1    **The anatomy of exhumed river-channel belts: Reconstructing bedform- to belt-scale river**

2    **kinematics from the Cretaceous Cedar Mountain Formation, Utah, USA**

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23   **ABSTRACT**

24         The geometry of river-channel belts is shaped by the migration, aggradation, and

25   avulsion of associated rivers, which is controlled by the motion of bedforms and barforms. The

26   motion of bedforms and barforms can be reconstructed using the geometry and preservation of

27   their associated strata. In order to reconstruct the kinematics of the ancient Cedar Mountain

28   rivers, preserved elements of the formative river channels, including bars, planform geometry,

29   and bed topography, are measured within exhumed fluvial strata exposed as ridges in the

30   Cretaceous Cedar Mountain Formation, Utah, USA. High-resolution ortho-images created by

31   Uncrewed Aerial Vehicles facilitated geologic field mapping, as were vertical and lateral

32   sections. The studied ridges are compound structures composed of the stacked deposits of

33   multiple, individual channel belts representing at least 5 to 6 channel re-occupations. Lateral

34   sections reveal sets of dune cross-beds constructing well-preserved barforms. Locally, the

35   paleo-topography of paleo-barforms defines the modern topographic surface of the outcrop.

36   Comparison of the channel-belt centerline to local paleotransport directions indicate channel

37   planform morphology was preserved through multiple re-occupations and minimal lateral

38   migration. The preservation of these formative channel elements is the product of the

39   kinematics of the depositional system. The avulsive character of the system preserved the state

40   of the active channel bed and its individual bars at the time of channel switching. Calculated

41   sedimentation durations for the preserved elements vary within a belt from a day to 11 days,

42   indicating most of the time recorded in these compound channel belts is represented by the

43   basal erosional surfaces at the bottom of channel belts. Avulsions in this system were frequent

44   with respect to lateral migration rates, which minimized lateral channel amalgamation.  The

45    well-preserved belts may be indicative of a larger-scale topography driving sedimentation at

46    the belt scale, much as bar topography can locally increase dune sedimentation rates and

47    preservation.

48    Keywords: fluvial sedimentology, channel belt, preservation, bar, sinuous ridge

49

50    **INTRODUCTION**

51        Fluvial channel belts are the time-integrated, channel-filling deposits of rivers. These

52    channel-filling deposits include strata associated with bedforms, such as ripples and dunes, as

53    well as free and forced bars. The vertical and lateral motion of a river is recorded by the motion

54    of these bedforms and bars, which in turn control the geometry of the resultant channel-belt

55    deposits and drive the belt's geometry away from that of the formative river (Van De Lagewag

56    et al., 2013). Therefore, in order to determine how ancient rivers migrated, aggraded, and

57    avulsed, it is necessary to understand the accumulation and preservation of the bedform and

58    barform strata within the associated channel belts (e.g., Reesink et al., 2015; Durkin et al.,

59    2018; Paola et al., 2018; Chamberlin and Hajek, 2019).

60        This work examines an erosionally exhumed, laterally and vertically amalgamated

61    complex of fluvial channel belts in the Cretaceous Cedar Mountain Formation, Utah, USA (Fig.

62    1). The goal of the project is to determine what aspects of ancient river topography and

63    geometry are partially-to-fully preserved within the channel belts. Paleo-environmental

64    properties such as water and sediment discharge associated with an ancient fluvial system can

65    be quantitatively estimated using geometric properties of its paleo-channels, if these

66    geometries can be accurately estimated from deposits representing the time-integrated record

67    of channel motion, deposition, and erosion (Wright and Parker, 2003; Parker et al., 2007;

68    Hayden et al., 2019). Here, new methodologies are developed for extracting the river-channel

69    kinematics from channel-belt deposits so that the most accurate picture of the formative paleo-

70    river channel and its kinematics, from bedform to planform scale, can be established. For

71    example, were these belts produced during incremental sedimentation, or do they represent a

72    snapshot of the channel topography and geometry at the time of avulsion? This topography

73    includes bed topography, such as barforms and bedforms, as well as planform channel

74    geometry. The measurements presented here cover the channel belts across a range of scales,

75    from local stacks of cross sets to the entirety of the outcrop, in order to understand the river

76    system that produced Cedar Mountain. The datasets analyzed in this work include aerial images

77    collected from a remotely piloted aircraft systems (RPAS, or "drone"), field maps, vertical and

78    lateral sections, modern river analogs, and another ancient example of fluvial stratigraphy in

79    the nearby Jurassic Morrison Formation. These methods and results provide a comparative

80    baseline for future studies of fluvial channel belts and their associated ancient fluvial systems.

81    The reconstruction of bed aggradation rates using these methods will also further the

82    community's understanding of how time in the rock record is recorded by preserved strata

83    versus erosional surfaces (Sadler, 1981; Paola et al., 2018).


84


85    **Background**

86    *Cedar Mountain Formation*

87        The rivers associated with the deposition of the lower Cretaceous Cedar Mountain

88    Formation drained uplifted areas of modern-day western Utah northeastward into the Mowry

89    Sea and its successor, the Western Interior Seaway. Ultimately, basin subsidence led to the

90    burial of the Cedar Mountain Formation by late Cretaceous coastal and marine deposits of the

91    Dakota Sandstone and the Mancos Shale (Currie, 1998). A regional unconformity separates the

92    base of the Cedar Mountain Formation from the top of the upper Jurassic Morrison Formation

93    (Peterson and Ryder, 1975; Kowallis et al., 1986).  The Cedar Mountain Formation has been

94    interpreted as consisting of river channel-filling sandstones and conglomerates, and overbank

95    mudstones and paleosols (Stokes, 1961; Currie, 1998; Garrison et al., 2007; Nuse, 2015; Hayden

96    et al., 2019).

97        Much of the recent work regarding the Cedar Mountain Formation has focused on the

98    geomorphology of exhumed channel-filling deposits. The channel belts are more resistant than

99    the surrounding floodplain material, resulting in the preferential erosion of floodplain strata

100    and preservation of the channel fill which defines the modern topography (Williams et al.,

101    2007; 2009; Hayden et al., 2019). The landscape within the eroding Cedar Mountain is defined

102    by ridges as tall as 35 m and 60 to 90 m wide on average, composed of exhumed channel belts

103    that are exposed in three dimensions (Fig. 1). Recent interest in these landforms and other

104    exhumed channel fills (e.g., Hayden et al., 2019; in Oman, Maizels 1987, 1990; Maizels and

105    McBean, 1990; in Egypt, Zaki et al., 2018) has partially been driven by high-resolution images of

106    morphologically similar 'fluvial sinuous ridges' on Mars (e.g., Burr et al., 2009; Davis et al., 2016;

107  Cardenas et al., 2018; Hughes et al., 2019). Hayden et al. (2019) have provided an important

108  comparison between field- and remote-sensing-based paleo-hydraulic reconstructions for the

109  Cedar Mountain Formation ridges, but a detailed sedimentologic workup provides additional

110  information for any paleo-environmental analysis.

111

112  *Dune, bar, and channel belt strata*

113      The dip direction of a dune cross stratum records the orientation of the formative dune

114  lee face, and reflects local dune migration direction (Allen, 1970; Rubin and Hunter, 1982). This

115  relationship is complicated for dunes with sinuous crestlines and variably deep troughs, which

116  create trough cross-stratification (McKee and Weir, 1953; DeCelles et al., 1983; Rubin, 1987).

117  Because of the crestline sinuosity, the local dip direction of a set of trough cross-stratification

118  may represent the mean migration direction of the associated dune plus or minus as much as

119  90° (Dott Jr., 1973). In plan-view exposures where local variability in individual dune cross strata

120  can be seen within the context of an entire set, the central axis of the set of trough-cross strata

121  can be clearly seen and migration direction can be unambiguously determined (Dott Jr., 1973).

122      Dipping strata composed of superimposed sets of dune cross strata represent the

123  accretion surfaces of larger-scale barforms. In active rivers, barforms are either forced by

124  channel shape or are free to migrate downstream (Miall, 1977; Seminara and Tubino, 1989;

125  Ikeda, 1989; Hooke and Yorke, 2011). Bars fixed to the inner bank of a channel bend, i.e., point

126  bars, grow into the channel (Ikeda et al., 1981) and record lateral river migration. Point bars

127  have been identified in the rock record based on sigmoidal lateral accretion surfaces dipping at

128     approximately perpendicular angles to the local dip directions of dune cross strata (e.g.,

129     Edwards et al., 1983; Wu et al., 2015; Almeida et al., 2016). Free bars may preserve a wider

130     array of relationships between local dune migration direction and the bar surface dip direction

131     (e.g., Allen, 1983; Almeida et al., 2016). Although not often discussed in the literature, free bars

132     and forced (point) bars commonly coexist in channels (Fig. 2; Kinoshita and Miwa, 1974;

133     Whiting and Dietrich, 1993; Hooke and Yorke, 2011).

134         In net-depositional settings, aggradation of the riverbed is coupled with aggradation of

135     the channel margins that exceeds distal floodplain aggradation (Pizzuto, 1987). Over time, the

136     channel becomes elevated relative to the floodplain, and the difference between the two

137     elevations defines the channel's superelevation. Past studies have shown that a superelevation

138     of 60% of the flow depth appears to be a threshold for river avulsion (Mohrig et al., 2000), the

139     process by which flow abandons a channel in favor of a lower topographic position (Heller and

140     Paola, 1996; Mohrig et al., 2000; Hajek and Edmonds, 2014). Studies of both modern and

141     ancient avulsive rivers suggest that rivers tend to return to previously abandoned channels that

142     became gravitational attractors to flow following the aggradation of the adjacent floodplain

143     (Heller and Paola, 1996; Reitz et al., 2010; Edmonds et al., 2016). Such systems leave behind

144     channel-belt complexes, i.e., sedimentary deposits composed of stacked channel-belts (Friend,

145     1979; Mohrig et al., 2000; Jones and Hajek, 2007; Cuevas Martínez et al., 2010; Chamberlin and

146     Hajek, 2015; Hayden et al., 2019).

147

148     **METHODS**

149    **Field campaign**

150         For this study, several datasets were acquired during a field campaign at two ridges

151    formed from exhumed fluvial strata of the Ruby Ranch Member of the Cedar Mountain

152    Formation (Fig. 1A and B). Aerial photosurveys, collected with a DJI Phantom 2 Vision Plus

153    drone, imaged the top and side surfaces of both ridges with >75% along path and side overlap

154    in photos (Fig. 1C-D). Flights were conducted from 15 to 20 m above ground level (AGL). Ground

155    control point locations were determined using an Archer Field PC with an external GPS antenna,

156    producing horizontal position data with <0.3 m RMS accuracy. Orthophotomosaics were

157    generated with 5 cm spatial resolution and were orthorectified using the concurrently

158    produced digital elevation model (DEM) using Agisoft Photoscan Pro [It's customary to cite the

159    agisoft website for this]. Structure from DEMs and orthoimages cover an area of0. 213 km$^2$ over

160    the eastern and western ridges (Fig. 1C to D). These datasets were used to map the locations of

161    bounding surfaces of cross-sets of various scales and major erosional surfaces. Dip directions of

162    cross-strata identified on the photomosaics were measured in the field using compasses.

163    During this survey, each set was classified as either being composed of sandstone or

164    conglomerate.

165         In addition to the plan view data, several sets of vertical sections were logged. Around

166    the perimeters of each ridge, 59 vertical sections were measured covering the entirety of the

167    available vertical exposure of the ridge-capping rock, resulting in 276 total meters of section.

168    Collecting the thicknesses of units between major bounding surfaces was the primary goal of

169    this effort. An additional 31 two dimensional (2-D) panels ranging from 1 m to 10 m wide were

170    collected around the perimeters of both ridges in order to describe the smaller, cross-set scale

171    architectural elements of the channel belts. Architectural variability in the transport direction at

172    the scale of a few meters was recorded, including changes in set thickness and the dips of

173    bounding surfaces. Across all of these surveys, the thickness of 362 sets of cross strata were

174    measured, and grain size was measured for 75 of those sets in the field using a grain size card.

175    Using a geographic information system (GIS), field mapping results were merged with the

176    remote sensing measurements. Ridge-scale bounding surfaces were digitized as lines, and 1,071

177    sets of planform-exposed trough cross strata and 107 exposures of large-scale dipping strata

178    were digitized as polygons.

179

180    **Modern analogs and the transport anomaly**

181         To test how well the ridge outcrop centerlines represent original channel centerlines, a

182    metric named here as the transport anomaly, $\Theta_{TA}$, is defined for both modern rivers and the

183    exhumed channel fills.

184                        $$\Theta_{TA\text{-}CHANNEL} = \Theta_{CL\text{-}CHANNEL} - \Theta_{D\text{-}CHANNEL} \qquad (1a)$$

185                        $$\Theta_{TA\text{-}RIDGE} = \Theta_{CL\text{-}RIDGE} - \Theta_{D\text{-}RIDGE} \qquad (1b)$$

186    where $\Theta_D$ is the $0-359°$ orientation of a transport or paleotransport measurement from an

187    active dune ($\Theta_{D\text{-}CHANNEL}$; Eq. 1a) or cross set ($\Theta_{D\text{-}RIDGE}$; Eq. 1b), and $\Theta_{CL}$ is the orientation of the

188    centerline nearest to the location where $\Theta_D$ was measured (Fig. 3). Values of $\Theta_{TA}$ may be

189    positive or negative, and are calculated using the Circular Statistics Toolbox available for

190    MATLAB, which measures the shortest angular distance, positive or negative, between the two

191    directions such that no measurement exceeds 180° or is less than -180° (Berens, 2009).

