Faulty foundations: Early breakup of the southern Utah Cordilleran foreland basin

Gabriela A. Enriquez St. Pierre and Cari L. Johnson
University of Utah, Department of Geology and Geophysics, 115 S 1460 E, Salt Lake City, Utah 84112, USA

ABSTRACT

Anomalous features of Upper Cretaceous strata in southern Utah challenge existing tectonic and depositional models of the Cordilleran foreland basin. Extreme thickness variations, net to gross changes, and facies distributions of nonmarine to marginal marine strata of the Turonian–early Campanian Straight Cliffs Formation are documented across the Southwestern High Plateaus. Contrary to most traditional models of foreland basin architecture, regional correlations demonstrate abrupt stepwise thickening, with a punctuated increase in average grain size of key intervals from west to east, i.e., proximal to distal relative to the fold-thrust belt. Except in the most proximal sections, fluvial drainage systems were oriented predominantly subparallel to the fold-thrust belt. Combined, these results suggest that modern plateau-bounding faults may have had topographic expressions as early as Cenomanian time, and influenced the position of the main axial river system by creating northeast-trending paleotopography and sub-basins. Laramide-style tectonism (e.g., basement-involved faults) is already cited as a driver for sub-basin development in latest Cretaceous–Cenozoic time, but new data presented here suggest that this part of the foredeep was “broken” into distinct sub-basins from its earliest stages. We suggest that flexural foundering of the lithosphere may have caused early stage normal faulting in the foredeep. Regional implications of these new data indicate that both detachment-style and basement-involved structures were simultaneously active in southern Utah earlier than previously recognized. These structures were likely influenced by inherited Proterozoic basement heterogeneities along the edge of the Colorado Plateau. This interpretation suggests that tectonic models for the region should be reevaluated and has broader implications for understanding variability and geodynamics of foreland basin evolution.

INTRODUCTION

Global archives suggest that retroarc foreland basins are broadly defined by several common characteristics (Jordan, 1981; DeCelles and Giles, 1996; Catuneanu, 2004; DeCelles, 2012). These fundamental attributes include (Fig. 1): a basinward-propagating, thin-skinned (detachment-style) fold-thrust belt; concomitant migration of four main depositional zones (wedge-top, foredeep, forebulge, and backbulge); a strongly asymmetric, continuous foredeep with greatest subsidence most proximal to the flexural load; clastic material sourced largely from the fold-thrust belt (particularly in proximal sections); and general coarse to fine-grained facies transitions across the depozones from proximal to distal.

Many of these characteristics are well expressed in the North American Cordillera, particularly in the Cretaceous Sevier fold-thrust belt and foreland basin system of Utah (Armstrong, 1968; Jordan, 1981; Lawton et al., 1994; DeCelles, 1994; DeCelles and Lawton, 1996; DeCelles et al., 1996; Currie, 2002; DeCelles, 2004; Fig. 1). Variations on the broad conceptual themes of foreland basin evolution are also widely acknowledged. For example, proximal to distal fining grain size trends can be disrupted by complex sediment routing systems, such as large axial rivers and distributive fluvial systems (DeCelles and Cavazza, 1999; Hartley et al., 2010; Weissmann et al., 2010), longshore drift (Swift et al., 1987; Slingerland and Keen, 1999), progradation of major delta systems (Ahmed et al., 2014; Fielding, 2015), and tidal deposits (Chentnik et al., 2015; Van Cappelle et al., 2018). Similarly, punctuated and widespread deposition of gravel and sand sheets extending beyond the proximal foredeep are documented (Weissmann et al., 2011; Lawton et al., 2014), and these deposits likely reflect redistribution of sediment during isostatic rebound, and/or syntectonic deposition during active thrust phases (Burbank and Beck, 1988; Garcia-Castillanos, 2002; Horton et al., 2004; Thomson et al., 2017). Although flexure is considered the main subsidence control on foreland basin evolution and forebulge migration (Catuneanu et al., 1997; Currie, 1998; Houston et al., 2000), dynamic subsidence is also an important potential driving mechanism (Mitrovica et al., 1989; Liu and Nummedal, 2004; Romans et al., 2010; Painter and Carrapa, 2013; Fosdick et al., 2014). Finally, onset of Laramide-style (basement-involved) tectonism in latest Cretaceous time (ca. 75–50 Ma; Dumi- tru et al., 1994; Crowley et al., 2002; Saleeb, 2003; Tindall et al., 2010; Peyton et al., 2012; Weil and Yonkee, 2012; Yonkee and Weil, 2015) fundamentally changed Cordilleran foreland basin architecture, subdividing the foredeep into multiple isolated basins thus forming a “broken foreland” (Dickinson et al., 1988; Schmidt et al., 1993; Erслev, 1993; Strecker et al., 2011; Dávila and Carter, 2013). Evolution of the Sevier fold-thrust belt and foreland basin system in southern Utah (Fig. 1B) is poorly understood relative to other parts of the Cordillera. Major thrust structures appear to have been emplaced by the Cenomanian (DeCelles, 2004), with little evidence of major basinward propagation over time. Preliminary data from foreland basin strata suggest unusually high accommodation, sediment supply, and coarse-grained deposition in distal parts of the foredeep, with relatively stationary (non-migrating) depozones (Eaton and Nations, 1991; Goldstrand et al., 1993; Mulhorn and Johnson, 2016). Although flexure adjacent to the Sevier fold-thrust belt is thought to be a primary control on subsidence and sedimentation patterns in northern Utah and Wyoming before ca. 83 Ma (DeCelles, 2004), previous studies of southern Utah foreland basin stratigraphy indicate a more complicated history from at least Cenomanian time, long before onset of Laramide-style tectonics (Gustason, 1989; Eaton and Nations, 1991; Laurin and Sageman, 2001).

Cenozoic normal faults bound a series of plateaus in the study area (Southwestern High
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Plateaus, SWHP; Fig. 2), exposing relatively undeformed Upper Cretaceous strata ~200 km west to east (proximal to distal relative to the fold-thrust belt), and ~50 km of outcrop from south to north. Whereas fluvial architecture along the western Kaiparowits Plateau has been studied in some detail (Gooley et al., 2016; Ben-hallam et al., 2016; Primm et al., 2018; Koch et al., 2019), much less is known about fluvial systems to the west in the Markagunt and Paunsaugunt Plateaus, which preserve more proximal deposits of the Cretaceous foreland basin. Regional correlations to date focus mainly on biostratigraphy and marine deposits, but do not detail regional fluvial interactions (Eaton and Nations, 1991; Eaton et al., 1993; Goldstrand, 1992, 1994; Moore and Straub, 2001; Biek et al., 2015).

We completed a regional correlation across the southern margin of the Sevier foreland basin to investigate controls on foreland basin development, and in particular their effect on fluvial systems during Late Cretaceous time. Targeted measured sections document stratigraphic architecture, proximal-to-distal changes in thickness and average grain size, and paleocurrent indicators. Ultimately, these data allow us to infer controls on sedimentation and to test prevailing models for foreland basin development.

