1992; Keighley et al., 2002; 2003). Note in Figure 4 an alternate correlation solution is presented where C marker is correlated to R2 instead of R4. Red arrows highlight organic carbon isotope data trends. API—American Petroleum Institute gamma ray units; EECO—Early Eocene Climatic Optimum; GRF—Green River Formation; Mrkr.—marker; OM—organic matter; PETM—Paleocene/Eocene Thermal Maximum.
Figure 4. Correlation of phase 2, peak hyperthermal regime and phase 3, waning hyperthermal regime in outcrops across the southern edge of the basin. Orange siliciclastic intervals containing highly seasonal fluvial channels (F1.1) and mouthbars (F2.2) represent correlation of identified hyperthermal events. Blue and yellow colored intervals represent organic rich (R zones) carbonate dominated and organic poor (L zones) siliciclastic dominated facies intervals that are correlated across the basin. Siliciclastic dominated lean zones may represent later smaller hyperthermal events, with carbonate dominate R zone representing inter-hyperthermal periods. Note in Figure 3 an alternate correlation solution is presented where C marker is correlated to R4 instead of R2. Age constraints: \(^{1}\)Remy (1992), see Figure 1 for error estimate; \(^{2}\)Smith and Carroll (2015); \(^{3}\)Smith et al. (2010). PETM—Paleocene/Eocene Thermal Maximum.
**Climate and lake evolution, Eocene Green River Formation**

**TABLE 1. FACES AND FACES ASSOCIATIONS, GREEN RIVER FORMATION, UINTA BASIN, UTAH, USA**

<table>
<thead>
<tr>
<th>Facies Association 1: Fluvial deposits</th>
<th>Structures</th>
<th>Geometry</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F1.1 Strongly channelized sandstone; fine to medium sandstone</td>
<td>Dominated by upper flow regime and high deposition rate structures; planar lamination with minor normal grading, low angle lamination, scour and fill, convex upward bedforms, and climbing ripples with minor trough cross stratification (TCS); dominantly downstream accreting bedforms</td>
<td>3–20 m thick, multi-story laterally and vertically amalgamated bodies, 1–15 m downcutting, erosionally based, lateral extent 200 to 1 km.</td>
<td>Highly seasonal fluvial channels</td>
</tr>
<tr>
<td>F1.2 Channelized sandstone; very fine to medium sandstone</td>
<td>Dominated by TCS; dominantly laterally accreting bedforms</td>
<td>5–14 m thick, single to two-story, lateral extent up to 400 m</td>
<td>Temperate, perennial fluvial channels</td>
</tr>
<tr>
<td>F1.3 Paleosol; siltstone, commonly pedogenically modified</td>
<td>Poorly consolidated red to purple to greenish-grey siltstone, mottled with minor root traces and burrows, blocky pedds where present, minor sillinkenes</td>
<td>Tabular to channelized</td>
<td>Floodplain and/or abandoned channel fill</td>
</tr>
</tbody>
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<thead>
<tr>
<th>Facies Association 2: Deltaic deposits</th>
<th>Structures</th>
<th>Geometry</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F2.1 Strongly channelized sandstone; fine to medium sandstone</td>
<td>Lateral accretion sets, TCS, current ripples, rip ups at base, minor fish debris</td>
<td>1–5 m thick, erosionally based, dominantly laterally amalgamated with minor vertical amalgamation</td>
<td>Terminal distributary channels</td>
</tr>
<tr>
<td>F2.2 Weakly channelized/weakly erosive sandstone; very fine to medium sandstone</td>
<td>Downstream accretion sets, low angle laminations, current ripples, minor TCS, planar parallel to wavy laminations, soft sediment deformation, rip ups, some fine scale normal grading</td>
<td>0.5–5 m, slightly erosionally based, 0.5–1 m of downcutting, sheet-like, lobate to tabular</td>
<td>Mouth bar (multi-storied)</td>
</tr>
<tr>
<td>F2.3 Tabular sandstone; very fine to medium sandstone interbedded w/siltstone</td>
<td>Sandstone beds contain wave ripples, planar parallel laminations, low angle lamincations, minor burrows, shell debris, carbonate grains</td>
<td>0.5–3 m, sharp based sandstone beds, tabular to wavy bedding</td>
<td>Wave-influenced shoreface</td>
</tr>
<tr>
<td>F2.4 Interbedded and interlaminated siltstone and sandstone</td>
<td>Carbonate poor siltstone interbedded with fine sandstone, associated with F2.2 and F2.4, displays current ripples, climbing ripples, and normally graded beds indicating hyperpycnal flow</td>
<td>0.5–10 m, sharp based, tabular to wavy bedding</td>
<td>Distal mouth bar</td>
</tr>
<tr>
<td>F2.5 Grey blocky siltstone</td>
<td>Fault laminations, some normal grading</td>
<td>Tabular, cliff or slope former in outcrop</td>
<td>Prodelta</td>
</tr>
</tbody>
</table>

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<thead>
<tr>
<th>Facies Association 3: Siliciclastic lake flat deposits</th>
<th>Structures</th>
<th>Geometry</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F3.1 Weakly channelized sandstone; very fine to sandstone</td>
<td>Low angle laminations, current ripples with minor trough cross bedding, plane parallel laminations, wavy laminations and soft sediment deformation</td>
<td>1–3 m thick, erosionally based, sheet-like, lobate to tabular</td>
<td>Mouth bar (single story)</td>
</tr>
<tr>
<td>F3.2 Heterolithic channel; mudstone and very fine sandstone and siltstone</td>
<td>Current ripples, low angle laminations, minor wave ripples</td>
<td>1–2 m, erosionally based</td>
<td>Lake flat channels</td>
</tr>
<tr>
<td>F3.3 Calcareous siltstone; green, calcareous matrix</td>
<td>Plane parallel laminations, current ripples and wavy ripples, soft sediment deformation, burrows, synneresis cracks</td>
<td>1–8 m, tabular, flabby weathering</td>
<td>Lake flat deposits</td>
</tr>
<tr>
<td>F3.4 Paleosol; mudstone and siltstone</td>
<td>Mottled, pedogenic, slicks, current ripples in siltstone</td>
<td>10–50 cm, tabular to lenticular</td>
<td>Floodplain/lake flat</td>
</tr>
</tbody>
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<tr>
<th>Facies Association 4: Low siliciclastic sediment supply, littoral to subtidal carbonate deposits</th>
<th>Structures</th>
<th>Geometry</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F4.1 Microbialite</td>
<td>Stromatolite with minor thrombolite, minor fish debris</td>
<td>0.01–1 m, wavey to tabular</td>
<td>Carbonate ramp, littoral (high energy)</td>
</tr>
<tr>
<td>F4.2 Coarse grained carbonate; wackestone and grainstone</td>
<td>Massive, planar parallel to wavy laminations, minor current/wave ripples, minor burrows, ooid, fish and ostracod debris, carbonate and microbialite rip up clasts, minor low angle lamincations</td>
<td>10 cm–2 m, tabular to wavy</td>
<td>Carbonate ramp, littoral (high energy)</td>
</tr>
<tr>
<td>F4.3 Organic poor carbonate mudstone</td>
<td>Massive to planar parallel to wavy laminations, varves, occasional rip ups, soft sediment deformation, siltstone interbeds, disseminated dolomite, algal pellets</td>
<td>20 cm–2 m, tabular to wavy</td>
<td>Carbonate ramp, sublittoral (low energy)</td>
</tr>
<tr>
<td>F4.4 Gastropod and bivalve bearing wackestone</td>
<td>Massive with dispersed and bedded gastropods and bivalves. Locally, shells densely packed enough to form conquomas.</td>
<td>10 cm–1 m</td>
<td>Fresh water carbonate ramp; littoral to subtidal (low energy)</td>
</tr>
</tbody>
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<thead>
<tr>
<th>Facies Association 5: Siliciclastic sediment starved, profound carbonate deposits</th>
<th>Structures</th>
<th>Geometry</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F5.1 Organic rich carbonate mudstone; oil shale; with fish fossils and thin microbialites</td>
<td>Massive to planar parallel to wavy laminations, soft sediment deformation, commonly associated with fish scales/bones, less organic rich, more clay-rich and less calcium rich than F5.2.</td>
<td>20 cm–2 m, tabular to wavy</td>
<td>Open water deposits, profundal</td>
</tr>
<tr>
<td>F5.2 Organic rich carbonate mudstone; oil shale; very few fossils</td>
<td>Massive to planar parallel to wavy laminations, soft sediment deformation, minor evaporites, marcasite, precipitated carbonate nodules, disseminated dolomite; very few fossils present, with the exception of stratigraphically limited &quot;botfly&quot; larvae and plant fossils; more organic rich, less clay rich and more carbonate rich than F5.2</td>
<td>2 cm–2 m, tabular to wavy</td>
<td>Open water deposits, profundal</td>
</tr>
<tr>
<td>F5.3 Oil shale breccia</td>
<td>Massive to planar parallel to wavy laminations, soft sediment deformation, minor evaporites, marcasite, precipitated carbonate nodules, disseminated dolomite</td>
<td>20 cm–2 m, tabular to wavy</td>
<td>Open water deposits, profundal</td>
</tr>
<tr>
<td>F5.4 Grey to brown calcareous mudstone</td>
<td>Finely laminated, lacks fossils, organic richness varies; associated with F4.1, F4.2, and F4.4</td>
<td>2 cm–10 m, tabular</td>
<td>Open water deposits, profundal</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Facies Association 6: Evaporite deposits</th>
<th>Structures</th>
<th>Geometry</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F6.1 Large nahcolite nodules; evaporite deposits</td>
<td>Large white to grey crystal nahcolite nodules that displace surrounding laminations in organic rich carbonate mudstone and/or large voids of dissolved nahcolite with displacive growth</td>
<td>Tabular bed occurrence, found in depocenter of lake</td>
<td>Depocenter of density stratified lake with saline bottom water</td>
</tr>
<tr>
<td>F6.2 Small dispersed nahcolite; evaporite deposits</td>
<td>Small dispersed crystals of nahcolite in planar laminated organic rich carbonate mudstone</td>
<td>Tabular bed occurrence; found in halo around depocenter of lake</td>
<td>Depocenter of density stratified lake with saline bottom water</td>
</tr>
<tr>
<td>F6.3 Fracture fill shortite; evaporite deposits</td>
<td>Fractured organic rich carbonate mudstone with fractures filled with shortite, a hydrothermal secondary mineral</td>
<td>Cross cuts bedding planes; fracture fill</td>
<td>Depocenter of density stratified lake with saline bottom water</td>
</tr>
</tbody>
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<table>
<thead>
<tr>
<th>Facies Association 7: Volcanic deposits</th>
<th>Structures</th>
<th>Geometry</th>
<th>Depositional environment</th>
</tr>
</thead>
<tbody>
<tr>
<td>F7.1 Volcanic tuffs</td>
<td>Massive to stratified, visible crystals, typically white to orange</td>
<td>1 cm–15 m, wavy to tabular</td>
<td>Lake ashfall and reworked ashlaff</td>
</tr>
<tr>
<td>F7.2 Volcaniclastic sandstone</td>
<td>Massive to convolute bedded buff colored, with mudstone rip up clasts, pyrite nodules and local small evaporite crystal nodules</td>
<td>1–2 m thick, tabular to loaded bases</td>
<td>Volcanic flows into lake</td>
</tr>
</tbody>
</table>
Figure 5. (A) Outcrop image of stacked F1.1 highly seasonal fluvial channels from Main Canyon, Douglas Creek Member, Uinta Basin, Utah, USA, package 2, phase 2 recording hyperthermal events. (B) Interpretation of (A). Example of thick accumulations upper flow regime-dominated sandstone, F1.1 highly seasonal fluvial channels cut or capped by heterolithic channels. Note visible accretion sets in event #2, and promi- nent low angle convex-upward bedforms in event #7. (C) Outcrop example of trough cross stratified sandstone bodies, F1.2, interpreted as temperate, perennial fluvial channels. (D) A floodplain succession (F1.3) that overlies and is overlain by upper flow regime dominated, highly seasonal fluvial sandstone (F1.1). Note the 1.8-m-thick red-tan siltstone (sltstn), interpreted to represent well-drained floodplain conditions transitions to <0.3-m-thick beds of gray-green siltstone, interpreted to record moderately to poorly drained floodplain conditions. This floodplain succession is interpreted to represent a shallowing water table level and increased fluvial input from base to top. (E) Rhizoliths within green-gray siltstone and overbank fluvial sandstone deposits interpreted as moderately to poorly drained floodplain deposits (F3.1). Rock hammer for scale.
Facies Association 1 (FA1) consists of fluvial facies that are interpreted to have been deposited in ancient river systems (Figs. 5A and 6). These systems are relatively quiescent and capable of eroding, transporting, and depositing highly seasonal fluvial channel deposits. The channelized, erosionally based, discontinuous sandstone bodies are typically multistoried, 5–10 m thick, with lateral amalgamation surfaces that define F.2.5 are typically cliff-forming carbonates, size, and stratigraphic architecture. The first (F.1.1) are multi-story, laterally and vertically amalgamated sandstone channel bodies dominated by upper flow regime structures including planar lamination with minor normal grading, low angle lamination, scour and fill features and convex upward bedforms with minor trough cross bedding (average 15%, up to 30%), as well as high deposition rate climbing ripples (Figs. 5A and 6). Downcutting ranges from 1 m up to 15 m. Thicknesses of amalgamated sandstone channel bodies range from 3 m to 20 m and are dominantly downstream accreting. Lateral extent ranges from 200 m to >1 km (Fig. 5A). The second type (F.1.2) are single to two story fluvial channel deposits characterized by trough cross bed sets, with sandstone channel body thickness ranging from 5 m to 14 m and laterally extent up to 400 m (Fig. 5B).

