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Reconstructing the morphologies and hydrodynamics of ancient rivers from source to sink: Cretaceous Western Interior Basin, Utah, USA

Sinéad J. Lyster[1]*, Alexander C. Whittaker[1], Gary J. Hampson[1], Elizabeth A. Hajek[2], Peter A. Allison[1] and Bailey A. Lathrop[1]

[1] *Department of Earth Science and Engineering, Imperial College London, London, UK.*

[2] *Department of Geosciences, The Pennsylvania State University, Pennsylvania, USA.*

**s.lyster17@imperial.ac.uk*

(A) Abstract

Quantitative reconstruction of palaeohydrology from fluvial stratigraphy provides sophisticated insights into the response, and relative impact, of tectonic and climatic drivers on ancient fluvial landscapes. Here, field measurements and a suite of quantitative approaches are used to develop a four-dimensional (space and time) reconstruction of palaeohydrology in Late Cretaceous palaeorivers of central Utah, USA — these rivers drained the Sevier mountains to the Western Interior Seaway. Field data include grain-size and cross-set measurements and span 5 parallel fluvial systems, 2 of which include up-dip to down-dip transects, across 7 stratigraphic intervals through the Blackhawk Formation, Castlegate Sandstone and Price River Formation. Reconstructed palaeohydrological parameters include fluvial morphologies (flow depths, palaeoslopes, palaeorelief, and planform morphologies) and various hydrodynamic properties (flow velocities, water discharges, and sediment transport modes). Results suggest that fluvial morphologies were similar in space and time; median flow depths spanned 2–4 m with marginally greater flow depths in southerly systems. Meanwhile palaeoslopes spanned 10^{-3} to 10^{-4}, decreasing downstream by an order of magnitude. The most prominent spatio-temporal change is an up to 4-fold increase in palaeoslope at the Blackhawk–Castlegate transition; associated alluvial palaeorelief is 10s of metres during Blackhawk deposition and >100 m during Castlegate Sandstone deposition. Unit water discharges do not change at the Blackhawk–Castlegate transition, which argues against a climatically driven increase in palaeoslope and channel steepness. These findings instead point to a tectonically driven palaeoslope increase, although one limitation in this study is uncertainty in palaeochannel widths, which directly influences total water discharges. These reconstructions complement and expand on extensive previous work in this region, which enables us to test the efficacy of quantitative reconstruction tools. Comparison of results with facies-based interpretations indicates that quantitative tools work well, but inconsistencies in more complex reconstructions (e.g. planform morphologies) highlight the need for further work.

(A) Introduction

The stratigraphic record is a fundamental physical archive of Earth surface processes in space and time (Wobus et al., 2006; Allen, 2008a, 2008b; Armitage et al., 2011; Whittaker, 2012). A key research challenge is to decode this archive to reconstruct the movement of water and sediment across Earth's surface in the geological past (Castelltort & Van Den Driessche, 2003; Jerolmack & Paola, 2010; Ganti et al., 2014; Romans et al., 2016; Straub et al., 2020) — effective quantification of palaeohydrology from fluvial stratigraphy is crucial to achieve this goal. Constraints on the morphologies and hydrodynamics of palaeorivers can be used to: resolve the size and scale of ancient catchments (Bhattacharya & Tye, 2004; Bhattacharya et al., 2016; Eide et al., 2018; Lyster et al., 2020); quantify sediment transport capacities and the magnitudes of sediment exported to oceans (Allen et al., 2013; Holbrook & Wanas, 2014; Lin & Bhattacharya, 2017; Sharma et al., 2017); decipher fluvial response to perturbation (Foreman et al., 2012; Foreman, 2014; Colombera et al., 2017; Chen et al., 2018); and reconstruct local palaeogeographies (Li et al., 2018). Importantly, these constraints can be used to investigate hydrological response to long-period forcing (>10^6 yrs) as river behaviour is intrinsically linked to tectono-climatic boundary conditions over geological timescales (Duller et al., 2010; Whitchurch et al., 2011; Whittaker et al., 2011; Castelltort et al., 2012; Hampson et al., 2013).

However, palaeohydrology is limited by incomplete (or absent) records of palaeorivers (Sadler, 1981; Jerolmack & Sadler, 2007), uncertainty as to what information fluvial stratigraphy actually preserves (Castelltort & Van Den Driessche, 2003; Jerolmack & Paola, 2010; Romans et al., 2016; Straub et al., 2020), and uncertainties associated with data type, data measurement, and reconstruction tools (e.g. Bridge & Tye, 2000). Where it is possible to overcome these challenges, the ability to decipher palaeohydrological information with high fidelity can enable sophisticated insights to be drawn about the sensitivity and response of ancient fluvial systems to tectonic and climatic drivers.

Here, a quantitative framework is used to reconstruct the palaeohydrological evolution of well-known source-to-sink systems of Late Cretaceous central Utah, USA. The focus of this study is the Blackhawk Formation–Castlegate Sandstone–Price River Formation fluvial succession as outcrops are extensive and well-documented (Kauffman, 1977; Kauffman & Caldwell, 1993; Cobban et al., 2006). These strata represent eastward flowing palaeorivers that drained the Sevier orogenic fold-and-thrust belt to the Western Interior Seaway (WIS). Previous work has primarily focused on *qualitative* inferences of palaeohydrology in these systems (Miall, 1994; Miall & Arush, 2001; Adams & Bhattacharya, 2005; McLaurin & Steel, 2007; Hampson et al., 2012; Flood & Hampson, 2014), which are sometimes complimented by simple quantitative reconstructions (e.g. Hampson et al., 2013). Meanwhile, *quantitative* work has mostly focused on architectural-scale elements in these systems, including preservation of channelized bodies and bars and associated autogenic processes, such as avulsion and backwater dynamics (Hajek et al., 2010; Hajek & Wolinsky, 2012; Flood & Hampson, 2015; Trower et al., 2018; Chamberlin & Hajek, 2019; Ganti et al., 2019a). The palaeohydrological evolution of these rivers at the system scale has not been comprehensively addressed using quantitative tools — this study addresses this outstanding research challenge to shed new light on these ancient systems.

Palaeohydrological field data were collected for 5 parallel transverse fluvial systems (spaced ~20–25 km apart) across 7 stratigraphic intervals within the Campanian stage (83.6±0.2 to 72.1±0.2 Ma) of the Late Cretaceous, which spanned 11.5 Myr (Figs 1, 2). These data allow for high resolution spatio-temporal reconstructions of these systems, both up-dip to down-dip and along depositional strike (Fig. 1). Reconstructed palaeohydrologic parameters include: flow depths; palaeoslopes and palaeorelief (specific to the alluvial domain); hydrodynamic properties, including flow velocities, water discharges and sediment transport modes; and planform morphologies. First and foremost, results show how the morphologies and hydrodynamic properties of these palaeorivers varied in space and time. Moreover, reconstruction of palaeoslopes and palaeorelief in the alluvial domain enable evaluation of the competing roles of tectonic and climatic drivers on the evolution of these ancient rivers. Finally, the results provide new insights regarding the extent to which quantitative palaeohydrologic methods (which are increasingly borrowed from the field of engineering) can be reconciled with sedimentological observables.

(A) Research background

(B) Palaeohydrology

Palaeohydrological interpretations traditionally derive from analysis of facies associations in fluvial strata, particularly of architectural-scale elements (Miall, 1994; Miall & Arush, 2001; Adams & Bhattacharya, 2005; McLaurin & Steel, 2007; Hampson et al., 2012; Hampson et al., 2013; Flood & Hampson, 2014), and increasingly take advantage of high-resolution remote imagery and three-dimensional outcrop models (Hajek & Heller, 2012; Rittersbacher et al., 2014; Chamberlin & Hajek, 2019). However, a combination of empirical, theoretical and experimental work has led to the development of fluid and sediment transport models that are applicable to geologic questions (e.g. van Rijn, 1984b; Ferguson & Church, 2004; Parker, 2004; Wright & Parker, 2004; Mahon & McElroy, 2018), enabling more sophisticated inferences of palaeohydrology from the rock record.

Recent quantitative research has focused on maximising the ability to accurately reconstruct the evolution of fluvial landscapes in the geologic past. Some efforts have centred on connecting landscape surface kinematics to stratal preservation (Paola & Borgman, 1991; Castelltort & Van Den Driessche, 2003; Jerolmack & Mohrig, 2005; Jerolmack & Paola, 2010; Hajek & Wolinsky, 2012; Ganti et al., 2013; Ganti et al., 2014; Reesink et al., 2015; Romans et al., 2016; Ganti et al., 2020; Leary & Ganti, 2020; Straub et al., 2020) and a number of these studies have focused on Late Cretaceous fluvial strata in central Utah (Flood & Hampson, 2015; Trower et al., 2018; Chamberlin & Hajek, 2019; Ganti et al., 2019a). Meanwhile, other quantitative work has applied fluid and sediment transport models to stratigraphic field data, with an overarching goal of constraining the characteristics of catchments, regional systems or entire fluvial landscapes in the geological past (Ganti et al., 2019b; Lapôtre et al., 2019), or even on other planetary bodies (Lamb et al., 2012; Buhler et al., 2014; Hayden et al., 2019; Lapôtre et al., 2019). This includes using quantitative palaeohydrological tools to reconstruct water and sediment discharges within mass balance frameworks (Holbrook & Wanas, 2014; Lin & Bhattacharya, 2017; Sharma et al., 2017), decipher local palaeogeographies (Bhattacharyya et al., 2015; Li et al., 2018), characterise pre-vegetation rivers (Ganti et al., 2019b), and reconstruct fluvial response to climatic perturbations for well-preserved fluvial strata straddling events such as the Paleocene–Eocene Thermal Maximum (PETM) (Foreman et al., 2012; Foreman, 2014; Colombera et al., 2017; Chen et al., 2018; Duller et al., 2019).

Despite the breadth of quantitative palaeohydrological tools available, previous applications to fluvial stratigraphic field data have typically centred on individual catchments and instantaneous or short-period intervals (i.e. individual discharge events and mean annual discharges) (Holbrook & Wanas, 2014; Lin & Bhattacharya, 2017; Sharma et al., 2017), or reconstructions across stratigraphic boundaries and short-period tectono-climatic events, such as the PETM (Foreman et al., 2012; Foreman, 2014; Colombera et al., 2017; Chen et al., 2018; Duller et al., 2019). Far fewer studies have focused on long-period intervals, such as the evolution of source-to-sink systems across geologic timescales (>10^6 yrs). This outstanding opportunity can be exploited in Late Cretaceous fluvial systems of central Utah, where outcrop availability supports a four-dimensional (space and time) study in a region subject to active tectonics, spanning both Sevier and Laramide deformation.

(B) Tectono-geographic setting and palaeodrainage

Input of sediment to the Late Cretaceous WIS was dominated by the western margin, where rivers draining the active Sevier fold-and-thrust belt eroded and transported huge volumes of clastic sediments eastwards into the foreland basin (Spieker, 1946; Armstrong, 1968; Kauffman, 1977; Hay et al., 1993; Kauffman & Caldwell, 1993) (Fig. 1b,c). This led to the deposition and progradation of a large, asymmetric clastic wedge on the western WIS margin. This study focuses on Campanian non-marine clastic sediments of this wedge in central Utah, USA (Figs 1–3), where palaeodrainage is relatively well-constrained (Bartschi et al., 2018; Pettit et al., 2019). Multiple transverse fluvial systems drained the Sevier thrust belt in this area (Fig. 1b). Several studies have additionally interpreted an axial, or longitudinal, fluvial system that drained north–northeast from the Mogollon Highlands (present day central Arizona) and Cordilleran magmatic arc, which interacted with transverse systems of the Sevier thrust belt (Lawton et al., 2003; Jinnah et al., 2009; Szwarc et al., 2015) (Fig. 1b) and led to downsystem sediment mixing (Bartschi et al., 2018; Pettit et al., 2019). Detrital zircon (DZ) data (Bartschi et al. (2018) indicate that these fluvial systems were dominated by a thrust-belt source in close proximity to the Sevier thrust front, but that more southerly transverse systems may have additionally featured a longitudinal component of drainage (Bartschi et al., 2018; Pettit et al., 2019). Herein, focus is on transverse fluvial systems that predominantly drained the Sevier mountains (Fig. 1).

Tectonic forcing in this region is well studied (DeCelles, 1994, 2004; DeCelles & Coogan, 2006) and palaeoclimate has been reconstructed from a variety of palaeontological, geochemical-proxy and modelling studies (e.g. Wolfe & Upchurch Jr., 1987; Fricke et al., 2010; Miller et al., 2013; Sewall & Fricke, 2013; Foreman et al., 2015). In central Utah, eastward propagation of the Sevier thrust belt (due to eastward subduction of the Farallon plate) resulted in thin-skinned deformation and movement on the north–south trending Canyon (~145−110 Ma), Pahvant (~110−86 Ma), Paxton (86−75 Ma) and Gunnison (75−65 Ma) thrust systems (DeCelles, 1994, 2004; DeCelles & Coogan, 2006). Associated exhumation created substantial topographic relief in the Sevier mountains, which has been described as "Andean" in scale with mean elevations approaching near 4000 m (Sewall & Fricke, 2013; Foreman et al., 2015). Modelling results and stable isotope evidence suggest a strong monsoonal precipitation along the eastern flank of the Sevier mountains and seasonal flooding across low-relief regions (Roberts, 2007; Roberts et al., 2008; Fricke et al., 2010; Sewall & Fricke, 2013). The tectono-geographic set-up of the Western Interior was particularly conducive to a monsoonal climate — the proximity of a warm sea to high elevation mountains commonly results in strong seasonal precipitation and convective circulation (e.g. Zhisheng et al., 2001). A seasonal temperate-to-subtropical climate therefore prevailed throughout Campanian deposition (L. R. Parker, 1976; Kauffman & Caldwell, 1993; Roberts & Kirschbaum, 1995). The Campanian onset of thick-skinned deformation as the subducting Farallon plate transitioned to lower-angle, or flat-slab, subduction (DeCelles, 2004) began to manifest as basement-cored Laramide uplifts (e.g. San Rafael Swell, central Utah, and Uinta Mountains, northern Utah), which partitioned the Sevier foreland basin and disrupted patterns of both regional subsidence and drainage (Bartschi et al., 2018; Pettit et al., 2019).

(B) Stratigraphic framework

Establishing a consistent stratigraphic framework in space and time is crucial for system scale palaeohydrological reconstructions. Here, focus is on the Upper Cretaceous Mesaverde Group and up-dip equivalents (Figs 1, 2) in central Utah, USA, specifically fluvial sediments situated less than ~100 km from the Sevier orogenic front (DeCelles & Coogan, 2006) in the flexurally subsiding foredeep (Fig. 3). These sediments include the Blackhawk Formation, Castlegate Sandstone and Price River Formation along the eastern front of the Wasatch Plateau (Figs 1–3). Up-dip, on the western Wasatch Plateau, the Blackhawk–Castlegate–Price River succession is correlated with the Sixmile Canyon Formation (Indianola Group) and the Price River Conglomerate (following Robinson and Slingerland (1998); Horton et al. (2004); Aschoff and Steel (2011b, 2011a)) (Figs 1–3). Up-dip to down-dip, these sediments encompass the entire alluvial domain of these palaeorivers draining the Sevier highlands. A broad summary of field sites and the stratigraphic framework (Figs 1, 2) is given below — extended information regarding regional stratigraphy and correlations is provided in the Supplementary Material.