GEOLOGIC BACKGROUND

Structural Evolution of Southern Utah

The Sevier fold-thrust belt trends northeast across Utah, deforming and exhuming Paleozoic and Mesozoic strata which provided sediment to the adjacent foreland basin to the east (DeCelles and Coogan, 2006). At its southwestern extent, the Sevier fold-thrust belt formed a high-angle junction with the Mogollon Highlands (Fig. 1) in northwestern Arizona. The geometry of these features allowed for incursion of the Utah Bight, an embayment filled by the Cretaceous Western Interior Seaway (Fig. 2 inset; Zapp and Cobban, 1960; McGookey et al., 1972).

The Sevier fold-thrust belt in northern and central Utah shows evidence for eastward propagation of thin-skinned thrust sheets, punctuated periods of out-of-sequence thrusting, and growth of structural culminations, all of which influenced subsidence and sediment supply (Fig. 1B; DeCelles et al., 1995). The Canyon Range thrust system represents the first...
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major thrusting episode in central Utah, initiating during Barremian–early Aptian time (ca. 130–120 Ma; DeCelles and Coogan, 2006). The Pavant thrust system propagated east during Late Aptian–Cenomanian time. By the Cenomanian (ca. 100 Ma), the Paxton thrust sheet initiated growth of the Sevier culmination (DeCelles et al., 1995). This culmination included reacti-

Figure 2. Inset map shows present day location of the Southwestern High Plateaus (SWHP) in Utah including relevant structures from Late Cretaceous to Cenozoic time. WIS—Western Interior Seaway; AZ—Arizona; CA—California; CO—Colorado; NM—New Mexico; NV—Nevada; UT—Utah; WY—Wyoming. Main map of the study area shows the locations of measured sections and relevant stratigraphy across the SWHP. Sections of the Straight Cliffs and Iron Springs Formations completed for this study are shown in orange. Generic lithologies are represented by brick (mixed limestone and sandstone) and stippled (dominantly sandstone) patterns; non-stippled is a mixture of mudstone and sandstone lithologies. Sections from this study (orange) and previous studies include: (1) Pine Valley Mountains (Cook, 1957); (2) Parowan Gap west of the SWHP; (3) Parowan Canyon; (4) Cedar Canyon (Eaton et al., 2001); (5) Orderville Gulch in the Markagunt Plateau; (6) Glendale and (7) Tenny Canyon (Tilton, 1991); (8, 9) Willis Creek and Heward Creek from the Paunsaugunt Plateau; and (10) Shakespeare Mine from the northern Kaiparowits Plateau. Previously measured sections in the Kaiparowits Plateau include: (11) Buck Hollow (Mulhern and Johnson, 2016); (12) Main Canyon (Chentnik et al., 2015); (13) Left Hand Collet (Dooling, 2013); (14) Rogers Canyon (Allen and Johnson, 2010); (15) Kelly Grade (Gallin et al., 2010); (16) Tibbet Canyon (Peterson, 1969); (17) Bull Canyon (Gooley et al., 2016); and (18) Rock House Cove (Gooley et al., 2016).
Miller, 1966; DeCelles, 2004; Fig. 1B). The Iron Springs thrust system is the youngest and easternmost expression of the Sevier fold-thrust belt in southern Utah, and it is characterized by three east-vergent, forward-breaking thrust sheets. It is estimated to have been active in Santonian-Campanian time (ca. 84–70 Ma; Goldstrand, 1994), but drag folds in Paleocene-Eocene conglomerates indicate minor late stage displacement as well (Anderson and Dinter, 2010).

Present-day structure and topography in the study area is reflected in the Markagunt, Paunsaugunt, and Kaiparowits Plateaus, which together comprise the Southwestern High Plateaus. These modern features are bounded by Cenozoic normal faults (from west to east: the Hurricane, Sevier, and Paunsaugunt faults), which mark the transition between the Basin and Range and Colorado Plateau provinces (Fig. 2). These faults have offsets of up to ~1000 m, and are thought to have initiated during the early Miocene, with the Hurricane fault showing modern active seismicity (Maldonado et al., 1997; Davis, 1999; Lund et al., 2008; Biek et al., 2015).

Stratigraphic Evolution of Southern Utah

The SWHP preserve relatively undeformed Upper Cretaceous strata of the Cordilleran foreland basin system. These deposits include marine to fluvial-alluvial strata of the Natura (formerly Dakota), Tropic, Straight Cliffs, Wahweap, and Kaiparowits Formations (Fig. 3; Peterson, 1969; Gustason, 1989; Eaton and Nations, 1991). The Straight Cliffs Formation, of primary interest to this study, overlies the marine Tropic Shale and records the initial regression of the Western Interior Seaway (Fig. 3, Peterson, 1969). Depositional environments of the Straight Cliffs Formation transition from proximal fluvial-alluvial systems west of the Markagunt Plateau to coal-bearing marginal marine strata in the eastern Kaiparowits Plateau (Eaton and Nations, 1991; Little, 1997). Changes in fluvial and marginal marine stacking patterns within the Straight Cliffs Formation have been attributed to sea-level fluctuations (Shanley and McCabe, 1993) and/or interactions between axial and transverse drainages, which suggest tectonic and possibly climatic influences on deposition (Lawton et al., 2014; Szwarc et al., 2015; Gooley et al., 2016; Primm et al., 2018). Recent geochronologic data from the Straight Cliffs Formation indicate that deposition spans Turonian–early Campanian time (ca. 94–81 Ma; Primm et al., 2018; Fig. 3).

The Tibbet Canyon Member of the Straight Cliffs Formation is a thick shoreface sandstone deposited during shoreline regression overlying the marine Tropic Shale (Peterson, 1969; Shanley and McCabe, 1991, 1995; Fig. 3). It is overlain by the Turonian-Coniacian fluvial and paralic (marginal marine), coal-bearing Smoky Hollow Member (Bobb, 1991; Hettinger et al., 1993; Hettinger, 2000). This unit is capped by the Calico Bed, a distinctive regional marker unit composed of gravelly sheet-sands (Peterson, 1969; Bobb, 1991; Little, 1997). In the eastern Kaiparowits Plateau, a minor unconformity (ca. 87–89 Ma) divides the Calico Bed into two informal units: a lower amalgamated fluvial section and an overlying estuarine-influenced section (Primm et al., 2018).

The Coniacian–Santonian John Henry Member comprises the thickest part of the Straight Cliffs Formation and preserves fluvial-alluvial deposits in the west, with marginal marine and offshore deposits in the eastern Kaiparowits Plateau (Peterson and Waldrop, 1965; Eaton and Nations, 1991; Shanley and McCabe, 1995; Hettinger et al., 1996). Marine deposits record shoreface and barrier island–lagoon systems with initially mainly transgressive then regressive stacking patterns throughout the Santonian (Hettinger et al., 1993; Shanley and McCabe, 1995; Allen and Johnson, 2010; Mulhern and Johnson, 2016). Extensive coal zones occur within the John Henry Member of the central Kaiparowits Plateau, which are correlated to fluvial sections in the west that vary in terms of fluvial style and tidal influence (Shanley and McCabe, 1993; Hettinger et al., 2009). The upper boundary of the John Henry Member records a transition to higher-energy, coarser grained fluvial sheet-deposits of the basal Drip Tank Member (Lawton et al., 2003, 2014; Gooley et al., 2016).