Interpretation. We interpret sandstone bodies (F.1.1 and F.1.2) as fluvial channel deposits indicative of differing discharge, and hence climate, regimes. F.3.3 deposits are laterally adjacent floodplain deposits that experienced subaerial exposure, as evidenced by pedogenic modification. Amalgamated multi-story fluvial channel bodies dominated by a variety of upper flow regime structures (F.1.1) are interpreted to have been deposited in ancient river systems that were strongly affected by a highly variable seasonal discharge regime, akin to modern subtropical fluvial systems (Fielding and Alexander, 1996; Fielding, 2006; Fielding et al., 2009; Plink-Björklund, 2015; Gall et al., 2017). These systems are relatively quiescent and inactive for a large period of the year but are capable of eroding, transporting, and depositing large volumes of sediment in seasonal to interannual transient flood events, resulting in high deposition rate sedimentary structures. Modern systems, like the Burdekin River of Australia or the Ganges of India, deliver the vast majority of their discharge and sediment during the short monsoon season (Alexander et al., 1999; Goodbred, 2003; Plink-Björklund, 2015). The large seasonal and interannual precipitation range is the key control on river morphodynamics and resultant sedimentary facies observed in these systems (Fielding et al., 2009; Plink-Björklund, 2015). We further interpret these facies to have been deposited in a semi-arid setting, based largely on the development of well-drained red floodplain and paleosol deposits (F1.3), as well as the common occurrence of highly seasonal fluvial systems in semi-arid settings. A majority of previous and recent studies in the Green River Formation at Nine Mile Canyon (e.g., Keighley et al., 2002; 2003; Schomacker et al., 2010) support a fluvial channel interpretation, but at least one study that pre-dates publications on highly seasonal fluvial facies models prefers a deltaic distributary channel interpretation (Moore et al., 2012). Recent work by Gall et al. (2017) at Main Canyon further identifies highly seasonal fluvial channels in the Green River Formation.

In contrast, fluvial channel bodies dominated by trough cross-bedding (F1.2; Fig. 5B) are in line with classic fluvial facies models that were developed from temperate mid-latitude rivers, like the Mississippi or Platte rivers of North America (Smith, 1970; Smith, 1971; Smith, 1972; Sambrook Smith et al., 2006; Sambrook Smith et al., 2009). These perennial, persistent discharge river systems within more stable climates are characterized by much smaller seasonal fluctuations in discharge and sediment transport, allowing time for the formation of classic stable unidirectional flow features, like trough cross bedding (Gall et al., 2017; Plink-Björklund, 2015). We interpret these bodies to record more stable, less seasonal climate conditions.

FA2 Deltaic Deposits

Description. Facies Association 2 (FA2) consists of siliciclastic deposits that lack evidence of major subaerial exposure including two types of sandstone bodies (F2.1, 2.2), two facies that are composed of heterolithic interbedded and interlaminated sandstone and siltstone (F2.3, 2.4), and one siltstone facies (F2.5).

F2.1 consists of fine to medium grained, channelized, erosionally based, discontinuous sandstone bodies 1–5 m in thickness (Figs. 7B, 7C). Sandstone bodies exhibit 1–3 m of erosional downcutting (Figs. 7B, 7C). They are typically multistoried, 5–10 m thick, with lateral and vertical amalgamation. Lateral accretion sets, trough cross stratification, rip-up clasts, and fish debris are common toward the base of these channel bodies and generally grade into current ripples and climbing ripples toward the top of the channel bodies. Soft sediment deformation and planar parallel laminations are found throughout.

F2.2 is weakly channelized to tabular very fine to medium sandstone. F2.2 differs from F2.1 in that it is less erosive and is dominated by upper flow regime structures. These deposits exhibit sigmoidal clinofans that are downstream to obliquely accreting relative to paleoflow indicators. Individual sandstone beds are typically 1–3 m thick and 30–100 m in outcrop length (Fig. 7A). Amalgamated sandstone bodies (n = 2–9) average 9 m in thickness and 120 m in length, although they can extend up to 700 m in length. Amalgamation surfaces are commonly erosive and may contain a basal lag or rip-up clasts. Deposits are tabular to lobate or convex-upward, with a sharp, weakly erosional to subaerial boundary and typically less than a meter of erosional relief (Fig. 7A). Sandstone bodies also contain plane parallel laminations, low angle laminations, convex upward bedforms, minor trough cross bedding, current ripples, laminae scale normal grading, and soft sediment deformation. Both F2.1 and 2.2, and the enclosing fine grained facies (F2.4 and F2.5), lack evidence of prolonged subaerial exposure, in contrast to fluvial deposits of FA1 that display pedogenic modification. F2.1 and 2.2 are commonly found in vertical or lateral stratigraphic association with one another. However, in some regions, such as Evacuation Creek, F2.2 occurs without the presence of F2.1.

F2.3 consists of laterally continuous, very fine to medium grained, tabular sandstone bodies with planar parallel laminations, low angle laminations, wave ripples, and minor scouring throughout. The sandstone beds are thin (10s of cm typically, up to 1 m) and are interbedded with siltstone beds (Fig. 8A). Internally, packages coarsen upwards. Sandstone beds contain abundant shell debris and carbonates. F2.3 is not associated with F2.1, 2.2, or 2.4.

F2.4 is composed of interbedded and interlaminated sandstone and siltstone (beds are 10–50 cm thick), but contains current and secondarily some wave ripples, climbing ripples and normally graded beds. Sandstone beds are typically sharp based and tabular. F2.4 is found in vertical and lateral stratigraphic association with F2.2. Commonly, F2.4 overlies F2.4.

Carbonate-poor gray blocky siltstone beds that define F2.5 are typically cliff-forming and found stratigraphically below F2.2 and 2.4. Siltstone beds typically display no clear evidence of pedogenic modification or subaerial exposure, with local areas displaying minor pedogenic modification. Burrowing is absent. In core, siltstone beds are normally graded. Siliciclastic facies F2.5 and F2.4 form coarsening-upwards successions capped by F2.1 or F2.2.