Down-dip field sites were grouped spatially into 5 field areas that represent 5 parallel transverse fluvial systems draining the Sevier thrust front: Price Canyon, Wattis Road, Straight Canyon (including Joe's Valley Reservoir), Link Canyon and Salina Canyon (Figs 1, 3). These 5 field areas are approximately ~50 km from up-dip alluvial fan lobes (Figs 1, 3). Assuming typical outlet spacings of rivers draining orogenic fronts (~25 km) (Hovius, 1996), it is likely that these field areas represent 5 distinct palaeorivers and form a ~125 km transect along depositional strike. For the 2 up-dip to down-dip transects (Fig. 1), the northern transect included 4 field areas: Dry Hollow, Lake Fork, Bear Canyon, and terminating at Price Canyon (Fig. 3c–e), and the southern transect included 3 field areas: Mellor Canyon, Sixmile Canyon, and terminating at Straight Canyon (Fig. 3d–f). These transects follow those widely implemented in previous work, both along-strike (Hampson et al., 2012; Hampson et al., 2013; Flood & Hampson, 2014, 2015; Chamberlin & Hajek, 2019) and up-dip to down-dip (Robinson & Slingerland, 1998; Horton et al., 2004; Aschoff & Steel, 2011b, 2011a).

In addition to grouping field sites in space, they were also grouped in time. In this study 7 stratigraphic intervals were defined: 1 = lower Blackhawk Formation; 2 = middle Blackhawk Formation; 3 = upper Blackhawk Formation; 4 = lower Castlegate Sandstone; 5 = middle Castlegate Sandstone; 6 = upper Castlegate Sandstone (Bluecastle Tongue); 7 = (lowermost) Price River Formation (Fig. 2).

Down-dip, on the eastern front of the Wasatch Plateau, it is straightforward to assign sediments of the Blackhawk–Castlegate–Price River succession to the appropriate "space–time" interval by facies associations, following extensive work that has been undertaken in this region (Lawton, 1983, 1986b; Miall, 1994; van Wagoner, 1995; Yoshida et al., 1996; Miall & Arush, 2001; Lawton et al., 2003; Adams & Bhattacharya, 2005; Hampson et al., 2012; Hampson et al., 2013; Flood & Hampson, 2014; Hampson et al., 2014; Flood & Hampson, 2015). The lower–middle Campanian Blackhawk Formation represents deposition on coastal plains behind wave-dominated deltaic shorelines which, up-section, pass landward into alluvial and fluvial plains (Hampson, 2010; Hampson et al., 2012; Hampson et al., 2013). The size and abundance of channelized fluvial sand bodies (deposited by both single- and multi-thread rivers) increase from base to top of the Blackhawk Formation (Adams & Bhattacharya, 2005; Hampson et al., 2012; Hampson et al., 2013; Flood & Hampson, 2015). The middle–upper Campanian Castlegate Sandstone is situated atop the Blackhawk Formation and is an extensive, cliff-forming river-dominated deposit. The lower Castlegate Sandstone and upper Castlegate Sandstone (Bluecastle Tongue) comprise amalgamated braided fluvial channel-belt deposits, whereas the middle Castlegate Sandstone comprises less amalgamated, more meandering, fluvial channel-belt deposits with interbedded mudstones (Fouch et al., 1983; Lawton, 1986b; Miall, 1994; van Wagoner, 1995; Yoshida et al., 1996; Miall & Arush, 2001). The ledge-forming upper Campanian Price River Formation sits conformably atop the Castlegate Sandstone and comprises large channelized sand bodies with interbedded siltstones and mudstones — channelized sand bodies form ~75% of the formation (Lawton, 1983, 1986b). Fluvial sediments of the Price River Formation represent the end of Sevier thrusting; the late Maastrichtian–Eocene North Horn Formation unconformably overlies the Price River Formation.

Up-dip, on the western Wasatch Plateau, correlative strata include more proximal sediments of the Indianola Group and Price River Formation, which is now known to not be time-equivalent with the down-dip Price River Formation exposed near Price, Utah (Robinson & Slingerland, 1998; Horton et al., 2004; Aschoff & Steel, 2011b, 2011a). To avoid confusion, these up-dip strata are here referred to as the Price River Conglomerate, following Aschoff and Steel (2011b, 2011a). Up-dip to down-dip correlations are limited by incomplete exposure on the western Wasatch Plateau and difficulty in dating conglomerates (see Supplement). Nevertheless, Robinson and Slingerland (1998) used palynology to correlate these strata across a variety of localities on the Wasatch Plateau (Fig. 2), which can be traced in seismic reflection data (Horton et al., 2004). The up-dip Price River Conglomerate is time-correlative with the down-dip lower, middle, and upper Castlegate Sandstone, and Price River Formation (Robinson & Slingerland, 1998; Horton et al., 2004; Aschoff & Steel, 2011b, 2011a), and is characterised by quartzite-dominated synorogenic fanglomerates and few gravel–sand fluvial bodies (Robinson & Slingerland, 1998; Aschoff & Steel, 2011b, 2011a). Of the Indianola Group, the upper Sixmile Canyon Formation is time-correlative with the Blackhawk Formation (Lawton, 1982; Fouch et al., 1983; Lawton, 1986b) and is predominantly characterised by synorogenic gravel–sand fluvial facies, spanning polymictic fluvial conglomerates to medium–coarse-grained sandstones (Lawton, 1982, 1986a, 1986b). Here a conservative approach is taken to up-dip to down-dip correlations; the upper Sixmile Canyon Formation of the Indianola Group (intervals 1–3) is time-averaged, and the Price River Conglomerate (intervals 4–7) is also time-averaged, but exceptions were made where field sites were known to be situated at either the top of the upper Sixmile Canyon Formation or at the top/base of the Price River Conglomerate. A full description of these correlations, including new logging in Mellor Canyon, is presented in the Supplement.

Each depositional-dip transect is pinned at the most downstream location, i.e. it is assumed that the most down-dip sites in each transect (Price Canyon and Straight Canyon) are approximately parallel and at the same downstream distance. Transects then work upstream, such that the most up-dip field site (Dry Hollow; northern transect) is at a downstream distance of 0 km. Downstream distances follow Robinson and Slingerland (1998) —post-depositional extension is not corrected for. Alternatively, when reconstructing along-depositional-strike transects, transects are pinned at the most northern location (Price Canyon) with an along-strike distance of 0 km, meanwhile southern locations have along-depositional-strike distances up to 125 km.

(A) Methods

Data were collected from channel-fill stratigraphy (cross-stratified sandstone and gravel deposits are interpreted as channel floor deposits) and were time-averaged across each stratigraphic space–time interval (field sites are listed in Supplementary Table S2). These field data, including uncertainties, were propagated through a quantitative framework to reconstruct the morphologies and hydrodynamics (flow depths, palaeoslopes, river long profiles, flow velocities and discharges, sediment transport modes and likely planform morphologies) of palaeorivers in both space and time.

(B) Field observations

(C) Grain size

At each field site the coarse-fraction (>2 mm in diameter) and sand-fraction (<2 mm in diameter) grain-sizes of channel-fill deposits were established (Fig. 4a,b). For coarse-fractions, grain-size distributions were measured via Wolman point counts (Wolman, 1954) (Fig. 4a); this technique has been successfully used to decode spatio-temporal trends in grain-size (e.g. Whittaker et al., 2011; D'Arcy et al., 2017; Brooke et al., 2018). For sand-fractions, scaled photographs were processed in *ImageJ* software and, similarly, the long axis of a minimum of 50 randomly selected grains was measured to recover grain-size distributions (Fig. 4b). From each measured grain-size distribution, the median grain-size, D_{50}, and 84th percentile, D_{84}, were extracted. Where grain-size facies were disparate, e.g. gravel topped with sand, data were collected for each grain-size facies and the proportions of each were estimated (Fig. 4c).

In order to achieve representative sampling for spatio-temporal grain-size trends, multiple grain-size observations were collected at each field site. Not only were data collected for each grain-size facies (Fig. 4a–c), but depending on overall outcrop extent Wolman point counts were repeated and/or additional scaled photographs were taken for *ImageJ* processing at intermittent stratigraphic intervals (e.g. one count per 5–10 m of strata or per channelized body). The extent of each field site can be approximated as the extent of outcrop apparent in Fig. 3c–h. From these data an average grain-size was produced for both the sand-fraction and gravel-fraction at each field site. As each space–time interval includes multiple field sites, this results in multiple average sand- and gravel-fraction grain-sizes, capturing channel-fill deposits from several channelized bodies. Finally, a bulk-grain-size was produced for each space–time interval using the gravel-to-sand proportions at each field site — each site within a space–time interval was assigned equal weighting. Further information regarding grain-size data collection, including axis selection, sample size sufficiency and weighting, is presented in the Supplement.

(C) Cross-sets

Cross-set heights were measured as these data can be used to reconstruct original bedform heights and formative flow depths. Trough- and planar-cross bedding, which are inherently indicative of bedload transport, were present at nearly all field sites. They occurred predominantly in sand-grade deposits, but also in granule- to pebble-grade deposits (Fig. 4d–f). To establish mean cross-set heights, the sampling strategy of Ganti et al. (2019b) was followed. Cross-set boundaries (i.e. the lower, asymptotic bounding surface and the upper, erosional bounding surface) were delineated and then heights were measured at regular intervals along the entire width of the cross-set dip-section (Fig. 4g–i). Measurements were made to a precision of ±5 mm. This protocol was repeated for individual cross-sets within co-sets to establish a mean cross-set height for each individual cross-set. Subsequently, maximum cross-set heights (i.e. the maximum distance between lower and upper bounding surfaces) were measured for a representative sample across the exposed outcrop (usually n=25–50).

Cross-set distributions (n=470) were used to establish the mean, 84th percentile (P_{84}) and maximum height for each individual cross-set, and relationships between each were established for the field area. These new relationships were then used to estimate mean cross-set heights from all measured maximum cross-set heights (n=4053), and these estimates of mean cross-set heights were propagated through subsequent calculations.

(C) Channel geometry and architectural element data

Above grain- and bedform-scales, channel geometries and major architectural elements were also measured, where possible, using a Haglof Laser Geo laser range finder to a precision of ±5 cm. This included maximum channel body/story thicknesses and bar-scale clinoform heights. Previous work in this region has documented the dimensions and distributions of fluvial architectural elements using high-resolution imagery and 3D outcrop models (Hajek & Heller, 2012; Rittersbacher et al., 2014; Flood & Hampson, 2015; Chamberlin & Hajek, 2019). Field data collection therefore focused on grain-size and cross-set measurements, with compilation of published secondary data (alongside new data from this study) to augment field data and evaluate our palaeohydrological reconstructions (see Supplementary Tables S4, S5).

(B) Quantitative palaeohydrology

(C) Channel geometries

To calculate original bedform heights from cross-set measurements, the relation of Leclair and Bridge (2001) was used, which is based on theoretical work by Paola and Borgman (1991). Leclair and Bridge (2001) showed that mean bedform (i.e. dune) height, h_d, can be approximated as a function of mean cross-set height, h_{xs}, where

$$h_d = 2.9(\pm 0.7)\ h_{xs}.$$

Eq. 1

While bedform height generally scales with flow depth, the mechanistic explanation for this is not fully resolved. As such, many scaling relations simply relate bedform height and flow depth (e.g. Yalin, 1964), whereas some incorporate additional parameters such as Froude number, D_{50}, and transport stage (e.g. Gill, 1971; van Rijn, 1984a), however their incorporation does not improve predictive power. Bradley and Venditti (2017) revisited previous bedform height−flow-depth scaling relations and derived a new relation between h_d and median formative flow depth, H, based on >380 field observations:

$$H = 6.7h_d,$$

Eq. 2

with the 1st and 3rd quartiles estimated by H=4.4h_d and H=10.1h_d, respectively. Bradley and Venditti (2017) proposed that their relations for the 1st and 3rd quartiles of H offer useful probability bounds on palaeoflow depths. As such, the 1st and 3rd quartiles of H (carrying forward the error on Equation 1) were also calculated, and these values were carried throughout subsequent calculations to offer reasonable bounds for the likely spread of values for each parameter. Where cross-bedding was absent (i.e. the most up-dip field sites), channel-body thicknesses were used as a proxy for flow depth.

Similar to H, channel width, W, can be estimated using scaling relations as direct measurement is not normally possible from outcrop. Bridge and Mackey (1993) proposed the relation W=8.8$H^{1.82}$ for single-thread channels. Alternatively, widths of fully-braided channel systems can be approximated as, for example, W=42$H^{1.11}$ (Leopold & Maddock Jr, 1953). However, estimates of W from outcrop data and scaling relations are particularly tentative and, where systems are braided, subject to further uncertainty pertaining to the number of threads. As such, results in this study are reported per unit width.

(C) Palaeoslopes and palaeorelief

Palaeoslopes were estimated using 2 independent methodologies, adapted from Ganti et al. (2019a). First, Shields stress, τ^*, was estimated using the bedform stability diagram of Carling (1999), which expresses bedform stability in terms of τ^* and D_{50} (for D_{50} < 33 mm). Minimum and maximum bounds of τ^* for the stable existence of dunes were then identified for a range of D_{50} values. Where D_{50} exceeded 33 mm, and in the absence of bedforms, a range of possible τ^* values of 0.04–0.06 were assigned. Then, 10^6 uniformly distributed random samples of τ^* were generated between these bounds, as well as 10^6 uniformly distributed random samples of H (between the 1st and 3rd quartile). To reconstruct palaeoslope, S, bed shear stress, τ_b, was approximated as the depth–slope product ($\tau_b=\rho gHS$) and then S can be given as

$$S = \frac{RD_{50}\tau^*}{H} \, , \tag{Eq. 3}$$

where R is the dimensionless submerged specific gravity of sediment in water (1.65 for quartz) and H is the flow depth (ρ is density and g is acceleration due to gravity). Similarly, 10^6 values of S were recovered and the median S, as well as the 1st and 3rd quartile of S, were extracted.

For the second approach, the method of Trampush et al. (2014) was used, which is based on Bayesian regression analysis of bankfull measurements in modern alluvial rivers (n=541); here slope is expressed as

$$\log S = \alpha_0 + \alpha_1 \log D_{50} + \alpha_2 \log H \, , \tag{Eq. 4}$$

where the constants are given by $\alpha_0 = -2.08 \pm 0.036$, $\alpha_1 = 0.254 \pm 0.016$, and $\alpha_3 = -1.09 \pm 0.044$. Using 10^6 values of H, and 10^6 values of α_0, α_1, and α_3 (uniformly distributed random samples between the bounds of the standard errors), 10^6 values of S were similarly recovered, and the 1st, 2nd, and 3rd quartiles were extracted. Using Equation 3, estimates of S derived from Equation 4 can be corroborated.

Along up-dip to down-dip transects, palaeoslope estimates can be usedto infer the shape of the river long profile, and therefore palaeorelief, in the alluvial domain. Palaeorelief was reconstructed using median estimates of S from Equations 3 and 4. The local slope at downstream position x, S_x, can be related to its upstream contributing catchment area, A_x, (Hack, 1973; Flint, 1974; Whipple, 2004) as

$$S_x = k_s A_x{}^{-\theta} ,$$

Eq. 5

where k_s is the steepness index and θ is the concavity, typically between 0.4 and 0.7 (Tucker & Whipple, 2002). Given that the palaeo-concavity is unknown, a range of plausible concavities (0.4, 0.5, and 0.6) were tested to gauge the spread of possible results. Following Hack's law, local catchment length, L_x, is related to A_x by $L_x=c_H A_x{}^h$, where c_H is the Hack coefficient, commonly taken as near 2 when L_x and A_x are in units of km^2 (Castelltort et al., 2009), and h is the Hack exponent, commonly taken as 0.5 (Hack, 1957). Using Hack's law, local slope can instead be estimated as a function of downstream distance, where

$$S_x = k_s L_x{}^{-\theta/h} .$$

Eq. 6

k_s is calculated from field data using downsystem palaeoslope estimates and knowledge of catchment lengths at each downstream location. As this study solely focuses on the alluvial domain, this means that up-dip fan apexes would have a catchment length of 0 km. Here, the most up-dip field sites are set as having a catchment length of 5 km to allow for additional up-dip fan length. Knowledge of distance to the coeval palaeoshoreline from our most down-dip sites (Price Canyon and Straight Canyon) is also required. Based on previous studies, approximate distances to the palaeoshoreline are set as ~10 km for the lower Blackhawk Formation, ~35 km for the middle Blackhawk Formation, ~50 km for the upper Blackhawk Formation, ~110 km for the Castlegate Sandstone (Hampson et al., 2012; Hampson et al., 2013), and ~200 km for the Price River Formation (Hettinger & Kirschbaum, 2002; Aschoff & Steel, 2011a). A nonlinear least squares regression was used to find best fit palaeoslope profiles (Equation 6) for both the northern and southern transects at each time interval. Palaeoslope profiles were then transformed into river long profiles by summing elevation increments along the downstream length to the palaeoshoreline. This elevation decrease is indicative of the likely relief in the alluvial domain of these palaeorivers.