Unconformities throughout the Cretaceous sections are important considerations for interpreting thickness variations. Unconformities within the Straight Cliffs Formation are considered relatively minor sequence boundaries, with minimal erosion into underlying sections (Peterson, 1969; Biek et al., 2015; Primm et al., 2018). More significant unconformities across the SWHP occur in latest Campanian-Maastrichtian successions, between deposition of the Kaiparowits, Canaan Peak, and Claron Formations, and are principally located in the Paunsaugunt and Kaiparowits Plateaus (Fig. 3; Goldstrand, 1994; Biek et al., 2015; Beveridge et al., 2020). These unconformities are largely attributed to Laramide-style deformation and formation of structures like the East Kaibab and Dutton monoclines (Roberts, 2007; Hilbert-Wolf et al., 2009; Lawton and Bradford, 2011). To date, there is little evidence that sub-Claron-aged erosion significantly affected the Straight Cliffs Formation in this area (Lawton et al., 2003; Biek et al., 2015).

METHODS

Measured sections across the SWHP document stratigraphic architecture (Figs. 4, 5, and 6) as well as proximal-to-distal changes in thickness, grain size, and paleocurrent indicators (Fig. 7). Six new measured sections of the John Henry Member are combined with previous work from Tilton (1991) in the Paunsaugunt Plateau to understand basin-wide stratigraphic and thickness trends. Net to gross
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(NTG) is calculated by measuring sandstone thickness (net) relative to total section thickness (gross) and reported as a decimal number (0.0–1.0, net thickness/gross thickness). Paleocurrent data were collected from trough axes, ripples, and barforms where possible, and combined with previous work of Tilton (1991) to create a more holistic picture of paleoflow directions (Fig. 2).

Correlation across the SWHP from proximal to distal (west to east) is based on the formation descriptions of Peterson (1969) from the Kaiparowits Plateau, and extrapolated to the west with the aid of publications, theses, and regional reports (Tilton, 1991; Eaton and Nations, 1991; Moore and Straub, 2001; Lawton et al., 2014; Biek et al., 2015), as well as new measured sections presented herein. Regional marker beds used for correlation within the Straight Cliffs Formation include the Tibbet Canyon Member shelf and Calico Bed and Drip Tank Member gravelly sheet sands (Fig. 3). Although these widespread deposits are likely to be at least partly time-transgressive, previous studies and available age-control suggests they are reasonable datums relative to the age of the entire section and are suitable to guide regional correlations (Peterson, 1969; Lawton et al., 2014; Biek et al., 2015).

RESULTS

Lithofacies and Facies Associations

Thirteen lithofacies were identified through field observations (Table 1), and these were grouped into four facies associations (Table 2: FA-1 to FA-4) that are distinguished by distinct channel architectural patterns and recurring associations of lithofacies. Channel facies associations were distinguished based on channel architecture, bed boundary relationships, and grain-size distributions (e.g., Miall, 1996; Slingerland and Smith, 2004). Specific channel-fill elements include bedforms and individual bed boundary relationships that fill a single channel (e.g., scour-and-fill structures). Channel stories are defined as a single channel unit with related channel-fill elements, separated by conspicuous basal scour that are laterally continuous. Channel belts are defined as laterally and/or vertically coalescing channel stories that form a single elongate sandstone body (Gibling, 2006; Ford and Pyles, 2014).
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Figure 4. Representative stratigraphic column of the John Henry Member from Orderville Gulch (Markagunt Plateau, Fig. 2) showing the relative abundance of Facies Association 1 (FA-1) and FA-3 within the section (see Fig. 7 for key to stratigraphic columns). Outcrop photographs illustrating the differences between well-drained and poorly drained floodplain lithofacies of FA-1 and FA-2, and small, laterally restricted fluvial channels found within these facies. (A) Carbonaceous mudstone overlain by siltstone lithofacies from poorly drained floodplain deposits. (B) Variegated mudstone with blocky texture, (C) Siltstone to very fine-grained sandstone lithofacies. (D) Isolated asymmetric sandstone body with basal cross-bedding. (E) Example of poorly drained floodplain carbonaceous mudstones and coal, showing light red and yellow paleosol development.

Figure 5. Representative stratigraphic column featuring facies association 3 (FA-3) and FA-1 within the John Henry Member at Heward Creek (see Figure 7 for key to stratigraphic columns). Outcrop photographs show: (A) Channel thalweg with erosive scour surfaces outlined in black. Note heterolithic nature of channel margins; (B) Distinct ribbon “steer’s head” channel belt architecture in cross section with upper and lower boundaries outlined from Heward Creek section; (C) Well-developed gray mudstone lithofacies overlain by tan siltstone facies commonly found above and below isolated sandstone channels; (D) Soft sediment deformation typical within thick sandstone units; (E) Horizontal unlined burrow clusters found within thick sandstone bodies.

mottled coloring (Fig. 4B), and may contain plant rootlets.

Facies association 2 (FA-2; Fig. 4) consists of organic-rich fine-grained lithofacies (Fc) and coal (C) with minor components of isolated sandstones (Table 1: Fc and C; Fig. 4A, 4E). Fine-grained lithofacies of FA-2 are expressed as structureless beds ranging from red, yellow, light gray, and black to charcoal gray in color (Fm, Fl, Fc, C). Coal seams (sub-bituminous grade) are common and are up to

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1 m thick in the northern Kaiparowits Plateau (Fig. 4A). Darker carbonaceous shales are present, with laterally variable oxidized lamina, and siderite concretions (Fig. 4A). Plant, coal, and woody fragments exist throughout the carbonaceous mudstones.

Both FA-1 and FA-2 contain distinctive fine-to medium-grained, isolated sandstone units/bedsets (Fig. 4D). These sandstone bedsets are up to 4 m thick and less than ∼10 m wide laterally. FA-1 and FA-2 sandstones commonly exhibit a lenticular geometry (both symmetric and asymmetric) but can also be tabular, and are no more than 2 stories (Fig. 4D). Individual beds are composed of very fine- to medium-grained sand. Bedsets thin upwards from 0.1 to 0.5 m thick. Sedimentary structures within FA-1 typically include soft sediment deformation (Sc), low-angle cross-lamination (Sl), and asymmetric ripples, particularly at the tops of sandstone bedsets (Fig. 4D). Sedimentary structures within FA-2 include asymmetric ripples, planar to low-angle cross-bedding (Sl), with no visible structures in the thinnest beds (Sm). Bioturbation is variable and infrequent for both FA-1 and FA-2, and where present, individual Skolithos traces are observed. Vertical successions of sandstones within FA-1 and FA-2 include basal beds with erosive bases, and trough cross-stratification

Figure 6. Representative stratigraphic column of facies association 4 (FA-4) within the Iron Springs Formation at Parowan Gap (see Fig. 7 for key to stratigraphic columns). Note the prevalence of internal scouring, trough cross-bedding, and woody fragments. Outcrop photographs show: (A) sheet-like channel sandstones with erosive surfaces marked in black with scour surfaces (s); (B) soft sediment deformation features; (C) pebble lags at the base of channels; (D) a representative photo of FA-4 within the Calico Bed of the Smoky Hollow Member at Orderville Gulch, showing large trough cross-bedding, and internal scouring of sandstone units; and (E) preserved mudstone and siltstone fine-grained layers between sandstone channels.