Interpretation. FA2 is composed of siliciclastic dominated packages interpreted to be deposited in a littoral to sublittoral environment via river-dominated deltaic processes (F2.1, F2.2, F2.4, F2.5) or wave-influenced shoreline (F2.3) processes. One of the key distinctions between FA1 and FA2 is the dominantly subaqueous deposition of FA2 as compared to the subaerial conditions of FA1.
Figure 6. Characteristic sedimentary structures found in F1.1, upper flow regime dominated, highly seasonal fluvial channels. A–D; F–G from Nine Mile Canyon, Uinta Basin, Utah, USA. E from Main Canyon, Uinta Basin, Utah, USA.
Figure 7. Outcrop examples of stratal geometries associated with FA2, deltaic deposits. (A) Mouthbar facies (F2.2), from Texas Creek, Douglas Creek Member, Uinta Basin, Utah, USA, package 2, phase 2. Weakly erosive sandstone (F2.2) displays 5 m of downcutting into underlying sandstone and siltstone beds. Paleocurrent data indicate that clinothems are downstream to obliquely accreting. (B) Strongly channelized sandstone, terminal distributary channel (F2.1) from Nine Mile Canyon, central Uinta Basin, Carbonate Marker Unit, package 1, phase 1. (C) Strongly channelized sandstone, terminal distributary channel (F2.1) from Evacuation Creek, eastern Uinta Basin, Douglas Creek Member, package 2, phase 2.
Figure 8. (A) Tabular sandstone interbedded with siltstone (F2.3), wave dominated shoreface deposits from Gate Canyon, central Uinta Basin, Utah, USA, Horse Bench Sandstone, package 5, phase 5. (B–G) Outcrop examples of FA3, siliciclastic lake flat deposits from Willow Creek/Indian Canyon, western Uinta Basin, transitional interval, package 3, phase 3. (B) weakly channelized sandstone, mouth bar body (F3.1) overlying red/green paleosol deposits (F3.4). (C and D) Heterolithic-fill channel deposits (F3.2) with basal scour highlighted in dashed yellow line and (D) internal accretion sets highlighted in yellow solid lines. (E) Green, calcareous siltstone, interpreted as lake flat deposits (F3.3) and single story, weakly channelized sandstone (F3.1) interpreted as mouth bar deposits. (F) Calcareous siltstone, lake flat deposits (F3.3) with vertical burrows. (G) Paleosol, floodplain or lake flat deposits (F3.4) underlying a channel body.
The river-dominated deltaic deposits include terminal distributary channels with subaqueous erosion (F2.1; Figs. 7B, 7C), as evidenced by the lack of floodplain paleosols and desiccation features above or below these channel bodies (Olariu and Bhattacharya, 2006; Tanavssu-Milkeviciene and Sarg, 2012). Climbing ripples and soft sediment deformation indicate high sediment supply and rapid deposition, respectively, whereas basal channel rip-up clasts indicate rapid flow. F2.1 is found in association with proximal and distal mouthbar delta deposits (F2.2 and 2.4, respectively). F2.2 are interpreted as mouth bar deposits due to their tabular to lobate or convex-upward geometry. Internal current-dominated sedimentary structures, sharp erosive base, and downstream to obliquely accreting bars (Fig. 7A; Fielding, 2005; Schomacker et al., 2010). Mouth bar deposits are interpreted to be products of erosive hyperpycnal flows generated from high-sediment-yield, episodic river discharge events. Similarly, F2.3, though finer grained than F2.2 also contains normally graded beds, current and climbing ripples, indicative of hyperpycnal flow events in more distal mouthbar environment. Upper flow regime to transitional sedimentary structures, such as convex-up bedforms and parallel to low angle laminations are also indicative of deposition associated with seasonal river discharge (Fielding, 2006), indicating that episodic flooding events may have been important in generating hyperpycnal flows. Because the fluvial system (FA1) entered a very shallow lake (Ryder et al., 1976) that was likely accommodation limited, a full delta front wasn’t able to develop and instead the delta system developed as a series of long run out mouthbar complexes formed in shallow lake water (Jerrold et al., 2016). Silstone of F2.4 is interpreted as the most distal expression of hyperpycnal flow events on the prodelta, as evidence by minor normal grading in core. Prodelta deposits (F2.4) can be distinguished from floodplain deposits (F1.3) based on the relative lack of burrowing and pedogenesis in prodelta deposits (F2.4) relative to floodplain deposits (F1.3). Coarsening upwards packages record mouth bar complex progradation in the littoral to sublittoral zone of the lake. The characteristics of these deposits including moderate channelization, subaqueous erosion, sandy mouthbar deposits, and hyperpycnal flow characteristic support the interpretation of a deltaic lake margin.

F2.3 record wave influenced, siliciclastic deltaic deposits along the lake margin (Fig. 8A). Wavy laminations and symmetrical ripples are indicative of wave action (Pietras and Carroll, 2006). A delta delivered siliciclastic sediment that was then wave-reworked. The lack of hummocky cross stratification suggests fairweather wave action rather than storms were the main mechanism reworking deltaic sediment. Shell debris and carbonate grains represent deposition in a high energy environment and the coarsening upwards trends indicate overall shallowing (Renaut and Gierlowski-Kordesch, 2010). A shoreface interpretation is favored due to the absence of fluvial channels in outcrop associated with F2.3. One example of F2.3 occurs in the Horsebench Sandstone (upper Green River Formation) at Gate Canyon in the central Uinta Basin (Fig. 8A). At the same stratigraphic interval, river-dominated deltaic deposits occur ~30 km to the west in Willow Creek/Indiana Canyon (Tom’s, 2014), suggesting along shore delta influence.

**FA3 Siliciclastic Lake Flat Deposits**

**Description.** Overall, F3.1 is finer grained than F2.2 and is found exclusively in the transitional interval of the middle Green River Formation in the western portion of the basin. F3.1.3, 2.3, 3.3, and 3.4 are all laterally and vertically related to one another in outcrop exposures (Fig. 8).

F3.1 is composed of laterally discontinuous, very fine to fine grained, weakly channelized sandstone bodies (Figs. 8A and 8E). Channel bodies range from 1 m to 3 m and are typically single-storied. F3.1 is very similar to F2.2 but is single-storied rather than multi-storied like F2.2. The sandstone bodies are dominated by low angle laminations, minor downstream accretion sets, and minor trough cross stratification toward the base with plant debris, some rip-up clasts, and very minor fish debris. Current ripples, climbing ripples, planar parallel laminations, and wavy laminations can be observed toward the top of these channel bodies. These sandstone bodies can be lenticular or tabular and sheet-like with minor downcutting, similar to F2.2. The channels are overlain by greenish mudstone units that commonly display evidence of subaerial exposure.

F3.2 consists of laterally discontinuous, channelized, single story, heterolithic interbedded muddy siltstone and very fine sandstone deposits (Figs. 8C, 8D). Interbeds range from 5 cm to 20 cm thick. Channel deposits are 1–2 m thick. Sedimentary structures are scarce in the sandstone beds but where present include current ripples, low angle laminations, and soft sediment deformation. Channel fill ranges from sandstone- to mudstone-dominated (Figs. 8C, 8D). F3.1 and 3.2 are found in vertical and lateral stratigraphic association with one another.

F3.3 and F3.4 are both mudstone facies laterally and vertically interbedded with F3.1 and 3.2. F3.3 consists of 1–8-m-thick green calcareous siltstone that is commonly interbedded with greenish purple claystone. These deposits are dominantly massive and mottled but can also have plane parallel laminations, current ripples, and minor burrowing (Figs. 8E, 8F). F3.4 is composed of green and purple claystone with minor siltstone beds (Fig. 8G). Beds are mottled and contain slicks, both of which are evidence of pedogenesis. In some cases, blocky ped are evident. Current ripples are present within the siltstone beds. Units are 10–50 cm thick.

**Interpretation.** F3.1 is interpreted as mouthbar packages that were deposited from an updip fluvial system along a very shallow gradient (Figs. 8A and 8E). Climbing ripples and soft sediment deformation are indicative of high sediment supply and rapid deposition, similar to F2.2. However, because mouthbar packages are single storied rather than multi-storied, an even lower accommodation setting, shallower gradient, and/or lower fluvial sediment supply is interpreted for F3.3 relative to F2.2.

F3.2 is interpreted as littoral lake flat channels (Figs. 8C, 8D) and are typically laterally related to mouthbar deposits of F3.1 as well as paleosols of F3.4 (Fig. 8G). The sandstone dominated lenses of lake flat channels (F3.2) represent fluvial deposition in the channel body (Ryder et al., 1976). The mudstone dominated lenses represent muddy fluvial deposition in the channel body as well as fluvial channel avulsion and abandonment with subsequent fill by lake flat mudstones. Due to their heterolithic nature, the lake flat channels represent a lower energy fluvial regime than those fluvial deposits documented in FA1.

F3.3 is interpreted as lake flat deposits that are found along the lake margin and lagoon environments (Fig. 8E). The green color, as opposed to a gray or brown color indicative of organic matter, is interpreted to result from a shallow lake level which allowed for the destruction of primary organic matter by bacteria and bottom feeders. The mottled texture is a result of periodic subaerial exposure and pedogenic modification (Keighley et al., 2002; 2003). F3.4 is interpreted as poorly developed paleosols that developed on the floodplain or lake flat environment, similar to those observed by Keighley et al. (2003). These paleosols contain current rippled siltstones that are indicative of overbank deposits from the related heterolithic lake flat channels of F3.2.

**FA4 Low Siliciclastic Sediment Supply, Littoral to Sublittoral Carbonate Deposits**

**Description.** FA4 is composed of micrite-limestone (F4.1), carbonate grainstone (F4.2), organic-poor carbonate mudstone (F4.3) facies, and gastropod and bivalve bearing wackestone (F4.4) (Fig. 9). F4.1 includes laterally discontinuous and continuous laminated mats less than
Figure 9. (A) Microbialite (F4.1), stromatolite from Skyline 16 core, eastern Uinta Basin, Utah, USA, middle R6, package 3, phase 3. (B) Shrub-like stromatolite (F4.1) from Skyline 16 core, eastern Uinta Basin, Douglas Creek Member, package 2, phase 2. (C) Stromatolite mounds (F4.1) from Evacuation Creek, eastern Uinta Basin. (D) Columnar or shrub-like stromatolite (F4.1) from Texas Creek, eastern Uinta Basin, Douglas Creek Member. (E) Organic poor wackestone to mudstone (F4.3) from Skyline 16 core, eastern Uinta Basin, R5, package 3, phase 3. (F) Coated ostracod grains in ostracodal packstone (F4.2) from Main Canyon. (G) Ostracodal grainstone bed from Gate Canyon, central Uinta Basin, transitional interval, package 3, phase 3. (H) Organic poor carbonate mudstone bed (F4.3) overlain and underlain by calcareous siltstone (F3.3) at Willow Creek/Indian Canyon, western Uinta Basin, transitional interval, package 3, phase 3. (I) Organic poor carbonate mudstone bed (F4.3) at Willow Creek/Indian Canyon, western Uinta Basin, transitional interval, package 3, phase 3.
4 cm thick; isolated stromatolitic mounds between 5 cm and 0.5 m tall; laterally continuous, connected stromatolite mounds up to 1 m thick (Figs. 9A and 9C); packages of stromatolitic fingers/columns up to 1 m thick (Fig. 9D), and continuous to discontinuous shrubs (Fig. 9B), and thrombolitic heads up to 1 m thick. Ostracod carapaces and ooid grains are commonly found within microbialite fabrics (Fig. 9F). Microbialite beds (F4.1) are typically laterally extensive across an outcrop region, ranging from 0.5 km to several km in extent. In places, sandstone bodies (F1.1, 1.2, 2.1, 2.2) downcut into microbialite beds, preventing preservation.