(C) Hydrodynamics

Flow velocities, U, were calculated following Manning's Equation, where

$$U = \frac{1}{n} H^{\frac{2}{3}} S^{\frac{1}{2}} ,$$

Eq. 7

and n is Manning's constant, set as 0.03. In reconstructing hydrodynamics, palaeoslope estimates derived from the Shields stress inversion (Equation 3) were carried forward. Water discharges were then estimated by multiplying flow velocity by flow depth, to obtain discharge per unit width (Q=UH).

To determine dominant mode of sediment transport, the Rouse number, Z, was calculated as

$$Z = \frac{w_s}{\beta \kappa u_*}$$

Eq.8

where β is a constant that correlates eddy viscosity to eddy diffusivity, typically taken as 1, and κ is the von Karman constant, taken as 0.4. Sediment settling velocity, w_s, was calculated as a function of grain size following Ferguson and Church (2004),

$$w_s = \frac{RgD_{50}{}^2}{C_1 v + (0.75 C_2 R g D_{50}{}^3)^{0.5}},$$

Eq. 9

where v is the kinematic viscosity of water (1×10^{-6} m^2/s for water at 20°C) and C_1=18 and C_2=1 are constants associated with grain sphericity and roundness. With Z, dominant mode of sediment transport is typically wash load for $Z < 0.8$, 100% suspended load for $0.8 < Z < 1.2$, 50% suspended load (i.e. mixed load) for $1.2 < Z < 2.5$, and bedload for $Z > 2.5$. To corroborate inferred sediment transport modes, the particle Reynolds number, Re_p, was additionally calculated in line with previous work (cf. Parker, 2004) as

$$Re_p = \frac{\sqrt{RgD_{50}}D_{50}}{v}$$

Eq. 10

and plotted Re_p as a function of τ^*, following Dade and Friend (1998). This enables field results to be contrasted with data that are typical of either suspended, mixed, or bedload sediments (Leopold & Wolman, 1957; Schumm, 1968; Chitale, 1970; Church & Rood, 1983; Andrews, 1984), and to identify where these data are positioned among characteristic flow regimes (no sediment transport; ripples and dunes; upper plane beds) following Allen (1982a, 1982b).

(C) Fluvial style

Fluvial style (i.e. planform morphology) of Blackhawk–Castlegate rivers has been described qualitatively from outcrop architecture (Miall, 1994; Miall & Arush, 2001; Adams & Bhattacharya, 2005; Hampson et al., 2013). Here, a quantitative approach is implemented to decipher fluvial style to complement these works, check for consistency, and interpret the interplay between different planform morphologies and the tectono-geographic setting. This is carried out for field areas along the eastern Wasatch Plateau. First, Froude number, Fr, is calculated as

$$Fr = \frac{U}{\sqrt{gH}}$$

Eq. 11

and, then, depth/width ratios were plotted against palaeoslope/Froude ratios (G. Parker, 1976). Various flow widths were assigned to determine what depth/width ratios are required such that the data fall within the theoretical stability fields for single-thread and multi-thread fluvial planform morphologies. These flow widths are then contrasted with estimates of apparent maximum flow width from architectural analysis of channelized sandstone bodies (e.g. Flood & Hampson, 2015) and field interpretations of fluvial style (Miall, 1994; Miall & Arush, 2001; Adams & Bhattacharya, 2005; Hampson et al., 2013).

For all palaeohydrological parameters the median result is presented. In instances where results additionally include the 1st and 3rd quartiles, these are the results when the 1st and 3rd quartiles of palaeoflow depth (and therefore palaeoslope, Shields stress, etc.) were propagated through the methodology.

(A) Results

(B) Channel geometries

Linear relationships between maximum cross-set height and both the mean and the P_{84} cross-set height were established from measured cross-set distributions (n=470) for our field area (Fig. 5a,b). Maximum and mean cross-set heights are very well-correlated (R^2=0.88) and 95% of observed mean cross-set heights fall within ~3 cm of the predicted mean cross-set height. Using these new relationships, mean cross-set heights were estimated for all (n=4053) measured maximum cross-set heights (Fig. 5c–e; Supplementary Table S3).

Maximum cross-set heights typically span 0.1–0.35 m — these field data are comparable to the results of previous work (e.g. Adams & Bhattacharya, 2005). From maximum cross-set heights, mean cross-set heights spanning 0.07–0.25 m are estimated, which correspond with original bedform heights of 0.2–0.75 m. Flow depths for the along-depositional-strike transect suggest that, in both space and time, these 5 transverse fluvial systems maintained median flow depths of 2–4 m, with 1st–3rd interquartile ranges spanning 1–7 m (Fig. 6). Overall, flow depths do not change across the Blackhawk–Castlegate transition but exhibit a marginal decrease during middle Castlegate Sandstone deposition of <0.5 m. Flow depths are also projected to be overall <1 m greater in southern fluvial systems (Fig. 6). However, these observed differences all lie within the interquartile range of calculations, suggesting these systems were similar to each other.

Reconstructed palaeoflow depths are consistent with independent palaeoflow depth proxies (Supplementary Table S4), which demonstrates applicability of cross-set scaling relations in the absence of well-preserved macroforms. Bar heights, where available, are consistent with projected flow depths of 2–4 m across field sites. For instance, Chamberlin and Hajek (2019) reported mean bar heights of 2.6 m, 3.6 m and 3.9 m for the entire Castlegate Sandstone at Price Canyon, Straight Canyon and Salina Canyon, respectively. At Price Canyon, both Lynds and Hajek (2006) and Hajek and Heller (2012) reported greater mean bar heights of 4.1 m specifically for the lower Castlegate Sandstone, with a typical span of 1–8 m (Lynds & Hajek, 2006; McLaurin & Steel, 2007) — we note that the 1st–3rd interquartile range of our reconstructed palaeoflow depths is typically 1–7 m and therefore agrees with this range. Meanwhile, channelized fluvial sandstone bodies are more extensively documented for the Blackhawk Formation and their heights offer a maximum limit on palaeoflow depths. Flood and Hampson (2015) recovered mean apparent heights for channelized sandtone bodies of 6–8 m across the entire Blackhawk Formation between Straight Canyon and Salina Canyon. As maximum bounds on palaeoflow depth, these values are also in good agreement with the upper bounds (3rd quartile) of estimated palaeoflow depths.

(B) Palaeoslopes and river long profiles

Palaeoslope estimates for our northern (Fig. 7a–f) and southern (Fig. 7g–m) transects and results from each method (Equations 3 and 4) were compared (Fig. 7). Palaeoslopes are presented as y/x — a palaeoslope of 0.001 results in an elevation decrease of 1 m per 1000 m and is equivalent to 0.057°. Maximum (up-dip) palaeoslopes of 5×10^{-3} are equivalent to slopes of ~0.3°; these magnitudes of palaeoslope are comparable with the slopes of modern rivers, including the Savannah and North Loup (USA) (Carlston, 1969; Crowley, 1983; Mohrig & Smith, 1996; Fotherby, 2009). Minimum (down-dip) palaeoslopes of ~5×10^{-5} are equivalent to slopes of ~0.003°; palaeoslopes in the range 10^{-5} to 10^{-4} are characteristic of lowland/low-slope rivers, such as the Niobrara, Platte and Mississippi (USA) (Carlston, 1969).

Up-dip, palaeoslopes are consistently of order 10^{-3} (Fig. 7), with the exception of the Blackhawk Formation in the southern transect where 1st–3rd interquartile ranges extend down to palaeoslopes of 7×10^{-4} (Fig. 7k–m). Importantly, an order of magnitude decrease in palaeoslope is reconstructed between a down-system distance of 10 and 25 km; this occurs in all stratigraphic intervals, at the same downstream distance, for both the northern and southern transects (Fig. 7). Down-dip, from ~25 km onwards, palaeoslopes are flatter and typically span 5×10^{-5} to 5×10^{-4}. In these lower gradient regions, there is an apparent down-dip increase in palaeoslope in Fig. 7b,c,i–m. However, this apparent increase is within the 1st–3rd interquartile range. Up-dip to down-dip palaeoslope estimates derived from Equations 3 and 4 are broadly consistent with one another — they are the same order of magnitude and the 1st–3rd interquartile ranges either overlap with, or are within a factor of 2–3 of, one another. However, Equation 3 overpredicts and underpredicts palaeoslope relative to Equation 4, such that palaeoslope estimates derived from Equation 3 imply higher topographic relief and estimates derived from Equation 4 imply lower topographic relief (Fig. 7).

To constrain temporal changes in palaeoslope, the evolution of the the most up-dip locations of both the northern and southern transects can be compared (Fig. 8). Palaeoslopes increase at the onset of Castlegate Sandstone deposition (intervals 4–6) and the magnitude of this increase differs between the north and the south (Fig. 8). In the north, the initial palaeoslope is higher (~2×10^{-3}) and increases by a factor of 1.5 to ~3×10^{-3} (Fig. 8a), whereas, in the south, the initial palaeoslope is lower (~1×10^{-3}) and increases by a factor of up to 4, to ~4×10^{-3} (Fig. 8b). This implies a coeval increase in palaeoslope at the onset of Castlegate Sandstone deposition which was more pronounced in the south. Again, estimates derived from Equation 4 dampen this increase relative to estimates derived from Equation 3.

With up-dip to down-dip palaeoslope estimates for both the northern and southern transects, best-fit palaeoslope profiles were derived as a function of downstream distance (Equation 7; Supplementary Table S6). Palaeoslope profiles generally fit reconstructed palaeoslopes well, with typical R^2 values >0.85, and it is noted that of 3 reference concavities, θ, used, the higher value of θ=0.6 typically recovered the best fits (Supplementary Table S6). A notable exception to this is palaeoslope profiles reconstructed from Shields stress palaeoslope estimates for the Castlegate Sandstone in the northern depositional-dip transect — the lower θ=0.4 value generates the best fit and this fit is relatively poor (R^2 of 0.35–0.6). However, palaeoslope profiles for these same space–time intervals derived from alternative palaeoslope estimates (Equation 4) fit well (R^2 >0.9; Supplementary Table S6).

In reconstructing palaeoslope profiles steepness index, k_s, values were recovered for each stratigraphic interval (for θ=0.5), which were mostly between ~5 and 35 m (Supplementary Table S6). There is an increase in reconstructed k_s values across the Blackhawk–Castlegate transition for both methods of palaeoslope estimation. For estimates derived from Equation 3, k_s values increase across the Blackhawk–Castlegate transition by a factor of ~2–3 in the northern transect, and by a factor of ~4–5 in the southern transect. In contrast, for estimates derived Equation 4, k_s values increase across the Blackhawk–Castlegate transition by a factor of <1.5 in the northern transect, and by a factor of ~2 in the southern transect (Supplementary Table S6).

Palaeoslope profiles were transformed into river long profiles, which are indicative of the palaeorelief in the alluvial domain, or depositional reaches, of Blackhawk–Castlegate–Price River fluvial systems only (Fig. 9). Given that the concavities of these ancient rivers are not known, implementing plausible concavities of 0.4, 0.5 and 0.6 enabled a likely spread of values for palaeorelief to be constrained (Fig. 9). Results indicate that different concavities recover similar values for palaeorelief; total estimates vary within a factor of ~2, between a concavity of 0.4 and 0.6 (Fig. 9).

Using palaeoslope estimates derived from Equation 3, palaeorelief during Blackhawk deposition was estimated as ~40–60 m in the northern transect (Fig. 9e,f) and 15–25 m in the southern transect (Fig. 9k–m). During Castlegate Sandstone deposition, palaeorelief increased by a factor of 1.5–2.5 in the northern transect, to an estimated 65–145 m of palaeorelief, whereas it increased by a factor of 5–6 in the southern transect, to an estimated 90–130 m of palaeorelief. Alternatively, using palaeoslope estimates derived from Equation 4, palaeorelief during Blackhawk Formation deposition was estimated as ~30–50 m in the northern transect (Fig. 9e,f) and 15–25 m in the southern transect (Fig. 9k–m). During Castlegate Sandstone deposition, palaeorelief increased by a factor of ~1.8 in the northern transect, to an estimated 55–90 m of palaeorelief, whereas it increases by a factor of 2 in the southern transect, to an estimated 30–50 m of palaeorelief. In detail, palaeorelief implied by Equation 3 (Shields) is up to a factor of 2 greater than the palaeorelief implied by Equation 4 (Trampush). This higher palaeorelief during Castlegate Sandstone deposition is sustained into Price River Formation times. It is stressed that these estimates refer to the alluvial domain only.

(B) Hydrodynamics and sediment transport

Median flow velocities of 0.8 m/s, with an interquartile range of 0.4–1.6 m/s are deduced across all field data (Fig. 10a), as well as median unit discharges of 2.5 m^2/s with an interquartile range of 1–10 m^2/s (Fig. 10b). Using plausible single-thread channel widths of 100–500 m at down-dip locations (see Planform morphologies), this would imply median total discharges between 250–1250 m^3/s, which is comparable with total discharges of well-known North American rivers such as the Platte, Hudson, Colorado, Arkansas and Susquehanna. However, if multi-thread rivers are assumed to possess >1 branch/braid, total discharges would have been several times greater. With a reconstructed increase in palaeoslope at the Blackhawk–Castlegate transition, a coeval increase in flow velocities and unit water discharges is expected analytically. Here, across all up-dip field areas, flow velocities are overall greater during Castlegate Sandstone deposition, up to a factor of 2 to 3 (Fig. 10c), relative to Blackhawk Formation deposition, whereas down-dip flow velocities are broadly the same through time (Fig. 10d). Both up-dip and down-dip, unit water discharges overall do not change at the Blackhawk–Castlegate transition (Fig. 10e,f). To offer a specific example for the Blackhawk–Castlegate transition (intervals 3 and 4), at Mellor Canyon, median flow velocity, U, increased from 1.9 to 3.0 m/s, and median unit water discharge, Q, only increased marginally from 4.4 to 4.6 m^2/s.

Reconstructed Rouse numbers, Z, indicate that dominant transport modes of bed-material varied in space and time (Fig. 11). Up-dip field sites consistently exhibit high Z values for both the median and 1^{st}–3^{rd} interquartile range, indicating predominant bedload transport (Fig. 11). Median Z values then decrease by a downstream distance of 30 km, indicating local transition to predominantly mixed load systems, however the likely spread of values indicated by the interquartile range implies that dominant transport modes at this downstream distance may have spanned both mixed load and a near entirely suspended load (Fig. 11). A crucial exception to this observation is for Castlegate Sandstone deposition in the southern transect (intervals 4–6) where, at a downstream distance of 30 km, median Z values suggest bedload remains the most important transport mode (Fig. 11g–i). At downstream distances associated with the most down-dip field sites, median Z values have further decreased, however 1^{st}–3^{rd} interquartile ranges mostly still span both the mixed load and entirely suspended load domains.