192         By measuring $\Theta_{D\text{-}CHANNEL}$ from dunes in modern rivers and $\Theta_{TA\text{-}RIDGE}$ from planform-

193    exposed cross sets in the Cedar Mountain, hundreds of measurements of $\Theta_{TA}$ (Eq. 1a to b)

194    between the ancient and modern systems were compared to test whether the transport

195    anomalies for the Cedar Mountain outcrop is systematically distinct from transport anomalies

196    observed in a modern river system, which in turn is used to understand how well the

197    centerlines of formative Cedar Mountain rivers are preserved in the exhumed channel belts and

198    represented by ridge geometry (Fig. 4A and B). For example, if the mean and standard deviation

199    ($\sigma$) of $\Theta_{TA\text{-}RIDGE}$ *(Eq. 1b)* approximately equal those of $\Theta_{TA\text{-}CHANNEL}$ (Eq. 1a), then the transport

200    anomaly of the Cedar Mountain is no greater than the variability in a modern river, and is

201    consistent with ridge planform preserving the formative channel planform (Fig. 4A). If lateral

202    migration and reworking has greatly widened the belt and reduced its overall sinuosity from

203    that of the formative channels, $\sigma$ should be greater in the Cedar Mountain, as well as a more

204    random distribution of $\Theta_{TA}\text{-}_{RIDGE}$ (Fig. 4B). An example of the latter case comes from visual

205    inspection of point bar strata of the Ferron Sandstone in Wu et al., (2015, their Fig. 13; 2016,

206    their Fig. 8), which shows a general northwest-curving paleotransport trend along a northeast

207    trending exposure, making their study location a high paleo-transport anomaly zone. Note that

208    the deviation angle in Wu et al. (2016) is calculated relative to an interpreted channel-form, not

209    the exhumed channel-belt shape. Wang and Bhattacharya (2017) show a more pronounced

210    example linked to point bar growth in their Figure 10A. Durkin et al. (2018) show examples of

211    this lateral amalgamation in ancient (McMurray Formation) and the modern (Mississippi and

212    New Madrid Rivers). In a third scenario where erosion patterns have not exhumed the belt

213    evenly from all directions, neither of the prior scenarios would be observed.

214         Measurements from active dune fields in a braided reach of the North Loup River,

215    Nebraska, USA, and the meandering Trinity River, Texas, USA, were used as the comparative

216    modern systems in this analysis. The North Loup is a sand bed river that has been used as a

217    modern analog to ancient fluvial strata (Mohrig et al., 2000; Mahon and McElroy, 2018). A UAV

218    photomosaic collected by Swanson et al. (2018) was used that images a 760 m reach of the

219    river in which downstream migrating bars and dunes, as well as the channel banks, are clearly

220    identifiable. The evolution of the Trinity River has recently been studied in detail using

221    topographic and bathymetric datasets (Mason and Mohrig, 2019; in review). These datasets

222    provide 32 river km to map dune crests on the river bed, as well as dunes formed during

223    bankfull flow and currently stranded on subaerially-exposed point bar surfaces.

224         Points defining the centerlines of each river were calculated using the series of points

225    defining each channel bank. For each point on one bank, the distance to the nearest point on

226    the opposite bank is calculated and taken as a local width measurement, and a centerline point

227    is placed at the location exactly between the two points. The North Loup banks, and therefore

228    the centerline, are defined by points spaced about 1 m apart, approximately that of a

229    representative dune wavelength in the North Loup River (Swanson et al., 2018). In a GIS, the

230    brink lines of 2,871 dunes covering a 763-m reach of the river were mapped using an

231    orthorectified UAV photomosaic. The orientation of each dune was estimated by a best-fit line

232    to a series of mapped brink points. The normal to each brink line, in turn, was taken as the local

233   transport direction for that dune, $\Theta_{D\text{-}CHANNEL}$. $\Theta_{D\text{-}CHANNEL}$ was then tied to a point located at the

234   average XY coordinate of all XY coordinates defining that particular dune brink line.

235       The same process was applied to 2,190 bedforms over the imaged 32 km reach of the

236   Trinity River imaged using sonar profiles of dunes on the channel bed (dataset from Mason and

237   Mohrig, in review), as well as dunes frozen on subaerially exposed point bar surfaces formed

238   during the previous bankfull flood imaged in a 2015 lidar survey (Mason and Mohrig, 2018).

239       In the Cedar Mountain, ridge centerlines were calculated using mid-points between the

240   left- and right-hand ridge edges, and smoothed using a spline method in the MATLAB curve

241   fitting toolbox. Centerlines are ultimately defined as points spaced ~1 m apart along the

242   smoothed line. Ridge width was measured at 10 m intervals along the centerline. Values for $\Theta_{D\text{-}RIDGE}$

243   $_{RIDGE}$ are taken from field measurements of planform trough cross strata across the top surfaces

244   of the two ridges (n = 1,071), and assigned associated XY coordinates at the center of the

245   corresponding mapped set.

246

247   **RESULTS**

248   **Vertical sections**

249       The measured vertical sections from around the perimeters of each ridge were

250   composed of over 99% cross-stratified sandstones and conglomerates. Where mudstones are

251   incorporated within the vertical sections they are associated with ridge-scale erosional surfaces,

252   and have thicknesses that range from pinched out to a maximum of 0.60 m. Mudstone

253     thickness can vary over short, meters-scale distances due to erosion from overlying channel

254     elements, but the erosional surfaces these mudstones were deposited on top of are laterally

255     persistent across ridges. These persistent erosional surfaces are used to define and separate

256     individual channel-belt stories (Fig. 5A and B and 6A and B; Friend et al., 1979). Any given

257     vertical section exposes one to four stacked stories which locally vary in thickness from 0.10 m

258     to 8.60 m, with a mean of 3.10 m ± 0.22 m (the calculated standard error of the mean), median

259     of 2.80 m ± 0.27 m (the calculated standard error of the median) and σ of 2.03 m ± 0.15 m (the

260     calculated standard error of the standard deviation; n = 89; Fig. 6C). These story-bounding

261     surfaces are also exposed along the top surfaces of the ridges. Five of these surfaces have been

262     mapped across the western ridge, and four have been mapped across the eastern ridge (Fig.

263     5C).

264

265     **Sedimentary structures and architecture**

266         The most common sedimentary structures preserved in plan view and vertical

267     exposures of the Cedar Mountain Formation ridges were trough cross sets (Fig. 7A to D) with

268     median grain sizes ranging from upper-fine sand to medium pebbles (Fig. 8A to C). The mean

269     thickness of these sets was 0.12 m ± 0.005 m, with a standard deviation of 0.09 m ± 0.003 m,

270     and a coefficient of variation of 0.79 ± 0.04 ($c_v$ = σ/mean, with propagated errors; n = 350).

271     Along the top ridge surfaces where these structures were exposed and mapped in plan view

272     (Fig. 9), the dominant dip direction of these sets was clearly identified and representative of the

273     associated bedform's migration direction. The polygons outlining these planform exposed sets

274    (n = 1,071) sum to a total area of 5,019 $m^2$. Of the 1,071 sets mapped in planform, 269 were

275    identified as conglomerate, representing 25.1% of sets and 26.5% (1330 $m^2$) of total set area.

276    The remaining 802 sets were identified as sandstone, representing 74.9% of sets and 73.5%

277    (3689 $m^2$) of total set area. Larger scale thicker compound cross-sets (n = 12), with a mean of

278    1.28 m ± 0.05 m and a σ of 0.19 m ± 0.04 m measured at preserved rollovers, and are also

279    exposed in plan view (Fig. 10A-D). The locations of plan view measurements of both types of

280    sedimentary structures are shown in Figure 11. The summed planform exposure area of these

281    sets (n = 103) is 520 $m^2$, or covering 10.3% of the planform area of trough cross-sets. Within

282    individual channel belt stories, shingled trough cross-sets record transport up and down larger-

283    scale topography (Fig. 12A-D).

284        Four arrangements of cross sets were observed, and are referred to as Type A, B, C, and

285    D. Type A featured a thick basal set of compound strata scoured along its top by an upstream-

286    dipping surface, and overlain by a thinner coset composed of smaller cross sets with a mean

287    thickness and standard deviation of 0.12 m ± 0.01 m and 0.07 m ± 0.01 m (Fig. 13). The

288    upstream dips of the scour surface range from 5° to 13° (mean = 7°, n = 6). In this case,

289    orientation of small cross sets roughly parallel the dip direction of the larger cross sets.

290        Type B featured a thick basal set of compound cross-sets that change both dip and

291    thickness in the downstream direction (Fig. 14). Individual cross sets thickened by as much as

292    300% over 1.5 meters in the downstream direction (E.g. 0.08 m to 0.23 m, 0.07 m to 0.23 m,

293    and 0.06 m to 0.19 m). Correspondingly, the bounding surfaces separating these cross sets

294    shallow downstream from as steep as 26° to as shallow as 5°, and the upper bounding surface

295     transitions from being markedly erosional to conformable (Fig. 14). Similar to type A, the

296     smaller cross sets roughly parallel the dip direction of the larger cross sets. The mean thickness

297     and standard deviation of these sets at shallowly dipping sections wers 0.13 m ± 0.01 m, and

298     0.08 m ± 0.01 m, respectively.

299         Type C was also composed of compound strata, but in these cases the dip direction of

300     the smaller foresets were roughly transverse to the dip direction of the larger cross sets (Fig.

301     10D). Type C sets were identified in plan view exposures, so set thickness measurements were

302     not made. Strata composing Types A, B, and C are sandstones.

303         Type D featured no compound cross-stratification, and bounding surfaces were sub-

304     horizontal or showed local variable curvature associated with trough geometry. Type D strata

305     featured pebble conglomerates and a ~ 90° scatter of transport directions, apparent by the

306     juxtaposition of trough and dip-normal exposures (Fig. 15). The mean thickness and σ of type D

307     sets was 0.19 m ± 0.02 m and 14.7 m ± 0.02 m, and sections contain up to ten stacked sets.

308

309     **Transport anomalies**

310         Maps of transport anomalies ($\Theta_{TA}$) for the Cedar Mountain Formation and North Loup

311     River are presented in Fig. 16A to C. The associated $\Theta_{TA}$ histograms and statistical moments for

312     these systems and the Trinity River are presented in Figure 17A to D. Significantly, all datasets

313     have mean values ranging between -12° and +6°, and standard deviations ranging from 25° to

314     35°. In the North Loup, anomalies were clearly controlled by local bar topography. However,

15

315    measured transport anomalies approach the reach mean when assembled over a downstream

316    distance of ~3 bar lengths (Figs. 16C and 18A to C). In the Trinity, as expected for meandering

317    rivers, both the magnitude of the mean and standard deviation of the transport anomalies are

318    the smallest (Fig. 17D). Transport anomalies that are observed are largely due to the deflection

319    of flow obliquely down point bar surfaces (Dietrich and Smith, 1984). In the Cedar Mountain,

320    areas with concentrated high anomalies were found to be located at ridge bends (Fig. 16A to B).

321

322    **DISCUSSION**

323    **Dune, bar, belt, and overbank strata**

324        A distinction is drawn between cross sets on either side of the break in scale shown in

325    Figure 8A. The thinner-bedded trough cross strata (Fig. 7A to C) are interpreted as forming via

326    the migration of 3-D dunes with variably deep troughs (Rubin, 1987). In planform and vertical

327    sections, these are clearly distinct from larger-scale dipping strata (Fig. 10A to D), which do not

328    show the same bounding-surface curvature and, significantly, feature cross strata defined by

329    compound cross-sets (Figs. 10D and 14A and B). These larger-scale strata are interpreted as

330    accretionary river-bar deposits (Edwards et al., 1983). The population of dip direction vs.

331    centerline trend anomalies for the bar strata feature a larger spread of values and modes

332    situated far from zero (compare Fig. 17A to D against Fig. 17E to F). The range of values

333    suggests the formative bar types included point bars with primarily cross-stream accreting

334    surfaces (Fig. 10D), and free bars which can feature cross-stream-, downstream-, and upstream-

335    dipping accretion surfaces (Skelly et al., 2003). Point bar structures are best observed at the

16

336    ridge scale, where clusters of bar surfaces dip towards the convex sides of ridge bends (Fig. 11;

337    note that the western-most point bar strata define a convex-north bend, Fig. 1C). Together,

338    these dune- and bar-scale cross strata are interpreted as deposits filling river channels during

339    episodes of active sediment transport within that channel reach. The mudstones associated

340    with ridge-scale erosional surfaces are interpreted to represent sedimentation during periods

341    of channel abandonment, which indicates a system that experienced multiple avulsions and

342    channel reoccupations (Mohrig et al., 2000; Jones and Hajek, 2007; Cuevas Martínez et al.,

343    2010). However, not every reoccupation of a channel during avulsion preserves a mudstone

344    layer connected to the channel-belt boundary.