F4.2 is composed of grainstone to packstone beds that are laterally continuous and commonly contain ooids, ostracods, oncolites, carbonate intraclasts, and/or stromatolite fragments (Figs. 9E–9G). They exhibit sparry cement in grainstone or a carbonate mud matrix in packstone units. Thin section analysis indicates ooids are the most common grains, although ostracods are also common and form the nucleus of some coated grains (Figs. 9F, 9G). F4.1 and F4.2 are usually interbedded and microbialites (F4.1) commonly cap grainstone deposits. Beds of F4.1 and 4.2 tend to be very laterally continuous over several hundred meters, and carbonate deposits over 1 m in thickness are used as marker beds that extend for several kilometers in outcrop regions.

Carbonate mudstone (F4.3) deposits are white to gray and are typically massive to faintly laminated (Figs. 9H, 9I). Carbonate mudstone or wackestone contain sparse ostracod carapaces and ooid grains. In outcrop, beds are laterally continuous and typically less than 0.5 mm thick, though may be up to 2 m thick. F4.3 is organic poor, with only a few occurrences of moderately organic rich deposits occurring in thin beds less than 5 cm thick. F4.2 and F4.3 commonly grade into each other laterally and stratigraphically, or occur in beds separated by a non-erosive contact.

F4.4 is composed of gray massive wackestone beds, typically 0.1–0.5 mm thick, that contains dispersed gastropods and bivalves. Locally, bivalves are densely packed, forming coquinas. F4.4 is commonly interbedded with F5.4. This facies occurs only in the lower Green River Formation.

Interpretation. FA4 is interpreted as a carbonate-dominated litoral to sublittoral lacustrine environment. Carbonate mudstone and wackestone (F4.3) are interpreted as deposited in lower energy environments where carbonate mud precipitated in the water column and settled on the lake bottom, likely in the sublittoral zone. The coated grains and lack of carbonate mud in grainstones to packstones (F4.2) indicate a high energy, wave-influenced environment, such as a lacustrine bar, shoal, or shoreline where most mud was winnowed out (Milroy and Wright, 2002; McGlue et al., 2010). The presence of microbialites (F4.1) indicates deposition within the photic zone in shallow water (Awrakam and Buchheim, 2015) and represents a mixture of depositional energy conditions (Frantz et al., 2014; Awramik and Buchheim, 2015). Gastropod and bivalve bearing wackestones indicate freshwater littoral to sublittoral conditions, based on the fossil assemblage.

Littoral to sublittoral carbonate deposits (FA4) are generally found where siliciclastic littoral to sublittoral deposits are spatially absent, laterally and stratigraphically. Lacustrine carbonate precipitation occurs when lake water is supersaturated with respect to calcium carbonate and depends on inflow of calcium-rich water (Gierlowski-Kordesch, 2004; Renault and Gierlowski-Kordesch, 2010). Paleozoic and Mesozoic limestone units along the major paleo-California fluvial system from the south likely supplied the necessary ions for ongoing carbonate precipitation within Lake Uinta. Additionally, fluvial input of siliciclastic sediment must be low or sufficiently spatially distant so that they don’t dilute carbonate deposition. This occurs during river avulsion or a siliciclastic deposition hiatus between river discharge events. Therefore, accumulations of carbonate facies in the succession are interpreted as low siliciclastic sediment supply conditions locally (or regionally depending on the lateral extent) in the lake. This process results in an alternation of siliciclastic and carbonate facies dependent on fluvial sediment input to Lake Uinta.

FA5 Siliciclastic Sediment Starved, Profoundal Carbonate Deposits

Description. FA5 consists of organic rich carbonate mudstone (F5.1 and F5.2), colloquially referred to as “oil shale” as well as oil shale breccia (F5.3; Figs. 10A–10C and 10J). Organic rich carbonate mudstone or oil shale facies (F5.1 and F5.2) both display massive to planar or wavy laminations and soft sediment deformation. Deposits are thinly laminated between darker colored (black to dark brown), organic rich and lighter colored (gray to brown), relatively organic poor laminations (Figs. 10A, 10C, and 10J). Laminations are parallel to low angle, with no strong evidence of mud ripples. Outcrop expressions typically weather white and form ledges. At the Skyline box cut near Evacuation Creek, low angle beds and relief combined with soft sediment deformation are observed at the several meter length scale. Facies of FA5 are generally much more organic rich relative to facies of FA4. Organic richness of F5.1 is moderate to high, with beds ranging from 10–30 gallons per ton (gpt) (Vanden Berg, 2008). F5.1 contains numerous fish scales and bones and it is commonly associated with thin (0.1–0.3 m) microbialite layers (F2.2; Fig. 10G).

As compared to F5.1, fish fossils are rare in F5.2. Fossil abundance varies by zone, but fossilized “botfly” larvae (found in specific stratigraphic intervals within the upper R6 and lower R8), insects, and plants are common. Also in contrast to F5.1, microbialites are not typically associated with this facies. F5.2 locally contains minor evaporite mineral precipitation in cores from the basin center in selected stratigraphic intervals, marcasite mineralization, precipitated carbonate nodules, disseminated dolomite, and lenses of organic matter (Fig. 10E).
Figure 10.
Most of these detailed characteristics are only observed in core and thin section. F5.2 is generally more organic rich than F5.1, with individual beds reaching up to 70 gpt in the Mahogany zone (Vanden Berg, 2008). F2.1 and F2.2 are similar, but exhibit a clear elemental geochemical difference based on core XRF data. F5.2 is more calcium rich than F5.1 (Rosenberg, 2013; Rosenberg et al., 2015). Soft sediment deformation is common, and some particularly organic rich intervals between a few cm and several meters thick are brecciated, containing oil shale rip-up intraclasts and extensive deformation, forming oil shale breccias that comprises F5.3 (Fig. 10B). Deformation and brecciation zones are laterally extensive up to 50 m in outcrop. F5.3 grades laterally into F5.2 (Figs. 10A, 10B). F5.4 is gray to brown mudstone that is not as organic rich as the other facies in F5.5. It is typically laminated and fissile, lacks fossils and is associated with F4.1, 4.2, and 4.4.

**Interpretation.** F5.5 is interpreted as low energy carbonate ramp deposits. F5.1 is interpreted as deposited in the sublittoral to profundal zone of the lake, based on the presence of thin microbialites that require episodic photic conditions to form. Based on the higher organic richness and carbonate content, F5.2 was deposited in the profundal zone of the lake. Organic rich carbonate mud can be deposited in a range of lake depth conditions. These finely laminated carbonate mudstones are indicative of a relatively deep, open water lacustrine environment in which there was minimal siliciclastic input, along with high plankton and algal production and preservation rates that outpaced carbonate production overall (Renaut and Gierlowski-Kordesch, 2010). Most mud was precipitated in the water column and deposited from suspension settling, though low angle laminae and low angle beds in outcrop suggest bottom water currents may have played a role in redistributing mud on the lake floor. The absence of mud ripples indicates traction flows were not the main deposition mechanisms. The organic richness of and lack of fish fossils in the carbonate mudstone F5.2 may be indicative of an anoxic environment, enhanced by density stratification of the lake, where bacteria and other organisms that degrade the organic matter were not able to survive and thus allowed for the preservation of organic matter (Bohacs et al., 2005). The geochemical distinction between F5.1 and F5.2 suggests a transition from proximal environment with higher detrital clay content to a more distal organic rich environment with higher carbonate production and lower detrital input within the profundal zone. Water saturated organic rich carbonate experienced various levels of early lithification correlated positively to inorganic carbonate content. F5.3 oil shale breccias may have formed through dewatering, subsequent deformation and failure, followed by downslope migration, causing brecciation (Törö and Pratt, 2015). Alternatively, oil shale breccia facies may represent early post-depositional bedding-plane-parallel melanges created by shear and liquefaction and formed in place (Olsen and Kinney, 2016).

**FA6 Evaporite Deposits**

**Description.** FA6 is marked by the occurrence of large (cm to m scale) nahcolite nodules (F6.1) and small (mm to cm scale) disseminated or laminated white to clear evaporite mineral crystals (F6.2) that are largely stratigraphically restricted to the upper Green River Formation. In F6.1, large white to clear to dark gray nahcolite nodules displace surrounding beds (Figs. 10F and 10I). Where dissolved from past and present groundwater and surface water interactions, nahcolite nodules form voids and cavities in the formation (Fig. 10I). Large nahcolite nodules (F6.1) are found in basin center localities and are haloed stratigraphically and geographically by small evaporite mineral crystals (F6.2). Vertical to subvertical vein or fractures that cross cut bedding are filled with white to clear to gra salts (F6.3; Fig. 10D). Evaporite minerals are most commonly hosted in organic rich carbonate mudstone beds (F5.2; Fig. 10H).

**Interpretation.** By association with F5.2, organic rich carbonate mudstone, FA6 is interpreted as also deposited in the profundal lake zone, where the lake waters can be supersaturated with ions as a result of density water column stratification (Tanavsuu-Mikeviciene and Sarg, 2012; Johnson and Brownfield, 2015). Large nahcolite nodules (F6.1) and small evaporite mineral crystals (F6.2) precipitated at the lakewater-sediment interface from density stratified, supersaturated lower water column conditions (Vanden Berg and Birgenheier, 2017; Dyni et al., 1985; Dyni, 1996). Note that the use of “evaporite” is misleading as it involves an image of subaerial evaporation on the edge of the lake, which is not the interpreted mechanism in this case. Instead the interpreted mechanism is supersaturated precipitation at the sediment-water interface in the depocenter, or deepest portion, of the lake. The stratigraphic and geographic distribution of evaporite crystals provides insight to the duration (episodic versus long-lived) and extent (geographically limited or basin-wide) of water column density stratification. We interpret the occurrence of large nahcolite nodules (F6.1) to record higher salinity bottom water conditions relative to the occurrence of small evaporite mineral crystals (F6.2). Fracture fill shortite is interpreted to have been generated displacively through hydraulic fracturing during a later, secondary hydrothermal fluid migration event (Jagiecki et al., 2013).