The inferred dominant sediment transport modes are corroborated with results in Fig. 12, in which Shields stress, τ^*, is plotted as a function of particle Reynolds number, Re_p, for each field site. These data are plotted alongside observed data that are characteristic of suspended load, mixed load and bedload regimes (Leopold & Wolman, 1957; Schumm, 1968; Chitale, 1970; Church & Rood, 1983; Andrews, 1984). Up-dip field sites (Dry Canyon, Lake Fork, Mellor Canyon) plot among secondary data that are typical for bedload rivers, meanwhile all other field sites plot in the mixed-load realm (Fig. 12). Of field sites dominated by a mixed load, data from Sixmile Canyon and Straight Canyon plot closest to the bedload realm, which is consistent with observations in Fig. 11, whereresults suggest that bedload transport remained important in the southern transect during Castlegate Sandstone deposition (intervals 4–6). Overall, results in Fig. 12 suggest that, down-dip, field sites are firmly in the mixed load range — it is unlikely that bed-material loads were predominantly suspended. In contrast, the 1^{st}–3^{rd} interquartile ranges in Fig. 11 suggest that dominant sediment transport modes may have spanned the mixed load/predominantly suspended domain. Down-dip, all field sites straddle the bounds between the stability fields for ripples and dunes and upper-stage plane beds (Fig. 12), which implies unidirectional flow and high sediment transport rates (both suspended transport and bedload transport).

(B) Planform morphologies

Finally, these data provide insights into the implied planform morphology of these ancient fluvial systems. However, to do this effectively estimates of palaeochannel widths are needed. Widths are difficult to constrain with confidence from field observations, and estimates from empirical scaling relations are tentative. Assuming single-thread channels, reconstructed median flow depths of 2–4 m might suggest channel widths of order 30–110 m and, using the upper bound of the 1–7 m interquartile range, widths up to ~300 m (following Bridge and Mackey (1993)). In contrast, if multi-thread channel belts are assumed, then channel belt widths of order 90–200 m, and up to ~400 m, might be expected (following Leopold and Maddock Jr (1953)).

For a range of possible widths, palaeoslope/Froude ratios were plotted against channel depth/width ratios (cf. G. Parker, 1976; Ganti et al., 2019b) (Fig. 13). Results imply that, for Blackhawk–Castlegate–

Price River fluvial systems, single-thread planforms would be stable at channel widths <1 km; channel and channel-belt widths >1 km would have been required to instigate formation of bars and support transition to multi-thread systems, forming vast channel-belt complexes (Fig. 13a–d). However, planform reconstructions are very dependent on grain-size, a factor which is often not evaluated systematically. Bulk grain-sizes were used in initial calculations (Fig. 13a–d; see Methods). However, when using gravel-fraction grain-sizes, which can be associated with tectonic or climatic perturbations (e.g. increased palaeoslope or high-magnitude low-frequency discharge events), the results show that multi-thread planforms were more likely (Fig. 13e–h). For gravel-fraction grain-sizes, results imply that single-thread planforms were likely stable at channel widths <500 m, and that channel and channel-belt widths >500 m would have supported transition to multi-thread systems (Fig. 13b).

Further, of Blackhawk–Castlegate–Price River fluvial systems, field results for the Castlegate Sandstone plot closest to the single-thread–multi-thread transition, whereas field results for the Price River Formation plot furthest from this transition (Fig. 13). This indicates higher propensity of Castlegate fluvial systems to braiding, relative to Blackhawk and Price River systems.

(A) Discussion

(B) What did Campanian palaeorivers look like?

These analyses provide new insights that build on previous work characterising ancient rivers in the Campanian of central Utah as a series of distinct parallel transverse systems draining the Sevier front (Robinson & Slingerland, 1998; Bartschi et al., 2018; Chamberlin & Hajek, 2019; Pettit et al., 2019). These rivers traversed a low-gradient landscape; alluvial relief was 10s of metres to c. 100 m, and the length scale of the alluvial domain (i.e. the distance from fan apexes to the palaeoshoreline) varied from as little as ~70 km during lower Blackhawk Formation deposition, up to and in excess of 250 km during Price River Formation deposition (Hettinger & Kirschbaum, 2002; Aschoff & Steel, 2011a; Hampson et al., 2012; Hampson et al., 2013). Relief was 10s of metres during Blackhawk deposition, when the length scale of the alluvial domain was at its narrowest. At the onset of Castlegate Sandstone deposition an increase in palaeoslope is documented, with palaeorelief increasing to c. 100 metres, which persisted into Price River deposition (Figs 7–9). For comparative purposes, such values of palaeoslope and palaeorelief are characteristic of the Mississippi river and downstream reaches of its principal tributaries e.g. the Missouri, Tennessee, Arkansas and Red rivers (Carlston, 1969).

Results imply that palaeoriver morphologies were similar in space and time, with palaeoflow depths of order 2–4 m (Fig. 6). Previous DZ results suggest that northerly field sites (Price Canyon and Wattis Road) represent smaller transverse systems and that southerly field sites (Straight Canyon, Link Canyon and Salina Canyon) represent larger systems that include a longitudinal drainage component (Bartschi et al., 2018; Pettit et al., 2019). These results indicate that size disparities between these 5 systems were not statistically significant — reconstructed variations in palaeoflow depths are within the interquartile range. However, palaeoflow depths appear to have been marginally greater in southerly systems (Fig. 6). If true, this may be attributed to the possible longitudinal drainage component (Bartschi et al., 2018; Pettit et al., 2019).

Comparisons with modern rivers suggest that these 5 parallel palaeorivers (being ~25 km apart) were substantial systems. Reconstructed hydrodynamic properties, such as flow velocities and unit water discharges, are consistent with the ranges of values of modern systems with similar outlet spacings and similar distances to range fronts (Perry et al., 1996; Schulze et al., 2005; Milliman & Farnsworth, 2013; Global Runoff Data Centre). Notably, unit discharges are overall constant in time — there is no apparent increase in unit discharge at the Blackhawk–Castlegate transition (coeval with palaeoslope increase). This raises questions as to the nature of down-system width evolution and has implications for total discharge — plausible single-thread river widths of 100–500 m at down-dip locations would imply median total discharges of 250–1250 m^3/s.

Bedload transport was dominant at gravel-dominated up-dip localities, as expected, and suspended- and mixed-load systems prevailed further down-dip, with some localised variations (Figs 11, 12). For example, results highlight the importance of bedload transport during Castlegate Sandstone deposition in the southern transect (Figs 11, 12). With this information it is possible to map out how river behaviour varied spatially within catchments, and this informs best practices when it comes to reconstructing sediment discharges. This is especially important where interested in reconstructing the entire sediment load of an ancient system. For instance, channel palaeohydrologic approaches are often used to reconstruct sediment discharges in ancient source-to-sink systems (Holbrook & Wanas, 2014; Lin & Bhattacharya, 2017; Sharma et al., 2017), however these reconstruction tools solely reconstruct the bedload fraction and the suspended fraction of the bed material load (van Rijn, 1984b; Wright & Parker, 2004), i.e. the portion of the suspended load that interacts with the bed. As such, these reconstruction tools are not appropriate, by themselves, for reconstructing the total sediment load of a wash load-dominated system, for example. Knowledge of prevailing sediment transport modes is important for evaluating whether different sediment discharge reconstruction methods are consistent with one another, as studies that reconstruct sediment discharges often corroborate results with an independent approach (Lin & Bhattacharya, 2017; Watkins et al., 2018; Zhang et al., 2018; Brewer et al., 2020; Lyster et al., 2020).

Here, reconstructions of planform morphology, following G. Parker (1976), and assuming channel widths <1 km, imply that single-thread rivers would have prevailed throughout Blackhawk–Castlegate–Price River deposition. Localized or intermittent transitions to braided planforms may have been associated with tectonic or climatic perturbations, such as increased palaeoslope or high-magnitude, low-frequency discharge events (Fig. 13). In detail, these perturbations (which can be associated with the gravel-fraction grain-size) can support braiding at narrower channel/channel-belt widths of order 500 m. Of these fluvial systems, Castlegate systems had a higher propensity to braiding. At this point, it is important to flag that traditional bipartite classification of fluvial systems aims to define fluvial systems as either straight/meandering or braided/anabranching end members (Leopold & Wolman, 1957). However, these are not mutually exclusive; both straight/meandering and braided/anabranching planforms can co-exist at reach scales. These reconstructions can be contextualised by field evidence; however, field observations point to a discrepancy and this topic is returned to later.

To create a holistic view as to the nature of these ancient fluvial landscapes, various modern analogues can be considered. In the Amazon basin, several of the most up-system tributaries axially drain the central and eastern Andean cordillera. For example, the Huallaga river, Peru, is an axial river fed by transverse systems draining the eastern Andean range front. These transverse rivers have regular outlet spacings, channel-belt widths of order 100s of metres (up to 1 km), and combine both single- and multi-thread planforms which vary at reach-scales. In the eastern Himalayas, transverse systems draining the range front into the axial Brahmaputra (Assam Valley) provide another modern analogue for the pattern and style of these ancient fluvial systems, despite the larger scale of this system.

(B) What drove spatio-temporal changes in morphologic properties?

A key result in this study is quantification of an increase in palaeoslope at the Blackhawk–Castlegate transition by a factor of 1.5–4, as well as the associated increase in palaeorelief (Figs 7–9). Increased palaeoslopes have implications for the morphologic and hydrodynamic properties of these palaeorivers, including their flow velocities and unit discharges. In this study, the increase in palaeoslope and palaeorelief implies that rivers were actively responding to changes in uplift rate in the hinterland region.

At the Blackhawk–Castlegate transition, palaeorelief increased from 10s of metres to c. 100 m (Fig. 9). An important point to remember is that these estimates are specific to the alluvial domain only. Behind the Sevier front, existence of a high-elevation plateau known as “Nevadaplano” is inferred (Allmendinger, 1992; DeCelles, 1994, 2004; DeCelles & Coogan, 2006), which has been likened to the modern high-elevation plateau, Altiplano, of the central Andes. Palaeo-elevations in the Sevier highlands and Nevadaplano are argued to be 3 to >4 km — these values have been deduced from a combination of climate modelling studies (Sewall & Fricke, 2013; Foreman et al., 2015), kinematic reconstructions (DeCelles, 1994, 2004; DeCelles & Coogan, 2006) and other data, including palaeoflora (Chase et al., 1998). Here, alluvial palaeorelief of order 100 m is reconstructed. Given that the low-lying alluvial domain of these palaeorivers has a length scale of order 70–250 km, and given proximity to high-elevation Sevier highlands, the entire river long profile is inferred to have likely been highly concave. This is supported in part by the fact that, in reconstructing palaeoslope profiles, the best fits were recovered when using a higher reference concavity of 0.6 (Supplementary Table S6). If best-fit palaeoslope profiles were projected up-dip into the Sevier hinterland, palaeoslopes of 10^{-1} might be reached within as little as 10 km of the most up-dip field area, and therefore elevations in excess of 1 km might be reached within a further 10 km. To again offer the modern Andes as an analogue, if one were to plot an elevation profile from Peruvian shorelines, through the alluvial domain, and into the western Andean cordillera and Altiplano, one would traverse an alluvial domain of order 50–150 km, with 500 m to 1 km of relief, before crossing into the >3 km elevations of the western cordillera and Altiplano. With a similar tectono-geographic setting in Late Cretaceous Utah, this comparison can also be used to highlight the potential high concavity of these ancient river profiles.

In reconstructing palaeorelief, steepness indexes, k_s, were also recovered for northern and southern transects (Equations 5 and 6) (Supplementary Table S6). While k_s was solved for using field data and a nonlinear least squares regression, k_s values are often estimated (albeit tenuously) as a function of known uplift rate and erodibility in bedrock channels, but additionally (although less frequently) in downstream alluvial reaches (Kirby & Whipple, 2012; Pederson & Tressler, 2012; Stucky de Quay et al., 2019). Inversely, where k_s can be measured, and where erodibility is known, first-order estimates of uplift rate can be made. Steepness indexes recovered in this study were typically ~5–35 m (for a reference concavity, θ, of 0.5) and, despite unknown erodibility, global data compilations indicate that low uplift rates of order 0.01–0.1 mm/yr are generally associated with these kinds of values (Kirby & Whipple, 2012). Despite overall low k_s values, it is important to note the relative increase in k_s by a factor of <1.5 to 5 at the Blackhawk−Castlegate transition. While these are first-order estimates, and are derived solely for the alluvial domain, an increase in k_s (and palaeorelief) can be attributed to a relative increase in uplift rate in the hinterland region. Here, this increase might be attributed to frontal thrust migration, or thrust initiation in the Sevier highlands (DeCelles, 2004; DeCelles & Coogan, 2006). This includes Sevier shortening in the Charleston–Nebo Salient (CNS), an eastward convex portion of the Sevier thrust front in north-central Utah (Fig. 1b) (Bruhn et al., 1986; Bryant & Nichols, 1988; Constenius et al., 2003; Bartschi et al., 2018), which is commonly attributed to the influx of quartzite-dominated coarse-grained detritus associated with Castlegate Sandstone progradation (Robinson & Slingerland, 1998; Horton et al., 2004). For Castlegate Sandstone deposition in the northern transect, results show that palaeoslope profiles did not fit reconstructed palaeoslopes well and favoured lower concavities (which also did not fit well). Our interpretation is that shortening in the CNS, which has been structurally linked with coeval basement Laramide uplifts in northern Utah (Bruhn et al., 1986; Bryant & Nichols, 1988; Constenius et al., 2003; Bartschi et al., 2018), may have significantly influenced river long profiles associated with northerly Castlegate fluvial systems near Price, and locally lowered their concavities. Whereas ~60 km south in the southern transect, higher concavity values of 0.6 deliver best fitting palaeoslope profiles through all 7 stratigraphic intervals (Supplementary Table S6).

While tectonic drivers are commonly attributed to variations in channel steepness (Kirby & Whipple, 2001; Kirby et al., 2003; Wobus et al., 2006; Boulton & Whittaker, 2009; DiBiase et al., 2010), climatic drivers, especially precipitation rates, also play a crucial role but are notoriously difficult to disentangle from their tectonic counterpart (Wobus et al., 2010; DiBiase & Whipple, 2011; Champagnac et al., 2012; Whittaker, 2012; D'Arcy & Whittaker, 2014). The role of climate is important to consider here, given the assumed monsoonal climate and, therefore, highly seasonal discharge variability (Roberts, 2007; Roberts et al., 2008; Fricke et al., 2010; Sewall & Fricke, 2013). Previous work shows that precipitation rates have a discernible role on steepness indexes (Champagnac et al., 2012; D'Arcy & Whittaker, 2014); analytically, an increase in channel steepness and palaeoslope can be attributed to a decrease in precipitation rate (to maintain similar total water discharge) (D'Arcy & Whittaker, 2014). To reduce palaeoslopes by a factor of 2 precipitation rate must typically be quadrupled (D'Arcy & Whittaker, 2014). Despite the supposed warm and wet climate (L. R. Parker, 1976; Kauffman & Caldwell, 1993; Roberts & Kirschbaum, 1995), few workers have argued for, or investigated, the possibility of increased aridity at the Blackhawk–Castlegate transition (van Wagoner, 1995; Adams & Bhattacharya, 2005). In theory, increased palaeoslopes can be explained by decreased precipitation (D'Arcy & Whittaker, 2014), however, here, no decrease in either flow velocities or unit discharges is reconstructed at the Blackhawk–Castlegate transition (Fig. 10). Generally, in down-dip locations, flow velocities and unit discharges are constant across this interval (Fig. 10d,f). At up-dip field sites, however, flow velocities are overall slightly greater during Castlegate Sandstone deposition relative to Blackhawk Formation deposition, but unit discharges remain similar for both.

