Effects of contemporaneous orogenesis on sedimentation in the Late Cretaceous Western Interior Basin, northern Utah and southwestern Wyoming

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Abstract
Three drivers of subsidence are recognized in the Western Interior Basin: Mesozoic–early Cenozoic flexure adjacent to the thin-skinned, eastward propagating Sevier Orogeny, Late Cretaceous–Eocene flexure associated with thick-skinned Laramide Uplifts and Late Cretaceous dynamic subsidence. This study combines outcrop lithofacies, palaeocurrent measurements, detrital zircon geochronology, biostratigraphy, stratigraphic correlations and isopach maps of Coniacian–Maastrichtian (89–66 Ma) units to identify these subsidence mechanisms impact on basin geometry and stratigraphic architecture in the northern Utah to southwestern Wyoming segment of the North American Cordillera. Detrital zircon maximum depositional ages and biostratigraphy support that the Maastrichtian Hams Fork Conglomerate was deposited above the Moxa unconformity in the wedgetop and foredeep depozones. The Moxa unconformity underlies the progradational Ericson Formation in the distal foredeep. The Hams Fork, however, is younger than the Ericson Formation, and instead equivalent to upper Almond Formation. Therefore, the hiatus associated with the Moxa unconformity continued for several million years longer in the fold belt and proximal basin than in the distal foredeep, with Ericson Formation-equivalent strata onlapping the Moxa unconformity towards the west. Regional thickness patterns record and constrain the timing of the transition from Sevier to Laramide-style tectonic regimes. From 88 to 83 Ma (upper Baxter Formation) a westward-thickening stratigraphic wedge characterized the foredeep developed by lithospheric flexure by thrust-belt loading. Nevertheless, the presence of >500 m of subsidence >200 km from the thrust front suggests a long-wavelength subsidence mechanism consistent with dynamic subsidence. By 83 Ma (Blair Formation) the long-wavelength depocentre shifted away from the thrust belt, with no evidence of a Sevier foredeep. This depocentre continued migrating eastward during the early-mid Campanian (ca. 81–77 Ma). The late Campanian–Maastrichtian

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In the Late Cretaceous the Western Interior Basin was transitioning from a simple foreland to a complex suite of partitioned sub-basins, separated by Laramide-style uplifts. The stratigraphic relationship between proximal sediments within and adjacent to the Sevier fold-thrust belt with the more distal deposits has been difficult to define; this is because of structural deformation, unconformities, incomplete chronostratigraphic control and a lack of continuous outcrop exposures. In this study, we have progressed our understanding of the tie between proximal and more distal strata by integrating existing biostratigraphic control, new detrital zircon data, subsurface (well log) control and measured sections. The purpose is to evaluate the stratigraphic response to an evolving tectonic context.

The North American Cordillera includes two deformational styles and phases: the thin-skinned Sevier and the thick-skinned Laramide orogenies (DeCelles, 2004; Dickinson, 2004). Cretaceous–Palaeogene subsidence in the retroarc is driven by a combination of flexure due to Sevier and Laramide loads and dynamic subsidence associated with mantle convection (Lawton, 2008; Liu & Nummedal, 2004; Liu et al., 2014; Mitrovica et al., 1989; Painter & Carrapa, 2013). The Sevier orogeny is characterized by the development of a contractional retroarc in the Late Jurassic, which led to subsidence along the thrust front and the formation of a foreland basin; the Western Interior Basin (Coney & Reynolds, 1977; Constenius et al., 2003). By the end of the Late Cretaceous subsidence was localized adjacent to Laramide uplifts far inland of the plate margin due to the shallowing of the Farallon Plate subduction angle under North America (Dickinson & Snyder, 1978; Dumitru et al., 1991; Jordan & Allmendinger, 1986; Saleeby, 2003).

The coeval deformation in the Sevier fold-thrust belt and Laramide uplifts altered the geometry of the Western Interior Basin and affected sedimentation patterns (Figure 1; DeCelles, 2004; Dickinson & Snyder, 1978; Painter & Carrapa, 2013; Rudolph et al., 2015). Shallowing of the subduction angle associated with the transition from Sevier to Laramide orogenesis would also have changed the locus and magnitude of dynamic subsidence concurrent with the migration of the leading edge of the flat portion of the subducted plate (Eakin et al., 2014; Heller & Liu, 2016; Liu et al., 2014; Mitrovica et al., 1989). As these orogenic events and subsidence mechanisms overlap in time, and much of the structural growth strata are eroded during ongoing exhumation, the sedimentary basin response can be employed to infer the geodynamics of orogenesis. The Late Cretaceous Western Interior Basin is therefore an ideal laboratory to investigate changes in the basin sedimentary record in response to changing plate margin and subduction dynamics.

The goal of this research was to investigate how the tectonic events of the Sevier and Laramide-style deformation affected sedimentation and basin geometry in the Late Cretaceous Green River Basin in northeastern Utah and southwestern Wyoming (Figure 2). In this area, the chronostratigraphic relationship between wedgetop conglomerates in the proximal (i.e. western) basin and marginal marine facies in the distal (i.e. eastern) foreland is largely unknown due to lack of robust biostratigraphic control in non-marine facies and poor exposure of intermediate strata. The second-order (10s of millions of years) stratigraphic cyclicity documented in the distal areas...
has been related to changes in subsidence rate (Rudolph et al., 2015). In proximal areas, the large-scale pattern is simpler, with overall progradation from the upper Frontier Formation (Turonian, ca. 90 Ma) to the Hams Fork Conglomerate (Maastrichtian, ca. 71 Ma). These differences in stacking patterns are likely due to very different rates of accommodation in the two areas.

We employ several approaches to address the question of proximal to distal correlation. Detrital zircon U-Pb geochronology provides maximum depositional ages (MDAs) for Upper Cretaceous stratigraphy. Well log correlations integrated with measured sections, biostratigraphy and detrital zircon MDAs illuminate spatial and temporal changes in depositional environment and basin geometry, providing control points between the proximal and distal basin locations.

2 | GEOLOGIC SETTING

2.1 | Tectonic setting

Subduction of the Farallon and Kula oceanic plates beneath the North American plate took place from the Jurassic to Palaeogene, with the type and distribution of deformation changing through time in response to variable convergence.
rate and subduction angle (Oldow et al., 1989). The Sevier fold-thrust belt and magmatic arcs began forming in the Jurassic, and Laramide deformation began in the Late Cretaceous (Oldow et al., 1989). The Sevier fold-thrust belt and magmatic arcs began forming in the Jurassic, and Laramide deformation began in the Late Cretaceous (Dickinson et al., 1988; Lawton, 1986). Deformation resulted in three primary phases of subsidence in the Wyoming area. From approximately 120–85 Ma, a westward-thickening siliciclastic sedimentary wedge developed parallel to Sevier fold-thrust belt (Jordan, 1981). By 84 Ma regional subsidence had a long-wavelength expression (Liu & Nummedal, 2004; Liu et al., 2011; Painter & Carrapa, 2013; Pang & Nummedal, 1995). In the Coniacian–Santonian (ca. 88.7–83.5 Ma) DeCelles (2004) and Roberts and Kirschbaum (1995) interpret the effects of dynamic subsidence as the widening of the retroarc depocentre. Finally, by 72 Ma, depocentres formed adjacent to basement-cored uplifts of the Laramide province (Dickinson et al., 1988; Jones et al., 2011).

2.1.1 | Sevier orogeny

The Sevier orogeny is characterized by a zone of thin-skinned thrust faults and folds that span a north-south distance of over 2,000 km, with a maximum width of 200 km (DeCelles, 2004; Gans & Miller, 1983). The thrust front migrated eastward through time, and regional shortening, estimated to be 100 km, was in the east–west direction (Constenius et al., 2003; Coogan et al., 1992; DeCelles, 1994, 2004; Lamerson, 1982; Royse Jr. et al., 1975; Yonkee, 1992). The Sevier fold-thrust belt west of the study area consists of several primary thrust systems: the Ogden, Willard-Paris-Mead, Crawford, Medicine Butte, Absaroka, Darby and Hogsback thrusts, from the west to east (DeCelles, 2004). The Ogden thrust system is the structurally lowest thrust and is responsible for the doubly plunging antiformal basement duplex known as the Wasatch culmination, which exposes Palaeoproterozoic crystalline rocks (DeCelles, 1994; Schirmer, 1988; Yonkee, 1992). The Willard-Paris-Mead thrust system carries thick units of Proterozoic and Palaeozoic rocks, with a minimum of 35 km of displacement recorded along the Willard segment of the system (DeCelles, 1994, 2004). The Crawford, Medicine Butte, Absaroka, Darby and Hogsback thrusts detach from a regional décollement in the lower Cambrian shale (Coogan et al., 1992; DeCelles, 2004; Royse, 1993; Yonkee et al., 1997). Thin-skinned deformation associated with the Darby-Hogsback system is interpreted to be active until early Eocene (Dorr & Gingerich, 1980; Lamerson, 1982; Warner & Royse, 1987).