**FA7 Volcanic Deposits**

FA7 contains the volcanic tuffs (F7.1) that are observed throughout the Green River Formation and most commonly in the R8 Interval and the Mahogany zone of the upper Green River Formation. F7.1 consists of volcanic ash with soft sediment deformation (tuff). Color is typically gray, orange, or red with thicknesses ranging from 1 cm to 50 cm. Bed geometries are wavy to tabular. Buff colored volcaniclastic units (F7.2) of mixed sandstone and mudstone are found on the eastern side of the basin, in the upper Green River Formation. Beds greater than 3 m thick are internally massive to soft sediment deformed with dispersed partially lithified organic rich mudstone intraclasts.

The tuffs (F7.1) present in the Green River Formation most likely originate from the Absaroka Volcanic province, Challis volcanic field and Lowland Creek Volcanics (Smith et al., 2008b). Where the tuffs are massive, they are thought to be ash fall deposits. Internal features, such as wavy laminations and soft sediment deformation, may indicate that the tuff was reworked by lake processes. Volcaniclastic deposits (F7.2) are interpreted as lacustrine debris flows.

**Stratigraphic and Geographic Facies Distribution**

The Green River Formation is divided into five stratigraphic intervals, which represent five phases of Lake Uinta evolution (Figs. 3 and 4).

**Phase 1: Post-Paleocene Eocene Thermal Maximum**

**Description.** The Phase 1 stratigraphic interval includes the Lower Green River Formation, which contains the Uteland Butte Member, the Colton Tongue, and the Carbonate Marker Unit in stratigraphic order. Phase 1 interval is dominated by carbonate facies with minor siliciclastic facies (Fig. 3). It notably lacks microbialite facies (F4.1). The dearth of siliciclastic deposits in the Phase 1 interval basin wide suggests low siliciclastic sediment supply with deposition on a carbonate ramp (Fig. 3).

The Uteland Butte Member is composed entirely of carbonate facies. Central Uinta Basin outcrop exposures in Nine Mile Canyon and cores to the north deeper in the basin display interbedded gastropod and bivalve bearing wackestone (F4.4), gray to brown calcareous mudstone (F5.4), and dolomitic grainstone, packstone, and mudstone beds (Johnson et al.,

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2016; Logan et al., 2016). The Uteland Butte member was measured in detail at one outcrop locality and one core in this study, and characterization is bolstered by recent studies of this interval (Johnson et al., 2016; Logan et al., 2016). On the eastern side of the Uinta Basin, the PR-15-7C core and surrounding outcrops display ostracodal packstones to grainstones (F4.2) interbedded with gray to brown mudstone (F5.4), representing deposition in a low siliciclastic sediment supply, low gradient carbonate ramp environment (Logan et al., 2016).

The Colton Tongue contains predominantly fluvial deposits (FA1) with small sandstone channels (0.5–15 m, 5 m average thick; 20–300 m, 100 m average in lateral extent) and adjacent floodplain paleosol deposits (F1.3). The unit is thicker (~100 m) in the south central Uinta Basin at Nine Mile Canyon, and thinner to the east (Hay Canyon, Texas Creek, and PR-15-7C core; Figs. 3 and 4). The unit is ~25 m thick in the PR-15-7c core, and lateral facies variation across the basin is limited (Figs. 3 and 4).

The Carbonate Marker Unit is carbonate dominated with secondary siliciclastic facies. In the south central Uinta Basin at Nine Mile Canyon, the unit is composed of interbedded microbiotalite (F4.1), ostracodal packstone to grainstone (F4.2) and gray to brown calcareous mudstone (F5.4). Fluvial channels (F1.2 and F2.1) and mouthbars (F2.2) are present but minor in volume. Sandstone channel body height ranges from 4–10 m thick (8 m average) and lateral extent ranges from 75–275 m (129 m average). Channel bodies contain dominantly trough cross stratification and ripples, with minor plane parallel lamination, indicative of perennial discharge fluvial systems (F1.2). The Carbonate Marker Bed, an ~10 m thick unit at the top of the Carbonate Marker Unit, is composed of organic rich carbonate mudstone (F5.1). On the east side of the Uinta Basin, the Carbonate Marker Bed is absent, but the Carbonate Marker Unit is of similar thickness (Figs. 3 and 4). It is composed largely of F5.1, organic rich carbonate mudstone and F5.4, gray to brown calcareous mudstone, with interbedded ostracodal wackestone to grainstone (F4.2). Overall, the Carbonate Marker Unit represents deposition in a carbonate ramp environment, with low siliciclastic sediment supply.

Interpretation. The underlying Wasatch/Colton Formation is thought to record the PETM, as documented in the Piceance Basin by Foreman et al. (2012) and corroborated by a multi-phase ~6‰ negative organic carbon isotope excursion documented in the Uinta Basin (Birgenheier et al., 2017). Phase 1, records the post-PETM recovery stage of the lower Green River Formation (Figs. 3 and 4). Because it is characterized by the dominance of carbonate facies basin wide that uniquely contain freshwater gastropods and bivalve valves, this interval records the first major fresh water transgression of Lake Uinta in the Uinta Basin. Siliciclastic sediment supply, in the form of fluvial channels and mouthbars, is present only locally, largely on the south central side of the basin where the fluvial system (Sunnyside Delta) was most active throughout deposition. Siliciclastics otherwise are relatively minor in Phase 1 strata (Figs. 3 and 4), suggesting relatively minor, stable, less seasonal fluvial sediment supply to the lake system.

Phase 2: Peak Eocene Hyperthermals Description. The Phase 2 stratigraphic interval consists of the lower middle Green River Formation, specifically the Sunnyside Delta Interval in the south central portion of the basin and the laterally equivalent lower Douglas Creek Member to the east (Figs. 3 and 4). Within the Sunnyside Delta Interval, distinct multi-lateral packages of fluvial-deltaic sandstone (F1.1, 1.3, 2.2, and 2.4) and floodplain deposits (F1.3) alternate with carbonate lacustrine-dominated intervals (F4.1, 4.2, and 4.3) containing minor channels (FA1) (Figs. 3 and 4; see also Keighley et al., 2003). Siliciclastic-dominated intervals, marked as Phase 2 orange stratigraphic intervals in Figures 3 and 4, are characterized by large amalgamated upper flow regime structure-dominated, highly seasonal fluvial channels (F1.1) and mouthbars (F2.2), as well as floodplain deposits (see also Gall et al., 2017 for more detailed stratigraphic architecture description and interpretation). Carbonate deposits (F4.1, 4.2, and 4.3) are present but minor. In the south central Uinta Basin at Nine Mile Canyon and Main Canyon, paleocurrent data from fluvial channels indicate flow from south to north, in keeping with regional paleogeographic reconstructions of the Sunnyside Delta system (Gall et al., 2017). Highly seasonal fluvial sandstone channel bodies (F1.1) at Nine Mile Canyon range in thickness from 0.5 to 17 m thick (7 m average) and 20–480 m in lateral extent (350 m average). Phase 2 strata contain the largest sandstone bodies within the basin (Figs. 3 and 4). Note that sandstone bodies are not physically connected or amalgamated across the basin. Instead, highly seasonal fluvial sandstone channel bodies (F1.1) are found within particular stratigraphic intervals that can be correlated across the basin margin (orange intervals, Figs. 3 and 4).

At least four negative shifts in δ13Ccarb sampled from fluvial floodplain-derived siltstone and paleosol units are associated with regionally extensive highly seasonal fluvial sandstone channel body packages (F1.1) in the Sunnyside Delta Interval and the Douglas Creek Member from widely spaced (~80 km) outcrop localities at Nine Mile and Hay Canyons (Figs. 4 and 11). Negative δ13Cexcursions range from 2.5‰ and 5‰ in magnitude (Figs. 3, 4, and 11).

Siliciclastic-dominated intervals contain a variety of siliciclastic facies that change along the basin margin and downdip. Siliciclastic dominated intervals containing fluvial-deltaic deposit dominated intervals are correlated across the basin margin from west to east (see localities 10. Nine Mile Canyon, 16. Hay Canyon, and 24. Texas Creek, Fig. 4). Mouth bar complexes (F2.2, 2.4, and 2.5) dominate on the eastern side of the basin (e.g., Texas Creek, Figs. 2 and 3; Rosencrans, 2015). Paleocurrent data mainly from current ripples in mouth bar deposits show a dominant flow direction to the northeast in the lower half of the Douglas Creek Member at Texas Creek (Rosencrans, 2015).

In addition to correlating across the basin margin, siliciclastic-dominated intervals also correlate downdip with expected proximal to distal siliciclastic facies changes. Channel sandstone and mouthbar facies at the margin correlate downdip to intervals that contain distal mouth bar to prodelta facies in the basin center (Fig. 12).

Carbonate-dominated intervals within package 2 on the basin margin contain littoral to sublittoral carbonate deposits (FA4) with minor profundal deposits (FA5) and minor siliciclastic facies (e.g., FA1 and FA2) (Fig. 2). Littoral and sublittoral carbonate deposits (FA4) on the basin margin transition to sublittoral to profundal carbonate deposits in the basin center (Fig. 12). The thickness of carbonate-dominated intervals ranges by stratigraphic interval and geographic location from 5 m to 40 m, with an average of 25 m.

Interpretation. The negative organic carbon isotope excursions found in association with siliciclastic-dominated intervals, specifically amalgamated, highly seasonal fluvial channels (F1.1), are interpreted to be associated with individual early Eocene hyperthermal events (Figs. 3 and 4). Specifically, the globally negative shifts in δ13C of the atmosphere at the onset of hyperthermals were recorded in photosynthesis of plant matter that was preserved as dispersed sedimentary organic matter in the fluvial succession. We interpret these packages of upper flow regime structure dominated fluvial channels (F1.1) to be deposited in highly seasonal, semi-arid subtropical climate conditions associated with early Eocene hyperthermals, during peak hyperthermal regime conditions. Positive shifts of δ13C follow, recording the recovery of the δ13C of the atmosphere between hyperthermals, constituting an excursion. The
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Figure 11. Detailed view of two negative organic carbon isotope excursions interpreted as hyperthermals, with the associated measured section detailing facies present. Record from Nine Mile Canyon, Uinta Basin, Utah, USA. vf—very fine; f—fine; m—medium; c—coarse.

Magnitude of organic carbon isotope excursions associated with early Eocene hyperthermal events in the Uinta Basin is similar in magnitude to those documented from neighboring basins (e.g., Abels et al., 2012; Abels et al., 2016).