With unit discharges constant in space and time, the crucial unknown is palaeochannel width. At minimum, channel widths can be considered as broadly the same across the Blackhawk–Castlegate transition. During Blackhawk Formation deposition, channelized sandbody widths of order 350–420 m offer a maximum limit on palaeochannel widths (Hampson et al., 2013; Flood & Hampson, 2015). Meanwhile, during Castlegate Sandstone deposition, bar package widths are between ~60–180 m (Chamberlin & Hajek, 2019); assuming 2–3 threads, these bar widths might imply channel belt widths of order half a kilometre. However, planform stability estimates based on G. Parker (1976) indicate that these rivers could have possessed anywhere between 1–10 threads (Fig. 13), which could result in channel-belt widths up to and in excess of 1 km. At maximum, this implies increased channel widths at the Blackhawk–Castlegate transition. Unless a significant decline in river widths is projected, then field results do not directly support a climatic driver. Consequently, our interpretation is that increased channel steepness and palaeoslope at the Blackhawk–Castlegate transition is due to tectonically driven uplift in hinterland regions.

(B) Effectiveness of palaeohydrological and palaeomorphological reconstructions

While quantitative reconstructions have led to significant advances in both the quantity and level of detailed information that can be extracted from fluvial strata (e.g. Ganti et al., 2019a), it is unclear how accurately these tools characterise ancient systems. Addressing this question is particularly important as sedimentology becomes increasingly numerical and it becomes easier to apply quantitative tools to stratigraphy (Duller et al., 2010; Whittaker et al., 2011; Holbrook & Wanas, 2014; Ganti et al., 2019b). With extensive existing work on Late Cretaceous fluvial systems of central Utah, results in this study offer a unique opportunity to highlight consistencies and discrepancies between quantitative interpretations of fluvial palaeohydrology and more qualitative field-based facies and architectural interpretations.

To first-order, whether or not point reconstructions of various morphologic and hydrodynamic parameters agree with qualitative interpretations can be evaluated using independent proxies (derived from field measurements or facies interpretations). As previously mentioned, reconstructed flow depths agree with several secondary observations of bar heights (Adams & Bhattacharya, 2005; Lynds & Hajek, 2006; McLaurin & Steel, 2007; Hajek & Heller, 2012; Chamberlin & Hajek, 2019) (Supplementary Table S4), which can be used as a direct proxy for flow depth (Bridge & Tye, 2000; Hajek & Heller, 2012). This agreement indicates that cross-set heights can be used to reconstruct reasonable flow-depth constraints and are useful as a bedform-scale approach. Such an approach is particularly useful in core data, locations with limited outcrop exposure, or deposits where the degree of bar preservation is poor. It is noted that scaling relations that relate cross-set heights with original bedform heights (and subsequently formative flow depths) are derived from theory and experiments that assume statistical steady state, in which flow is constant (Paola & Borgman, 1991; Leclair, 2002; Jerolmack & Mohrig, 2005). As such, agreement of flow depth reconstructions with bar heights might therefore imply that these dunes were formed in steady flow conditions (Ganti et al., 2020). This contrasts with literature that alludes to the preferential preservation of dunes in unsteady flow conditions (Reesink & Bridge, 2007; Reesink & Bridge, 2009; Reesink et al., 2015; Leary & Ganti, 2020), and merits further work regarding the kinematic controls on dune preservation in this region.

For more complex palaeohydrologic reconstructions, such as palaeoslopes and palaeorelief (Figs 7–9), it is not possible to directly corroborate estimates with independent proxies derived from field data. Nevertheless, it is still possible to evaluate reconstruction tools by contrasting commonly used methods. In this study the first approach used a theoretically-based Shields stress inversion (Equation 3), whereas the second approach used the empirically-derived model (Equation 4) of Trampush et al. (2014). Palaeoslope estimates derived from each approach are in broad agreement with one another. Each method typically recovers estimates of the same order of magnitude — in many cases the interquartile ranges of estimates overlap and in all cases the extent of the extremes overlap (i.e. the whiskers). These point comparisons between the 2 methods are promising, and in line with comparisons made elsewhere (e.g. Ganti et al., 2019a). However, there are implications when larger spatial scales are concerned, imparting uncertainty that must be carried forward in interpretation of palaeorelief in the depositional reaches of these systems. Along the northern and southern transects, Shields stress inversion estimates consistently show higher differences in palaeoslope (i.e. higher slopes up-dip and lower slopes down-dip) relative to palaeoslopes derived from the Trampush et al. (2014). This difference is likely an outcome of the Trampush et al. (2014) method using a continuous function to estimate slope, whereas the Shields stress inversion relies on a step-change empirical estimate for gravel or sand-bed rivers. Regardless of the method used, palaeoslope reconstructions are dependent on grain-size and flow-depth estimates. Because flow depths did not appreciably change in Blackhawk and Castlegate palaeorivers, variations in reconstructed slopes and derivative estimates (e.g. water and sediment discharge) are largely driven by observed differences in grain-size.

Despite the differences of the 2 methodologies on palaeorelief, estimates of palaeorelief can be compared with relief in modern systems possessing similar tectono-geographic set-ups. Palaeorelief estimates between 50 and 100 m in depositional reaches of these ancient fluvial systems are reasonable when compared with relief in modern systems with a similar tectono-geographic setting. For example, one can return to the Andean analogue, but cross over to the eastern Andean cordillera and into the foreland basin and low-lying plains of the Amazon river. For most of its course, the Amazon long profile has a relief of less than 100 m (Milliman & Farnsworth, 2013) — relief only exceeds 100 m in proximity to the range front (Milliman & Farnsworth, 2013).

Finally, these results complement field evaluation of the nature of Blackhawk Formation and Castlegate Sandstone planforms, but also raise new questions. Channelized sandstone bodies of the Blackhawk Formation are typically 350–420 m wide (Adams & Bhattacharya, 2005; Hampson et al., 2013; Flood & Hampson, 2015), although a small proportion are much larger and some exceed 1 km (Flood & Hampson, 2015). These sandstone bodies offer a maximum cap on palaeoflow width. The Blackhawk Formation is considered to mostly represent single-thread systems, which results in this study agree with. However there is significant field evidence that many channelized sandstone bodies of the Blackhawk Formation represent multi-thread systems with mid-channel bars, based on bar facies observations (Adams & Bhattacharya, 2005; Hampson et al., 2013; Flood & Hampson, 2015). Field observations of multi-thread Blackhawk fluvial systems of order 100s of metres are inconsistent with our results, which suggest multi-thread systems would not have been stable (Fig. 13). Meanwhile, the Castlegate Sandstone is interpreted to be fully-braided from facies observations (Miall, 1993, 1994; Miall & Arush, 2001; McLaurin & Steel, 2007). Reported mean bar package widths of order 60–180 m for the Castlegate Sandstone (Chamberlin & Hajek, 2019) would imply total channel widths <1 km (assuming a few braids); our reconstructed planform stability estimates, which indicate that Castlegate systems should have been single-threaded, are again inconsistent with sedimentological facies and architectural interpretations. Other quantitative reconstructions of planform have contradicted traditional field-based facies observations (Ganti et al., 2019a), and these inconsistencies must be treated carefully. The main limitation to reconstructing ancient channel planforms is a lack of reliable methods for estimating palaeochannel widths. Interpreting palaeochannel planforms from facies associations and stratigraphic-architectural data is not trivial, particularly where outcrop is limited or where observations are equivocal. But, in this case, a number of workers have concluded that braided conditions prevailed at the time of Castlegate Sandstone deposition (Lawton, 1986b; Miall, 1994; van Wagoner, 1995; Miall & Arush, 2001) and occurred at times during Blackhawk Formation deposition (Adams & Bhattacharya, 2005; Hampson et al., 2013; Flood & Hampson, 2014, 2015). As such, it can be argued that further detailed work to test and reconcile facies-based and hydraulically derived interpretations of channel planforms is a pressing research goal.

(A) Conclusions

Here a four-dimensional reconstruction of palaeohydrology in Late Cretaceous palaeorivers of central Utah, USA, is presented, using field data and a well-established quantitative framework. Overall, fluvial morphologies were similar in space and time, although marginally greater reconstructions of flow depths in southerly systems likely reflect the contribution of a longitudinal drainage component. The most prominent spatio-temporal change is an increase in palaeoslope at the Blackhawk–Castlegate transition by a factor of 1.5–4; this reflects an increase in palaeorelief (for the alluvial domain) from 10s of metres during Blackhawk Formation deposition up to, and in excess of, 100 m during Castlegate Sandstone deposition, which persisted into Price River Formation times. The observation that unit water discharges do not change at the Blackhawk–Castlegate transition does not support a climatically driven increase in palaeoslope and channel steepness. Results therefore point to a tectonically driven palaeoslope increase. In deciphering the relative role of tectonic and climatic drivers, the main limitation in this study is uncertainty in palaeochannel widths, which directly affect total water discharges. Palaeochannel width reconstructions therefore remain a prominent research challenge.

Results complement and expand on extensive facies-based interpretations of these systems, which offers unique opportunity to evaluate the efficacy of quantitative palaeohydrological reconstruction tools. Bedform-scale palaeoflow depth reconstructions are in good agreement with observations of preserved barforms. Moreover, while different palaeoslope reconstruction methods produce results that broadly agree, the results show that at larger spatial scales they over- and under-predict relief relative to one another, which has implications for quantifying alluvial palaeorelief and, therefore, the magnitude of change in relief at the Blackhawk–Castlegate transition. Finally, quantitative hydraulic reconstructions of planform somewhat disagree with facies-based interpretations. While this discrepancy ties back to uncertainty in palaeochannel widths, these results highlight that further work is required to reconcile hydraulically- and facies-based approaches in order to facilitate their application in the geological past.

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Author Contributions

SJL and ACW designed the study. SJL, ACW and BAL conducted field data collection. SJL processed field data and results. SJL, ACW, GJH and EAH analysed and interpreted results. SJL wrote the manuscript. ACW, GJH, EAH and PAA all contributed significantly to the manuscript.

Data Availability

Field data available in article supplementary material.

Figures

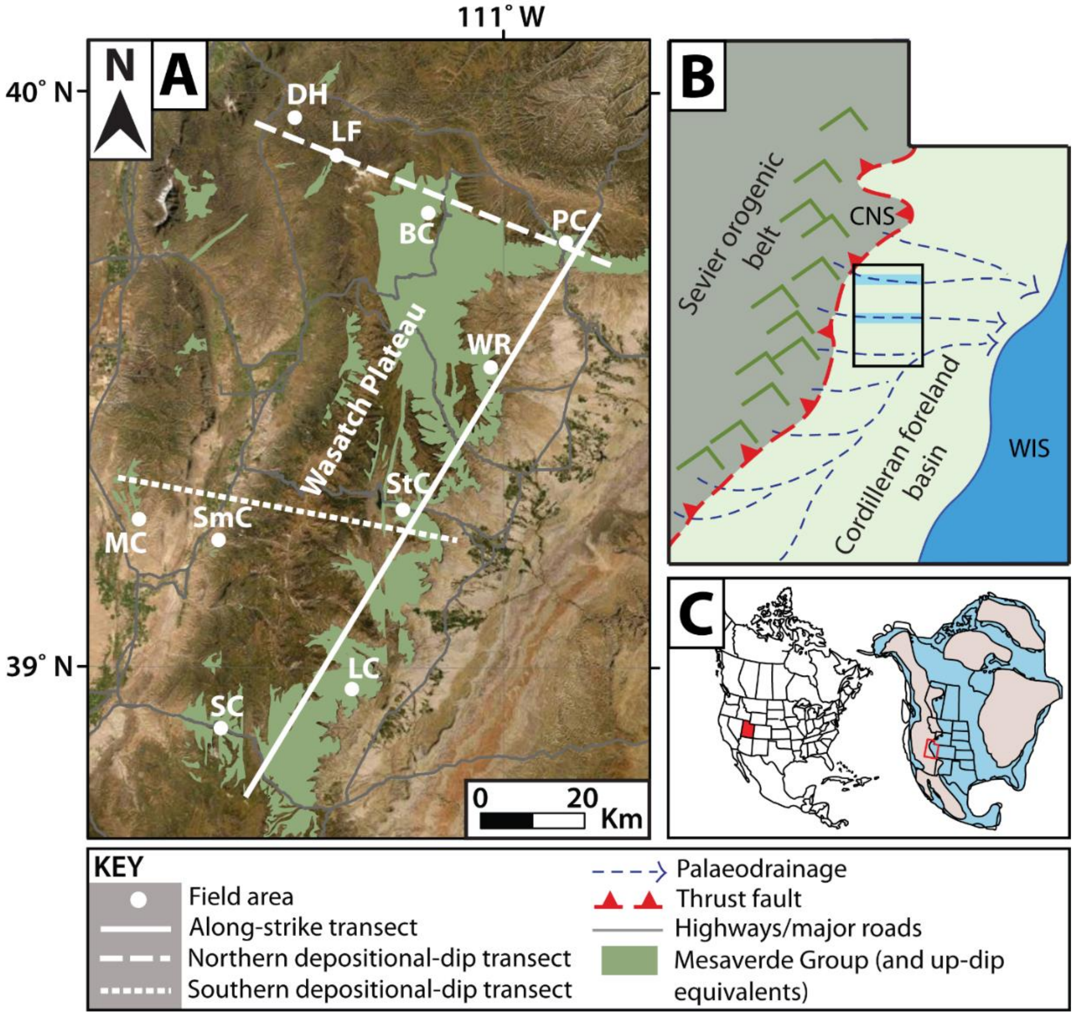

Figure 1: Study area. Part A) Field areas in central Utah, USA, which include Bear Canyon (BC), Dry Hollow (DH), Lake Fork (LF), Link Canyon (LC), Mellor Canyon (MC), Price Canyon (PC), Salina Canyon (SC), Sixmile Canyon (SmC), Straight Canyon (StC) and Wattis Road (WR). The solid white line indicates the along-depositional-strike transect defined in this study, the dashed white line indicates the northern depositional-dip transect defined in this study, and the dotted white line indicates the southern depositional-dip transect defined in this study. Part B) A conceptual diagram of Utah palaeogeography and palaeodrainage in the Campanian (Late Cretaceous). Likely configurations of drainage toward the Western Interior Seaway (WIS) are indicated by dashed blue lines. CNS = Charleston–Nebo Salient. The black outlined box indicates the study area (i.e. part A), and the two highlighted drainage routes (shaded blue) represent the northern and southern depositional-dip transects defined in this study (see part A). Part C) The location of Utah relative to the modern North American continent (left) and the Late Cretaceous North American continent (right), which features the Western Interior Seaway (blue). Utah is highlighted as a red box.