345        The four cross-stratal types observed in the Cedar Mountain ridges document the

346    interaction of the ancient dunes and bars. Type A architectures are characteristic of free bars,

347    and possess a bar-scale bounding surface, shown in red in Figure 13, separating bar lee strata

348    below from deposits of the bar stoss surface above. As such, this bounding surface preserves

349    the characteristic dip of the stoss side of the bar form. At first glance, it might seem surprising

350    to accumulate a coset of stoss-side deposits, but theory (Paola and Borgman, 1991) and a

351    recent morphodynamic bedform model (Swanson et al., 2019) show that set stacking can occur

352    even under conditions of net bypass and net erosion because of variability in dune scour

353    depths.

354        Type B architectures highlight change in compound dune strata imparted by free bar

355    migration (Fig. 14). The steepest 26° cross strata represent lee construction in a direction

356    approximately normal to average transport. The observed shallowing of bounding-surface dips

357    and thickening of sets in the downstream direction records the planform deformation of the

358    bar crest over time, where steep downstream-accreting surfaces gradually begin to accrete

359    laterally. As evident from the compound nature of these sets (Fig. 14), this bar growth is driven

360    by dune accretion in front of the bar. At the two locations where A and B type architectures are

361    adjacent (Figs. 13 and 14), the transition of the stoss scour surface to the conformable

362    bounding surface of a cross-stratum represents the delivery of sediment mined from the bar

363    stoss up and over the crest of the bar, and onto the bar lee. Taken together, these two

364    architectures preserve the processes associated with bar migration via the mining and delivery

365    of sediment by a surface veneer of smaller dunes compound to a larger free bar. These are the

366    'form sets' of Reesink et al. (2015), but observed in bar strata rather than dune strata. One

367    lateral section shows the stacking of lee strata on stoss strata, recording the aggradation and

368    migration of a bar (Fig. 19).

369         Type C compound strata define bar growth at an oblique angle to the net transport

370    direction, and define the lateral migration of a bank-attached bar form (Fig. 10D). The coarser,

371    non-compound type D architectures are interpreted as thalweg deposits (Fig. 15). Together,

372    these four architectural types describe the construction of channel-bottom topography within

373    individual channel belts via the migration and growth of dunes (both on bars and in the

374    thalweg), free bars, and point bars.

375

376    **Channel bed topography**

377        If preserved, bar form topography is interpreted to record the moment of channel

378    abandonment (Fig. 12A to D). Two lines of evidence support this and, significantly, these lines

379    of evidence are true for the Cedar Mountain Formation, as well as nearby outcrop of the

380    Jurassic Morrison Formation, indicating these may be consistent recorders of preserved bar

381    topography (Fig. 12E; location shown in Fig. 1B). First, in both cross section and map view, the

382    compound relationship between dune and bar strata informs us that entire bar forms are

383    preserved, complete with bar rollover (Figs. 12, 13, 14, and 19). Second, the stoss-positioned

384    dune sets are restricted to a surface veneer composing less than the upper 25% of the bar, with

385    the remainder composed of steeply-dipping bar-scale strata. If deflation of ridge surfaces

386    commonly broke through the surface veneer of dune sets, large bar scale strata would

387    constitute a greater percentage of sedimentary structures exposed on ridge surfaces. Instead,

388    dunes occupy approximately an order of magnitude more surface area of the outcrop. The

389    preservation of the river-bottom topography at the time of avulsion is interpreted to be the

390    consequence of a relatively rapid channel abandonment coupled with minimal erosion of the

391    channel-filling deposit by the subsequent channel reoccupation. Additionally, if this paleo-bar

392    topography can be detected using remote sensing, it could be used to better constrain flow

393    depths of ancient rivers from fluvial channel belts exposed at the surface of Mars.

394

395    **Channel planform geometry**

396        The near zero mean values and the high kurtosis of the Cedar Mountain Formation

397    paleo-transport anomaly measurements, coupled with the similarity of the standard deviations

19

398    measured in the ancient and the modern, are interpreted to indicate that the outcrop geometry

399    preserves the formative river centerline in a reliable way (Fig. 17A to D). Regions of the channel

400    belts showing concentrations of high transport anomaly measurements are associated with

401    point bar lateral accretion surfaces (Figs. 1C, 11, and 16A to B), supporting the hypothesis that

402    lateral point bar migration is a cause of high anomaly measurements (Fig. 4B). However, these

403    regions do not represent a majority of the ridge area.

404         The studied ridges are composed of several vertically-stacked channel-fills. The

405    preservation of the formative river channel centerlines through multiple re-occupations of the

406    channel is expected in fluvial settings with high rates of vertical aggradation within the channel

407    relative to lateral migration rates (Gibling, 2006; Jerolmack and Mohrig, 2007). As a result,

408    there is a general lack of centerline distortion, even though the ridge represents a channel-belt

409    complex. An interesting corollary is that the sinuous planform of the lowest river must have

410    been at least partially inherited from the antecedent surface (e.g., Lazarus and Constantine,

411    2013), as the point bars seem to have been forced by pre-existing curvature rather than driving

412    the formation of channel sinuosity.

413

414    **Channel-belt thickness**

415         Because avulsions are likely to occur when a channel bed has aggraded to a sufficient

416    level of superelevation, the thickness of a preserved channel belt, on average, is posited to

417    equal paleo-channel depth plus an aggradational component.  The thickness of a free bar

418    deposit from rollover to bottom is assumed to be a measure of local channel depth (Mohrig et

20

419    al., 2000). Bar measurements reported in Fig. 8A suggest an overall, mean channel depth of

420    1.28 ± 0.05 m. The mean belt thickness of 3.10 m ± 0.22 m (Fig. 6C) is then composed of an

421    aggradational component consisting of 1.82 m ± 0.20 m. This indicates that, on average, a

422    channel aggraded to a height of 1.53 ± 0.22 times its original depth before avulsing, creating a

423    channel belt with a total thickness of 2.42 ± 0.19 times its flow depth.

424

425    **Reconstructing river-bed kinematics**

426    *Dune motion on bars and in the thalweg*

427            The theoretical model of Paola and Borgman (1991) demonstrates that in the case of no

428    bed aggradation, bedforms with gamma-distributed heights created cross-sets with thicknesses

429    that follow an exponential distribution. Bridge and Best (1997) and Jerolmack and Mohrig

430    (2005) emphasize the importance of bed aggradation as a control on the distribution of set

431    thickness, showing that increased aggradation rates decrease the relative control of variable

432    scour depth on set thickness. Jerolmack and Mohrig (2005) showed that the coefficient of

433    variation ($c_v$) of set thicknesses decreases from a value of 0.88 in the case of no aggradation, to

434    values approaching the $c_v$ of the formative bedform heights with significant bed aggradation.

435    Coupled with this change in $c_v$ is a gradual shift from the predicted exponential distribution of

436    set thicknesses, to a gamma distribution mirroring the distribution of the formative bedform

437    heights. Significantly, this analysis has been shown to be general enough to apply to ancient

438    fluvial (Jerolmack and Mohrig, 2005) and aeolian strata (Swanson et al., 2019; Cardenas et al.,

439    2019). Therefore, the reporting and analysis of set thickness distributions should be considered

440    a significant part of any quantitative reconstruction of clastic sedimentary systems where there

441    is an interest in understanding the kinematics and transport within the ancient system.

442          When taken together, all measured dune set thicknesses (n = 350) have a $c_v$ of 0.79 ±

443    0.04. While not in the realm of true bypass ($c_v$ = 0.88 ± 0.3; Bridge 1997), the value implies

444    variable scour is the dominant control on stratification. The scour-dominated case also creates

445    laterally discontinuous sets (Jerolmack and Mohrig, 2005; Cardenas et al., 2019). This scour

446    dominance appears to be at odds with the preservation of bar 'form sets' described above (Figs.

447    13 and 14). To understand the construction of the channel belt, measurements must be

448    subdivided by environment and locally normalized. The blind application of these analyses,

449    without considering local architecture, can result in inaccurate interpretations.

450          Set-thickness analysis performed separately for bar lee sets, bar stoss sets, and thalweg

451    sets yields a different result then the bulk average. Set thicknesses were first normalized by

452    their local average value. Normalized cumulative distribution functions (CDFs) are shown in

453    Figure 20. Coefficients of variation for the normalized distributions are 0.29 ± 0.04 for lee sets,

454    0.47 ± 0.07 for stoss sets, and 0.67 ± 0.10 for thalweg sets. Although $c_v$ values as low as 0.29

455    were not examined by Jerolmack and Mohrig (2005), interpolation of their Figure 4B leads to a

456    ratio of aggradation rate to migration rate for lee sets of ~ 0.1 (climb angle from 5 to -6°). Stoss

457    sets have a ratio of ~3.2 × $10^{-2}$ (climb angle from 1° to 2°), and thalweg sets have a ratio of ~3.2

458    × $10^{-3}$ (climb angle from 0.1° to 0.2°). The lee sides of downstream-migrating barforms, where

459    the most sediment accumulation is expected (Reesink et al., 2015), have the highest ratio of

460    aggradation rate to migration rate. This significant aggradation is supported by a Kolmogorov-

461    Smirnov statistical test comparing the measurements to fitted exponential and gamma curves

462    (Fig. 20A to C). For lee sets, the exponential curve is rejected at a significance level of 0.05 (p <

463    0.001), and the gamma curve is not (p = 0.46). This is consistent with the observed stacking and

464    downstream thickening of sets in lee-type architectures (Fig. 14). Even though thalweg sets are

465    rejected as being exponentially distributed (p = 0.02) and not rejected as gamma distributed (p

466    = 0.17), the two fitted curves are more similar than in the lee and stoss cases.

467         A non-trivial amount of climb is recorded by stoss sets, given the $c_v$ of 0.47 ± 0.07, and is

468    supported by the rejection of an exponential fit (p = 0.01), and non-rejection of a gamma fit (p =

469    0.90). This conflicts with the interpretation that these sets are the record of erosive dunes

470    removing sediment from bar stoss slopes and transporting it to bar lee slopes, driving bar

471    migration. The upstream-dipping basal scour surfaces and the presence of few stacked sets

472    support the stoss interpretation (Fig. 13), but a significantly higher $c_v$ of 0.88 ± 0.3 is expected

473    (Paola and Borgman, 1991; Bridge, 1997). Bar stoss slopes are erosional most of the time, as

474    required for bar migration, but by chance, what has been preserved is the record of short-lived

475    episodes of upstream stoss-slope accretion. Using ground-penetrating radar cross sections,

476    Skelly et al. (2003) interpreted upstream accretion in recent fluvial channel-fills of the Niobrara

477    River, Nebraska, USA.

478

479    *Constraints on the time recorded by individual channel belts*

480         How is time distributed through Cedar Mountain channel belts? Backing out

481    sedimentation rates from these strata would provide information on the kinematics of the

23

482    Cedar Mountain rivers, as well as how local controls might dictate the construction of the rock

483    record (Sadler, 1981; Jerolmack and Sadler, 2015; Paola et al., 2018). The distribution of cross-

484    set thicknesses, in conjunction with assumed bedform migration rates, can provide some sense

485    of how much time is preserved within individual channel belts. Given that the accumulation of

486    dune sets at the bar lee is the process through which these bars migrated (Fig. 14), it follows

487    that

488    $$r_{lee} \,/\, c_{lee} = s_{bar} \,/\, m_{bar} \qquad (2)$$

489    where $r_{lee}$ is the aggradation rate of the bed, $c_{lee}$ is the migration rate of dunes, $r_{lee}\,/c_{lee}$ is

490    associated with a particular $c_v$ and does not vary significantly with the calculated standard error

491    (Jerolmack and Mohrig, 2005, their Fig. 4B), and $m_{bar}$ is the dune migration distance required to

492    stack cross sets up to the average bar thickness, $s_{bar}$ (Fig. 21A-C). Both sides of Equation 2 can

493    be thought of as equaling the climb angle of a compound barform. Solving for $m_{bar}$, the only

494    unknown, yields

495    $$m_{bar} = 1.28 \text{ m} \pm 0.05 \text{ m} \,/\, 10^{-1} \qquad (3)$$

496    which equals 12.8 m ± 0.5 m of bar migration. In the North Loup River, downstream-migrating

497    bars migrate ~10 m per day (Mohrig and Smith, 1996). Assuming this is a comparable rate to

498    the ancient Cedar Mountain fluvial system, the observed lee architectures are a record of only

499    ~1.28 ± 0.05 days of sedimentation. This suggests the bar strata and associated compound dune

500    strata do not record the gradual aggradation of the channel bed leading up to avulsion, but

501    rather record the higher frequency modification of the channel bed via bar migration. Instead,

502    it is hypothesized that the gradual aggradation of the channel bed is recorded in thalweg strata.