Intervening carbonate dominated packages (FA4) with minor fluvial channel bodies (FA1) and mouthbar bodies (F2.2 and F2.3) are interpreted to record periods between hyperthermals (inter-hyperthermals) with more stable, non-seasonal or persistent discharge river systems that transported less sediment overall than the highly seasonal systems (Figs. 3 and 4). The overall lower siliciclastic input and sediment dilution allowed precipitation of carbonates to dominate sedimentation between hyperthermal events.

The Sunnyside Delta Interval and equivalent Douglas Creek Member were deposited before the EECO, when early Eocene hyperthermal events were frequent and relatively higher magnitude (Fig. 1). At this time Lake Uinta was very shallow, as evidenced by widespread littoral facies, even toward the basin center where they are interbedded with sublittoral and profundal facies. In contrast to Phase 1, Phase 2 does not contain fresh water mollusks and contains microbivalves, suggesting Phase 2 salinity increased as compared to Phase 1.

The basin wide occurrence of siliciclastic dominated intervals alternating with carbonate dominated intervals is interpreted to represent temporal change between high and low siliciclastic sediment supply periods across the basin. An alternate interpretation is that siliciclastic dominated intervals represent lake level lowstand periods and carbonate dominated intervals represent lake level highstand. If so, it would follow that siliciclastic deposits were deposited updip of carbonate deposits along a single depositional profile. However, from all the data sets examined, the authors have never observed updip siliciclastic deposits transition downdip to carbonate deposits. Instead, siliciclastic and carbonate dominated intervals display along strike and dip continuity in mineralogy dominance across the basin (Figs. 3 and 4). Predictable down dip and along strike facies changes occur, but are constrained within the mineralogically dominant stratigraphic interval. Within siliciclastic dominated intervals, facies change along dip and down dip, according to the siliciclastic environments outlined (Fig. 12). Additionally, within carbonate dominated intervals, facies change predictably downdip from littoral to sublittoral to profundal carbonate deposits (Fig. 12).

**Phase 3: Waning Hyperthermals, EECO**

**Description.** Stratigraphic package 3 is made up of the upper middle Green River Formation, specifically the transitional interval (above the C marker, sensu Remy, 1992) in the western and central Uinta Basin and the equivalent upper Douglas Creek and lower Parachute Creek members to the east (Gall et al., 2017; Figs. 3 and 4). Siliciclastic fluvial (FA1), deltaic (FA2), and/or siliciclastic lake flat (FA3) dominated intervals alternate with carbonate dominated intervals. In the basin center, this alternation is expressed as siliciclastic organic lean (L) zones and carbonate dominated organic rich (R) oil shale zones, respectively. Within Phase 3 strata, alternating R (carbonate) and L (siliciclastic) zones display an overall long term upward transition toward more distal facies and increased tuff occurrence.

Siliciclastic intervals (FA1, FA2, or FA3 dominant) contain minor littoral to sublittoral carbonate deposits (FA4; Figs. 3 and 4). Along the southern outcrop margin of the basin, siliciclastic facies type changes from west to east. In the western Uinta Basin, from Willow Creek/Indian Canyon, siliciclastic lake flat deposits (FA3), which are more mud rich and lower energy than FA1 and FA2, are dominant. Paleocurrent data from F3.1 and F3.2 in the Willow Creek/Indian Canyon region indicate a north-northeast flow direction (Toms, 2014). In the south central Uinta Basin at Gate Canyon, siliciclastic intervals display a mix of fluvial channels (FA1) and deltaic deposits (F2.1, 2.2, 2.4, and 2.5). Paleocurrent measurements from the channels indicate dominant flow direction to the northwest with sediment derived from the south, similar to Phase 2 strata below (Toms, 2014). Fluvial channels present at Gate Canyon are dominated by trough cross stratification (F1.2) or contain a mix of trough cross stratification (F1.2) and upper flow regime structures (F1.1) and are markedly smaller than those found in Phase 2 strata below. Fluvial channels from Main Canyon display trough cross stratification exclusively (Gall et al., 2017). Outcrop and cores on the eastern margin of the basin display a predominance of northeast directed mouthbar deposits (F2.2, 2.4, and 2.5) with little to no evidence of subaerial exposure, suggesting a more distal fluvial source to the southwest (Rosenblum, 2015). Toward the basin center, moving into core to the north and east, siliciclastic dominated intervals or L zones, transition from...
fluvial channels and proximal mouth bar facies into mud to very fine sandstone dominated distal mouth bar (F2.4) and prodelta (F2.5) deposits, representing a proximal to distal facies transect along depositional dip (Fig. 12). In core, distal mouth bar packages contain normally graded beds stacked successively, forming prograding mouth bar complexes, indicating sedimentation outpaced accommodation.

Carbonate dominated packages on the southern basin margin in outcrop contain dominantly littoral to sublittoral carbonate deposits (FA4; Figs. 3 and 4). This includes outcrop on the southern margin from west to east: Willow Creek/Indian Canyon, Gate Canyon, Hay Canyon, Evacuation Creek, and Texas Creek. In the western Uinta Basin at Willow Creek/Indian Canyon, microbialite facies (F4.1) are notably absent. This suggests the lake water chemistry and environmental factors were not favorable for microbialite growth, perhaps due to the low fluvial input and hence possible elevated salinity to the west (Toms, 2014). Toward the basin center, in carbonate dominated packages from cores to the north and east of the southern outcrop margin, littoral, sublittoral, and profundal deposits are present (Fig. 12). This reflects expected lake margin to lake center carbonate facies changes (i.e., deepening) along the ramp profile. Carbonate dominated packages contain 1–5-m-thick shallowing upwards cycles or parasequences that contain profound or sublittoral facies at the base that grade upwards into sublittoral to littoral facies at the top of the cycle (Rosenberg et al., 2015). These are analogous to lake expansion and contraction cycles as described and interpreted from the Wilkins Peak Member, Green River Formation, Wyoming by Pietras and Carroll (Pietras and Carroll, 2006). In R zones within Phase 3 strata, parasequences stack aggradationally to retrogradationally, indicating accommodation was equal to or outpaced sediment supply.

**Interpretation.** Phase 3 in the upper middle Green River Formation is characterized by waning hyperthermal regime conditions in a transgressing lake system (Figs. 3 and 4). Phase 3 strata document a long term trend in rising lake level recording mostly littoral and sublittoral facies (FA2, FA3, FA4, and FA5), with limited evidence of prolonged subaerial exposure.
Siliciclastic packages alternate with carbonate dominated packages, similar to Phase 2 strata, but facies express higher lake level overall. For example, Phase 3 strata largely lack floodplain fine deposits (F1.3) observed in Phase 3 strata. Additionally in Phase 3 strata, carbonate dominated intervals are composed of littoral to sublittoral (FA4) and include profound carbonate deposits (FA5), but similar intervals in Phase 2 strata lack profound deposits (FA5) and are dominated by littoral deposits (F4.1 and 4.2), providing evidence of overall lake transgression between Phase 2 and 3. Organic carbon isotope data from select Phase 3 core intervals were collected, but results are not shown here due to a change in organic matter source upwards through the stratigraphy from terrestrial floodplain sedimentary organic matter in Phase 2 strata to lacustrine algal sedimentary organic matter in Phase 3 strata. Changes in organic matter source rather than carbon cycle fluxes are the main driver of isotopic changes between Phase 2 and 3 strata and preclude comparison in the context of global hyperthermals. Available age constraints indicate Phase 3 was likely deposited during a series of later, lower magnitude or less severe hyperthermal events and during the peak of the EECO (Figs. 1 and 4). The mix of sedimentary structures in fluvial channels found in Phase 3 strata (F1.1 and F1.2) indicate a long term shift from subtropical highly seasonal conditions in Phase 2 (peak hyperthermal regime) to less arid and less seasonal conditions in Phase 3 (waning hyperthermal regime). The alternating siliciclastic (lean zones) and carbonate (rich zones) stratigraphic motif from the transition interval and lower Parachute Creek Member provides strong evidence of continued cyclic, less severe, smaller hyperthermal events upward through the stratigraphy. Similar cyclic eccentricity paced forcing of later, weaker early Eocene hyperthermals has been documented in the marine realm (Sexton et al., 2006) and the neighboring Green River Basin, Wyoming (Aswasereelert et al., 2013; Smith et al., 2014). Phase 3 also likely marks the transition from a semi-arid, highly seasonal climate (Phase 2) to a relatively humid, less seasonal climate (Phase 4).

**Phase 4: Post-hyperthermal**

**Description.** Phase 4 strata include the lower upper Green River Formation, specifically the Mahogany zone and lower R8 interval (Figs. 3 and 4). Basin center cores and outcrop exposures along the basin margins are dominated by profundal deposits (FA5), predominantly organic rich carbonate mudstone or oil shale (F5.2) with minor oil shale breccia (F5.3) and sublittoral organic poor carbonate mudstone (F4.3). Tuffs are common, including the dated Curly tuff at the base of the Mahogany Zone and Wavy tuff at the base of the lower R8. Particularly organic rich beds have been named and correlated around the basin (Johnson et al., 2010).

Evaporite deposits (FA6) occur in three cores in a geographically restricted area of the eastern central basin. These record the first phase of evaporite deposition in the Green River Formation, where stratification of the lake water lead to evaporate precipitation in hypersaline lake bottom waters in the paleo-depocenter of the lake (Vanden Berg and Birgenheier, 2017). On the south central basin margin at Gate Canyon, outcrop exposures of the Mahogany zone display several cm thick beds that contain wave modified current ripples, reflecting sediment input from the Sunnyside Delta system to the south, and episodically shallower conditions. The A groove is composed of dolomite with stylolites in the basin center from core. The stratigraphically equivalent S2 sandstone at Gate Canyon in the south central portion of the basin is composed of wave-modified shoreface deposits. At Willow Creek/Indian Canyon farther to the west, the S2 sandstone is expressed as a distal mouth bar deposit (F2.4) and displays basinward migrating clinoform sets.