Figure 7. Correlation of John Henry Member strata across the Sevier and Paunsaugunt faults, showing the relative sandstone to mudstone rations (net to gross, NTG), grain size, and thickness across the Southwestern High Plateaus (SWHP). Inset map shows cross section line across SWHP; see Figure 2 for detailed section locations. Upper datum of the correlation is the base of the Drip Tank Member, the top Calico Bed (upper Smoky Hollow Member) is the lower correlation marker. KP—Kaiparowits Plateau; MP—Markagunt Plateau; PP—Paunsaugunt Plateau.

Correlation of the John Henry Member

| Key | | Modern plateau-bounding normal faults (schematic) |
|---|---|---|
| | | |
| Mudstone dominated lithofacies | *Wood casts | FA-1: Well-drained floodplain |
| Sandstone dominated lithofacies | Concretions | FA-2: Poorly-drained floodplain |
| Convolute bedding | Paleocurrent data from this study | FA-3: Major channels in well-developed floodplain |
| Pebble lags | | FA-4: Fluvial sheet deposits |

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that transitions upwards to planar or low-angle cross-stratified beds, capped by ripple-laminated sand beds.

**Fine-grained Facies Associations (FA-1 and FA-2) Interpretation.** FA-1 represents well-drained floodplain deposition between active fluvial channels. Siltstone and sandstone units represent overbank fines deposited during flooding events when bankfull discharge is exceeded and channel levees are breached (Slingerland and Smith, 2004). The blocky mudstone structures, motled coloring, and plant rootlets represent weakly developed paleosols and indicate brief periods of subaerial exposure between active channels and well-drained floodplains (Kraus and Wells, 1999; Kraus, 2002). Sandstone lithofacies represent crevasse splay deposits with crevasse-channel development in some areas; these formed as the suspended and bedload material within a channel spills into the floodplain during levee-breaching flooding events (Smith et al., 1987; McCabe and Shanley, 1992). This rapid plant accumulation occurred under humid tropical conditions, as was common during deposition of the John Henry Member (Wolfe and Upchurch, 1987; McCabe and Shanley, 1992). Sandstone lithofacies are interpreted as overbank splay deposits in which channels were breached during times of high discharge or flooding events (Slingerland and Smith, 2004; Donselaar et al., 2012). Lenticular and tabular sandstone lithofacies are interpreted as minor channel-fill successions, abandoned during channel migration or avulsion (Slingerland and Smith, 2004).

**Channel-Fill Facies Associations (FA-3 and FA-4)**

Two channel facies (FA-3 and FA-4) were distinguished based on the geometry of the sandstone body, specifically narrow and discrete channel belts versus broader sheet-like units.

**Description FA-3.** Facies association 3 (FA-3) consists of sandstone-rich lithofacies (Fig. 5), including very fine- to coarse-grained sandstones (St, Sr, Sc; Table 1) with infrequent gravels (Gt). Sandstone beds are 0.5–3 m thick, and form bedsets that are 5–7 m thick. Observed sedimentary structures include trough-crossbeds that have a cross-bed height of 0.25–25 cm (St) and frequent soft sediment deformation, more commonly present in the thicker sandstone beds, including convolute bedding and dewatering structures (Sc; Figs. 5D). The margins of the larger channel belts (>3 m) preserve lateral accretion elements. Bioturbation is scarce and varies by location, with the most common identifiable traces being Skolithos and Asthenopodichnium (Moran et al., 2010). Areas with higher bioturbation intensity are typically unlined burrow clusters within sandstone channels (Fig. 5E).

Overall, channels within FA-3 exhibit lens-shaped architecture and are composed of a sandstone-rich core with wings that thin laterally.

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**TABLE 1. LITHOFACIES OF THE SMOKY HOLLOW AND JOHN HENRY MEMBERS**

| Code | Lithofacies | Description | Interpretation |
|------|-------------|-------------|---------------|
| Gt   | Trough cross-stratified gravel and sandstone | Matrix-supported conglomerate with trough cross-stratification; clasts range from ~1 mm to 5 cm and are subangular to subrounded; matrix grain sizes range from fine- to very coarse sand; cross-set height ranges from 10 cm to 0.5 m | Migration of 3D unidirectional dunes |
| Gm   | Massive mudstone conglomerate | Matrix-supported conglomerate; clasts range from 2 mm to 5 cm and are subangular to subrounded; matrix grain sizes range from very fine-very coarse sand | High strength flow (debris flows) |
| St   | Trough cross-stratified sandstone | Fine- to coarse-grained sandstone with cross-stratification; occasional pebble-sized clasts; cross-set height ranges from 3 cm to 1 m | Migration of 3D unidirectional dunes |
| Sr   | Ripple-laminated sandstone | Very fine- to medium-grained sandstone with current ripples; ripple heights are <0.5 cm | Current ripples under lower flow regime currents |
| Sc   | Convolute-bedded sandstone | Fine- to coarse-grained sandstone with convolute and deformed bedding; dewatering structures visible; height up to 1 m | Dewatering structures or post-depositional slumping |
| Sp   | Planar cross-bedded sandstone | Very fine- to medium-grained sandstone with planar cross-stratification; cross set height up to 0.5 m | Upper flow regime |
| Se   | Erosional sandstone | Fine- to very coarse-grained sandstone with cross-stratification; cross-stratification are up to 0.5 m thick and <15° angle | Rapid deposition of coarse bed-load |
| Sl   | Low-angle cross-bedded sandstone | Very fine- to medium-grained sandstone with planar to very low angle cross-stratification | Upper flow regime |
| Sm   | Massive sandstone | Very fine- to medium-grained sandstone with no visible features | Rapid sedimentation in channel-fill and splay deposits |
| Fl   | Laminated sandstone, siltstone, and mudstone | Interlamination of mudstone, siltstone and mudstones; may have small-scale ripples | Suspension settling during weak traction currents |
| Fm   | Massive mudstone/siltstone | Structureless clay to variegated mudstone and siltstone; may contain fragments of plant and coal material | Waning-flow sedimentation |
| Fc   | Carbonaceous mudstone/siltstone | Structureless dark gray to black mudstone and siltstone with red and yellow nodules; abundant plant material and coal fragments | Suspension settling with high occurrence of organic matter |
| C    | Coal | Black coal beds up to 1 m thick | Floodplain proximal to peat swamps |

**TABLE 2. FACES ASSOCIATIONS OF THE SMOKY HOLLOW AND JOHN HENRY MEMBERS**

| Facies associations | Lithofacies | Architectural Elements | Description |
|--------------------|-------------|-----------------------|-------------|
| FA 1: Well drained floodplain | Fl, Fm, St, Sm, Sr, Sh | Asymmetric–symmetric lenticular to tabular sandy bedforms; accreting barforms | Massive mudstones and siltstones dominate; mudstones variegated in places with pedogenic structures; fine- to medium-grained 0.5–3 m thick tabular to lenticular sandstones; sandstones up to 7 m thick; beds thin and fine upward to fine-grained 0.1–0.5 m thick beds; convolute bedding present; individual Skolithos burrows and clusters present |
| FA 2: Poorly drained floodplain | C, Fc, Fm, Sr, Sh | Tabular to lenticular isolated sandy bedforms, common convolute bedding | Massive mudstone and siltstone; carbonaceous shale and 0.1–1 m thick coal beds; very fine- to medium-grained sandstone beds <3 m thick; less common than in well-drained floodplain |
| FA 3: Major channel within well-developed floodplain | St, Sc, Se, Fl | Complex major sandy ribbon channels and channel fill, often multistory | Very fine- to medium-grained sandstone; tabular-lenticular beds range from 1–5 m thick; 5–7 m thick bed sets; trough cross-beds and soft sediment deformation common; mud clasts up to 4 cm wide at base of erosive scours; channel fill includes barforms and channel lags |
| FA 4: Gravelly to sandy fluvial sheet deposit | Gt, Gm, St, Sc, Sr, Fl, Fm | Tabular to lenticular laterally extensive gravelly and sandy sheets | Medium- to very coarse-grained, fining-up sandstone and sandy conglomerate; 0.1–2 m thick bed sets; mud clasts and pebbles common at base of erosive scours; fluvial channel fill includes channel lag and barform deposits |