2.1.2 | Laramide orogeny

Laramide deformation was coeval with the late stages of Sevier thrusting, with the early stage of deformation beginning by 80 Ma, and potentially as early as 100 Ma (Carrapa et al., 2019; Clinkscales & Lawton, 2015;
DeCelles et al., 1987, 1991; Dickinson et al., 1988; Fan & Carrapa, 2014; Lawton, 2008; Painter & Carrapa, 2013; Schwartz & DeCelles, 1988; Shuster & Steidtmann, 1988; Tindall et al., 2010). The overall direction of shortening differed slightly from that of the Sevier orogenesis, with regional shortening in the west-southwest–east-northeast directions (Dickinson & Snyder, 1978). By the end of the Late Cretaceous, Laramide uplifts partitioned the foreland basin into more than 20 smaller basins, with Laramide orogenic features distributed throughout Utah, Wyoming and Colorado, and extending as far south as New Mexico and Texas (DeCelles, 2004; Dickinson, 1988; Dickinson & Gehrels, 2008; Engebretson et al., 1984). Deformation is thick-skinned, or basement involved, and characterized by basement-cored arches, monoclines and reverse-fault bounded uplifts that formed as a result of the buckling and shear of the continental crust (DeCelles, 2004; Dickinson et al., 1988; Dickinson & Snyder, 1978; Roehler, 1965). The partitioning of the basin by Laramide uplifts greatly affected sediment distribution patterns into the Palaeogene, as the location of depocentres shifted from adjacent to the thrust belt to adjacent to local uplifts (Brown, 1988; Dickinson et al., 1988).

2.1.3 | Flat subduction

The change in deformation style and subsidence history in the Late Cretaceous is attributed to the flattening of the Farallon slab beneath North America (Coney & Reynolds, 1977; Constenius et al., 2003; Dickinson & Snyder, 1978; Saleeby, 2003). During the Sevier orogeny, the angle of slab subduction is interpreted to have been fairly steep, leading to abundant arc magmatism (Dickinson & Snyder, 1978; Engebretson et al., 1984). Subsidence was confined to a narrow zone adjacent to the thrust belt due to flexure of the lithosphere caused by crustal shortening and tectonic loading and a trench-proximal (i.e. western) position of the locus of maximum dynamic subsidence (Currie, 1988, 2002; Gentry et al., 2018; Heller et al., 1986; Jordan, 1981; Painter & Carrapa, 2013; Pang & Nummedal, 1995). The wavelength of subsidence gradually increased with the initiation of the Laramide orogeny, thereby widening the basin by five to ten times compared to the basin produced by flexural loading of the Sevier Orogeny (Liu et al., 2014; Painter & Carrapa, 2013; Roberts & Kirschbaum, 1995). This widening is attributed to eastward migration of dynamic subsidence caused by shallowing of subduction (Liu et al., 2014). Slab flattening is also indicated by the attenuation of arc magmatism (Dickinson & Snyder, 1978; Engebretson et al., 1984; Humphreys, 2009). While the cause for the initiation of flat slab subduction is still debated, several mechanisms have been proposed. These include the anchoring of the subducting slab in the deep mantle, slab suction due to low dynamic pressures in the mantle wedge, and subduction of a buoyant oceanic plateau (Humphreys, 2009; Liu & Currie, 2016). The flattening of the slab not only shifted the locus of subsidence, but also played a role in seaway transgressions and regressions, and the ultimate eastward migration of the palaeoshoreline across North America (Smith et al., 1994; Spasojevic et al., 2009). These changes in dynamic topography are recorded in the sedimentation patterns throughout the Green River Basin.

2.2 | Stratigraphy

Upper Cretaceous strata of the greater Green River Basin document a west-to-east transition from alluvial-plain to coastal-plain, deltaic, marine shoreline, to marine-shelf and finally slope environments (Figure 3; Roehler, 1990). Major and minor sedimentary cycles can be identified on a regional scale, with the stacking patterns of the facies responding to changes in sea level and tectonism (Rudolph et al., 2015). Two major (>10 Myr duration) cycles of transgression and regression are observed in the Upper Cretaceous in distal settings. The stratigraphy is described below by basin location in order of most to least studied.

2.2.1 | Distal Basin: Rock Springs, Wyoming

The distal greater Green River Basin east of ca. 110°W (Figures 2 and 3) is one of the most studied and best understood portions of the Late Cretaceous basin due to a relatively complete record of sedimentation, good biostratigraphic control and an accessible suite of wells and outcrops. Proximity to the palaeo-marine shoreline also makes this location ideal for recognizing sequence stratigraphic patterns. The base of the interval of interest is within the Baxter (or Hilliard) Shale. The transition from the underlying Frontier Formation to the Baxter Shale is a second-order marine transgression that has been related to a basin-wide increase in subsidence that may mark the onset of dynamic subsidence in this area (Rudolph et al., 2015). Above the Baxter Shale, the Campanian Mesaverde Group records minor periods of progradation and retrogradation (Roehler, 1990). The Mesaverde Group is approximately 1,500 m thick and contains, from base to top, the Blair, Rock Springs, Ericson and Almond formations (Roehler, 1990).
Blair and Rock Springs formations have progradational stacking patterns and are capped by a second-order sequence boundary, the Moxa unconformity (Figure 3; Rudolph et al., 2015). The Blair and Rock Springs formations are comprised of coastal plain sediments with some fluvial channel deposits, as well as a series of marine tongues (Roehler, 1990). In the southeastern Rock Springs Uplift outcrops, marine shoreline sandstones include the Basal Blair, Chimney Rock, Brooks and McCourt members which are intercalated with marine offshore mudstones (Plink-Björklund et al., 2008; Rudolph et al., 2015).

Deposition of the Mesaverde Group is interrupted by the regional Moxa unconformity, which eroded thick intervals of Mesaverde strata over the course of several million years (Roehler, 1990). The Moxa unconformity was associated with Laramide uplifts and corresponds to a basinward shift in facies (Gomez-Veroiza & Steel, 2010). In distal areas, such as the Rock Springs Uplift, the Moxa unconformity splits into two surfaces (Figure 3; Rudolph et al., 2015).
The proximal portion of the basin is moderately well understood. However, the stratigraphic record is incomplete due to removal of thick stratigraphic intervals by the Moxa and Laramide unconformities. Moreover, the stratigraphy is discontinuously preserved in synforms within the Sevier fold-thrust belt. Lastly, the sequence stratigraphic expression and stacking patterns are more difficult to interpret because of the proximal nature of the deposits and associated lack of ammonite biostratigraphic control. Near Coalville, Utah, the base of the section of interest begins with deposition of the Henefer Formation (DeCelles, 1994). The Henefer Formation contains mostly fluvial sandstone and mudstone. However, the upper Henefer Formation contains Inoceramus fossils, indicating some marine influence even in the westernmost outcrops near Echo Reservoir (DeCelles, 1994). The Henefer Formation is overlain by the Echo Canyon and Weber Canyon conglomerates. The Echo Canyon and Weber Canyon conglomerates were mostly deposited in the proximal foredeep and related to movement on the Crawford thrust (DeCelles, 1994).

In the Echo, Utah area, the Hams Fork Conglomerate lies on top of the Echo Canyon Conglomerate and is bound by the Moxa unconformity below and the Laramide unconformity above. The unconformity at the base of the Hams Fork Conglomerate cuts as deep as the Silurian in the fold belt (Coogan & King, 2016). The Hams Fork Conglomerate is composed primarily of a clast-supported pebble to cobble conglomerate with minor amounts of sandstone (DeCelles & Cavazza, 1999). The Hams Fork Conglomerate was deposited in a wedge-top position and formed as a result of thrusting of the Absaroka fault (DeCelles & Cavazza, 1999). Sediment is thought to be sourced from the Willard sheet of the Sevier fold-thrust belt, which may have been reactivated during thrusting on the Absaroka system (DeCelles & Cavazza, 1999). The temporal equivalent of the Hams Fork Conglomerate in the distal basin is unclear due to poor palynological control and limited outcrop exposure between Coalville and Rock Springs (Jacobson & Nichols, 1982). Determining the chronostratigraphic relationship between the proximal and distal basin is a goal of this study.