**Interpretation.** Phase 4 represents the deepest, most expansive expression of the lake. It is the highest lake level period with density and oxygen stratification, as evidenced by the extensive organic rich carbonate profundal mudstones (FA5) of the Mahogany zone and lower R8 interval (Figs. 3 and 4). Localized hypersaline conditions in the lake paleo-depocenter further support significant density stratification (Fig. 3). Radiogenic dates of tuffs that bound the Mahogany zone (Smith et al., 2008b; Smith et al., 2010) indicate that Phase 4 post-dates all known early Eocene hyperthermal events (Fig. 1). The stratigraphic progression from alternating rich and lean zones recorded in Phase 3 to the overlying sediment starved, organic rich Mahogany and R8 zone in Phase 4 marks the global transition out of the pulsed hyperthermal climate regime. The A groove, above the Mahogany Zone, is thought to represent a period when sediment supply outpaced accommodation, perhaps due to a transient lake shallowing event.

**Phase 5: Post-EECO**

**Description.** Phase 5 strata in the upper Green River Formation includes the R8, the Horsebench Sandstone and saline zone (Fig. 3; Vanden Berg and Birgenheier, 2017). Following localized evaporite deposition during Phase 4 (the first phase of evaporate deposition), the second phase of evaporite deposition occurred during the deposition of the middle to upper R8 zone (Fig. 3). It is characterized by evaporite deposits (FA6) in the eastern portion of the basin (Fig. 3, localities 12–15 and 18–22). The evaporative deposits at the sediment water interface are the result of hypersaline conditions that characterized the closing of Lake Uinta in the east (Vanden Berg and Birgenheier, 2017). The Horsebench Sandstone outcrops in the western and south central portions of the basin at Willow Creek/Indian Canyon and Gate Canyon, respectively. At Gate Canyon, it is composed of wave-influenced shoreface deposits (F2.3) (Toms, 2014). At Willow Creek/Indian Canyon to the west it contains river dominated delta deposits (F2.1 and 2.2) (Toms, 2014).

The third and final phase of evaporite deposition (FA6) as bedded evaporite minerals occurred in the western portion of the basin (Fig. 3, localities 2–4, 6, 8, 9, 11), representing the final phases of lake deposition in the west, while the eastern portion of the lake was infilled with coeval fluvial-deltaic deposits of the Uinta Formation (Fig. 3; Vanden Berg and Birgenheier, 2017).

**Interpretation.** Phase 5 is characterized by post-EECO hypersaline lake conditions and subsequent evaporite mineral deposition (FA6) at the sediment-water interface in the paleo-depocenter (Fig. 3, localities 12–15 and 18–22; Vanden Berg and Birgenheier, 2017). Salinity reflects hydrologic interactions between basins and between the updip catchment area and the downdip basin. Widespread hypersalinity developed in Lake Uinta as the Uinta Basin became the terminal basin with evaporation the only water outlet. Unlike modern sabka or evaporite flat models, the presence of evaporite precipitation in Lake Uinta does not necessarily reflect arid lake margin conditions or climate aridity versus humidity, but instead is more indicative of hydrologic conditions. The Uinta Basin became the terminal basin at this time when the Piceance portion of Lake Uinta was filled with sediment from the north (Tanasu-su-Milkeviciene and Sarg, 2012), pushing saline water and sediment from east to west over Douglas Creek Arch into the Uinta Basin (Johnson et al., 2010). During Phase 5, Lake Uinta in the Uinta Basin also began to fill with sediment from the east to west (Fig. 3), shifting the depocenter in the same direction (Dyce et al., 1985; Vanden Berg and Birgenheier, 2017). The radiogenic age of the Fat tuff that is correlated through Phase 5 strata on the western side of the basin constrains Phase 5 to a post-EECO timeframe (Figs. 1, 2, and 3). The presence of thickly bedded evaporites in the western Uinta Basin suggests high evaporation rates and aridity characterized the lake during this final phase.
DISCUSSION

Early Eocene Climate Records

Early Eocene marine (Zachos et al., 2010; Sexton et al., 2011) and western U.S. terrestrial climate proxy records, such as those from the Bighorn Basin (Abels et al., 2012; Abels et al., 2016) and Greater Green River Basin (Aswasereelert et al., 2013; Smith et al., 2014), have been converging on a 100 k.y. short-eccentricity cycle pacing of early Eocene hyperthermal events. Marine records indicate hyperthermals occurred during eccentricity maxima and lasted ~40 k.y., developing rapidly over the first 5–10 k.y. and decaying more slowly over the subsequent 30 k.y. (e.g., Cramer et al., 2003; Lourens et al., 2005; Zachos et al., 2010; Sexton et al., 2011). However, a terrestrial lithologic and δ13C record from the nearby Greater Green River Basin indicates hyperthermals occurred during eccentricity minima (Smith et al., 2014). Limited age constraints on the early Eocene succession in the Uinta Basin hamper our ability to resolve this debate.

The stratigraphic record of the Green River Formation in the Uinta Basin corroborates lithologic and climate proxy models from nearby basins in two aspects: (1) fluvial expansion is recorded during hyperthermals and (2) lacustrine carbonate dominated facies record inter-hyperthermal periods. The identification of orbitally paced siliciclastic sedimentation alternating with lacustrine sedimentation from the Green River Formation in Wyoming generally matches the stratigraphic motif identified here in the Green River Formation of the Uinta Basin, Utah and further substantiates the hypothesis that climate was a major driver of fluvial sedimentation during the early Eocene (Smith et al., 2014). Other records of the expansion of fluvial facies during the hyperthermal events of the EECO include river avulsion in the Big Horn Basin in Wyoming during orbitally paced warming events (Abels et al., 2012; Abels et al., 2016). In a carbonate isotope record from pedogenic carbonate in the Tornillo Basin, carbon isotope excursions associated with Eocene hyperthermals are not captured, but their predicted stratigraphic positions coincide with thick fluvial sandstone bodies, and are interpreted to record fluctuations in precipitation seasonality and intensity driven associated with global temperature changes (Bataille et al., 2016). Examples showing an expansion of fluvial facies during the PETM from the Spanish Pyrenees and the Piceance Basin, Colorado (Schmitz and Pujalte, 2007; Foreman et al., 2012) provide further evidence for an increase in siliciclastic sedimentation during global warming events. In addition to the documented fluvial expansion, our study of the Green River Formation of the Uinta Basin indicates that seasonally controlled fluvial depositional regimes correspond to hyperthermal events. Because we are able to interpret the stratigraphy basin wide, we can demonstrate the basin wide extent of fluvial-deltaic deposition, both along depositional strike and dip, allowing for expected facies changes related to the paleogeographic position of fluvial input in the lake. We also capture the inter-hyperthermal periods that record lacustrine carbonate dominated facies basin-wide, again allowing for expected updip and downdip as well as lateral changes in facies with geographic changes in lake depth and salinity. The stratigraphic motif of alternating siliciclastic hyperthermal intervals and carbonate inter-hyperthermal intervals characterizes Phase 2 of the lake evolution, with the stratigraphic motif continuing into Phase 3. This is compared to the later portion of the record that lacks hyperthermal events and transitions to perennial, persistent discharge trough cross stratified fluvial systems (Gall et al., 2017).

The magnitude of early Eocene hyperthermal events was small relative to the PETM, recording only ~1% negative δ13C excursions in the marine record (Lourens et al., 2005; Nicolo et al., 2007; Zachos et al., 2010; Sexton et al., 2011); and 3–5% negative δ13C excursions in the terrestrial record (Abels et al., 2012; Abels et al., 2016). The PETM was likely triggered by a large perturbation to Earth’s carbon reservoirs, specifically the release of methane from gas hydrate deposits causing atmospheric warming (Thomas et al., 2002). However, the relatively modest carbon fluctuations associated with early Eocene hyperthermal events point more likely to mass release of sedimentary carbon or to redistribution of carbon between reservoirs at Earth’s surface, specifically astronomically paced release of marine dissolved organic carbon from the anoxic Southern Ocean, as advocated by Sexton et al. (2011). The fluvial-lacustrine record from the Uinta Basin indicates low organic carbon preservation along with increased fluvial sediment delivery during hyperthermal events. This is coupled with organic rich deposition and significant lacustrine carbon sequestration between hyperthermal events. Carbon sequestration between hyperthermal events offers a negative feedback mechanism for warming events. The mass balance and timing of early Eocene lacustrine organic carbon sequestration globally should be further evaluated quantitatively, particularly as Eocene lacustrine source rocks are globally widespread (e.g., China continental interior basins). Recent research that compares marine, northern hemisphere terrestrial and southern hemisphere terrestrial carbon isotope records indicate that EECO was characterized by different responses in the marine and terrestrial realm globally (Hyland et al., 2016).

Advancing Lacustrine Models

The three-part lake classification scheme developed by Carroll and Bohacs (1999, 2001) provides a useful starting point for basin-wide characterization but fails to characterize the complexity and detail of the evolution of Lake Uinta in the Uinta Basin. In this model overfilled, underfilled, and balance filled conditions are primarily defined based on the facies association present, which predict the sediment and water fill to accommodation ratio (Carroll and Bohacs, 1999, 2001). Using this scheme the lower Green River Formation Phase 1 and 2 strata contain the fluvial-lacustrine facies association and record overfilled conditions, where sediment and water fill outpaced accommodation (Fig. 3). Phase 3 strata contain the fluctuating profound facies association, recording balance filled conditions where sediment and water fill approximately balanced accommodation. Finally, Phase 4 and 5 strata contain the evaporative facies association and record underfilled conditions, where accommodation outpaced sediment and water fill.

Tectonic Controls

Our data indicate that, in the Uinta Basin, steady subsidence driven accommodation was overlain by varying rates of sediment supply through time. Long-term tectonic subsidence and flexure created by Laramide orogenic features, like the Uinta Mountains to the north and San Rafael Swell and Uncompaghre Uplift to the south, was responsible for creating lake basin accommodation through time. This shift from overfilled (Phase 1 and 2) to balance filled (Phase 3) conditions was a result of increased accommodation driven by tectonic subsidence and/or tectonic silt uplift. This was coupled with an overall decrease in siliciclastic sediment supply between phases 2 and 4, as interpreted from the stratigraphic record of fluvial-lacustrine deposits herein. As a function of these two controls, maximum lake depth increased between phases 2 and 4.

In fact, our paleocurrent data support the emergence and uplift of the Douglas Creek Arch, the north-south–trending Laramide structure that acted as a topographic silt between the Uinta and Piceance basins, between Phase 2 and 3 strata. Phase 2 strata from Texas Creek display consistent paleocurrent flow to the north-northeast, showing no influence from an emergent hinterland to the east (Fig. 2; Rosencrans, 2015). In contrast Phase 3 strata from the Evac-
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Evolution Creek region display a paleocurrent flow to the west-northwest, suggesting an emergent fluvial source to the southeast from the Douglas Creek Arch.