503     To test this hypothesis, Equation 2 is redefined in terms of the thalweg sets and the average

504     channel belt thickness:

505                                    $r_{thalweg} / c_{thalweg} = s_{belt} / m_{thalweg}$          (4)

506     Solving for the amount of necessary dune migration to construct the thalweg component of the

507     mean channel belt thickness,

508                                    $m_{thalweg} = 1.82 \text{ m} \pm 0.20 \text{ m} / 10^{-2.5}$          (5)

509     which equals 576 m ± 63 m of necessary dune migration (Fig. 21A-C). The average

510     instantaneous dune migration rate in the North Loup, independent of flow size or dune depth,

511     is 60 m per day (Mohrig and Smith, 1996). Although dune migration rates have not been

512     reconstructed in the ancient Cedar Mountain rivers, the North Loup is at least a sand-bed river

513     with interacting dunes and bars (Mohrig and Smith, 1996), and should provide a reasonable

514     order of magnitude estimation. Assuming steady construction at this rate, only 9.7 days ± 1.1

515     days are required to accumulate the thalweg strata reported here. This rejects the hypothesis

516     that the thalweg strata record the gradual aggradation of the channel bed over its total

517     occupation. For most rivers, occupation may last anywhere from years to thousands of years

518     (Stouthamer and Berendsen, 2001; Slingerland and Smith, 2004). It is unlikely these channels

519     were only occupied for 9.7 days. Instead, these strata may only represent the aggradation that

520     occurred during the final episode of sedimentation that preceded avulsion and channel

521     abandonment. This episode is likely to coincide with the final flood prior to avulsion.

522     This result suggests that the channel was instead in a state of bypass for most of its

523     occupation. Had channel abandonment not prevented it, the aggradation recorded by each

524     channel belt would likely have been completely reworked. The complete reworking of the

525     channel bed has, in fact, been observed in modern net-depositional rivers (Nittrouer et al.,

526     2011a). This also suggests that floodplain deposits might more completely record successive

527     episodes of flood-stage deposition than channel fills, as presumably an episode of floodplain

528     deposition is not immediately followed by reworking and removal.

529     Alternatively, the aggradation of the thalweg may represent 9.7 days ± 1.1 days of

530     accumulation spread intermittently over a longer time period, as much as 335 days ± 55 days if

531     a commonly assumed intermittency, the fraction of time per year spent at bankfull flow, of 0.02

532     is used (Parker et al., 1998). However, there is no stratigraphic indicator that the accumulations

533     within a single channel belt are the product of multiple small depositional episodes separated

534     by hiatuses (e.g., aeolian reworking, Carling and Leclair, 2019). Accounting for this would

535     require the assumptions that there was no reworking of the channel bed between depositional

536     episodes (e.g., significant channel bed reworking shown in Nittrouer et al., 2011b; Shaw and

537     Mohrig, 2014; Mason and Mohrig, in review), and that dune migration perfectly resumed right

538     where the previous episode of sedimentation left off. It is more likely that such gradual bed

539     aggradation would produce a scour-and-fill style architecture and a higher $c_v$ of 0.88 ± 0.03

540     (e.g., Paola and Borgman, 1991; Bridge and Best, 1997; Bridge, 1997; Jerolmack and Mohrig,

541     2005; Cardenas et al., 2019). This is not to say that an intermittency of 0.02 is impossible or

542     unlikely for these rivers; it simply cannot be constrained using only the accumulation rates of

543     the final episode of sedimentation along a continuously reworked channel bed.

544

545    **Channel vs. floodplain accumulation**

546          On average, a Cedar Mountain river constructed a channel belt of thickness 2.42 ± 0.19

547    times its original depth before avulsing, based on the ratio of channel-belt thickness to flow

548    depth. This is a slightly higher value than reported for the Guadalope-Matarranya ancient fluvial

549    system in Spain (Mohrig et al., 2000). Mohrig et al. (2000) report an average superelevation of

550    0.61 times flow depth before avulsion occurs from the Guadalope-Matarranya ancient fluvial

551    system in Spain.  In their Table 2, the sum of the incision depth and superelevation is equal to

552    the belt thickness measurements presented here. Dividing their belt thickness measurements

553    by local channel depth yields a smaller amount of average total aggradation before avulsion of

554    1.84 times channel depth. This suggests the ancient Cedar Mountain fluvial system was able to

555    support more channel bed aggradation before avulsing, suggesting a higher ratio of floodplain

556    aggradation rate to channel bed aggradation rate than in the Guadalope-Matarranya system

557    (Mohrig, 2000). This is consistent with the punctuated style of channel-bed aggradation

558    interpreted from the time needed to accumulate the thalweg deposits. In this scenario, most of

559    the sediment rapidly accumulated in the channel during a bankfull episode is reworked to zero

560    net accumulation shortly after. This is coupled with a contrasting steady, gradual levee and

561    floodplain aggradation, assuming these overbank environments are less likely to be reworked

562    to a state of zero net aggradation between floods. That is, in order for the channel bed to reach

563    the threshold superelevation to avulse, more channel bed aggradation was required during the

564    final depositional episode to catch up with steadily aggraded levees and floodplain (Fig. 22).

565

566     **Comparison to other fluvial systems and strata**

567          The Cedar Mountain rivers may represent an end-member behavior, opposite to a

568     channel bed undergoing steady aggradation and frequent but partial reworking, such as

569     described in Carling and Leclair (2019). In that work, an entrenched reach of the Luni River,

570     India, is shown to repeatedly partially rework and scour its bed during floods, but optically-

571     stimulated luminescence age dates show a trend of increasing age with depth, topping out at

572     1.9 ka (their Table 2). This is a record of slow, steady bed aggradation over the last 1.9 ka with

573     each flood, rather than the rapid sedimentation frozen in place following channel abandonment

574     in the Cedar Mountain. As another counter example, the Cretaceous McMurray Formation,

575     seismically imaged in the subsurface of Alberta, Canada, has recorded multiple channel widths

576     of lateral accretion and meander bend cutoffs within the channel belt, which has been

577     demonstrated to drive the erosion of formative channel elements (Durkin et al., 2018), as well

578     as alter the geometry of the channel belt relative to the formative river channel (their Fig. 12).

579     Durkin et al. (2018) estimate a minimum of 4,000 years to produce the observed reworking.

580     Indeed, heavily reworked examples of fluvial strata exist, and the given examples require

581     contrasting processes to the Cedar Mountain ancient fluvial system. Their accumulations

582     represent longer time scales, as well. River avulsions into local topographic lows formed

583     between abandoned, raised channel beds has been shown to locally increase sedimentation

584     rates and prevent reworking, leading to well-preserved channel belts at the km scale (Swartz et

585    al., in prep; Cardenas et al., in prep). Given the total extent of the Cedar Mountain ridges and

586    the relatively rapid sedimentation rates, this process may also be at work here.

587

588    **CONCLUSIONS**

589         The belts of the Cedar Mountain Formation represent rivers that avulsed frequently

590    enough to limit lateral migration of any single channel. This is similar to the kinematics of

591    megafan channels, which can avulse on the scale of years and don't generally migrate to any

592    significant degree (e.g., Horton and DeCelles, 2001; Chakraborty et al., 2010), although the time

593    required to aggrade the thalwegs required an environment capable of sustaining sediment-

594    transporting flows for at least 9 days continuously. In addition to capturing what may be a

595    common bankfull event, 41% of any belt represents a day's worth of free bar migration. This

596    further supports the 'strange ordinariness' (Paola et al., 2018) of much of the stratigraphic

597    record, as well as the frequently observed phenomenon that most of the time represented in a

598    stratigraphic section is recorded by erosional or hiatal surfaces between punctuated

599    depositional episodes, rather than the accumulations (Sadler, 1981; Miall, 2015; Sadler and

600    Jerolmack, 2015; Paola et al., 2018). This methodology provides a framework for the

601    quantitative analysis of other modern and ancient fluvial systems. The distribution of dune

602    cross-set thicknesses within the context of bar topography is the dataset at the root of this

603    analysis, and should be emphasized in future field campaigns, including the exhumed fluvial

604    strata that will be examined during the 2020 Mars rover mission to Jezero crater (Goudge et al.,

605    2018).

606

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615

## DATA AVAILABILITY

617    The data that support the findings of this study are available from the corresponding

618    author upon reasonable request.

619

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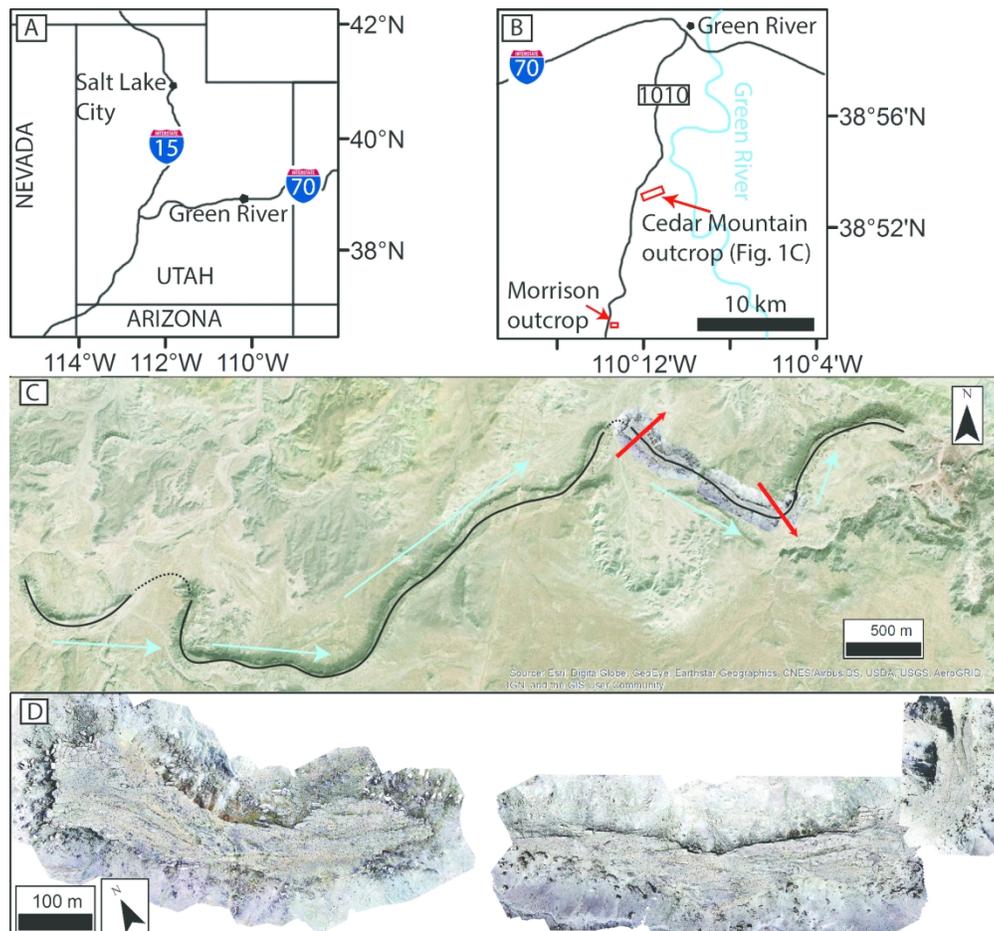
Figure 1 – (A) Map of Utah showing major highways, Salt Lake City and Green River. (B) Zoom in near Green River, showing the location of the town as well as the studied ridges of the Cedar Mountain and the Morrison Formation south of town along 1010. (C) Zoomed out view showing the Cedar Mountain ridges beyond the study area. Black line maps out a ridge centerline for several km, with interpreted dashed segments bridging erosional discontinuities. Teal arrows show the general direction of paleoflow. Red arrows mark the two major bends bounding the studied part of the ridge. The arrows point away from the center of curvature, and match with the general dip directions of local dipping bar strata. (D) Drone ortho-images of the studied eastern and western ridges of the Cedar Mountain Formation ridges. The photomosaics are rotated slightly to fit the panel, but are correctly co-located.

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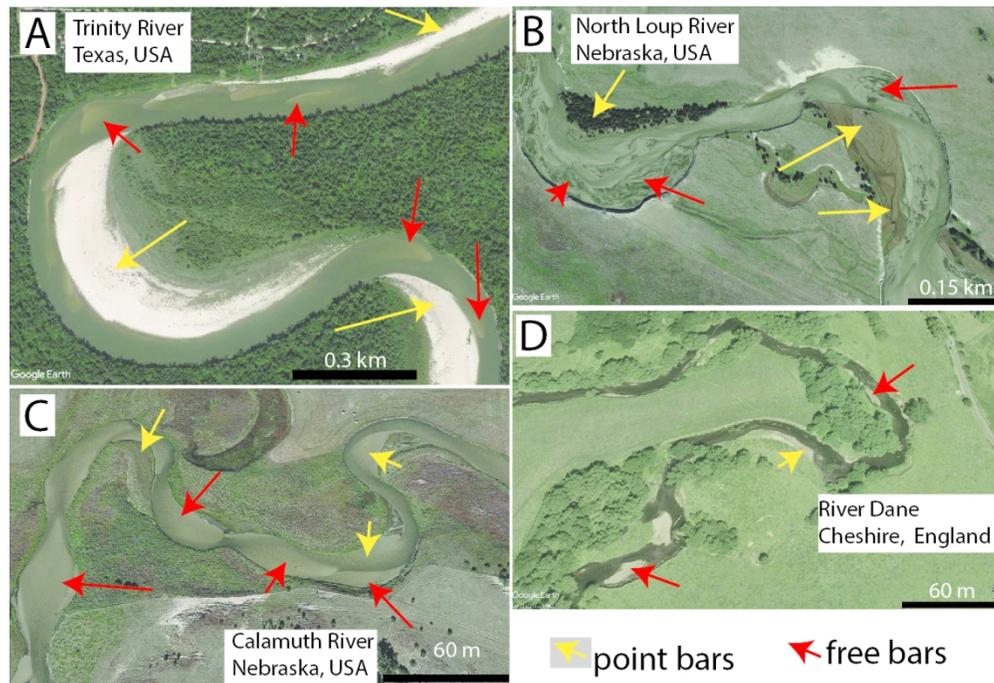
Figure 2 – Free bars (red arrows) and point bars (yellow arrows) commonly coexist in rivers, both in straight reaches and bends. (A) Trinity River, Texas, USA. Image centered at 30.134° N, -94.815° E. (B) North Loup River, Nebraska, USA. Image centered at 42.019° N, -100.098° E. (C) Calamuth River, Nebraska, USA. Image centered at 42.083° N, -99.649° E. (D) River Dane, Cheshire, England. Image centered at 53.183° N, -2.259° E.