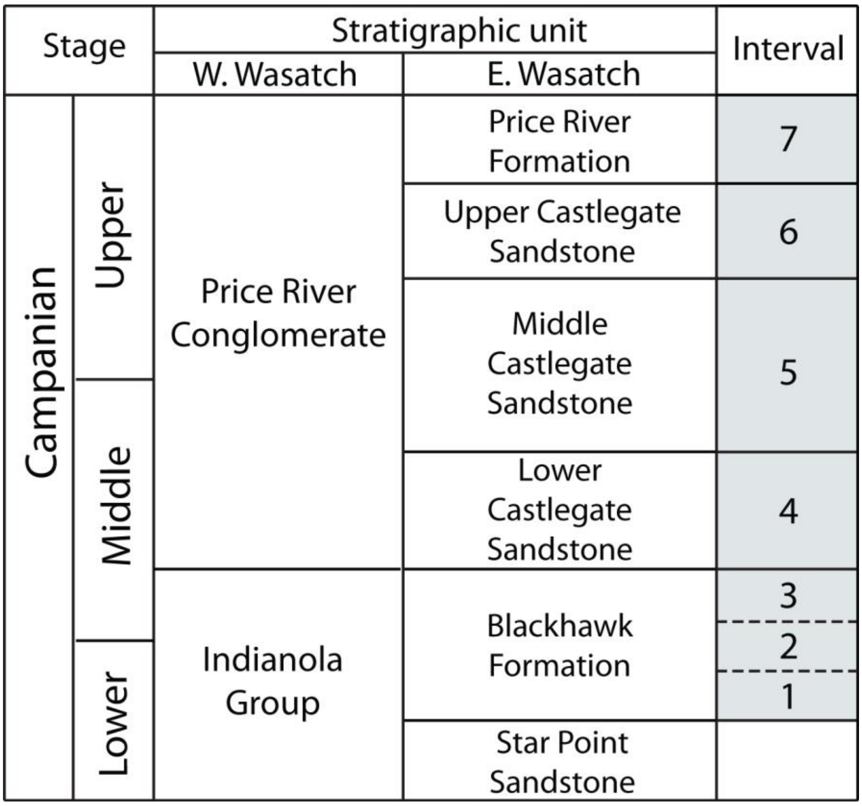

Figure 2: Regional stratigraphy and up-dip (western Wasatch Plateau) to down-dip (eastern Wasatch Plateau) stratigraphic correlation followed in this study. Shaded intervals indicate the stratigraphic intervals used in this study (note that they are not of equal duration). 1 = lower Blackhawk Formation; 2 = middle Blackhawk Formation; 3 = upper Blackhawk Formation; 4 = lower Castlegate Sandstone; 5 = middle Castlegate Sandstone; 6 = upper Castlegate Sandstone (Bluecastle Tongue); 7 = (lowermost) Price River Formation. Dashed lines indicate an approximate interval boundary. Modified and compiled using data from Fouch et al. (1983); Robinson and Slingerland (1998); Miall and Arush (2001); Horton et al. (2004); Cobban et al. (2006); Aschoff and Steel (2011a, 2011b); Bartschi et al. (2018). Price River Conglomerate nomenclature follows Aschoff and Steel (2011a, 2011b).

Figure 3: An overview of fluvial strata from which palaeohydrological field data were collected. Data were collected for 5 parallel palaeorivers in Late Cretaceous central Utah, USA. These 5 palaeorivers cropped out in canyons on the eastern front of the Wasatch Plateau — parts A and B show typical exposure of the Blackhawk Formation, Castlegate Sandstone, and Price River Formation in these canyons. Specifically, part A shows strata in Salina Canyon and part B shows strata in Straight Canyon (see Fig. 1), and dashed white lines indicate lithostratigraphic boundaries. For two of these 5 palaeorivers, data were additionally collected upstream to downstream along defined depositional-dip transects (see Fig. 1). Parts C–E show deposits on the northern depositional-dip transect. From up-dip to down-dip, part C shows debris flow facies of the Price River Conglomerate, part D shows amalgamated fluvial gravels and sands of the Castlegate Sandstone near Bear Canyon, and part E shows amalgamated fluvial sands of the Castlegate Sandstone in Price Canyon. Parts F–H show deposits on the southern depositional-dip transect, for older sediments. From up-dip to down-dip, part F shows channelized fluvial gravel–sand bodies of the upper Sixmile Canyon Formation in Mellor Canyon, part G shows a small channelized sandstone body of the upper Sixmile Canyon Formation in Sixmile Canyon, and part H shows a large channelized sand body of the Blackhawk Formation in Straight Canyon (in the background the Castlegate Sandstone is visible).

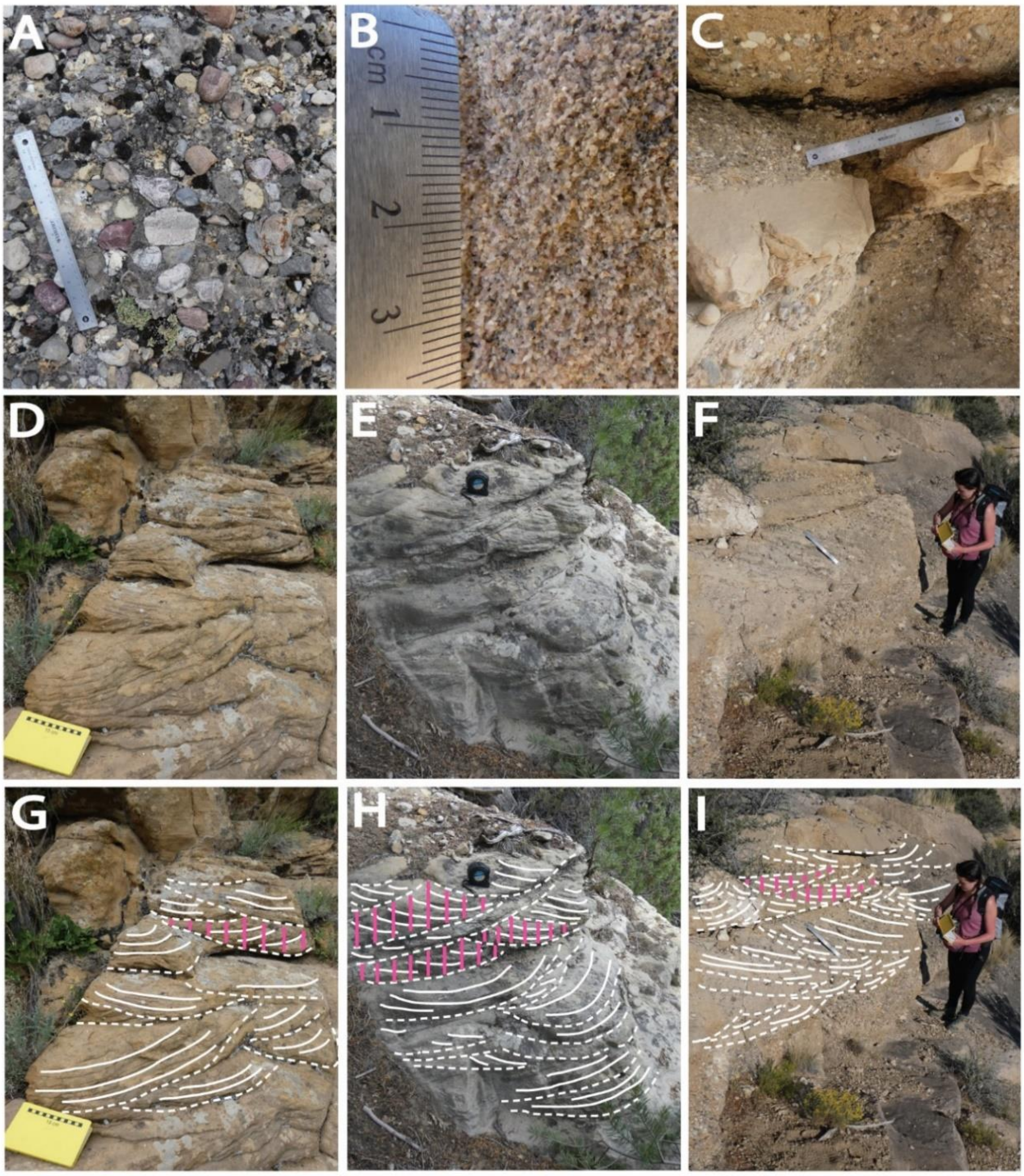

Figure 4: Field data collection included grain-size measurements for (part A) gravel and (part B) sand fractions, as well as (part C) estimates of the proportions of different grain-size facies. Parts D–F depict cross-bedding, and parts G–I depict interpreted versions of the same images. Dashed white lines indicate bounding surfaces of individual cross-sets and solid white lines indicate selected foresets within individual cross-sets. To exemplify sampling procedure when determining mean cross-set height, solid pink lines demonstrate how heights are measured for selected cross-set dip sections. Field notebook with 15 cm scale, tape measure, and 30 cm rule for scale.

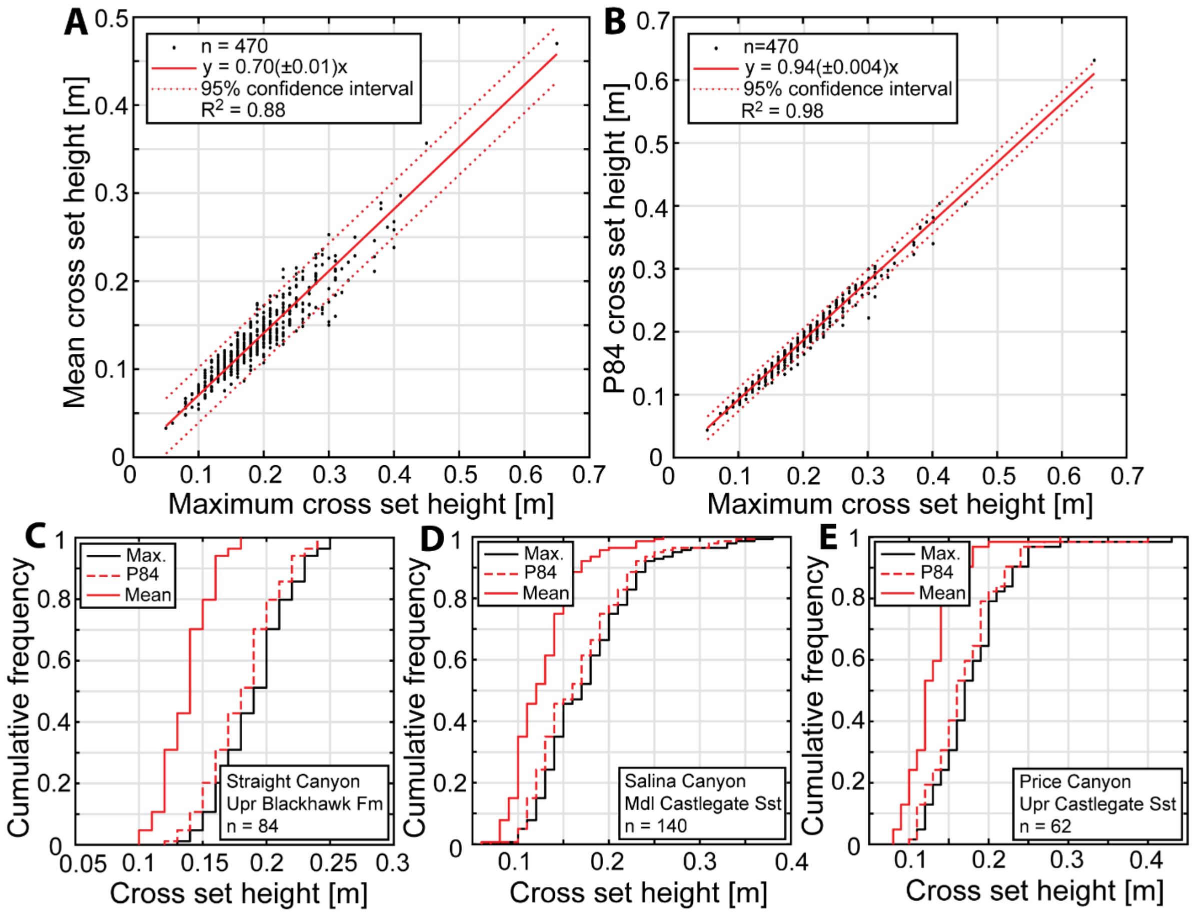

Figure 5: Part A) Relationship between maximum cross-set height and mean cross-set height. Part B) Relationship between maximum cross-set height and the 84th percentile (P_{84}) of cross-set height. Data are based on 470 measured cross-set distributions. Errors reported in the fits are 95% confidence intervals. Parts C–E) Examples of the use of these new relations (parts A and B) to predict the mean and P_{84} cross-set height from maximum cross-set heights. Examples are for the upper Blackhawk Formation in Straight Canyon (part C), the middle Castlegate Sandstone in Salina Canyon (part D), and the upper Castlegate Sandstone in Price Canyon (part E). In parts C–E, n indicates the number of maximum cross-set heights used to predict mean and P_{84} cross-set heights. Full cross set data for each field site, through each stratigraphic interval, are located in Supplementary Table S3.

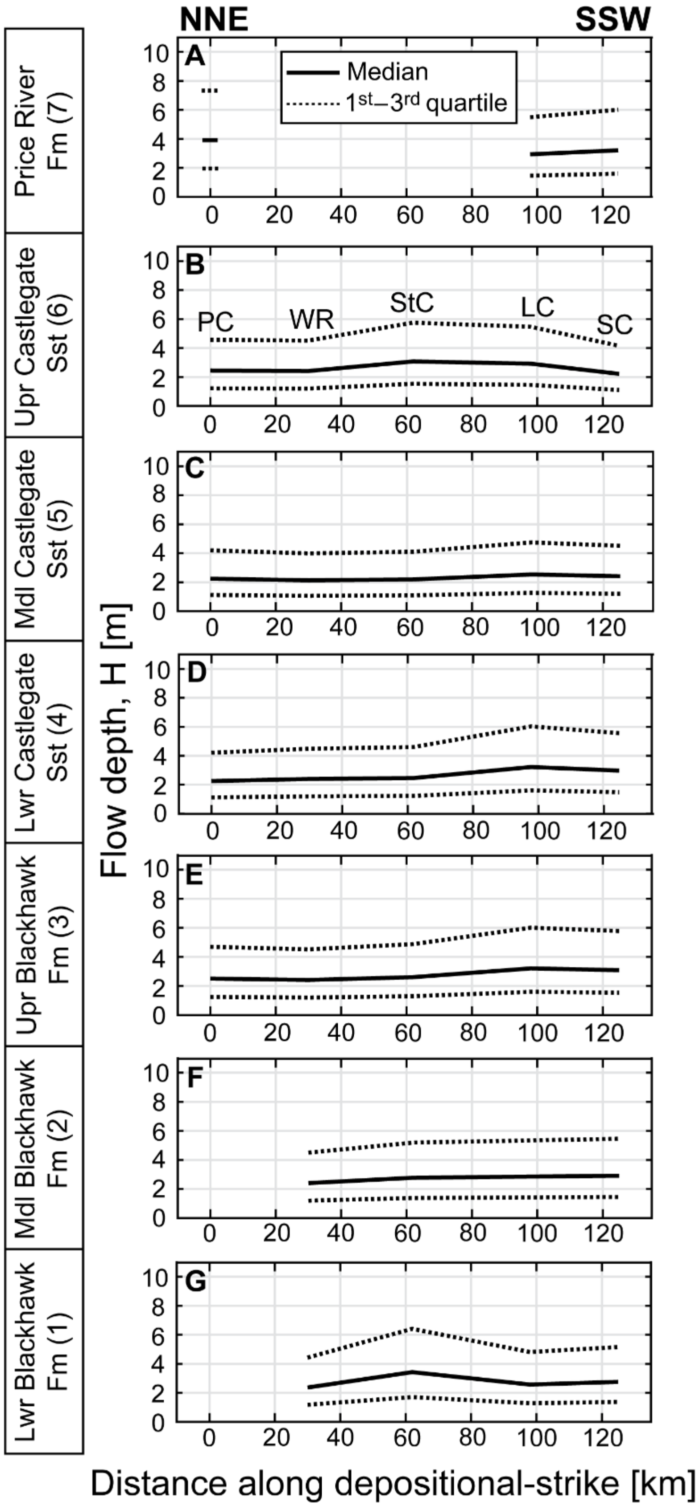

Figure 6: Reconstructed palaeoflow depths for the 5 parallel fluvial systems, for each stratigraphic interval (parts A–G), where possible, using mean cross-set heights. Results are presented as along-depositional strike transects from NNE (left; 0 km) to SSW (right; 125 km). Field sites span Price Canyon (PC), Wattis Road (WR), Straight Canyon (StC), Link Canyon (LC) and Salina Canyon (SC). Solid lines indicate median palaeoflow depths and dashed lines indicated the 1st and 3rd quartiles of palaeoflow depths. This figure is replicated in the Supplement alongside palaeoflow depths reconstructed from maximum cross-set heights (Supplementary Fig. S5).