### 2.2.3 Medial Basin: Kemmerer, Wyoming

The medial region of the basin is the most poorly understood, and the spatial and temporal relationships between the proximal and distal strata remain uncertain. The Adaville Formation is exposed in this location and is interpreted to have been deposited during a period of regression beginning in the late Santonian (ca. 83 Ma; Roehler, 1990). The lowest portion of the Adaville Formation, the Lazeart Member, is characterized by a nearshore marine sandstone that was deposited as part of a prograding deltaic system (Lawrence, 1992). The Lazeart Member interfingers up-section with the coastal plain deposits of the main Adaville Formation. The Adaville Formation consists of an undifferentiated section composed of coastal plain mudstone, coal and fluvial sandstone (Lawrence, 1992). The Little Muddy Conglomerate is also exposed near Kemmerer with deposition taking place during late Santonian (ca. 85 Ma) shortening on the Absaroka thrust (Pivnik, 1990). Similar to the proximal part of the basin, the Moxa and Laramide unconformities have removed thick sections of
Upper Cretaceous strata. The lack of continuous, preserved section, structural complexity and low-resolution biostratigraphic control make stratigraphic correlation of this area difficult.

3 | METHODS

3.1 | Measured sections

In order to examine changes in depositional environments through time from proximal to distal localities within the basin, 10 stratigraphic sections were measured in the southwestern Green River Basin in Utah and Wyoming. Key stratigraphic sections are located near Coalville (Utah), Kemmerer (Wyoming) and Rock Springs (Wyoming; Figure 2). Measured sections from this study were aggregated with or complemented by those from literature. These include those measured by Roehler (1990), Rudolph et al. (2015) and Leary et al. (2015) on the flanks of the Rock Springs Uplift. Furthermore, DeCelles and Cavazza (1999) and DeCelles (1994) measured several sections through the Hams Fork, Echo Canyon and Weber Canyon conglomerates near Coalville. Lawrence (1984) has measured numerous sections in the lowermost Adaville Formation near Kemmerer. Complete measured sections can be found in Appendix S1. Sections were measured using a standard 1.5 m Jacob staff and Brunton compass, and grain sizes were assigned using a standard grain size card.

Palaeocurrent measurements were taken in trough cross-stratified sandstones, three-dimensional exposures of current ripples and sigmoids, and using imbricated clasts in conglomerate (Miall & Middleton, 2003). Measurements were taken in trough cross-stratified sandstone using the biplanar method described in DeCelles et al. (1983) where 3D exposures were present.

3.2 | Well log correlation

Well log correlations were conducted using wells provided by the Utah Department of Natural Resources Division of Oil, Gas, and Mining, the Wyoming Oil and Gas Commission and the Colorado Oil and Gas Conservation Commission (Figure 4). Wells located in the thrust belt were restored according to estimates of thrust movement (Yonkee & Weil, 2015). Wells presently located on the Darby, Absaroka and Crawford sheets were restored 25, 50 and 75 km to the west respectively. Wells were correlated using sequence stratigraphic techniques such as identifying and correlating parasequence stacking patterns and interpreting sequences and component systems tracts (Catuneanu, 2002; Shanley & McCabe, 1994; Van Waggner et al., 1990). Key bounding surfaces of regional significance can be defined by

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FIGURE 4 Base map showing locations of well logs and measured sections used in this study. Well logs and measured sections located in the thrust belt have been restored. Present-day locations are shown in white; restored locations are shown in grey (wells) and red (measured sections). One hundred and fifty-one wells and 14 measured sections comprise the database.
recognizing progradational, retrogradational and aggradational stacking patterns. Specifically, maximum flooding surfaces, sequence boundaries and transgressive surfaces are defined by changes in stacking patterns. These can provide insight into sedimentary cyclicity that is related to local subsidence and uplift and global eustasy. A more complete treatment of the stratigraphic details and methodology as applied to this area can be found in Rudolph et al. (2015).

Correlations, as well as isopach and structure maps were made using the IHS-Markit software Petra™. Measured sections were also correlated with subsurface well log data (Figure 5). Age control from outcrop sections, including palaeontology and detrital zircon data, were used to refine the correlations. This integrated well log-outcrop framework was used to construct isopach and depositional environment and palaeogeography maps.

**FIGURE 5** Type composite section and well log of Upper Cretaceous strata from the southeast flank of the Rock Springs Uplift. Figure modified from Rudolph et al. (2015)
3.3 | Detrital zircon geochronology

Detrital zircon geochronology was used to constrain MDAs of the stratigraphic units. Where applicable, additional detrital zircon U-Pb data from Leary et al. (2015), Painter et al. (2014) and Laskowski et al. (2013) were aggregated with samples from this study to increase the number of ages determined from each unit (Figure 2; Pullen et al., 2014; Saylor & Sundell, 2016). Aggregated samples are located within close stratigraphic and geographical (≤2 km) proximity to one another.

Samples were prepared following standard methods outlined in Gehrels (2000) and Gehrels et al. (2008). Approximately 400 grains were mounted on double-sided tape on a glass slide for analysis. In situ U-Pb geochronology was conducted at the University of Houston via laser ablation-inductively coupled plasma-mass spectrometry (LA-ICP-MS). The $^{206}\text{Pb}/^{238}\text{U}$ ratio was used to determine the age for grains younger than 1.2 Ga, whereas the $^{206}\text{Pb}/^{207}\text{Pb}$ ratio was used to determine age for grains older than 1.2 Ga (e.g. Gehrels, 2012). We focus on the $^{206}\text{Pb}/^{238}\text{U}$ ratio because we are interested in ages that approach the Cretaceous depositional ages of the host strata.

Multiple methods of calculating MDAs have been suggested including the youngest single grain, the youngest graphical peak, the maximum likelihood method, or the youngest grain cluster that overlap at 1σ or 2σ uncertainty, and there is not a single universally accepted method (Barbeau et al., 2009; Coutts et al., 2019; Dickinson & Gehrels, 2009; Ludwig, 2012; Vermeesch, 2021). Therefore, for this study the MDA for each sample was calculated by four methods: the weighted mean of the youngest group of ages overlapping within 1σ and 2σ uncertainty with the youngest grain, the age of the youngest peak in the sample’s probability density plot, and the youngest single grain (Coutts et al., 2019; Dickinson & Gehrels, 2009; Gehrels, 2012). MDAs are calculated based on large sample sets (>300 grains/sample, Appendix S3) maximizing the likelihood of observing young zircons (Coutts et al., 2019; Sharman & Malkowski, 2020). These ages were then compared against inoceramid and ammonite intervals, radiometrically dated ash beds and palynology or other biostratigraphic constraints. The absolute ages of the biostratigraphic control are based upon the timescale of Ogg et al. (2016). The dated ash beds associated with the biozones are documented in Table 1. A more detailed description of detrital zircon methodology and data from analyses are provided in Appendix S3. Zircon-based provenance results are complex and tangential to the primary goals of this study and are therefore not discussed below.

4 | RESULTS

4.1 | Detrital zircon geochronology

Geochronology results are summarized first, as depositional age results were used to constrain the stratigraphic framework and construct palaeogeographic maps. Twenty-five sandstone samples were collected and analysed across 15 different formations with an age range of ca. 85–69 Ma (Table 1; Appendix S3, Table S3 1). The calculated MDA for fifteen samples were younger than or within biostratigraphic constraints and were used to constrain correlations (Table 1; Appendix S3). The youngest grain cluster that overlap at 2σ uncertainty was most consistent with independent chronostratigraphic data where biostratigraphic ages and radiometric ages from ash beds tightly constrain the true depositional age of the stratigraphy. We, therefore, extended the chronology to regions with poor chronostratigraphic control based on MDAs calculated from the youngest grain cluster that overlap at 2σ uncertainty. MDAs are discussed below by location, from the proximal to distal basin, and in order of oldest to youngest strata.