124x85mm (300 x 300 DPI)
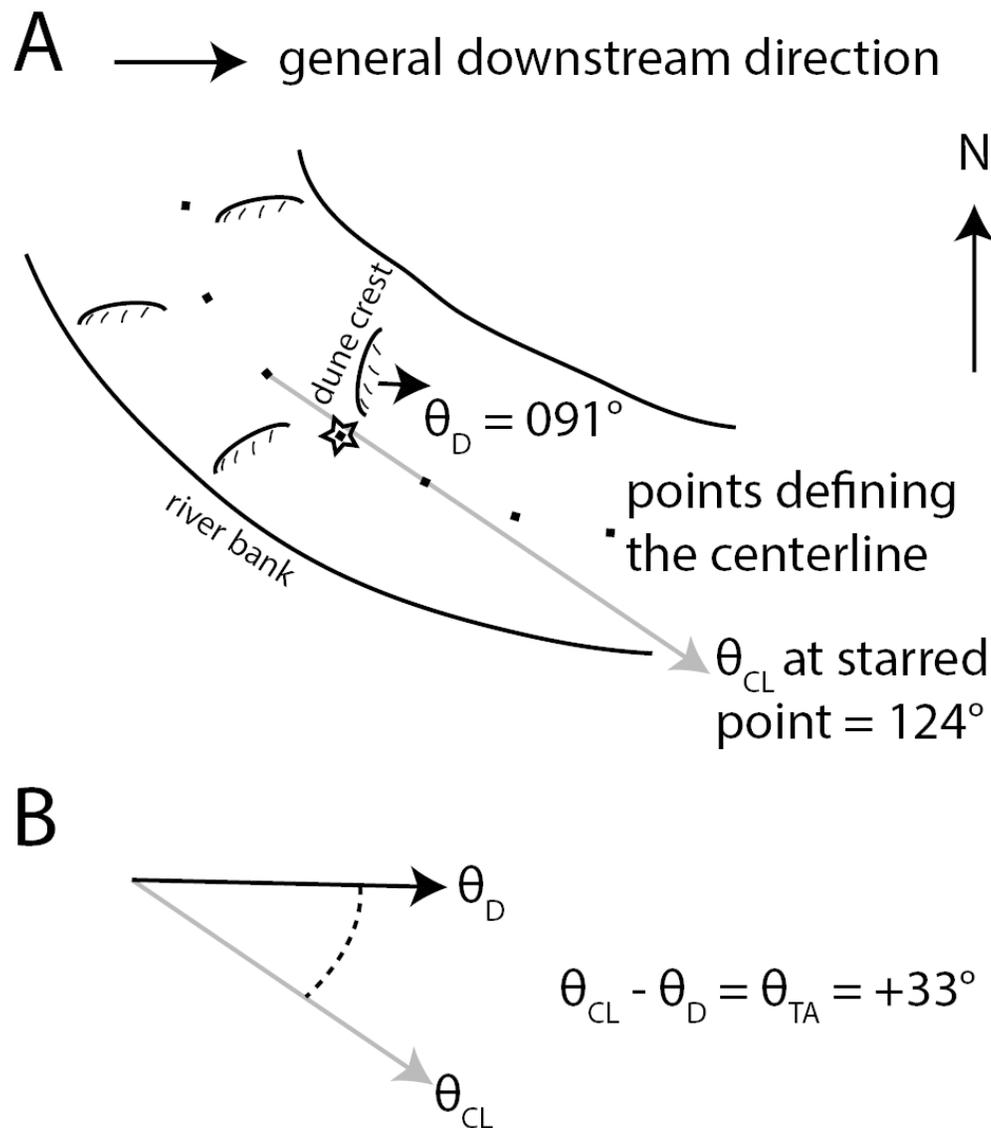
Figure 3 – (A) Diagram defining the components of the transport anomaly, ΘTA, for a modern river channel. A measurement of transport direction, ΘD, is made from the orientation of a dune crest (short black arrow; 091°). The centerline point closest to the measurement of ΘD is starred. The orientation of the starred centerline point, ΘCL, is defined as the trend of the ray originating at the adjacent upstream point and passing through the adjacent downstream point (gray arrow; 124°). (B) The transport anomaly, ΘTA, is defined as ΘCL – ΘD. It may be positive or negative, and is bound between -180° and positive 180°. In this scenario, ΘTA = 124° - 091° = 33°.

80x91mm (300 x 300 DPI)
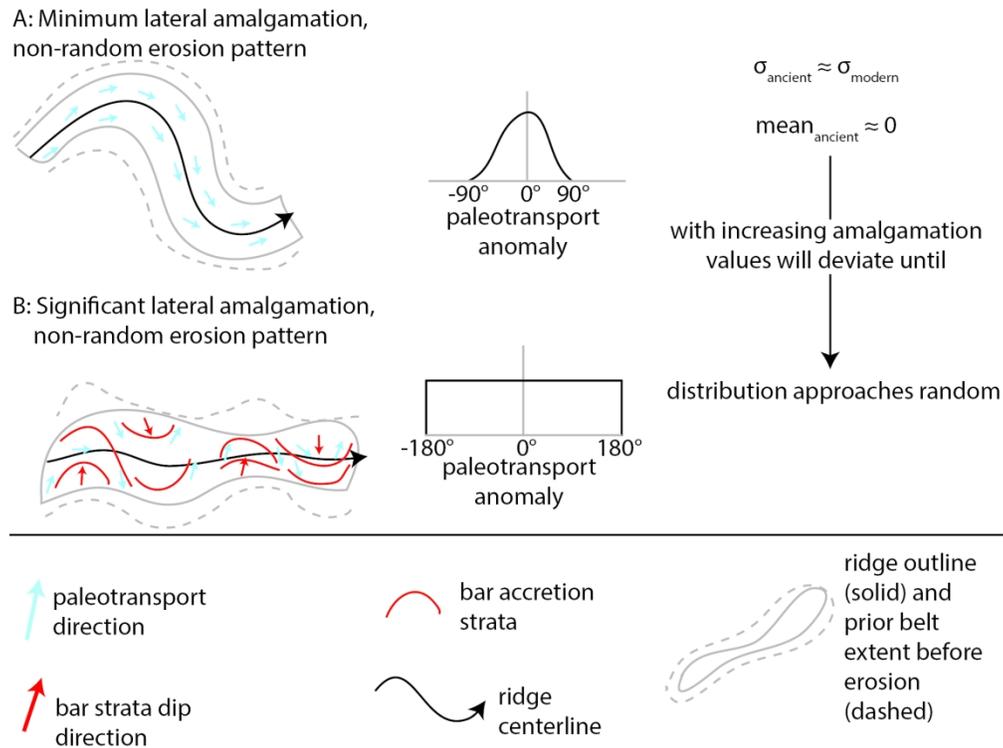
Figure 4 – Hypothesized scenarios guiding interpretations of paleotransport anomaly results. Schematic diagrams are on the left, the distribution of paleotransport anomaly measurements are in the middle, and relevant statistical moments are on the right. Standard deviation is shown by σ. Legend is at the bottom. (A) The ridge centerline represents well the formative channel centerline. With increasing lateral amalgamation, results will instead approach the scenario in panel B. (B) Lateral amalgamation of the channel-belt separates any formative channel centerline from the ridge centerline. Laterally-accreting bar strata are preserved. A random exhumation pattern not following the edges of the channel belt is unlikely to show any of these patterns.

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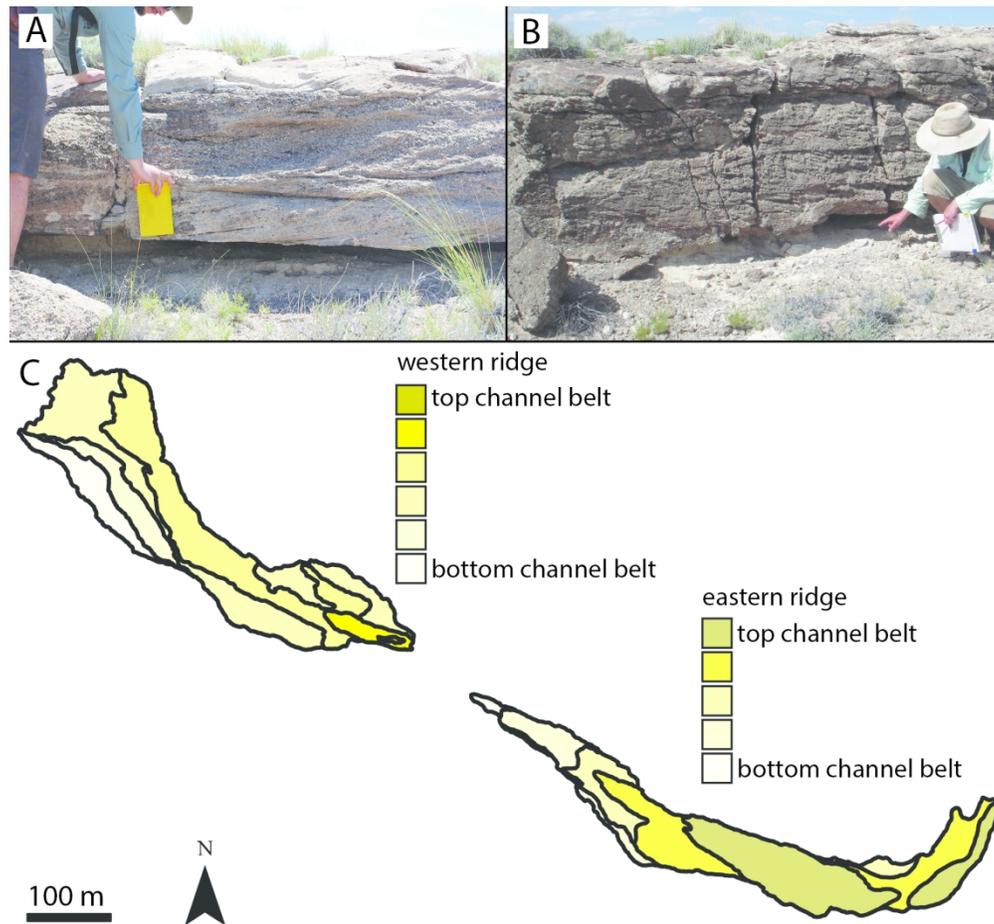
Figure 5 – (A-B) Beneath the yellow notebook (A) and at where the finger is pointing (B) are erosional surfaces associated with friable, recessed mudstones separating coarse-grained, cross-bedded packages. These erosional surfaces are interpreted to represent the contacts between stacked channel belts. Not all mudstones are associated with these surfaces. (C) Geologic maps showing the stacking patterns of channel-belts exposed at the surface of both ridges. Darker yellows indicate higher stratigraphic positions. There is no attempt to correlate individual channel belts between ridges.