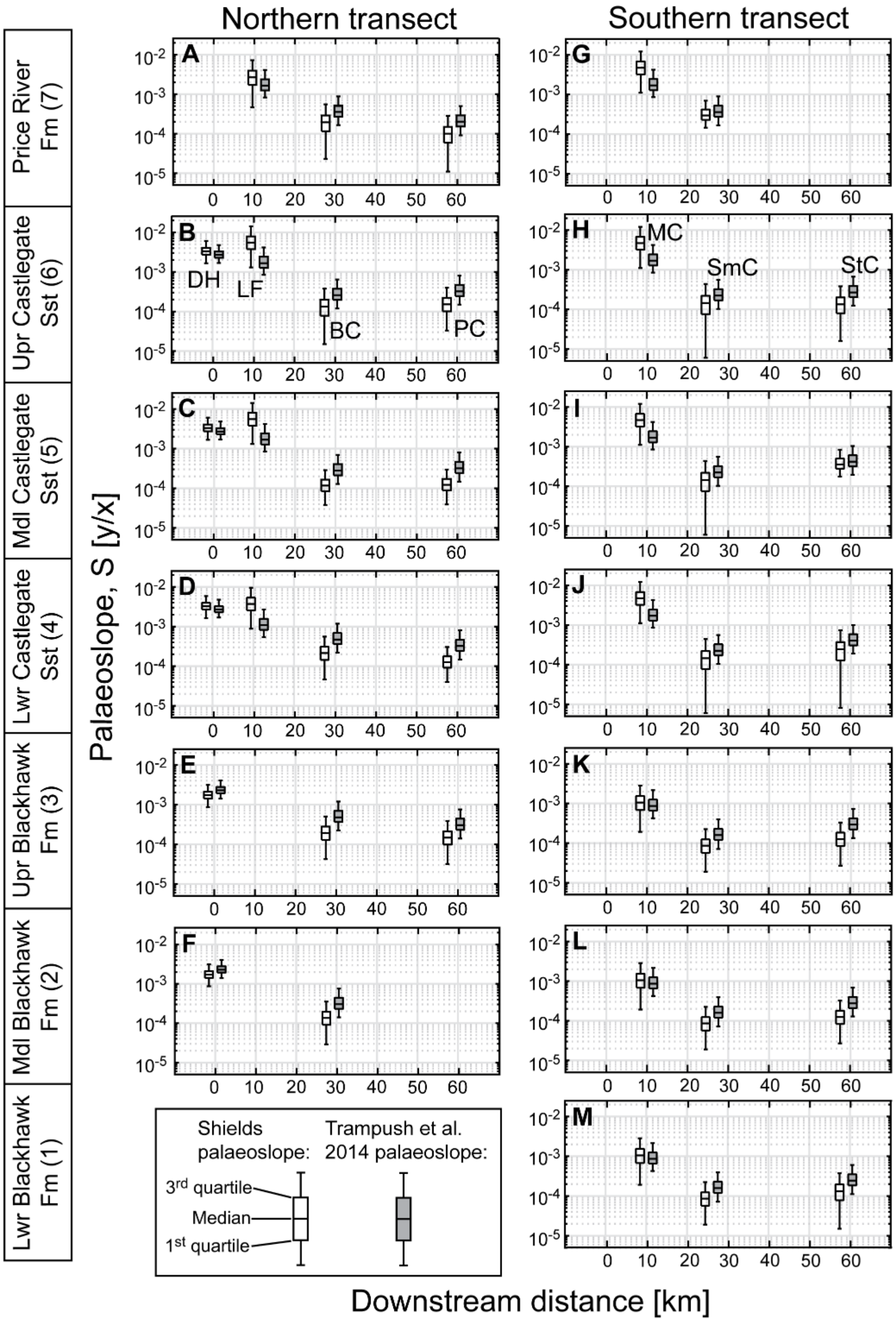

Figure 7: Up-dip to down-dip palaeoslope estimates for the defined northern and southern transects, using bulk grain-size data, for each stratigraphic interval (1–7), where possible. Parts A–F represent up-dip to down-dip palaeoslopes for the northern transect, from the middle Blackhawk Formation to the Price River Formation. Parts G–M represent up-dip to down-dip palaeoslopes for the southern transect, from the lower Blackhawk Formation to the Price River Formation. The central mark of each box indicates the median estimate, and the bottom and top edges of each box indicate the 1st and 3rd quartiles (or 25th and 75th percentiles), respectively. The whiskers extend to the most extreme estimates that are not considered to be outliers. Palaeoslope estimates are derived from 2 independent approaches; boxes with no fill indicate estimates of palaeoslope derived using a Shields stress inversion (Equation 3) and boxes with grey fill indicate estimates derived from the method of Trampush et al. (2014) (Equation 4). BC = Bear Canyon; DH = Dry Hollow; LF = Lake Fork; MC = Mellor Canyon; PC = Price Canyon; SmC = Sixmile Canyon; StC = Straight Canyon.

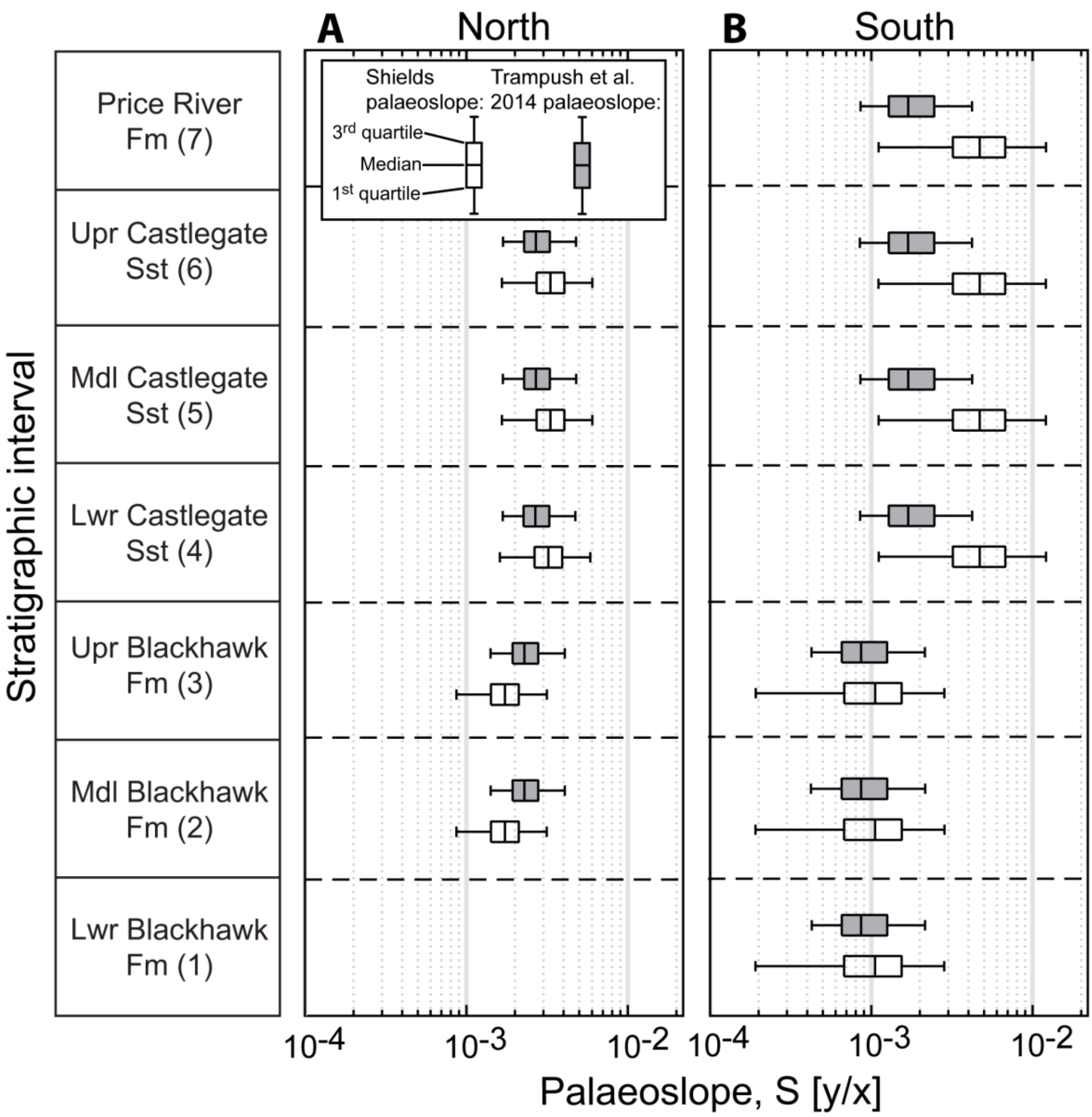

Figure 8: Palaeoslope estimates for the most up-dip location of the defined northern (part A) and southern (part B) depositional-dip transects, for each stratigraphic interval (1–7), where possible, using bulk grain-size data. The central mark of each box indicates the median estimate, and the edges of each box indicate the 1st and 3rd quartiles (or 25th and 75th percentiles) of estimates. The whiskers extend to the most extreme estimates that are not considered to be outliers. Palaeoslope estimates are derived from 2 independent approaches; boxes with no fill indicate estimates of palaeoslope derived using a Shields stress inversion (Equation 3) and boxes with grey fill indicate estimates derived from the method of Trampush et al. (2014) (Equation 4).

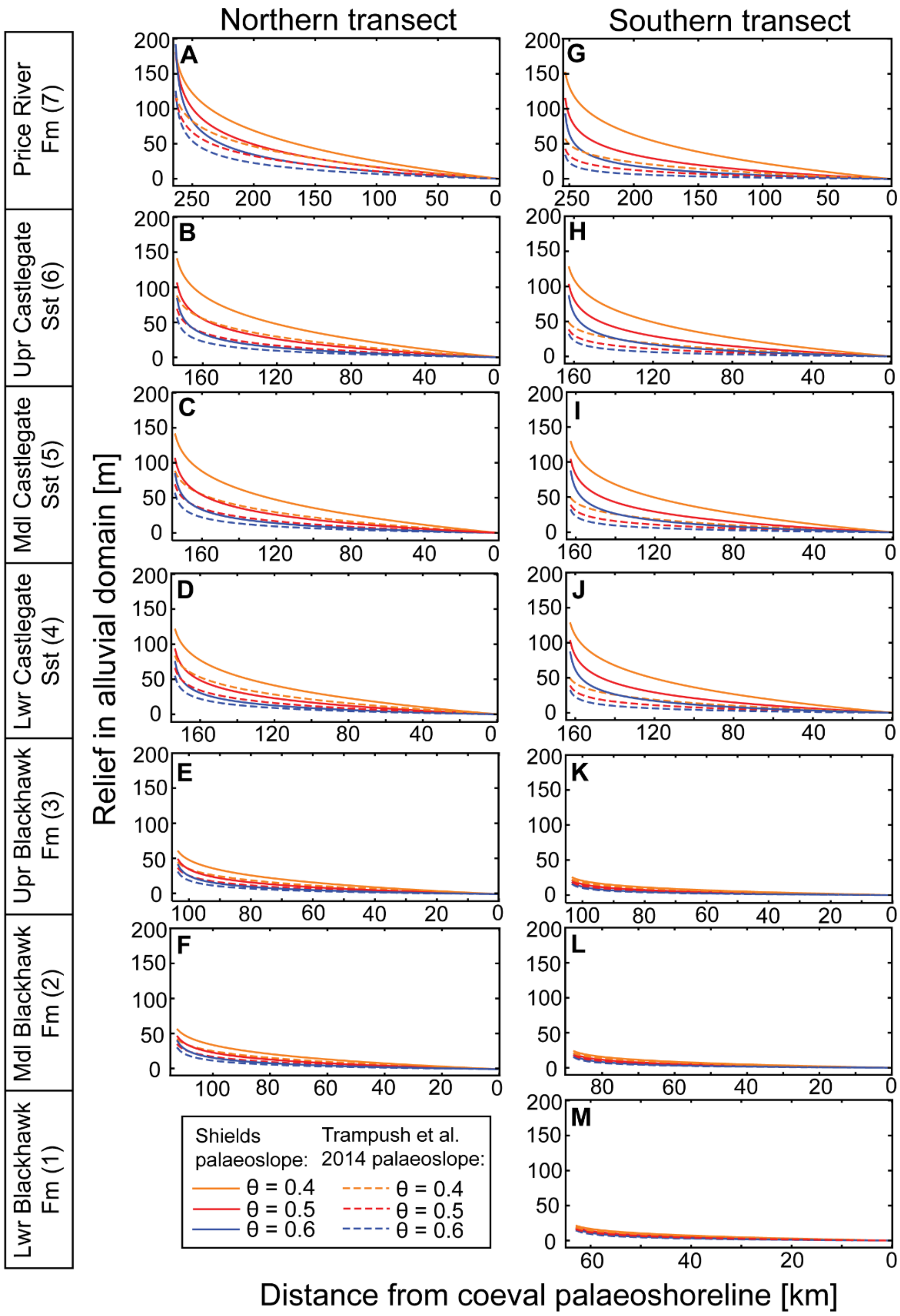

Figure 9: Estimated palaeorelief in the alluvial domain for the defined northern and southern transects, using bulk grain-size data, for each stratigraphic interval (1–7), where possible. Parts A–F depict estimated palaeorelief for the northern transect, from the middle Blackhawk Formation to the Price River Formation. Parts G–M depict estimated palaeorelief for the lower Blackhawk Formation to the Price River Formation. Palaeorelief estimates are derived using palaeoslope estimates from 2 independent approaches; palaeoslopes from a Shields stress inversion (Equation 3) and palaeoslopes from the method of Trampush et al. (2014) (Equation 4). In addition, palaeorelief is estimated using a plausible range of values for the concavity index, θ. Unlike other depositional-dip transects in this study, the x axis instead depicts distance from the coeval palaeoshoreline (following Hettinger and Kirschbaum (2002); Hampson et al. (2012); Hampson et al. (2013)).