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into adjacent overbank and floodplain deposits (Kjemperud et al., 2008). Channels extend laterally from 40 to 100 m (Figs. 5A, 5B). Channel-fill style varies from simple single channels to complex channel belts that comprise 6–10 stories separated by erosive scours (Ford and Pyles, 2014). Larger channel-fill elements can be strongly vertically amalgamated, incising into lower stories and preserving incomplete channel fill and barforms (cf. Chamberlin and Hajek, 2019). Mud rip-up clasts up to 4 cm in diameter are common at the base of these scour surfaces. In general, there is a decrease in grain size, bed thickness, and bedform height, with an increase in mudstone content moving from the channel axis toward the channel wings (Fig. 5B). Lateral accretion sets occur along the margins of channels (Hassanpour et al., 2013).

**Description FA-4.** Facies association 4 (FA-4; Fig. 6) typically consists of coarser-grained bedforms with fine to gravelly sandstone lithofacies (Table 2). Locally, gravel lithofacies are more common (Gt, Gm; Fig. 6D). Gravel clast sizes range from 2 mm to 4 cm, and commonly consist of subangular to angular gray to black chert, quartzite and carbonate lithic clasts, and mud rip-ups (Fig. 6D). These beds are often normally graded with larger grains and clasts at the base of trough cross-beds and erosional scours (Sc, St). Sandstone beds are 0.1–2.5 m thick and bedsets (greater than 2 m) typically exhibit a fining upwards trend. Sedimentary structures observed within the sandstone and gravel-prone bedsets include trough cross-beds and convolute bedding (Sc, St; Fig. 6B). Many bedsets contain thicker beds at the base that have trough-cross and convolute bedding, which are overlain by thinner, or “flaggy” bedding that is commonly ripple-laminated. Mud rip-up clasts, pebbles, and coarse-grained sandstones are common at the base of erosive scours and trough cross-beds, particularly at the base of channel bedsets. Siltstones and mudstones are infrequently preserved between channel stories (Fig. 6E).

Bioturbation is variable, ranging from scarce to common, with occurrences of *Skolithos, Asthenopodichnium* (Moran et al., 2010) and escape traces. Bioturbation intensity is variable across the plateaus and can be relatively high locally (~15% bioturbated). Wood casts (cm-scale to >1 m in length) are commonly preserved at the base of erosive scours. *Asthenopodichnium* traces are visible within wood casts. Casts of theropod and ornithopod tracks are preserved on the bases of some channel deposits (e.g., Parowan Canyon; Milner et al., 2006). Crocodilian, dinosaur, and turtle bone fragments erode out of sandstone and siltstones of FA-4 (Eaton, 1999). Concretions 0.3–1.0 m thick are commonly found in the larger sandstone beds.

Sandstone architecture is dominated by laterally extensive tabular sandstones, with erosive scours between stories (Figs. 6A, 6E). Tabular sandstones are often vertically and laterally amalgamated forming laterally extensive sheet complexes, but lateral and vertical channel amalgamation varies by location. Some sheet deposits form channel belts that are laterally extensive and can be mapped for 100s of meters with >5 stories identifiable, e.g., the Calico Bed (Fig. 6E). In contrast, some channel belts have a reduced lateral extent (10s of meters), but instead have high vertical amalgamation (typically between 1 and 3 stories) with rare fine-grained deposits preserved (Fig. 6A). Tabular sandy channel belts are commonly 1–3 m thick but can be up to ~5 m thick.

**Channel-Fill Facies Associations (FA-3 and FA-4) Interpretation**

**FA-3 Interpretation.** Channel-fill successions of FA-3 represent deposition of discrete channel belts within well-developed floodplain, similar to the “steers-head” channels of Kjemperud et al. (2008). Channel deposits of FA-3 represent fluvial systems with a higher sediment suspended load and lateral bank stability, interpreted as distributary channels draining into floodplain deposits (Figs. 5A, 5B). Distinct channel belt forms suggest bank stability, allowing for lateral and vertical aggradational deposition within the channel margins (Gibling, 2006).

Channel-fill facies are locally adjacent to lateral-accretion elements, although complete barform preservation is rare. The lateral wings of channels are typically thinner-bedded and finer grained with a higher preservation of interbedded mudstone and siltstones. These are interpreted as channel levees, where sediment spilled onto overbank regions (Allen, 1978; Friend et al., 1979). Channel levee development and preservation indicates a high level of bank stability that is concurrent with the higher frequency of vegetated floodplain deposition seen in FA-3 (Slingerland and Smith, 2004). Additionally, preservation of overbank and inter-channel fine-grained material suggests there was a higher volume of suspended sediment load relative to the sand-rich sheet deposits that comprise FA-4. Lateral accretion bedsets and barforms are infrequently preserved and vary by location, but where present indicate some level of reworking within channel margins (Mohrig et al., 2000; Hassanpour et al., 2013). These fixed-location channels could indicate a less mobile river system, with less frequent avulsion and stable, well-developed river channels (Straub and Esposito, 2013).

**FA-4 Interpretation.** Sheet-like sandstone deposits of FA-4 represent unconfined flow within poorly developed fluvial channels dominated by sandy to gravelly bedload. The multi-story character of these deposits records episodes of cutting and filling, likely a result of fluctuations in discharge causing incision during high discharge events, and subsequent deposition during waning flow (Holbrook, 2001). A lack of distinct channel boundaries and laterally extensive tabular geometries indicate lateral instability and a dominance of unconfined flow (Smith et al., 1989; Owen et al., 2017). The coarse-grained nature of the channel fills, numerous mud-rip-ups, and gravelly lags suggests high discharge with frequent bankfull flows (Gibling, 2006). The cut-and-fill patterns combined with coarser grained channel fill may also indicate some level of internal reworking, suggesting a fluvial system that migrated or avulsed more frequently (Straub and Esposito, 2013) or during relatively slow accommodation creation (Miial and Arush, 2001).

**Regional Stratigraphic Patterns**

Figure 7 shows a regional correlation composed of six new measured sections of the John Henry Member combined with two sections from Tilton (1991; Glendale and Tenny Canyon). Sections range from the northern Kaiparowits Plateau (Table Cliffs) region, to the east and southern flanks of the Paunsaugunt Plateau (PP; Fig. 2), and into the central and western Markagunt Plateau (MP; Fig. 2).