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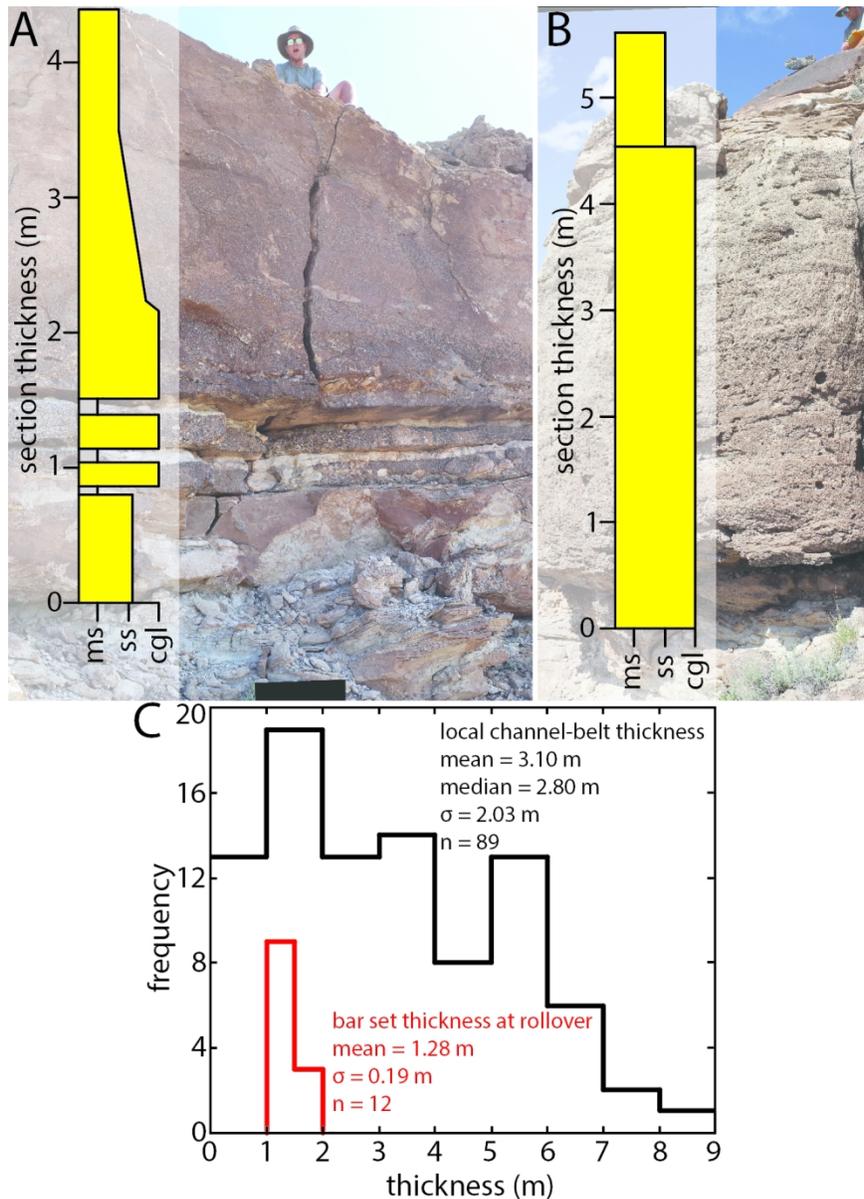
Figure 6 – (A) Vertical section showing story-bounding surfaces and associated mudstones. Stories in this section are of average to below-average thickness. (B) Two stories bounded by an erosional surface with no associated mudstone. The bottom story is above average thickness. (C) Histogram of local channel-belt (story) thicknesses measured from vertical sections, and the mean thickness of a bar set at the rollover (red line), which is used as a proxy for channel depth. The difference between channel depth and channel-belt thickness is due to aggradation of the channel bed.

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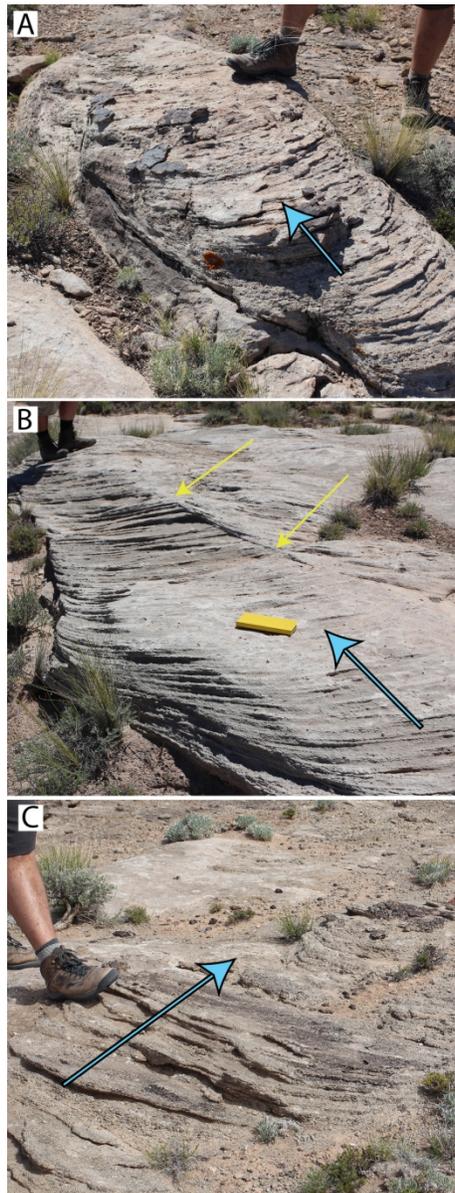
Figure 7 - Photos of cross strata exposed in planview along upper ridge surfaces. Teal arrows show the mean dip directions of cross strata. (A) This 3-D outcrop of a sandstone set shows the relationship between planform-exposed cross strata and vertically exposed cross strata. Boots for scale. (B) A 3-D exposure of a sandstone set. Yellow arrows point to the bounding surface separating two sets. This sort of amalgamation is common. Yellow field book for scale. (C) A conglomerate set showing minimal planform curvature. Boot for scale.

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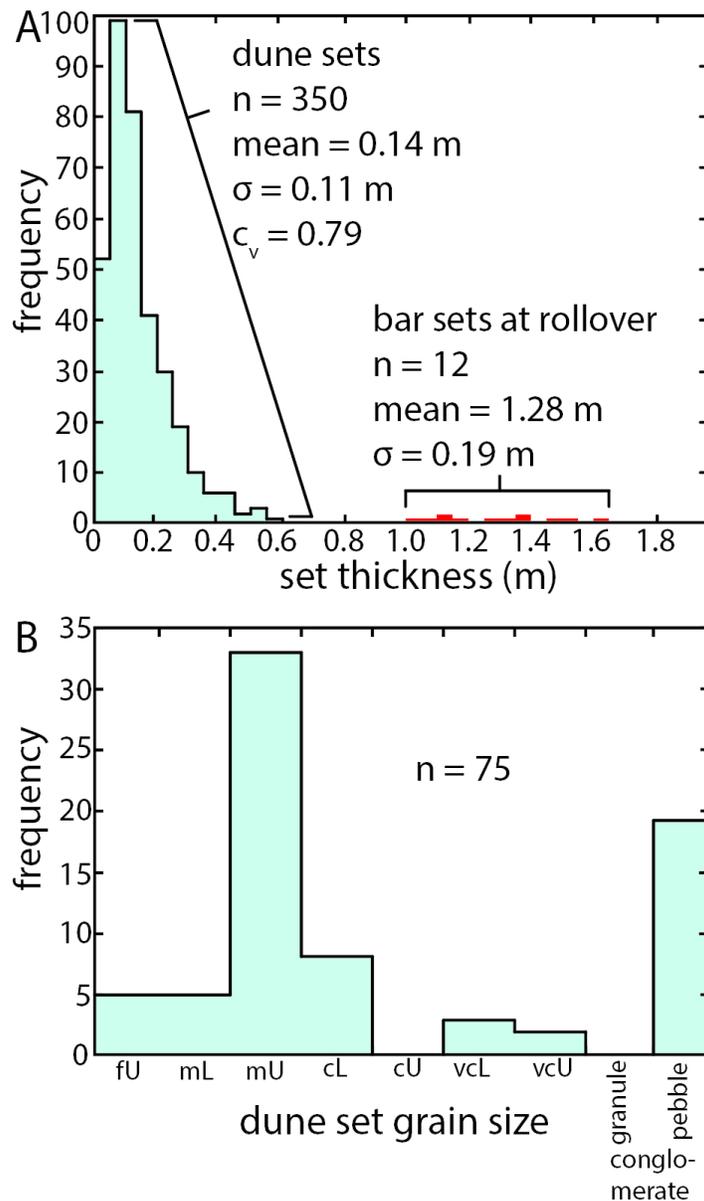
Figure 8 – (A) Teal histogram shows the distribution of the entire population of dune cross-set thicknesses, with statistical moments and the coefficient of variation (cv). The distribution of bar set thicknesses at the rollover is shown in the red histogram. When measuring bar sets at the rollover, there is a break in the distribution of dune sets and bar sets. (B) Distribution of grain sizes of dune cross-sets. Measurements made in the field with a grain size card and hand lens.

80x134mm (300 x 300 DPI)
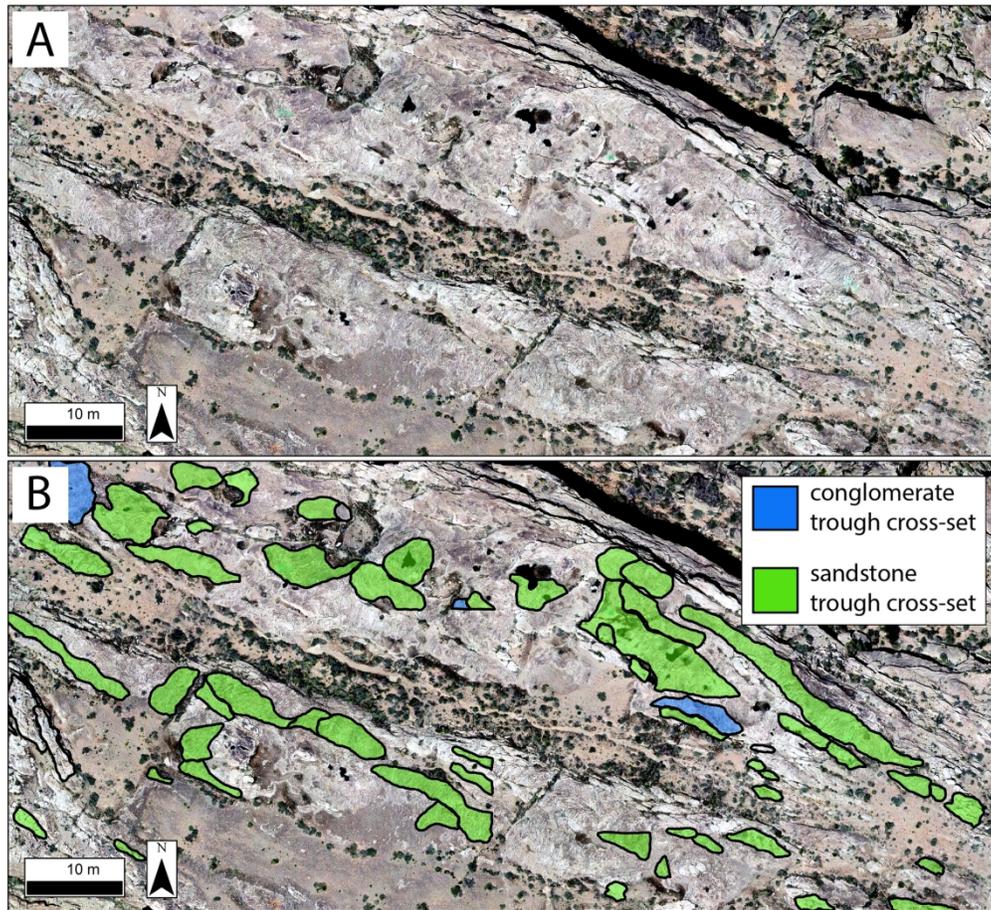
Figure 9 – (A) Example of uncrewed aerial vehicle photomosaics used as field base maps. (B) Digitized field map showing planform-exposed sets of cross strata outlined and filled in with green (sandstone) or blue (pebble conglomerate).

170x155mm (300 x 300 DPI)
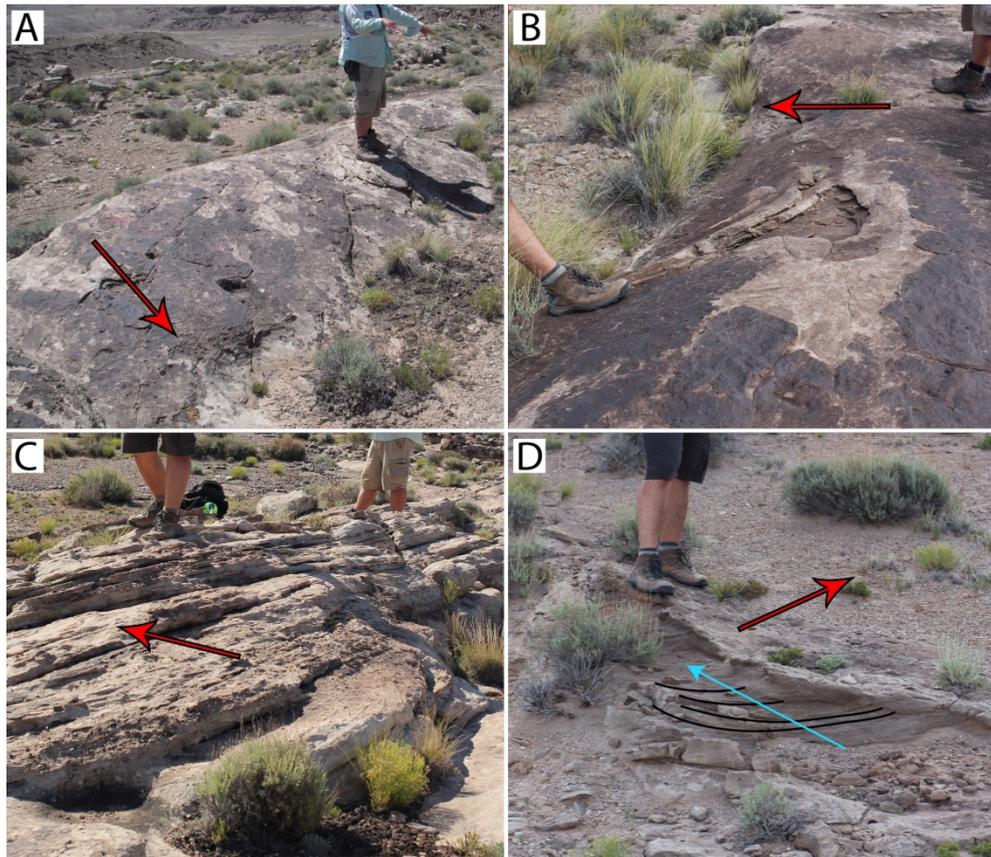
Figure 10 - Examples of larger-scale accretion strata in the field. Red arrows show the dip direction of the strata in each panel. (A) A typical example exposed for several meters. A lack of exposed bounding surfaces on this topographic surface suggests the topography itself represents a bounding surface. (B) Beneath the arrow, erosion exposes internal stratification parallel to the surface. (C) A 3-D outcrop of larger-scale dipping strata composed of smaller-scale stratification exposed by erosion. (D) Compound cross strata with a larger-scale accretion surface (red arrow) dipping obliquely to a smaller-scale dune set (teal arrow). A few dune cross strata are mapped in black lines.