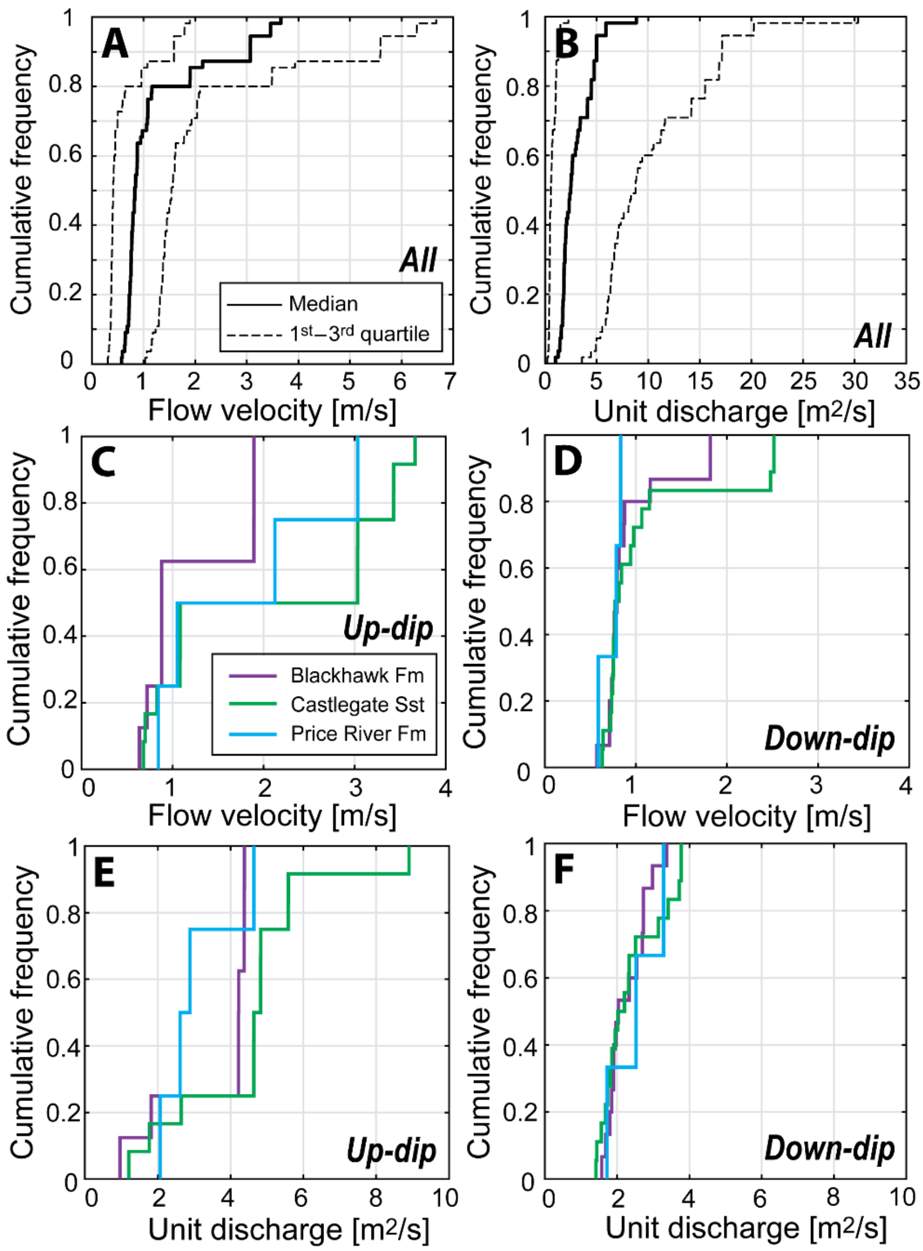

Figure 10: Cumulative frequency distributions of (part A) reconstructed flow velocities across all field areas and (part B) reconstructed water discharges, per unit width, across all field areas. Solid lines indicate median values and dashed lines indicates the 1st–3rd interquartile range. Flow velocities are derived using Manning's formula (Equation 7), as described in the Methods section. Parts C–F depict flow velocities and unit water discharges split into up-dip and down-dip field sites. Down-dip field areas include field areas on the along-strike depositional transect (Price Canyon, Wattis Road, Straight Canyon, Link Canyon and Salina Canyon), meanwhile up-dip field areas include all those that are relatively up-dip (Dry Hollow, Lake Fork, Bear Canyon, Mellor Canyon, Sixmile Canyon). Field areas were also split into the Blackhawk Formation (and up-dip equivalents, i.e. intervals 1–3), Castlegate Sandstone (and up-dip equivalents, i.e. intervals 4–6) and Price River Formation (and up-dip equivalents, i.e. interval 7). Parts C and D depict cumulative frequency distributions of reconstructed flow velocities for up-dip (part C) and down-dip (part D) field areas, respectively. Parts E and F depict cumulative frequency distributions of reconstructed unit water discharges for up-dip (part E) and down-dip (part F) field areas, respectively.

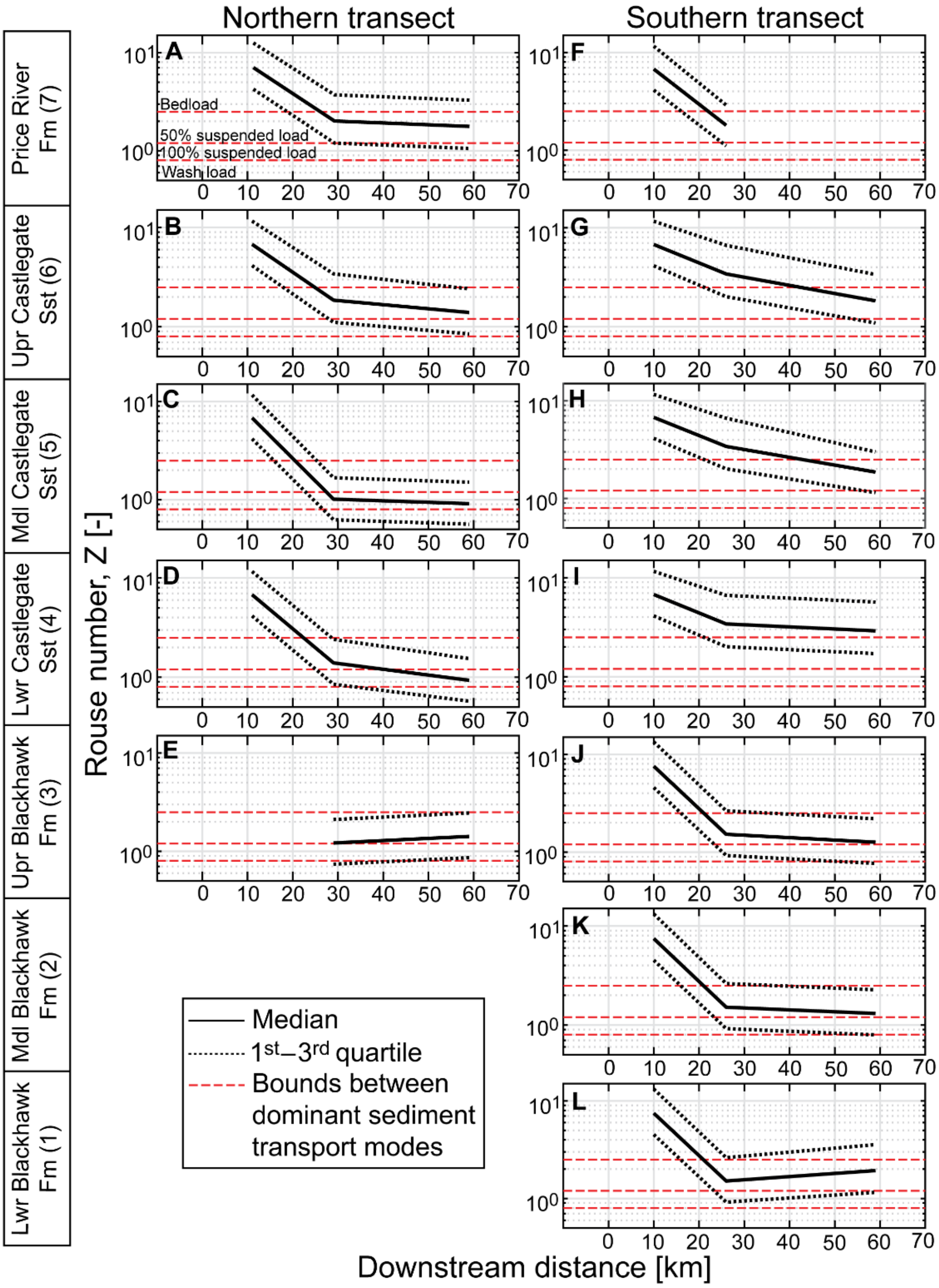

Figure 11: Estimated Rouse numbers, *Z*, for the defined northern and southern transects, using bulk grain-size data, for each stratigraphic interval (1–7), where possible. Dominant mode of sediment transport is typically wash load for $Z < 0.8$, 100% suspended load for $0.8 < Z < 1.2$, 50% suspended load (i.e. mixed load) for $1.2 < Z < 2.5$, and bedload for $Z > 2.5$. Parts A–E represent up-dip to down-dip Rouse numbers for the northern transect, from the upper Blackhawk Formation to the Price River Formation. Parts F–L represent up-dip to down-dip Rouse numbers for the southern transect, from the lower Blackhawk Formation to the Price River Formation. Solid black lines indicate the median estimate and dashed black lines indicate the 1st and 3rd quartiles. Dashed red lines indicate the bounds between differing dominant sediment transport modes, as labelled in part A.

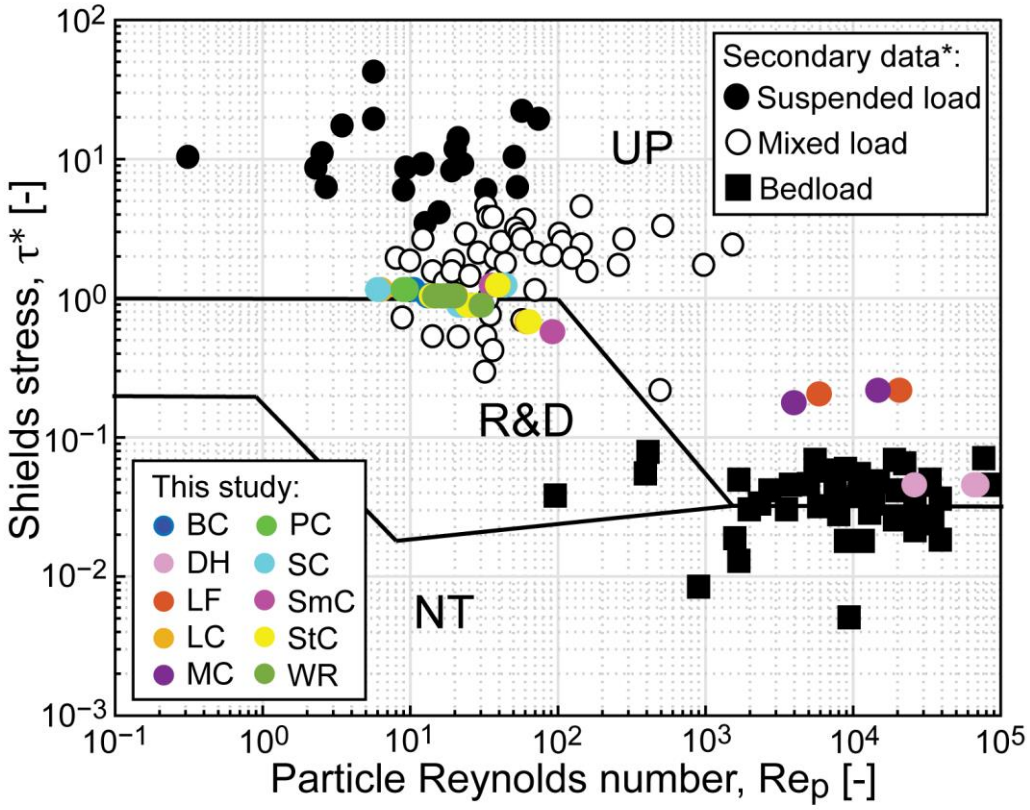

Figure 12: Shields stress, τ^*, plotted as a function of particle Reynold's number, Re_p, for all field sites and for each stratigraphic interval (1–7), where possible, using bulk grain size data. Colour-filled circles indicate field results from this study for Bear Canyon (BC), Dry Hollow (DH), Lake Fork (LF), Link Canyon (LC), Mellor Canyon (MC), Price Canyon (PC), Salina Canyon (SC), Sixmile Canyon (SmC), Straight Canyon (StC) and Wattis Road (WR). *For comparison, this plot includes secondary data, originally compiled by Dade and Friend (1998), from Leopold and Wolman (1957); Schumm (1968); Chitale (1970); Church and Rood (1983); Andrews (1984), for characteristic dominant transport modes. Black squares indicate bedload, white circles indicate mixed load, and black circles indicate suspended load. Solid black lines indicate stability fields of different flow regimes: no sediment transport (NT), ripples and dunes (R&D) and upper-stage plane beds (UP), in line with Allen (1982a, 1982b).

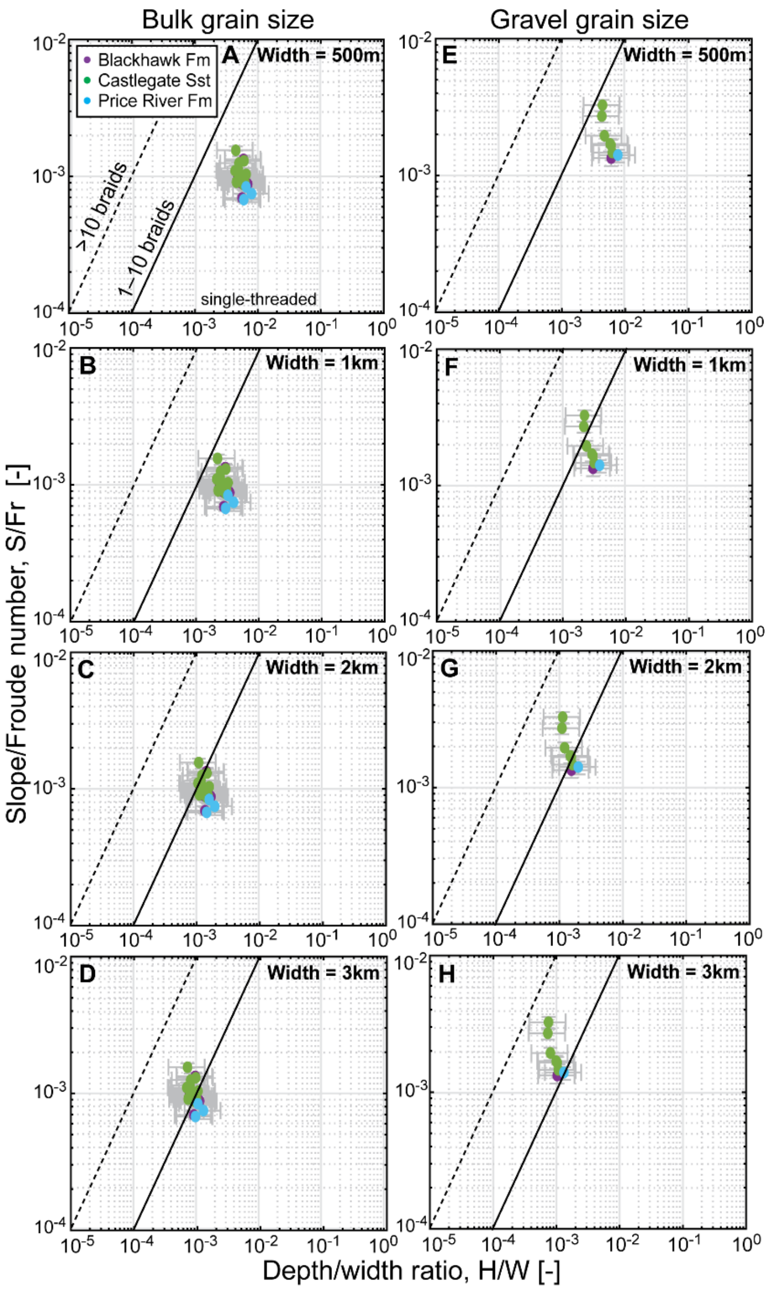

Figure 13: Theoretical stability fields of fluvial planform morphologies, i.e. single-thread and multi-thread planforms, for both bulk grain-sizes (parts A–D) and gravel fraction grain-sizes (parts E–H), where present (not all field localities possessed a gravel fraction). For both bulk and gravel grain-size fractions, a range of river widths are assumed (500 m, 1 km, 2 km and 3 km) and used to calculate the depth/width ratio. Data points are for all localities, in space and time, along the defined along-depositional strike transect, i.e. these data points represent the five parallel fluvial systems and do not consider up-dip localities. Data are further subdivided into the Blackhawk Formation (intervals 1–3), Castlegate Sandstone (intervals 4–6) and Price River Formation (interval 7). Coloured markers and error bars represent the median and the 1st–3rd interquartile range, respectively. Solid black lines indicate the bounds of each stability field, and therefore the predicted transition from single-thread (straight/meandering) to multi-thread (anabranching/braided) planform morphology. Dashed black lines indicate a potential transition from 1–10 threads to >10 threads, based on modern data (G. Parker, 1976).

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