**Thickness Trends**

**Markagunt Plateau**

Measured sections from the northwestern edge of the Markagunt Plateau (Parowan Gap and Parowan Canyon; Fig. 7) represent the most proximal strata in the study area. Near the Wah Wah and Blue Mountain thrust faults (Fig. 8), the Iron Springs Formation thickens from at least 350 m as measured in this study (Parowan Canyon), and up to ~900 m thick to the southwest (Fig. 8, Gunlock and Pine Valley Mountains; Cook, 1957; Hintze, 1986; Fillmore, 1991). The base of the Straight Cliffs Formation (Tibbet Canyon and Smoky Hollow Members) is not exposed in Parowan Gap. There, the partially exposed upper John Henry Member is ~80 m thick (Fig. 7), and this section thickens to ~360 m in Parowan Canyon (~15 km southeast; Fig. 7). In the central part of the Markagunt Plateau, the Smoky Hollow Member ranges from ~110–117 m thick (Orderville Gulch, Fig. 7). The Calico Bed has a well-developed ~17-m-thick lower bed but pinches out along the western margin of the Markagunt Plateau. The John Henry Member has an average thickness of ~200 m along the south-central margin of the Markagunt Plateau.
**Paunsaugunt Plateau**

Thinning of the John Henry and Smoky Hollow Members occurs across the Sevier and Paunsaugunt faults (Fig. 7). Within the Paunsaugunt Plateau, the Smoky Hollow Member is roughly half as thick as it is to the west (~30 m). John Henry Member strata thin across the Sevier fault from ~200 m to ~160 m (Glendale, Fig. 7; Tilton, 1991). This thickness is maintained along the western and southern flanks of the Paunsaugunt Plateau (Fig. 7; Tilton, 1991). Localized thickening occurs along the northeastern edge of the Paunsaugunt Plateau, in both the John Henry Member (~160 to ~270 m) and the Smoky Hollow Member (~30 m to ~60 m; Heward Creek, Fig. 7).

**Kaiparowits Plateau**

Across the Paunsaugunt fault, strata from the Smoky Hollow Member and John Henry Member strata thicken across the Paunsaugunt fault, in the northern Kaiparowits Plateau–Table Cliffs region (Fig. 7; Shakespeare Mine). Here, the Smoky Hollow Member is ~70 m thick, a slight thickness increase relative to the west. The John Henry Member, however, is ~456 m thick, a 69% increase in thickness compared to Heward Creek, across the Paunsaugunt fault (Shakespeare Mine; Fig. 7). These thickness changes continue into the paralic and shallow-marine John Henry Member strata to the east, which are ~460 m thick along the northeastern margin of the Kaiparowits Plateau in Buck Hollow (Mulhern and Johnson, 2016). The John Henry Member in the southern portion of the Kaiparowits Plateau is only ~200 m thick (Gallin et al., 2010; Allen and Johnson, 2010; Dooling, 2013; Pettinga, 2013; Gooley et al., 2016).

**Architectural Trends**

**Markagunt Plateau**

Fluvial architecture in the most proximal John Henry Member sections at Parowan Gap and Parowan Canyon is characterized by sheet deposits of FA-4 intercalated with minor components of FA-1 and some FA-3 (Fig. 7). These strata are typically composed of medium- to coarse-grained sandstones with pebble lags at the base of scours. Channel belts are vertically amalgamated with minor floodplain development; NTG in these proximal sections is ~0.64 and ~0.48 (Fig. 7).

In contrast, across the rest of the Markagunt Plateau, the John Henry Member consists of well-drained floodplain material (FA-1) with isolated sandstone channels of FA-3 (Fig. 7; Orderville Gulch). Instances of FA-4 within the middle John Henry Member can be traced laterally for 100s of meters. Sheet deposits of FA-4 within the John Henry Member are composed of fine- to medium-grained sandstone and lack pebble lags. NTG estimates of the John Henry Member decrease from ~0.56 along the western margin of the plateau to ~0.37 in the central plateau. Overall paleocurrent indicators are oriented north/northeast (average paleoflow ~0.55; n = 120; Fig. 7).

**Paunsaugunt Plateau**

Along the western and southern margins of the Paunsaugunt plateau (Glendale and Tenny Canyon; Fig. 7), the John Henry Member consists of laterally and vertically amalgamated sheet-like deposits (Fig. 6; FA-4) with minor components of isolated channel belts within well-developed floodplain deposits (FA-1 and FA-3). This architecture changes dramatically in the northeast margin near the Paunsaugunt fault (Fig. 7; Heward Creek), where the John Henry Member consists primarily of FA-3 with minor components of FA-1. Here, channel belts of FA-3 are typically ~8 stories high and situated within well-developed gray to red mudstone deposits (Fig. 5A, B). Typical bedforms are difficult to identify in many of the thickest channels due to the abundance of convolute and slumped bedding disrupting bedform structures (Fig. 5D). Overall, the average NTG of the John Henry Member in the south and southwest of the Paunsaugunt Plateau is ~0.64 (average of Tenny Canyon and Glendale), but drops to ~0.32 at Heward Creek. The average paleocurrent direction is to the northeast, with no significant changes in trends across the plateau (Tilton, 1991; Lawton et al., 2014; Fig. 7).

**Kaiparowits Plateau**

At Shakespeare Mine in the northern Kaiparowits Plateau (Table Cliffs) region, the lower John Henry Member is composed of the poorly drained floodplain and coals of FA-2 and defined channels of FA-3. Channel...
amalgamation increases both laterally and vertically up-section, transitioning from small, isolated channels of FA-2, to larger defined channels of FA-3, and to sheet-like sandstones of FA-4. Fluvial John Henry Member NTG in the northern Kaiparowits region is ~0.87. Overall paleocurrent direction is to the east-northeast (Fig. 7).

**DISCUSSION**

Stratigraphic data presented here show increasing thickness and NTG to the east in distal parts of the foredeep, which raises questions regarding controls on basin architecture and evolution of the Late Cretaceous southern Utah foreland. This discussion focuses primarily on thickness and facies variations of the fluvial John Henry Member of the Straight Cliffs Formation, and factors controlling sediment supply and accommodation within the foreland basin.

**Thickness Trends**

Regional stratigraphic correlations of Upper Cretaceous strata across the SWHP of southern Utah indicate abrupt, stepwise thickness variations in key parts of the section that coincide with modern plateau-bounding faults (Figs. 8, 9). The most proximal sections of the Iron Springs Formation (e.g., Parowan Canyon) are at least 350 m thick, whereas the age-equivalent Straight Cliffs Formation thins by 50% to the east across the Markagunt and Paunsaugunt Plateaus (average thickness ~175 m). This proximal eastward-thinning trend is reversed across the Paunsaugunt Plateau where John Henry Member sections thicken from ~270 m at Heward Creek to over ~450 m at Shakespeare Mine over a distance of ~25 km (Fig. 7). This thickening into the distal foredeep also occurs from south to north across the Kaiparowits Plateau in the John Henry Member, as well as the overlying Wahweap and Kaiparowits Formations (Fig. 9B; Lawton et al., 2003; Gallin et al., 2010; Gooley et al., 2016).

Present-day stratigraphic thicknesses are reported here, and thus are minimum sediment accumulation estimates. The observed eastward- and northward-thickening trends in the Kaiparowits Plateau (Fig. 9) would be exaggerated by decomposition, given the concentration of mudstone and coal deposits in paralic and marine sections within the John Henry Member (Hettinger, 2000; Hettinger et al., 2009). Mulfinn and Johnson (2016) estimated the entire Straight Cliffs Formation could be expanded from 456 m to over 800 m at Buck Hollow (Fig. 2), even with conservative decomposition estimates.