The transition to underfilled conditions during Phase 4 and 5 suggests accommodation increased again relative to sediment and water fill. Increased accommodation was likely driven by continued or increased tectonic subsidence. Stratigraphic evidence points to a water connection between the Uinta and Piceance basins during the deposition of the Mahogany zone (Tanavski-Milkeviene and Sarg, 2012, 2017). The Uinta Basin became the terminal basin during Phase 5, receiving water and sediment from neighboring basins. For these reasons, sill uplift was likely not the cause of increased lake basin accommodation during Phase 4 and 5. Low siliciclastic sediment supply characterized Phase 4 deposition. In Phase 5 strata, low sediment plus water fill conditions are evidenced by the abundance of profunidal mudstone hosted evaporite, suggesting relatively low sediment supply conditions plus high evaporative rates from aridity characterized the lake system.

Evidence of tectonically controlled major watershed capture events that have been documented in the Greater Green River Basins (e.g., Carroll et al., 2008; Smith et al., 2008b) were likely not characteristic of the Uinta Basin, mainly because the paleo-California river system that traveled to the Uinta Basin from the south was established and long-lived (late Paleocene to early Eocene; Davis et al., 2010; Dickinson, 2012).

Climate Controls

Classical models view climate controls through the lens of lake contraction and expansion associated with temporal or geographic changes in humidity and aridity conditions (Carroll and Bohacs, 1999; Pietras and Carroll, 2006). Though accurate and useful, the approach cannot capture the full range of possible climatic controls on the fluvial-lacustrine system or the high resolution facies and stratigraphic variation observed.

Instead, the lake model presented here is distinctive in two ways. First, it considers climatic controls beyond the lake footprint or lake level itself as an indicator of general aridity and humidity. Instead, the study focuses on sedimentologic indicators of climatically driven fluvial sedimentation style. It considers the integrated lake system with updip fluvial-deluvial sediment delivery and the impact on/response of the downdip lake system.

Secondly, the model herein is unique as compared to previous models in distinguishing longer timescales of climatic control. Fluvial-lacustrine system response to global warming and cooling occurred at least on two timescales: (1) hyperthermals (40 k.y. to 100 k.y. cycles) and (2) the longer term (my-scale) reflecting the waxing and waning of hyperthermals as well as the EECO.

During seasonal or episodic (not necessarily seasonal) flooding events that characterized hyperthermal events on the 40 k.y. to 100 k.y. timescales, the fluvial system delivered a large volume of water and sediment into the downdip lake system. This is typical of modern highly seasonal fluvial systems (Plink-Björklund, 2015), and has been demonstrated in the highly seasonal Burdekin River of Queensland, Australia (Alexander et al., 1999; Fielding, 2005; Fielding and Alexander, 1996). In many semi-arid to arid settings, because water deposited during flooding events is transient and evaporative, and discharge rates are quite low between sharply peaked hydrograph events (Alexander et al., 1999; Plink-Björklund, 2015). Similarly, because of the high evaporation potential in an overall semi-arid climate setting of Lake Uinta, the water delivered to the lake seasonally was likely transient and evaporative. In contrast, fluvial sediment flushed into the lake during seasonal flood events had no outlet, entered the lake basin sink, and remained as an archive of the event. In this sense, sediment and water fill were decoupled during hyperthermal regime conditions, operating on different time scales. This is an important distinction and is in contrast to classical models in which sediment and water fill are coupled over long geologic timescales (Carroll and Bohacs, 1999; Carroll and Bohacs, 2001). In contrast, modern perennial fluvial systems show a strong coupling between water and sediment delivery (Plink-Björklund, 2015). Between hyperthermal events, during periods of lower seasonality of fluvial sediment delivery, sediment and water input were likely coupled, with fewer high magnitude flood events. As a result, preserved siliciclastic sediment in the lake during these intervals was lower relative to the hyperthermal events. During these intervals, lacustrine carbonates and organic rich carbonate mudstone facies accumulated, relatively undiluted by siliciclastic sediment. Carbonate precipitation was an ongoing background process in Lake Uinta and record siliciclastic depositional pauses locally or regionally (where stratigraphically applicable). On million year timescales, the frequency and magnitude of documented hyperthermal events also waxed and waned. Changes in fluvial style between Phases 2, 3, and 4 follows a longer term transition from peak hyperthermal regime to waning hyperthermal regime to post hyperthermal regime conditions on million year timescales. Specifically, highly seasonal fluvial systems (Phase 2) transition to mixed seasonal and perennial discharge systems (Phase 3) to low siliciclastic sediment delivery (Phase 4).

The expansion of Lake Uinta during Mahogany Zone deposition (Phase 4) is typically attributed to tectonic mechanisms (Smith et al., 2008b), however, here we propose that regional to global climate transitions may have also played a fundamental role. A recorded, major stream capture event in the northern Greater Green River Basin, associated with increased volcaniclastic sediment delivery from the Absaroka and Challis volcanic field is thought to have caused freshening of Lake Gosiute in the Greater Green River Basin and the associated sediment fill is thought to have caused the closing of Lake Gosiute soon afterward (Smith et al., 2008b). Stream capture (increased fresh water delivery) combined with the ensuing sediment closure of Lake Gosiute caused water to overflow the lake margin of Lake Gosiute and spill over the topographic sill to the south into the Piceance Basin of Lake Uinta to the Piceance Basin and then over the silt to the Uinta Basin, resulting in lake expansion in the Piceance and Uinta basins during Mahogany deposition (Carroll et al., 2008). Because the Piceance Basin of Lake Uinta was significantly more saline than the Uinta Basin, saline water spilled over from the Piceance Basin to the Uinta Basin, which resulted in an overall increase in salinity in the Uinta Basin during Phase 4 deposition (Kelts, 1988).

We propose an alternate climate driven interpretation in which the stratigraphic progression upwards from alternating rich and lean zones (Phase 3) to the overlying, organic rich Mahogany zone and R8 zones (Phase 4) was driven by the global transition out of the pulsed hyperthermal climate regime, at which time the lake became deeper and stratified for prolonged periods. Notably, the most organic rich oil shale deposits (Mahogany zone) in the Uinta Basin post-date the last documented early Eocene hyperthermals as the main EECO (Fig. 1). Between R4 and the Mahogany zone (R7) in the Uinta Basin, the oil shale deposits become richer upwards (R4: 6.6 gpt average; R5: 7.4 gpt average; R6: 8.5 gpt average; Mahogany, R7: 17.1 gpt average; lower R8: 9.7 gpt average) (Vanden Berg and Birgenheier, 2016). Both increasing organic richness upward trends along with the transition from littoral to sublittoral carbonate facies to sublittoral to profundal carbonate facies record an overall lake transgression in the Uinta Basin. The end of the EECO was characterized by progressively weaker to absent intervening hyperthermal conditions (Sexton et al., 2006, 2011), concurrent with longer term transition from highly seasonal to less seasonal, increasing
stable climate conditions and fluvial regimes, as evidenced in the transition from Phase 2–3 to 4 in the Green River Formation. In fact, sediment and water delivery from highly seasonal fluvial systems active during hyperthermals may have progressively impacted lake levels through time on million year timescales. The less seasonal fluvial systems delivered a higher water to sediment ratio associated with more year round humidity and less seasonal climate conditions. As hyperthermals waned and climate became less seasonal and more humid, more water was delivered to the lake through time. As a result, lake levels rose.

CONCLUSIONS

Acknowledging long term tectonic controls on basin and lake system evolution is critical, but when the lens is shifted toward more complex climate drivers and interactions as understood from recent literature, we can improve on the existing interpretation framework and allow for a better understanding of the integration and evolution of linked fluvial-lacustrine systems. Specifically, this study provides an ancient example of fluvial expansion in a semi-arid climate and trends toward higher seasonal variability in water and sediment discharge associated with transient warming events that mark early Eocene hyperthermal events. This trend is reversed with the waning of hyperthermals, recording a shift toward humidity, and less seasonality in water and sediment discharge, along with lake expansion. Climate controlled the evolution of the fluvial-lacustrine system on at least two timescales, both the thousand and million year timescales, representing individual hyperthermals and the long term waxing and waning of the early Eocene hyperthermal regime, respectively. The evolution of Lake Uinta in the Uinta Basin is divided into five climatically-driven lake phases:

1. **A post-PETM phase** of stable non-seasonal climate recording extensive fresh water lake conditions.

2. **A peak hyperthermal phase** with large upper flow regime-dominated fluvial bodies deposited in a highly seasonal, semi-arid climate that recorded the early Eocene eccentricity paced hyperthermal events with intervening carbonate dominated periods of low siliciclastic sediment supply and less seasonal climate conditions recording inter-hyperthermals. The lake was relatively shallow exhibiting largely littoral to sublittoral carbonate facies on the margin, with profuse facies in basin depocenter. Siliciclastic facies transitioned from fluvial-deltaic on the margin to distal deltaic in the basin center in siliciclastic-dominated intervals.

3. **Waning hyperthermal, EECO phase** with continued alternation between siliciclastic fluvial-deltaic packages and carbonate/oil shale lacustrine packages that may reflect later weaker hyperthermal events and inter-hyperthermals, respectively. Carbonate facies in carbonate-dominated intervals ranged from littoral on the margin to profundal in the depocenter, with a shift toward a higher proportion of profundal facies, reflecting lake deepening. 

4. **A post-hyperthermal phase** with density-stratified and anoxic high lake level conditions where extensive organic rich oil shale deposits accumulated. Siliciclastic sediment supply was low. This phase is interpreted to represent the global transition out of the pulsed hyperthermal climate regime.

5. **A post-EECO phase** in which the Uinta Basin became the terminal lake basin for the Uinta-Gosiute system, recording hypersaline conditions and the closing of Lake Uinta from east to west.

The climatic control on fluvial-lacustrine deposition at various scales of resolution presented in this study shifts the perspective of lake deposition toward climatic control during the early Eocene. Specifically, the model here provides a link between climate controls on hinterland fluvial systems and the downwip response of the lake system to water and sediment, whereas previous models focus mainly on lake contraction and expansion associated with humidity and aridity conditions. This study provides a terrestrial palaeoclimate proxy model based largely on sedimentary indicators that can be used as the basis for future geochemical palaeoclimate proxy records to test more specific aspects of terrestrial early Eocene climate shifts.

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