In the proximal basin, no zircon grains approximating depositional age were obtained for the Echo Canyon and Weber Canyon formations (samples 3 and 4, Figure 3). However, Haque et al. (2020) interpret a depositional age range of >85 Ma and ca. 85–76 Ma, respectively, based on correlation to the geomagnetic polarity time scale. Detrital zircon analysis helped to constrain the timing of deposition of the Hams Fork Formation. MDAs from samples collected in Coalville and Kemmerer locations are within 2σ uncertainty. Sample #1 (HFLCR) from the Hams Fork Formation near Coalville was found to have an MDA of 71.5 ± 4.5 Ma (2σ) and is within uncertainty of the MDAs from samples 5 and 6. Triceratops cf. T. flabellatus Marsh (synonym for T. horridus) was identified in the Hams Fork Formation (Oriel & Tracey, 1970; Rubey et al., 1961) approximately 6.6 km south of sample #5 (HFK-01) and within the same, contiguous outcrop belt. The calculated MDAs for samples 5 and 6 (73.8 ± 1.8 Ma and 71.6 ± 0.9 Ma respectively) are slightly older than, but consistent with, the Triceratops zone (70.6–66.0 Ma), and refine published age estimates from the upper Aquilapollenites quadrilobus zone (78.5–69.5 Ma; Jacobson & Nichols, 1982; Lehman, 1987; Tracy & Oriel, 1959).

Calculated MDAs for the Adaville and Little Muddy formations in the Kemmerer locality were found to be within palynology interval zones. Sample #10 (LMK01) from the Little Muddy Formation has an MDA of 85.5 ± 1.1 Ma, which lies within the Proteacidites retusus interval zone 88.4–84.2 Ma (Jacobson & Nichols, 1982). Sample #8 (UAK01) from the Adaville Formation has an MDA of
| Sample                  | DZ age          | Paleon stratigraphic control | Ammonites correlated to ash beds | Source                                                                 |
|------------------------|-----------------|------------------------------|----------------------------------|------------------------------------------------------------------------|
| Coalville, Utah        |                 |                              |                                  |                                                                        |
|                        |                 |                              |                                  |                                                                        |
| HFLCR                  | 1σ 70.7 ± 2.1 (2; 0.0) | Triceratops zone 70.6–66.0 Ma (Hams Fork) |                                    | 1Lehman (1987); Tracy and Oriel (1959); 2Jacobson and Nichols (1982) |
|                        | 2σ 71.5 ± 4.5 (3; 1.3) | Aquilapollenites quadrilobus Interval Zone 78.5–69.5 Ma (Hams Fork) |                                    |                                                                        |
|                        | YGP 70.7 ± 4.9 |                              |                                  |                                                                        |
|                        | YSG 70.7 ± 3.1 |                              |                                  |                                                                        |
| Kemmerer, Wyoming      |                 |                              |                                  |                                                                        |
|                        |                 |                              |                                  |                                                                        |
| HFK02.25               | 1σ 70.4 ± 0.9 (7; 0.7) | Triceratops zone 70.6–66.0 Ma (Hams Fork) |                                    | 1Lehman (1987); Tracy and Oriel (1959); 2Jacobson and Nichols (1982) |
|                        | 2σ 71.55 ± 0.9 (15; 1.4) | Aquilapollenites quadrilobus Interval Zone 78.5–69.5 Ma (Hams Fork) |                                    |                                                                        |
|                        | YGP 73.5 ± 1.2 |                              |                                  |                                                                        |
|                        | YSG 68.6 ± 3.0 |                              |                                  |                                                                        |
| HFK01                  | 1σ 72.7 ± 2.3 (2; 0.2) | Triceratops zone 70.6–66.0 Ma (Hams Fork) |                                    | 1Lehman (1987); Tracy and Oriel (1959); 2Jacobson and Nichols (1982) |
|                        | 2σ 73.8 ± 1.8 (3; 1.1) | Aquilapollenites quadrilobus Interval Zone 78.5–69.5 Ma (Hams Fork) |                                    |                                                                        |
|                        | YGP 72.0 ± 3.0 |                              |                                  |                                                                        |
|                        | YSG 72.3 ± 3.1 |                              |                                  |                                                                        |
| LMK01                  |                 |                              |                                  |                                                                        |
|                        |                 |                              |                                  |                                                                        |
| LMK01                  | 1σ 83.7 ± 2.9 (2; 0.6) | Palynology: Pseudoplicapalis newmanni zone 84.2–81.1 (Adaville) |                                    | Nichols et al. (1982)                                                   |
|                        | 2σ 83.7 ± 2.8 (2; 0.7) |                               |                                  |                                                                        |
|                        | YGP 84.0 ± 4.4 |                              |                                  |                                                                        |
|                        | YSG 82.0 ± 5.0 |                              |                                  |                                                                        |
| LZAK01                 | 1σ 83.7 ± 1.4 (7; 0.7) | Between Palynology Zones: Pseudoplicapalis newmanni zone 84.2–81.1 (Adaville) and Proteacidites retusus Interval Zone 88.4–84.2 (Hillard) |                                    | Nichols et al. (1982)                                                   |
|                        | 2σ 84.8 ± 1.4 (10; 1.3) |                               |                                  |                                                                        |
|                        | YGP 96.0 ± 0.77 |                              |                                  |                                                                        |
|                        | YSG 82.0 ± 0.77 |                              |                                  |                                                                        |
| Rock Springs, Wyoming  |                 |                              |                                  |                                                                        |
|                        |                 |                              |                                  |                                                                        |
| CCERS02.110            | 1σ 72.7 ± 1.4 (7; 0.9) | Baculites reesidei 73.6–73.3 ± 1 (lwr Almond) | 72.94 ± 0.45 (B. reesidei) | 1Gill et al. (1970); 2Baadsgaard et al. (1993); Cobban et al. (2006) |
|                        | 2σ 72.5 ± 1.5 (6; 1.0) |                               |                                  |                                                                        |
|                        | YGP 75.0 ± 1.6 |                              |                                  |                                                                        |
|                        | YSG 70.0 ± 2.7 |                              |                                  |                                                                        |
Table 1 (Continued)

| Sample   | DZ age | Paleo constraints | Ammonites correlated to ash beds | Source |
|----------|--------|-------------------|---------------------------------|--------|
| CCERS01  | 1σ     | 72.5 ± 0.96 (11; 0.5) | B. reesidei 73.6–73.3 ± 1 (lwr Almond) | 72.94 ± 0.45 (B. reesidei) | Gill et al. (1970); Baadsgaard et al. (1993); Cobban et al. (2006) |
|          | 2σ     | 72.98 ± 0.4 (28; 0.5) |                                  |        |
|          | YGP    | 75.0 ± 0.9         |                                 |        |
|          | YSG    | 70.3 ± 2.7         |                                 |        |
| TERS01   | 1σ     | 77.0 ± 15 (2; 1.2)  | B. reesidei 73.6–73.3 ± 1 (lwr Almond), Baculites sp (weak ribs) 81.1–80.9 ± 1 (Black Butte) | 72.94 ± 0.45 (B. reesidei) | Gill et al. (1970); Smith (1965); Baadsgaard et al. (1993); Cooper (1994); Cobban et al. (2006) |
|          | 2σ     | 77.4 ± 4.2 (2; 1.2) |                                  |        |
|          | YGP    | 79.0 ± 11.9        |                                 |        |
|          | YSG    | 76.0 ± 3.3         |                                 |        |
|          |        |                    | **Depositional ages with good biostratigraphic control** |        |
| ALRS01   | 1σ     | 73.2 ± 2.0 (3; 0.6 ) | B. baculus 72.1–71.1 ± 1 (upr Almond), B. eliasi 72.7–72.1 ± 1 (upr Almond), Palynology: Erdtmanipollis ≤72.1 (mid Almond), B. reesidei 73.6–73.3 ± 1 (lwr Almond) | 71.98 ± 0.31 (B. eliasi) | Gill et al., (1970); Rudolph et al. (2015); Cobban et al. (2006); Baadsgaard et al. (1993) |
|          | 2σ     | 73.2 ± 2.0 (3; 0.7) |                                  |        |
|          | YGP    | 74.0 ± 3.5         |                                 |        |
|          | YSG    | 71.8 ± 3.3         |                                 |        |
| TERS02.51| 1σ     | 80.4 ± 1.3 (1)     | B. reesidei 73.6–73.3 ± 1 (lwr Almond), Baculites sp (weak ribs) 81.1–80.9 ± 1 (Black Butte) | 72.94 ± 0.45 (B. reesidei) | Gill et al. (1970); Smith (1965); Baadsgaard et al. (1993); Cooper (1994); Cobban et al. (2006) |
|          | 2σ     | 80.4 ± 2.5 (1)     |                                  |        |
|          | YGP    | 80.4 ± 2.5         |                                 |        |
|          | YSG    | 80.4 ± 2.5         |                                 |        |
| URSRS01.58| 1σ   | 80.8 ± 1.6 (4; 0.6) | B. baculus sp (weak ribs) 81.1–80.9 ± 1 (Black Butte), S. hippocrepis III 81.5–81.3 ± 1 (Chimney Rock and Upr Blair) | 80.58 ± 0.55 (Scaphites preventricosus) | Smith (1965); Baadsgaard et al. (1993); Cobban et al. (2006) |
|          | 2σ     | 82.5 ± 2.2 (7; 2.1) |                                  |        |
|          | YGP    | 80.6 ± 2.1         |                                 |        |
|          | YSG    | 79.9 ± 1.7         |                                 |        |
| BBRS02.7 | 1σ     | 80.7 ± 2.3 (4; 0.3) | S. hippocrepis III 81.5–81.3 ± 1 (Chimney Rock and Upr Blair), Desmoscaphites bassleri 84.6–84.2 ± 1 (upr Baxter) | 81.86 ± 0.36 (B. obtusus) | Smith (1965); Cobban et al. (2006) |
|          | 2σ     | 83.3 ± 2.5 (8; 1.7) |                                  |        |
|          | YGP    | 90.0 ± 2.0         |                                 |        |
|          | YSG    | 79.8 ± 4.5         |                                 |        |
|          |        |                    | **Depositional ages with poor biostratigraphic control** |        |
| PRERR-01 | 1σ     | 69.1 ± 1.56 (1)    | B. reesidei 73.6–73.3 ± 1 (lwr Almond), Baculites gregoryensis 78.4–77.6 ± 1 (Allen Ridge) | 72.94 ± 0.45 (B. reesidei) | Gill et al. (1970); Cobban (1958); Baadsgaard et al. (1993); Cobban et al. (2006) |
|          | 2σ     | 71.9 ± 8.5 (3; 3.3) |                                  |        |
|          | YGP    | 69.1 ± 3.07        |                                 |        |
|          | YSG    | 69.1 ± 3.07        |                                 |        |
83.7 ± 2.8 Ma, which is consistent with *Pseudoplicapalis nevmannii* zone 84.2–81.1 Ma (Nichols et al., 1982). Giant oysters of the species *C. cusseta* were also identified in the Adaville Formation and are consistent with a Campanian age (Sohl & Kauffman, 1964). The Lazeart Formation is bracketed by these two palynology zones, which is consistent with its detrital zircon MDA (84.8 ± 1.4 Ma). Samples #6, (HFK02.25) and #5 (HFK01) in Kemmerer were found to have MDAs of 71.55 ± 0.95 Ma and 73.8 ± 1.8 Ma respectively (Table 1).