As mentioned previously, intraformational unconformities and inferred sequence boundaries in Upper Cretaceous sections are thought to be relatively minor in terms of duration, potential eroded section, and angular stratigraphic relationships (Peterson, 1969; Gustason, 1989; Tilston, 1991; Shanley and McCabe, 1995; Lawton et al., 2003, 2014; Szwarc et al., 2015; Primm et al., 2018). While the sub-Claron unconformity is regionally significant, available data suggest that the Kaiparowits Formation was mainly deposited in an isolated depocenter located in the northern Kaiparowits Plateau (Fig. 9; Eaton et al., 1993; Roberts, 2007; Sampson et al., 2013; Biek et al., 2015) and that variations in thickness are not entirely a function of subsequent erosion. Thus, observed stratigraphic thicknesses of Cretaceous sections in the SWHPs likely represent paleo-accommodation variations.

**Facies Trends**

Using fluvial architecture and NTG as proxies for grain size and facies trends across the foreland basin (Fig. 9A), results show consistent “distal fining” only from the most proximal Iron Springs Formation (NTG 0.64) to the Straight Cliffs Formation in the Markagunt Plateau, with its well-developed floodplain deposits and isolated fluvial channel belts (NTG 0.37). In the Paunsaugunt Plateau, distal fining trends become more complex. Across the Sevier fault in the western and southern Paunsaugunt Plateau, channel belt amalgamation increases, becoming more sheet-like with less floodplain deposition (NTG 0.64). In contrast, to the northeast along the Paunsaugunt fault (e.g., Heward Creek; Fig. 7), the John Henry Member section becomes thicker with much higher floodplain deposition and well-developed isolated channels (NTG 0.32). Coincident with dramatic thickening across the Paunsaugunt fault, the John Henry Member sections in the northern Kaiparowits Plateau (Shakespeare Mine; Fig. 7) have a notably high NTG (~0.87) with highly amalgamated fluvial channel belts and a few well-developed coal zones primarily in the lower parts of the section (Fig. 4; Hettinger, 2000; Hettinger et al., 2009).

These Late Cretaceous facies trends can be partly explained by fluvial drainage networks that were aligned subparallel to the trend of the fold-thrust belt in southern Utah, thus implying an axial fluvial system (Fig. 10). Direct evidence for orogen-transverse drainage patterns is essentially limited to the most proximal deposits of the Iron Springs Formation (Gunlock and Pine Valleys; Fig. 8; Fillmore, 1991; Goldstrand, 1991). In addition, episodic progradation of transverse fluvial systems across the foreland is recorded in extensive “sheet-sands” that punctuate the Cretaceous stratigraphy in the area (e.g., Calico Bed, Drip Tank Member, and Wahweap Formation; Lawton et al., 2003, 2014; Lawton and Bradford, 2011; Szwarc et al., 2015; Primm et al., 2018). These distinctive and widespread units typically have east-directed paleocurrent indicators and provenance signatures that indicate a Sevier fold-thrust belt source.

Thus, evidence for orogen-transverse drainage systems appears to be localized and episodic relative to most of the Cretaceous foreland basin succession in southern Utah. Previous interpretations of major axial fluvial systems are partly based on facies distributions combined with paleocurrent analyses, which show dominantly northeast transport in Upper Cretaceous fluvial sections of the Kaiparowits Plateau (Szwarc et al., 2015; Gooley et al., 2016). Our data compilation confirms overall northeast-directed paleocurrent directions for the John Henry Member across the Markagunt and Paunsaugunt Plateaus (Fig. 7). Similarly, U-Pb detrital zircon geochronology of the Straight Cliffs Formation in the Kaiparowits Plateau (Szwarc et al., 2015; Primm et al., 2018) indicates relatively minor detrital zircon input from the Sevier fold-thrust belt. Rather, major zircon source areas lie to the south (e.g., 1.4 and 1.7 Ga zircon populations sourced from the Mogollon Highlands) and southwest (e.g., ca. 147 Ma zircon populations likely sourced from the Mojave Dike Swarm in southeastern California; Fig. 1).

In sum, fluvial sections in the Straight Cliffs Formation show distinct facies and thickness variations by location; these features are oriented sub-parallel to the fold-thrust belt and coincide with modern plateau-bounding faults. The interpretation of these basin-axial corridors suggests some allogenic control on local accommodation rather than a continuous foredeep throughout the Late Cretaceous.

**Accommodation Controls**

**Flexural Loading**

Previous studies have interpreted the Upper Jurassic-Lower Cretaceous Morrison, Carmel, and Cedar Mountain Formations to record forebulge migration and deposition, with the forebulge lying just east of the Kaiparowits Plateau by Cenomanian or Turonian time (Royse, 1993; Currie, 1998, 2002; DeCelles, 2004). An important implication of this interpretation is that high Late Cretaceous sediment accumulation rates of the Kaiparowits Plateau appear to have formed in the distal foredeep to forebulge depozones, 150-200 km east of the Wah Wah and Blue Mountain thrust systems. This pattern is particularly odd given thinning of time-equivalent
strata to the west in more proximal parts of the foredeep, where thicker sediment accumulation would be expected (Fig. 9).

Differential loading could have increased flexural accommodation in the northern Kaiparowits Plateau due to movement along the Pavant/Paxton thrust sheets of central Utah, ~75 km to the northwest (Fig. 1B). The thrusting style of the Pavant/Paxton system consisted of duplexing and culmination growth during the Coniacian-Santonian, simultaneous with deposition of the John Henry Member (DeCelles et al., 1995; Coogan and DeCelles, 1996). During this time, sediment supply into the central Utah foreland basin was likely high, but provenance data and
Blind Thrust Faults

Given evidence for early emplacement and a relatively stationary fold-thrust belt in the Wah Wah and Blue Mountain thrust systems, the possibility of buried Sevier-related thrust faults may be considered to explain observed foredeep sedimentation patterns. No subsurface thrusts have been reported across the SWHP during Late Cretaceous time (Cook and Hardman, 1967; Van Kooten, 1988), nor is there outcropping evidence of significant detachment horizons. Late-breaking thrust faults at Parowan Gap and Ruby’s Inn likely occurred after deposition of the Straight Cliffs Formation (Merle et al., 1993). Furthermore, we note that studies of wedge-top basins indicate very different stratigraphic patterns than those documented here, including common progressive unconformities and coarse-grained alluvial deposits with local provenance (Lawton and Trexler, 1991; Horton, 1998; Quinn et al., 2018).

Dynamic Load

Subsidence mechanisms across the southern Utah foreland can be attributed to flexural loading of the Sevier fold-thrust belt and dynamic processes (Jordan, 1981; Mitrovica et al., 1989; Pang and Nummedal, 1995; Liu and Nummedal, 2004; Liu and Gurnis, 2010). During the Coniacian to Santonian (i.e., during deposition of the John Henry Member), subsidence within the foreland is thought to transition from a narrow flexural profile to a longer-wavelength foreland basin (Roberts and Kirschbaum, 1995). Dynamic subsidence likely contributed to the expansion of the entire southern Utah foreland basin, a wavelength of 100s kilometers (Painter and Carrapa, 2013; Heller and Liu, 2016). This effect is on a much broader scale than the evidence for localized variable accommodation presented in this study (10s–100s of m per km). These abrupt thickness changes suggest that dynamic subsidence alone cannot explain evidence for discrete sub-basins in the distal foredeep.