170x146mm (300 x 300 DPI)
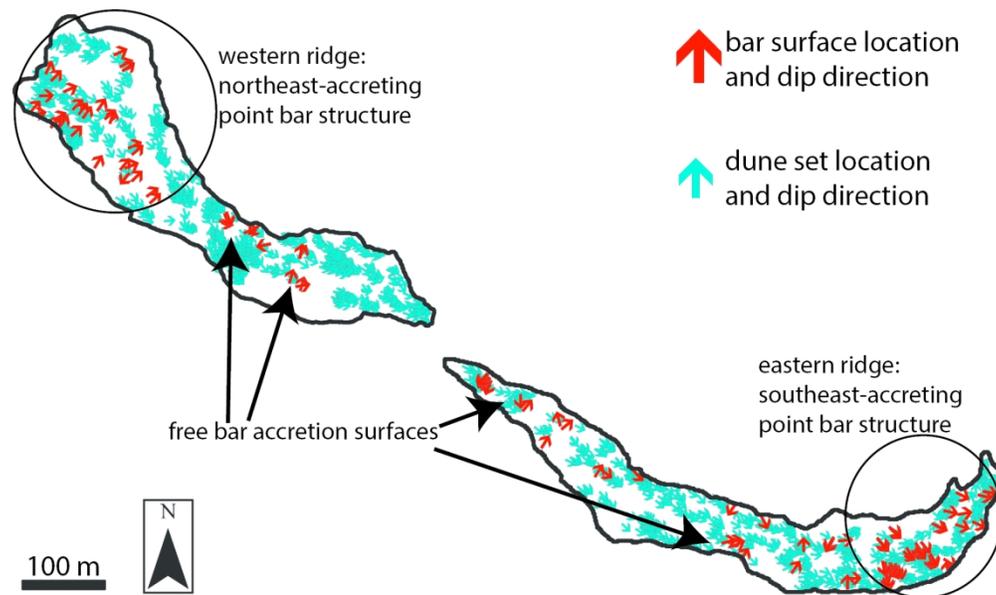
Figure 11 – Planform maps outlining the top surfaces of both studied ridges. Teal arrows show the locations of paleotransport measurements and point towards the measured direction. Red arrows show dip directions of large-scale dipping strata. Two locations with clusters of similarly-dipping bar accretion surfaces following ridge curvature are interpreted to represent point bars. The northeast-accreting point bar structure of the western ridge corresponds with a larger-scale ridge curvature beyond the extent of the study area (Fig. 1C). Bar accretion surfaces not clearly associated with a point bar are interpreted as free-bar accretion surfaces.

170x100mm (300 x 300 DPI)

Figure 12 – The preservation of bar topography on upper-ridge surfaces. (A) Fisheye view of a sandstone mound rising towards the downstream direction (left to right), with a surface defined by shingled cross-sets climbing with topography. Interpreted as the stoss surface of a downstream-migrating barform. Tape measure for scale in foreground, arrow pointing to person in background. (B) Interpretation of panel A. (C) Downstream end of sandstone mound featuring cross sets and topography falling in the downstream direction. Interpreted as the lee slope of a downstream-migrating barform. Person for scale. (D) Interpretation of panel C. (E) Preserved point bar topography in the nearby Morrison Formation (Fig. 1B). In the foreground, the vertical exposure cuts through dune and bar stratigraphy, and clearly does not follow the paleo-topography of the bar. Contrast this with the steeper surface just beyond the vertical cut filling most of the background, which exposes shingled sets of planview-exposed dune cross sets. This surface approximates the original topography of the bar.

170x203mm (300 x 300 DPI)

Figure 12 – The preservation of bar topography on upper-ridge surfaces. (A) Fisheye view of a sandstone mound rising towards the downstream direction (left to right), with a surface defined by shingled cross-sets climbing with topography. Interpreted as the stoss surface of a downstream-migrating barform. Tape measure for scale in foreground, arrow pointing to person in background. (B) Interpretation of panel A. (C) Downstream end of sandstone mound featuring cross sets and topography falling in the downstream direction. Interpreted as the lee slope of a downstream-migrating barform. Person for scale. (D) Interpretation of panel C. (E) Preserved point bar topography in the nearby Morrison Formation (Fig. 1B). In the foreground, the vertical exposure cuts through dune and bar stratigraphy, and clearly does not follow the paleo-topography of the bar. Contrast this with the steeper surface just beyond the vertical cut filling most of the background, which exposes shingled sets of planview-exposed dune cross sets. This surface approximates the original topography of the bar.

170x114mm (300 x 300 DPI)
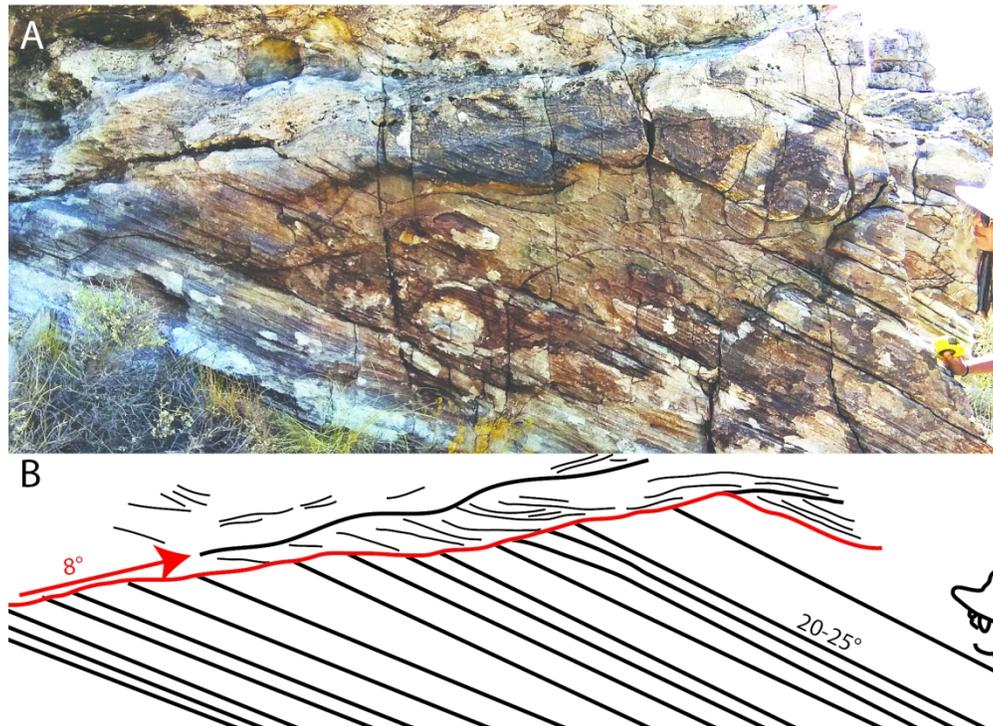
Figure 13 - Cross-sectional view of preserved strata from the stoss side of a bar, with an interpreted transition to the lee side. Flow was from left to right. In the sketch, the truncation surface separating lee-side (underneath) from stoss-side (above) strata is marked as a red line. This line is interpreted to preserve the upstream-dipping, stoss-side slope for the bar of 8°. Sets of dune strata are marked by thick black lines and dune cross strata by thinner black lines. Deposition on the stoss-side of the bar is best shown by the upward climbing dune sets at the left-side side of the panel. Beneath these stoss-side strata are steeply dipping beds recording the prior lee face of the bar. Towards the right-hand side of the photo, the truncation surface separating stoss and lee strata becomes conformable. The package of thick dune sets at this position is interpreted to be associated with higher sedimentation rates on the lee face of the bar. The observed spatial change in the dip angles of dune sets and the truncation surface document change in bar shape through time.

170x124mm (300 x 300 DPI)
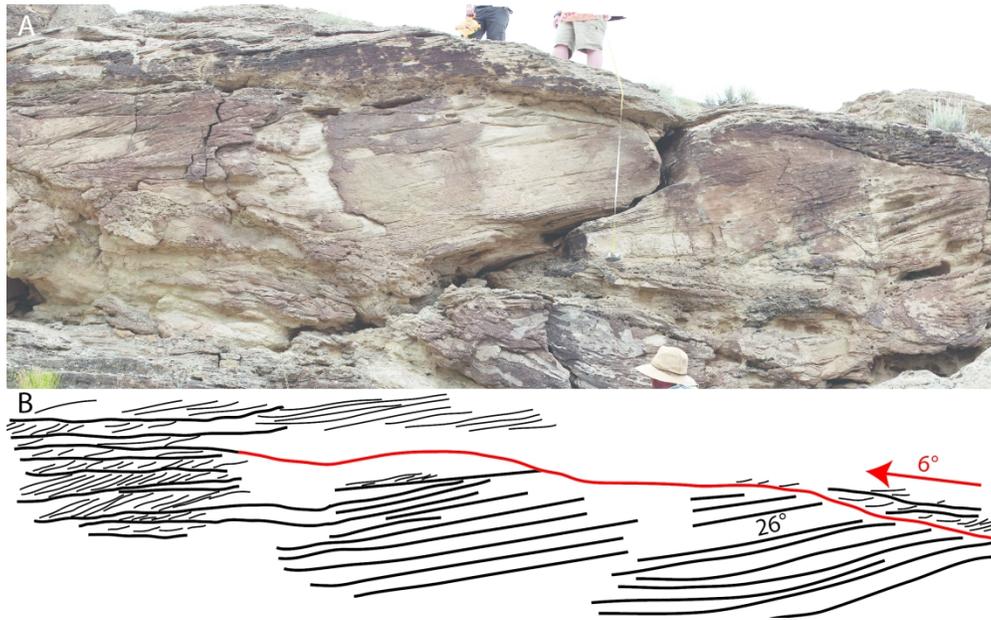
Figure 14 – Cross-sectional view of preserved strata from the lee and stoss sides of a bar form. Flow was from right to left. In the sketch, the truncation surface separating lee-side (underneath) from stoss-side (above) strata is marked as a red line. This line is interpreted to preserve the upstream-dipping, stoss-side slope for the bar of 6°. Sets of dune strata are marked by thick black lines and dune cross strata by thinner black lines. Deposition on the stoss-side of the bar is best shown by the upward climbing dune sets at the right-side side of the panel. Beneath these stoss-side strata are steeply dipping beds recording the prior lee face of the bar. At the left-hand side of the exposure it is particularly clear that the bar form is composed of compound strata in which the bar shape evolved through deposition of dune cross-beds. At this same location the truncation surface separating stoss and lee strata becomes conformable. The package of thick dune sets at this position is interpreted to be associated with higher sedimentation rates on the lee face of the bar. The observed spatial change in the dip angles of dune sets and the truncation surface document change in bar shape through time.