In the distal basin the MDA for the base of the Blair Formation was calculated to be 83.3 ± 2.5 Ma, which is consistent with the presence of *Scaphites hippocrepis III* (81.5–81.3 ± 1 Ma) in the upper Blair Formation, and *Desmoscaphites bassleri* (84.6–84.2 ± 1 Ma) in the upper Baxter Formation (Ogg et al., 2016; Smith, 1965). The MDA from the Trail Member of the Ericson Formation in Rock Springs (77.4 ± 4.2 Ma) is bracketed by the *Baculites reesidei* ammonite zone (73.6–73.3 ± 1 Ma) in the lower Almond Formation and the *Baculites sp* (weak ribs) ammonite zone (81.1–80.9 ± 1 Ma) in the upper Rock Springs Formation (Gill et al., 1970; Ogg et al., 2016; Smith, 1965). Samples #11 (CCERS02.110) and #17 (CCERS01) of the Canyon Creek Member in Rock Springs have ages of 72.5 ± 1.5 Ma and 72.98 ± 0.44 Ma respectively. The Canyon Creek Member is thought to be equivalent to the Pine Ridge Formation to the east (Rudolph et al., 2015), and the calculated MDA of Pine Ridge sample #22 (PRERR-01; 71.9 ± 8.5 Ma) is consistent with this.

### 4.2 | Stratigraphic framework

A regional cross-section was constructed across the study area using measured sections and well logs (Figure 6). Age control from ammonites, inoceramids, palynology and detrital zircon geochronology are displayed on measured sections and were used to constrain correlations. Age ranges associated with biostratigraphic control symbols can be found in Table 1.

Maximum flooding surfaces, transgressive surfaces (marine environments) and corresponding abandonment surfaces (non-marine settings), and sequence boundaries were identified. The lowest-order boundaries (second order) identified in the study interval define sequence sets. Second-order surfaces, from oldest to youngest, include the Frontier sequence boundary, Baxter maximum flooding surface, Moxa 1 unconformity, Moxa 2 unconformity, Lewis maximum flooding surface and Laramide unconformity (Figure 6; Rudolph et al., 2015). The Moxa 2 unconformity is considered the master unconformity, and the Moxa 1 unconformity ultimately converges with the Moxa 2 unconformity in the subsurface west of Rock
Springs (Rudolph et al., 2015). Second-order sequence sets, defined by the shaded intervals, include the early and late highstand, lowstand and transgressive sequence sets. Third-order and fourth-order sequences further subdivide the sequence sets, breaking up formations into higher-order (smaller scale) stacking patterns. Other correlations, usually flooding surfaces at the tops of parasequences and parasequence sets, are indicated in black. This stratigraphic framework has been extended throughout the entire well and outcrop section database (Figure 4).

The upper Baxter, Blair and Rock Springs formations constitute a second-order highstand sequence set marked by an overall progradation of facies. The third-order base of Blair sequence boundary subdivides the early and late highstand sequence sets. The highstand sequence set is primarily composed of marine strata in the distal basin, with the exception of the Gottsche Member of the upper Rock Springs Formation. The Moxa 1 unconformity marks the boundary between the second-order highstand and lowstand sequence sets. The fluvial Ericson Formation comprises the second--order lowstand sequence set, and has an aggradational stacking pattern (Rudolph et al., 2015). This is overlain by the retrogradational Almond Formation, which makes up the transgressive sequence set. The Almond Formation transitions from non-marine to marine up section and is ultimately capped by the second-order maximum flooding surface in the lowermost Lewis shale and associated with an organic-rich condensed section (Pyles & Slatt, 2007). The Lewis maximum flooding surface is the base of the highstand sequence set which is characterized by progradation. Much of the highstand systems tract has been removed by erosion represented by Laramide unconformity in the outcrops of the Rock Springs Uplift. In the Washakie Basin to the east, more of this interval is preserved and is represented by a thick Fox Hills Formation (shallow marine) and Lance Formation (non-marine); this can be observed in the easternmost well on the regional stratigraphic cross-section (Figure 6).

Campanian to Maastrichtian strata (ca. 83–71 Ma) thin dramatically from east to west. This thinning is evidenced by incremental depositional thinning, as well as the truncation of multiple surfaces beneath the Laramide, Moxa 1 and Moxa 2 unconformities, which merge to the west. The Moxa cuts into progressively older strata in the proximal area, ultimately juxtaposing rocks as old as the Lower Jurassic Nugget Formation (Lost Creek Reservoir exposure) and Silurian Laketown Dolomite (Causey Reservoir outcrops) against the overlying Maastrichtian Hams Fork Formation. Younger strata, such as the Canyon Creek Member of the Ericson Formation, are inferred to onlap the Moxa 2 unconformity towards the proximal basin. The Lewis, Fox Hills and Lance formations, which are >1 km thick to the east in the Washakie Basin, thin to <200 m near the thrust belt where it is eroded by the Laramide unconformity.

The structural cross section highlights the current structural relief of the Sevier fold-thrust belt as well as the Laramide Rock Springs Uplift (Figure 7). The Rock Springs Uplift is juxtaposed on either side by the regional lows of the West Green River Basin and the Washakie Basin.

4.3 Depositional systems

Measured sections and well log character were used to determine the evolution of depositional environments throughout the basin. Depositional environments were assigned according to the characteristics outlined in Table 2. A description of depositional elements can be found in Appendix S1, Table S1 1. Representative field photos of lithofacies associated with interpreted depositional environments are documented in Appendix S2. Four time slices were evaluated, spanning from the early Campanian (ca. 81 Ma) to the early Maastrichtian (ca. 71 Ma). On the resulting maps, the approximate restored positions of the coeval Crawford, Early Absaroka or Late Absaroka thrusts are indicated, conforming to the timing as interpreted by Yonkee and Weil (2015).

During the deposition of the Brooks Member of the Rock Springs Formation (ca. 81 Ma), the basin is characterized by fluvial/ coastal plain, shallow marine and offshore marine environments (Figure 8a). In general, the shorelines are oriented southwest-northeast. The western boundary of Brooks Member-equivalent strata is defined by its truncation beneath the Moxa 2 unconformity. Palaeocurrent measurements within fluvial equivalents of the Brooks Member indicate flow was to the east-southeast during this time.