Laramide-style Tectonics

The idea of sub-basin formation within the Cordilleran foreland basin is commonly cast in the context of the Laramide orogeny, or “broken foreland” of Dickinson et al. (1988). Laramide-style, basement-involved uplifts exist on the margins of the SWHP, including the Kanarra Fold to the west, and the East Kaibab monoclone and Circle Cliffs uplifts to the east (Fig. 1). The earliest evidence of this foreland break-up in southern Utah includes growth faults in the Campanian Wahweap Formation in the northern Kai parowits Plateau (e.g., Hilbert-Wolf et al., 2009; Simpson et al., 2014). One of the main drivers for Laramide tectonism is flat-slab subduction, which is thought to have initiated at the latitude of southern Utah during the earliest Campanian (Simpson et al., 2014). Inverse mantle convection models show the flat-slab segment, likely the conjugate of the Shatsky Rise, entering the subduction zone along the Mojave area of California at ca. 90 Ma, followed by northeast underthrusting beneath Arizona to Wyoming during the latest Cretaceous (Liu et al., 2010; Liu and Currie, 2016). Although this is a plausible driving mechanism, timing estimates would place onset of Laramide-style deformation up to ∼20 m.y. later than the suspected formation of sub-basins within the Straight Cliffs Formation and thickness changes in the Naturita Formation. This timing is similar to the emergence of basement-cored structures of the northern foreland in southwest Montana, where thermochronologic evidence shows exhumation of Laramide structures occurred prior to ca. 80 Ma, and as early as ca. 100–120 Ma (Schwartz and DeCelles, 1988; Carrapa et al., 2019; Garber et al., 2020; Orme, 2020). The early (Aptian-Albian) onsets of Laramide-style deformation is similar to our earliest estimates of foreland partitioning from the Campanian Naturita Formation (Gustason, 1989), and argues against migration of the Shatsky rise as a driver of Laramide deformation (Carrapa et al., 2019; Garber et al., 2020). Furthermore, classic Laramide-style “broken foreland” basins (e.g., the Uinta basin) are generally much larger than the sub-basins now exposed along the SWHP (Picard and High, 1968; Dickinson et al., 1988).

Structural Inheritance and Normal Faults

The location of modern plateau-bounding faults was likely influenced by preexisting crustal weaknesses. Inherited rift-related basement faults across southern Utah (Fig. 1A) are known to have been reactivated during multiple tectonic events throughout the Phanerozoic (Schwartz, 1982; Picha and Gibson, 1985). Evidence for Late Cretaceous–early Cenozoic activity along the Hurricane fault (Cook and Hardman, 1967) suggests that other SWHP-bounding faults may have been prone to structural inheritance controls and reactivation as well. We suggest that reactivation of Proterozoic basement features, possibly as normal faults, was a factor in early partitioning of the Cretaceous southern Utah foreland basin.
Structural variations within the crust of foreland basins, like those in southern Utah, have been shown to influence crustal deflection and evolution of the foredeep (e.g., Romans et al., 2010). In some cases, they may cause segmentation of the foreland (Waschbusch and Royden, 1992) or rapid migration of the foredeep depocenter (DeCelles et al., 1995). For example, eastward propagation of the central Utah thrust sheets may have been aided by the Ancient Ehramp fault and other basement structures (Fig. 1). The northeast-trending Paragonah linearament separates the southern and central Utah fold-thrust belt and is speculated to have offset the southern Utah thrust sheets, inhibiting their further propagation to the east (Picha, 1986). In the Magallanes foreland basin in Argentina, pre-foreland attenuated lithosphere led to atypical foredeep subside, amplified by sediment loading (Romans et al., 2010; Fosdick et al., 2014; Gianni et al., 2015).

Multiple mechanisms have been suggested for normal faulting associated with flexure. Extension observed along the outer rise in subduction zones may provide a simple conceptual analog (Ludwig et al., 1966). The Taconic peripheral foreland basin was accompanied by motion along normal faults due to inelastic extensional deformation on the convex side of the flexed plate, which suggests bending of the lithosphere beyond its elastic limit (Bradley and Kidd, 1991). A possible expression of flexural foundering has also been documented in the Apulian foreland basin in Italy, where systematic joints formed along the forebulge due to flexure-related fiber stresses (Billi and Salvini, 2003). This flexural foundering within the forebulge can alter typical basin depocenters by stopping forebulge migration (Waschbusch and Royden, 1992). Normal faulting due to extensional stresses within flexural foreland basins is likely more common than is recognized because it may present as subtle shifts in stratigraphic architecture that cannot be explained by other allogenic or autogenic signals (Bradley and Kidd, 1991; Waschbusch and Royden, 1992; Londoño and Lorenzo, 2004; Gianni et al., 2015).

Syndepositional Tectonic History—An Integrated Model

Traditional models of foreland basin evolution conflict with aspects of stratigraphic architecture and depocenters across the southern Utah Cretaceous foreland basin. Specifically, thickening and increases in NTG in the distal foredeep or forebulge area (the northern Kaiparowits Plateau) is unusual for classic foreland basin models. We propose that inherited structural heterogeneities within the crust restricted typical foredeep migration and may have favored the formation of normal faults on the bounding plate (Fig. 10), which ultimately triggered formation of isolated sub-basins within the southern Utah foreland (Fig. 9). Restricted foreland basin migration could have increased local stress within the plate allowing for faulting to occur along preexisting heterogeneities (e.g., Bradley and Kidd, 1991; Hudson, 2000).

Even given uncertainty regarding their orientation and kinematics, we infer that movement along inherited structures during Late Cretaceous time influenced sediment transport and accumulation within the foreland basin (Fig. 10). In some cases, changes in thickness and fluvial architecture are abrupt on either side of modern plateaux-bounding structures such as the Sevier and Paunsaugunt normal faults (Eaton et al., 1993). Sub-basin formation partitioned the axial fluvial system into distinct northeast-trending fluvial facies belts (Fig. 10) with variable NTG and channel belt architectures preserved. thinning of foreland basin strata across the Markagunt and Paunsaugunt Plateaux is reversed in the most distal parts of the foredeep, particularly in the northern Kaiparowits Plateau. Similar evidence for foredeep partitioning is reported as early as Cenomanian time (Gustason, 1989; Eaton and Nations, 1991), and we interpret these signals to have intensified by Santonian time during deposition of the John Henry Member.

CONCLUSIONS

New stratigraphic data from the Straight Cliffs Formation of the SWHP, including sediment thickness, facies architecture, and net to gross estimates, indicate that the Cretaceous southern Utah foredeep was divided into discrete sub-basins at least 10–20 million years prior to onset of known local Laramide-style deformation. Flexural foundering at the edge of the Colorado Plateau may have favored sub-basin formation within the foreland, thinning strata across the Paunsaugunt Plateau, and dramatically thickening strata in the northern Kaiparowits Plateau. We suggest that the Paunsaugunt, Sevier, and Hurricane faults are the modern expressions of these inherited features, which ultimately reflect lithospheric heterogeneities formed in the Proterozoic. These findings demonstrate that “broken forelands” evolve different basin geometries, sedimentary facies, and stratigraphic stacking patterns than those predicted by traditional models of foreland basin systems.

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