Deposition of the Trail Member of the Ericson Formation during the late Campanian (ca. 77 Ma) marks the transition from the late highstand to early lowstand sequence set (Figure 8b) and an associated basinward shift in facies. The Trail Member is entirely fluvial in character, transitioning from high net-to-gross to low net-to-gross sandstone thickness from west to east based on well logs (Figure 6). Palaeocurrent measurements from Leary et al. (2015) demonstrate palaeoflow was predominantly to the east/southeast on the northwest side of the uplift. On the eastern side of the uplift, palaeocurrent measurements vary considerably, with flow to the north, east and southeast. Both observations are consistent with a change from braided to meandering fluvial deposition. The orientation of the boundary between braided fluvial and
Integrated cross-section showing well logs and measured sections. Measured section located in Coalville, Utah is a composite section from DeCelles (1994) and DeCelles and Cavazza (1999). Measured section located in Kemmerer, WY is a composite section from this study and Lawrence (1984). Measured section in northwest Rock Springs is a composite section from this study, Leary et al. (2015) and Roehler (1990). Measured section in southeast Rock Springs is a composite section from Rudolph et al. (2015). Maximum flooding surfaces are shown in blue, transgressive and abandonment surfaces are shown in green and sequence boundaries are displayed in red. Age control from ammonites, inoceramids and detrital zircon geochronology is displayed on measured sections. Location of cross-section line is shown in Figure 4. Ammonite and palynology from Ogg et al. (2016), Hendricks (1983), Jacobson and Nichols (1982), Nichols et al. (1982), Roehler (1978), Cobban (1969), Smith (1965), Weimer (1961) and Hale (1950)
meandering fluvial deposits closely matches that of the palaeogeographical map of the Trail shown by Roehler (1990) and Minor et al. (2021). The Trail Member transitions eastwards to the high-sinuosity coastal plain deposits of the lower Allen Ridge Formation in the area of the Rawlins Uplift (Figure 3; Roehler, 1990; Rudolph et al., 2015). Further downdip, its equivalent is the deltaic shoreline sandstones of the Parkman Formation, in the Powder River Basin (Figure 3). Similar to the Brooks Member, the western edge of the Trail is defined by truncation beneath the Moxa 2 unconformity. Palaeocurrent measurements taken on the flank of the Uinta Mountains indicate flow to the north or northeast, consistent with the initial exhumation of the Uinta Mountains during this interval (Leary et al., 2015; Saylor et al., 2020).

Like the Trail Member, the Canyon Creek Member of the Ericson Formation (ca. 73 Ma) is predominantly fluvial (Figure 8c). High net-to-gross fluvial deposits characterize the Canyon Creek Member to the west of the Rock Springs Uplift transition to low net-to-gross fluvial deposits east of the uplift. There is also a decrease in grain size, with pebbly sandstones common in the outcrops of the northwest uplift. The depositional environment of the Canyon Creek in outcrop is most commonly interpreted as braided fluvial (Leary et al., 2015; López & Steel, 2015; Roehler, 1990; Rudolph et al., 2015). However, Martinsen et al. (1999) described the proximal Canyon Creek as a hybrid meandering-braided system. While the absence of significant preserved floodplain deposits hampers interpretation, the lack of point bar elements favours a low sinuosity interpretation, possibly near its transition to high sinuosity deposits, based on well logs to the east (Figure 6).

Downdip, the Canyon Creek Member is equivalent to the fluvial Pine Ridge and fluvo-deltaic Teapot formations (Figure 3). Distal equivalents show increasing thickness, but lower net-to-gross ratio and limited connectivity of channels (López & Steel, 2015). In the Sand Wash Basin in northwestern Colorado, the Canyon Creek Member is thought to be equivalent to the Williams Fork Formation, based on ammonite control, as well as sanidine and zircon...
dating (Cobban et al., 2006; Minor et al., 2021; Walker et al., 2021). The Williams Fork Formation contains coastal plain deposits as well as the shoreline sandstones of the Trout Creek and Twentymile members (López & Steel, 2015). The orientation and location of the shoreline is consistent with palaeogeographic maps from Roehler (1990). Alluvial fan deposits are interpreted based on well log character along the western Uinta Mountains and adjacent to the thrust front, where the Canyon Creek Member onlaps onto the Moxa 2 unconformity, consistent with detrital zircon dating. Palaeocurrent measurements from Leary et al. (2015) and López and Steel (2015) illustrate that flow direction is highly variable, however, this is likely due to the aggregation of palaeocurrent measurements over a large vertical interval and area. Most measurements are directed towards the east-southeast. Palaeocurrent measurements taken adjacent to the Uinta Mountains indicate flow to the north-northwest.

Deposition of the Hams Fork Formation (ca. 71 Ma), and its distal equivalent the Almond Formation, marks the beginning of a transgressive sequence set and a landward shift in facies (Figure 8d). In its most proximal location, the Hams Fork Formation is characterized by alluvial fan deposits, which transition to braided fluvial near Kemmerer. Palaeocurrent measurements from DeCelles and Cavazza (1999) taken on imbricated clasts in alluvial fan deposits, as well as those from this study in braided fluvial deposits are oriented to the east and south. The shift from braided fluvial to coastal plain deposits marks the eastern extent of the Hams Fork Formation. The Almond Formation contains both marine and non-marine deposits. West of Rock Springs the Almond is deposited in a non-marine coastal plain setting, with some tidal influence present in the lower coastal plain deposits. The shoreline is oriented roughly N–S along the western flank of the Rock Springs Uplift, slightly curving to the east near the Uinta Mountains. Palaeocurrent measurements from the Almond were taken only within the fluvial elements of what is interpreted as an overall shallow marine section and indicate flow to the north and northeast (Keift et al., 2011). East of the shoreline, the upper Almond Formation transitions to the offshore and distal offshore marine facies characteristic of the Lewis Shale (Roehler, 1990).

| Gross depositional environment | Depositional elements | Lithology/grain size | Ss+Cgl/gross |
|-------------------------------|-----------------------|---------------------|-------------|
| Alluvial fan                  | Fluvial channel complex, debris flows (minor) | Clast-supported pebble to cobble conglomerate; medium to coarse red sandstone lenses | 85%–100% |
| Braided fluvial               | Channel/bar complex, overbank deposits, abandoned channel | Medium to coarse tan sandstone, thin interbeds of brown siltstone and mudstone | 75%–100% |
| Meandering fluvial            | Point bars, levee/floodplain | Fine to medium tan sandstone, intervals of brown siltstone and mudstone | 40%–75% |
| Coastal plain                 | Small fluvial systems, floodplain, crevasse splays | Predominantly brown siltstone and mudstone with very fine brown sandstone lenses, presence of dark brown to black carbonaceous material and coals | 10%–30% |
| Lower coastal plain/Back-barrier | Interdistributary bay, coals, tidal channels/bars, tidal flat, crevasse splays, floodplain | Very fine to fine lower tan sandstone, grey mudstone and siltstone | 20%–45% |
| Shallow marine                | Foreshore, upper shoreface, proximal lower shoreface, tidal bars | Thickly bedded fine upper to medium buff sandstone, interbedded grey mudstone and siltstone with fine lower tan sandstone | 50%–80% |
| Offshore marine               | Distal lower shoreface, offshore mudstones | Interbedded grey mudstone and siltstone with tan fine lower sandstone | 5%–20% |
| Distal offshore marine        | Offshore | Dark grey clay and mudstone, very fine tan sandstone lenses | 5%–15% |

Abbreviations: Cgl, conglomerate; Ss, sandstone; Ss+Cgl/Gross, is equal to the thickness of sandstone and conglomerate facies divided by the gross thickness.

### 4.4 Regional thickness mapping

Seven isopach maps were generated in order to determine the change in shape and locus of depocentres during the transition from Sevier to Laramide tectonic regimes (Figure 9). Isopach maps span from the late Coniacian to
the late Maastrichtian (ca. 88–66 Ma). Because the correlations are being extended into proximal settings, many of the intervals are bounded below and/or above by unconformities (sequence boundaries). The surfaces used to create the isopach maps were pragmatically chosen, based on the reliability of the correlations and their preservation below either the Moxa or Laramide unconformities. Also, we did not want to use intervals that span major changes in sedimentary accommodation/preservation patterns, especially the Moxa unconformities. It should also be noted that the resulting maps of interval thicknesses do not always reflect original depositional thicknesses due to erosion below the unconformities, and onlap above; areas of significant erosion are noted on the maps and is an essential objective of the isochore mapping.

The first interval begins at the Baxter maximum flooding surface and extends to the base of Blair sequence boundary (late Coniacian–early Campanian, ca. 88–83 Ma; Figure 9a). This interval is thickest against the thrust front (ca. 1,200 m), thinning gradually towards the distal basin (ca. 600 m). Based on biostratigraphy (Cobban, 1969; Jacobson & Nichols, 1982), the upper Baxter is equivalent to the proximal and wedge top conglomeratic Echo Canyon Formation and the upper part of the Henefer Formation. At this location, Late Cretaceous strata have been heavily eroded by the Moxa 2 unconformity, and the upper Baxter equivalents have been uplifted and thinned to as little as 100 m. Only pockets of the Henefer and Echo Canyon formations are preserved, such as in the Stevenson Canyon Syncline (DeCelles, 1994).

The next interval extends from the base of Blair sequence boundary to the top of Blair sequence boundary (early Campanian–mid Campanian, ca. 83–81 Ma; Figure 9b). The Blair Formation displays incremental, depositional thickening from the west (350 m) to southeast (750 m) north of the Uinta Mountains. West of the Moxa Arch, the Blair Formation is heavily eroded and

FIGURE 8  Paleogeographic map of the Green River Basin during (a) the deposition of the Brooks member of the Rock Springs Formation (early Campanian, ca. 81 Ma). Paleocurrent measurements are plotted on a rose diagram. (b) Trail member of the Ericson Formation (late Campanian, ca. 77 Ma). Paleocurrent measurements are plotted on a rose diagram and sourced from Leary et al. (2015). (c) Canyon Creek member of the Ericson Formation (late Campanian, ca. 73 Ma). Paleocurrent measurements are plotted on a rose diagram and sourced from Leary et al. (2015) and López and Steel (2015). Boundary of coastal plain and location of shoreline in Colorado referenced from López and Steel (2015). (d) Hams Fork member of the Evanston Formation and upper Almond Member of the Mesaverde Formation (early Maastrichtian, ca. 71 Ma). Paleocurrent measurements are plotted on a rose diagram. Paleocurrents in Coalville, Utah locality sourced from DeCelles and Cavazza (1999). Paleocurrents on the southeast side of the Rock Springs Uplift sourced from Kieft et al. (2011). Only wells and measured sections used to construct map are shown
thinned (down to 0 m) by the Moxa unconformity. The top of Blair sequence boundary to the Moxa 1 unconformity (mid Campanian, ca. 81–77 Ma) shows similar thinning along the Moxa Arch and is entirely eroded over the crest of the arch (Figure 9c). This interval thickens dramatically to 1,300 m in the southeast.

The interval from the Moxa 1 to Moxa 2 unconformity (mid Campanian–late Campanian, ca. 77–74 Ma) further demonstrates depositional and erosional thinning to the west (Figure 9d). The absence of a preserved foredeep depozone for this and younger intervals is consistent with the isopach maps of Painter and Carrapa (2013) and Li

**FIGURE 9** Isopach maps from the (a) late Coniacian–early Campanian, ca. 88–83 Ma; (b) early Campanian–mid Campanian, ca. 83–81 Ma; (c) mid Campanian, ca. 81–77 Ma; (d) mid Campanian–late Campanian, ca. 77–74 Ma; (e) late Campanian, ca. 74–73 Ma; (f) late Campanian–early Maastrichtian, ca. 73–71 Ma; (g) early Maastrichtian–late Maastrichtian, ca. 71–66 Ma. Coloured polygon outline thickening is attributed to different modes of subsidence. Magenta: thickening along Sevier foldbelt; black: thickening in the distal position; blue: thickening against Laramide uplifts.
and regional correlations from this study, however, indicate the Canyon Creek Member is older than the Hams Fork Formation, augmented with detrital zircon geochronology. Published age estimates from the upper Almond, or younger strata. Ammonites found in the upper Almond Formation, such as Baculites baculus (72.1–71.1 ± 1), are consistent with the detrital zircon MDA from the Hams Fork (ca. 71 Ma). These additional age constraints have allowed for the creation of palaeogeographic maps throughout the deposition of the Canyon Creek and Hams Fork formations.

Palaeogeographic maps of the late Campanian (ca. 73 Ma) and the early Maastrichtian (ca. 71 Ma) are shown in Figure 10. Deposition of the Canyon Creek Member was synchronous with late movement on the Absaroka thrust as well as the initial uplift of the Uinta Mountains, albeit of modest elevation (Saylor et al., 2020). This deformation produced alluvial fan deposits, which fed into coarse-grained, low sinuosity fluvial systems that comprised a broad braid plain (Figure 10a; Yonkee & Weil, 2015). These rivers flowed mostly to the southeast from the fold belt, but also locally northward off of the Uinta Mountains. Braided rivers transition to meandering rivers east of the Rock Springs Uplift, ultimately becoming small streams and crevasse splays deposited on the coastal plan. Locations of the coastal plain and shoreline are based on López and Steel (2015).

Deposition of the Hams Fork Formation and its distal equivalent the upper Almond Formation marks a significant transgression with a landward shift in facies (Figure 10b). The Hams Fork Formation is typically a pebble to cobble conglomerate in the proximal wedge-top depozone. Near Kemmerer the Hams Fork Formation changes character and represents deposition by medium-grained braided rivers and is often very poorly exposed. Braided rivers may have been present north of the western Uinta Mountains; however, these deposits are not preserved due to erosion by the Laramide unconformity. East of Kemmerer the Hams Fork Formation transitions to coastal plain and marine deposits of the Almond Formation. The shoreline is thought to have changed orientation from roughly N–S to E–W near the Wind River Mountains, which, based on isopach mapping, were being uplifted at this time (Figure 8c,d). The change in orientation of the shoreline is consistent with observations by Roehler (1990). The deposition of thick marine deposits in the distal basin is attributed to flexure from the Laramide uplift of the Granite Mountains (Johnson et al., 2004a). The shorter wavelength and localized nature of the subsidence suggests flexure is the dominant mechanism rather than dynamic subsidence (Figure 10a; Saylor et al., 2020).

5.1 Palaeogeographic reconstructions

The lack of exposure near Kemmerer has previously made it difficult to determine the distal equivalent of the Hams Fork Formation, and the chronostratigraphic relationship between the proximal and the distal basin. The Canyon Creek Member of the Ericson formation was previously thought to be equivalent to the Hams Fork Formation (DeCelles & Cavazza, 1999). Published vertebrate palaeontology, augmented with detrital zircon geochronology and regional correlations from this study, however, indicate the Canyon Creek Member is older than the Hams Fork Formation, and onlaps onto the Moxa 2 unconformity to the west as it approaches the Sevier fold-thrust belt.

More definitively, Triceratops cf. T. flabellatus Marsh (synonym for T. horridus) was identified in the Hams Fork Formation (Oriel & Tracey, 1970; Rubey et al., 1961) approximately 6.6 km south of sample #5 (HFK-01) and within the same, contiguous outcrop belt. Zircon MDAs from HFK-01 (73.8 ± 1.8 Ma) are slightly older than, but consistent with, the Triceratops zone (70.6–66.0 Ma), and refine published age estimates from the upper Aquilapollenites quadrilobus zone (78.5–69.5 Ma; Jacobson & Nichols, 1982; Lehman, 1987; Tracy & Oriel, 1959). The numerous occurrences of Triceratops in North America are dated as Maastrichtian. Therefore, the Hams Fork is interpreted to be correlative to upper Almond, or younger strata.
5.2 Regional depositional patterns

Regional cross sections, depositional environment maps and isopach maps illuminate the effects of tectonics on deposition and basin architecture. Isopach maps clearly demonstrate three distinct styles of subsidence (Figure 9). These include flexural subsidence associated with the Sevier Orogeny, long-wavelength subsidence attributed to mantle convection, and flexural subsidence associated with local Laramide uplifts. Flexural subsidence due to the Sevier Orogeny is demonstrated by a foredeep depozone adjacent to the thrust front. Similarly, Laramide flexural subsidence can be identified by a local increase in accommodation adjacent to uplifts.

Long-wavelength, dynamic subsidence, on the other hand, is best demonstrated as a thickening of deposits away from the thrust front and a widening of the basin. Note that dynamic subsidence, as previously described in the Western Interior (Liu & Nummedal, 2004; Painter & Carrapa, 2013), occurs over wavelengths of about 500 km. Therefore, this study area is too small to fully document its distribution. Flexural models also help to validate the interpreted origin of the Baxter foredeep and most of the Laramide flexures interpreted (Rudolph, 2018; Saylor et al., 2020).

The isopach map from the Baxter maximum flooding surface to the base of Blair sequence boundary (late Coniacian–early Campanian, ca. 88–83 Ma) shows wedge-thickening towards the contemporaneous Crawford Thrust, indicating a flexural foredeep depozone (Figure 9a). In the distal basin, however, this interval is still in excess of 500 m thick. This thickening in the distal basin is not observed in previous time slices (e.g. Frontier Formation; DeCelles, 2004; Liu & Nummedal, 2004; Liu et al., 2005, 2011, 2014; Rudolph et al., 2015). The presence of thick deposits of upper Baxter in the distal basin, which extend hundreds of kilometres from the thrust belt, could be a result of the initiation of flat slab subduction and eastward migration of long-wavelength subsidence (Painter & Carrapa, 2013; Roberts & Kirschbaum, 1995). Thus, this interval is interpreted as reflecting dynamic subsidence superimposed on a shorter-wavelength flexural foredeep.

By the early–mid Campanian (ca. 83–81 Ma), long-wavelength subsidence was focused in eastern areas, creating a broader depocentre. The locus of deposition also shifted basinward, with deposits thickening distally to the southeast (Figure 9b) relative to the previous time slice, as observed in previous regional interpretations (Liu et al., 2014; Painter & Carrapa, 2013). In the western (Green River) basin, extensive erosion above the Moxa Arch has removed hundreds of metres of the Blair Formation. This interval, however, does not appear to thicken towards the thrust belt prior to its erosion (Figure 6). Instead, the Blair Formation depositionally thins by several hundred metres (>300) from the easternmost well to those west of Rock Springs before it is ultimately truncated (Figure 6).

This suggests a traditional foredeep was not developed, and rather, that the proximal area was being supported and uplifted by the flattening of the slab (Eakin et al., 2014; Heller & Liu, 2016). While we propose that there was no preserved foredeep in Campanian to Maastrichtian strata, the lack of preservation up dip due to the erosion at the base of the Moxa and Laramide unconformities and the lack of chronostratigraphic control make it difficult to distinguish
between erosional and depositional thinning. It should be noted that post-Moxa strata (<81 Ma) also do not indicate a foredeep and Campanian-Maastrichtian thickening associated with a proximal foredeep depozone have not been documented along strike from northern Wyoming to northern New Mexico, (Li & Aschoff, 2021; Painter & Carrapa, 2013). To the north, there is evidence of Campanian-Maastrichtian foredeep development in the Crazy Horse Basin (Johnson et al., 2004a, 2004b), central Montana (Dyman et al., 1995) and the Western Canada Basin (Dawson et al., 1994). This approximately coincides with a decrease in Laramide-style deformation towards the north.

Mid–late Campanian strata (ca. 81–74 Ma) are truncated beneath the Moxa 2 unconformity to the west, and the Rock Springs and lower Ericson formations are no longer preserved in the proximal basin (Figure 6). The locus of deposition continues to migrate eastward during this time, suggesting the continued shift of long-wavelength subsidence. This thickening to the southeast is consistent with previous, more regional interpretations (Liu et al., 2014; Painter & Carrapa, 2013; Roberts & Kirschbaum, 1995).

By the late Campanian (ca. 73 Ma), thickening is observed against the northern flank of the Uinta Mountains (Figure 9e). The modest wedge-thickening, occurrence of coarse pebbly sandstones in outcrop, blocky well log character in nearby wells, and northward palaeocurrent measurements (Leary et al., 2015), suggest the Uinta Mountains were actively being uplifted at this time. Flexural modelling from stratigraphic cross sections infers 200–400 m of topographic uplift of the Uintas during this period (models 16–18, Saylor et al., 2020). Tectonic loading associated with this initial Laramide uplift drove flexural subsidence along the flank of the Uinta Mountains, providing additional accommodation space. The Canyon Creek Member of the Ericson Formation is limited to distal positions whereas in proximal locations this interval is characterized by bypass/erosion, with onlap of the interval to the east. Therefore, the Moxa unconformity must extend several million years younger in the fold belt (Figure 6). Note that the Moxa 2 unconformity is a correlative, yet diachronous, surface throughout Wyoming and is associated with a regional fluvio-deltaic regression, that includes the Canyon Creek, Pine Ridge and Teapot sandstones (Figure 3). This profound degradational to progradational succession is attributed to a period of uplift (updip) and low tectonic subsidence (downdip) as documented by Rudolph (2018). Thick deposits (ca. 200 m) are also present east of the Uinta Mountains, which could represent the westernmost remnant of long-wavelength subsidence. This suggests that long-wavelength subsidence persisted at least until ca. 74 Ma.

Maps from the late Campanian–late Maastrichtian (ca. 73–66 Ma) show similar thickening against the Uinta Mountains, indicating their continued emergence. More drastic thickening is observed to the north and northeast adjacent to the Wind River and Granite mountains respectively (Figure 9f,g). Increased thickness of deposits along the Wind River and Granite mountains are also documented on a more regional scale by Johnson et al. (2004a) and Lynds and Lichtner (2016). These thickening patterns represent local flexural accommodation caused by the tectonic loading and uplift of the Wind River and Granite mountains (Johnson et al., 2004a; Rudolph et al., 2015). Thermochronology from the Wind River and Granite mountains suggest exhumation from 70 to 60 Ma and 71 Ma respectively (Cerveny, 1990; Peyton et al., 2012; Stevens et al., 2016). Both the Wind River and Granite mountains show exhumation in the Maastrichtian, which, due to the lag time between exhumation and the reaching of closure temperature, is consistent with Maastrichtian uplift. This is consistent with the observed isopach patterns. Thinning west of the Rock Springs Uplift, on the other hand, is indicative of lower accommodation and erosion by the Laramide unconformity.

6 | CONCLUSIONS

The integration of detrital zircon geochronology and biostratigraphy indicates the Canyon Creek Member of the Ericson Formation is older than the Hams Fork Formation, though previously thought to be temporally equivalent. Rather, the Canyon Creek Member is interpreted to onlap onto the Moxa 2 unconformity in the proximal basin. Thus, the Moxa 2 unconformity extends for several million years younger in the fold belt. Therefore, while lowstand Canyon Creek fluvial sediments were being deposited, the proximal area was eroded. This eroded area extends from the fold belt to approximately 50 km east of the coeval (late Absaroka) Sevier deformation front. Moreover, it is coincident with the initial uplift of the Uinta Mountains and approximately 750 m of uplift and erosion on the Moxa Arch (Roehler, 1965; Rudolph et al., 2015). Thus, the initiation of the Laramide deformation in western Wyoming occurs by upper Campanian time (ca. 74 Ma).

The Hams Fork Formation is equivalent to the upper Almond Formation in the distal basin. These additional timing constraints allow for the correlation between the proximal and distal basin. Paaleogeographical maps of the late Campanian (ca. 73 Ma) display the westward onlap of the Canyon Creek Member and early Maastrichtian (ca. 71 Ma) maps show the westward (i.e. landward) shift in facies during the deposition of the Hams Fork Formation. Thus, the proximal fluvial/alluvial fan sediments represent transgressive deposits. Moreover, because of variability in subsidence/uplift
and source-to-sink considerations, the stacking patterns seen in the distal basin record are different than those seen in the more incomplete updip stratigraphic record.

Regional thickness patterns help to record and constrain the timing of the transition from Sevier to Laramide tectonic regimes. The Sevier flexural foredeep is preserved from the Baxter maximum flooding surface to the base of the Blair sequence boundary (ca. 88–83 Ma). The presence of deposits with thicknesses >600 m in the distal basin, in a position east of the expected forebulge, indicates an additional mode of subsidence is present. We interpret this as the initiation of long-wavelength subsidence in the area and attribute it to onshore migration of dynamic subsidence associated with flattening of the Farallon slab beneath North America.

Beginning with the Blair Formation (ca. 83–81 Ma), deposits are observed to thicken away from the thrust belt, and there is no preserved evidence of a significant foredeep. Campanian strata (ca. 83–74 Ma) show a similar pattern, with the locus of deposition migrating eastwards. Campanian strata are truncated beneath the Moxa belt, and there is no preserved evidence of a significant forebulge. The presence of deposits with thicknesses >600 m in the distal basin, in a position east of the expected forebulge, indicates an additional mode of subsidence is present. We interpret this as the initiation of long-wavelength subsidence in the area and attribute it to onshore migration of dynamic subsidence associated with flattening of the Farallon slab beneath North America.

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CONFLICT OF INTEREST
The authors of this work have no conflict of interests to declare.

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Measured sections and detrital zircon geochronology data are provided in the Supporting Information.

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SUPPORTING INFORMATION

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