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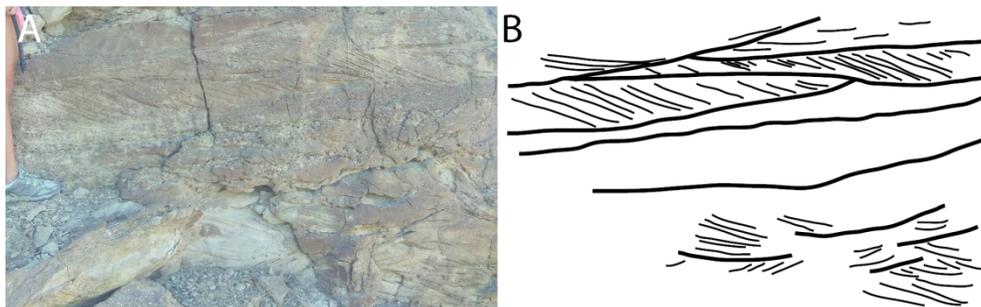
Figure 15 – Cross-sectional view of strata featuring a ±90° spread in transport direction, conglomerates, and a lack of bar architecture. This type of architecture is interpreted as a thalweg environment due to the coarser grains driven by higher velocity flow, and a larger spread in transport driven by changes in steering due to bar growth. The sketch shows bounding surfaces and cross strata.
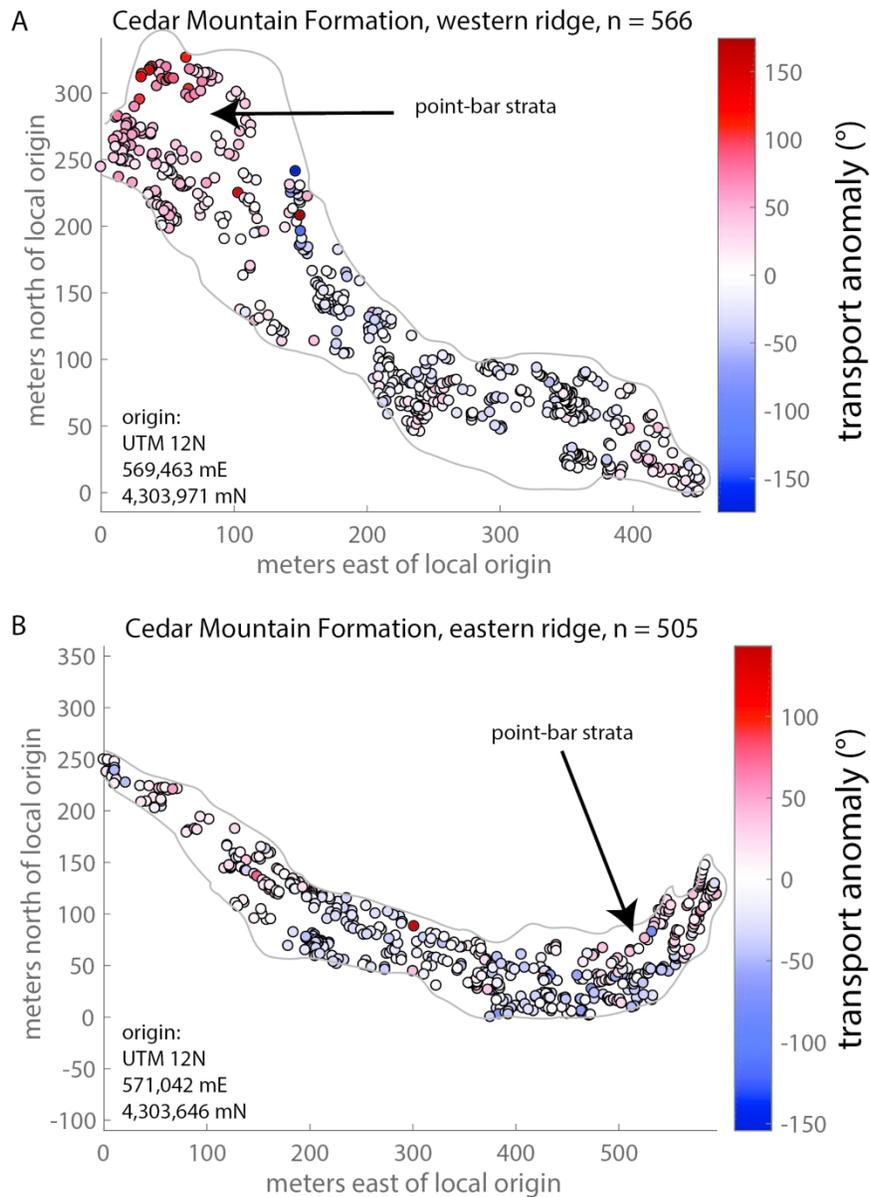
170x51mm (300 x 300 DPI)

Figure 16 – Transport anomaly maps of the western ridge of the Cedar Mountain Formation (A), the eastern ridge (B), and the North Loup River (C). X and Y coordinates are relative to a different local datum in each map, shown in the bottom left corner of each panel. Circles show the location of paleotransport or modern transport direction measurements. The color at each point represents the paleotransport or transport anomaly (Fig. 3A-B). Gray lines represent ridge outlines and the banks of the North Loup River. Black arrows in panels A and B point to regions recording point bar accretion, and are associated with relatively high anomaly values, particularly in the western ridge (Figs. 1C and 11).

170x230mm (300 x 300 DPI)

Figure 16 – Transport anomaly maps of the western ridge of the Cedar Mountain Formation (A), the eastern ridge (B), and the North Loup River (C). X and Y coordinates are relative to a different local datum in each map, shown in the bottom left corner of each panel. Circles show the location of paleotransport or modern transport direction measurements. The color at each point represents the paleotransport or transport anomaly (Fig. 3A-B). Gray lines represent ridge outlines and the banks of the North Loup River. Black arrows in panels A and B point to regions recording point bar accretion, and are associated with relatively high anomaly values, particularly in the western ridge (Figs. 1C and 11).

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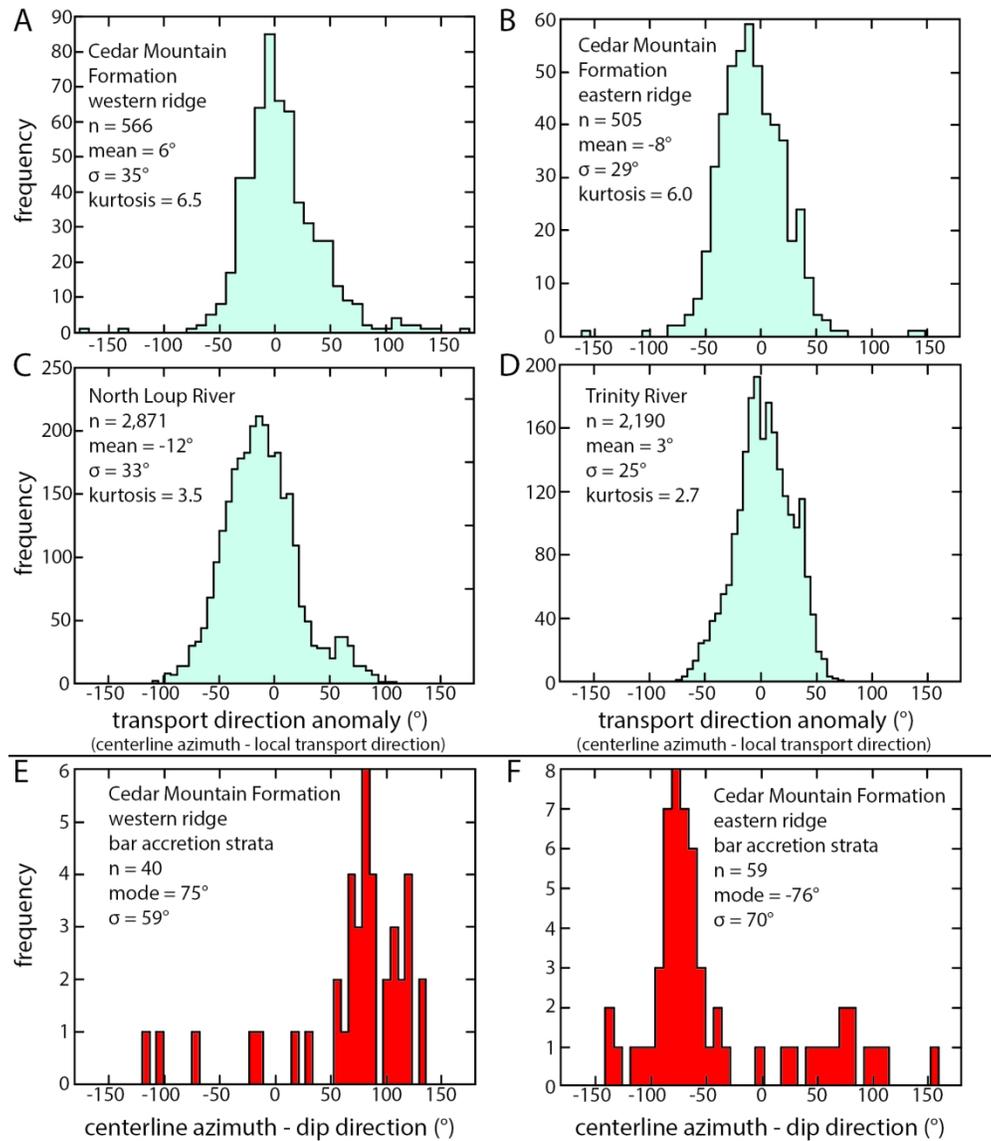
Figure 17 – Histograms showing the distribution of paleotransport/transport anomalies of the western (A) and eastern (B) ridges of the Cedar Mountain Formation, and the modern North Loup River (C) and Trinity River (D). The number of measurements, mean, and standard deviation are reported in each panel. Note the similarity in mean and standard deviation between the ancient and modern datasets. Histograms (D) and (E) show the difference between dip directions of bar accretion strata exposed along the upper surfaces of ridges and the centerline. Both histograms show a wide distribution of values with peaks approaching perpendicular to the centerline trend.

170x193mm (300 x 300 DPI)
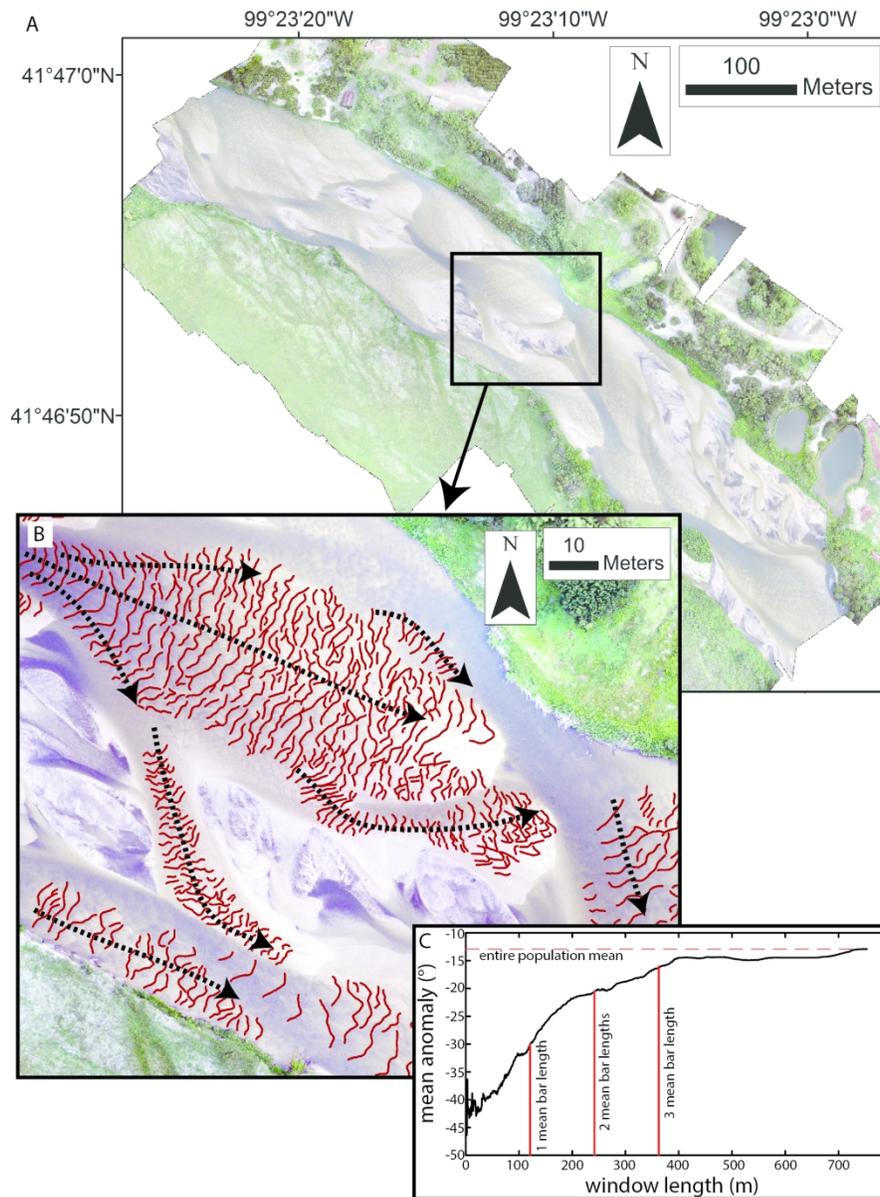
Figure 18 – (A) Drone photomosaic of the North Loup River near Taylor, Nebraska, USA (Swanson et al., 2018). Brighter tan colors within the channel are subaqueous and represent higher portions of downstream-migrating bars beneath shallow water. Darker reaches of the channel represent deeper water. Mixed white and black areas with no crestlines mapped are subaerially exposed bar tops that are not currently undergoing fluvial transport. The location of panel B is shown in the black box. (B) Zoom in showing dune crestlines (short red lines) interpreted as perpendicular to dune transport direction. Black dashed arrows show general trends in local transport directions due to the steering of flow around bars. (C) Window length vs. the mean transport anomaly within the window. As the window length approaches that of the ~ 3 barforms, the sampled mean approaches the mean of the entire dataset. Changes in curvature of this line happen near multiples of mean bar length (red vertical lines), supporting topographic steering as the source of the transport anomaly.
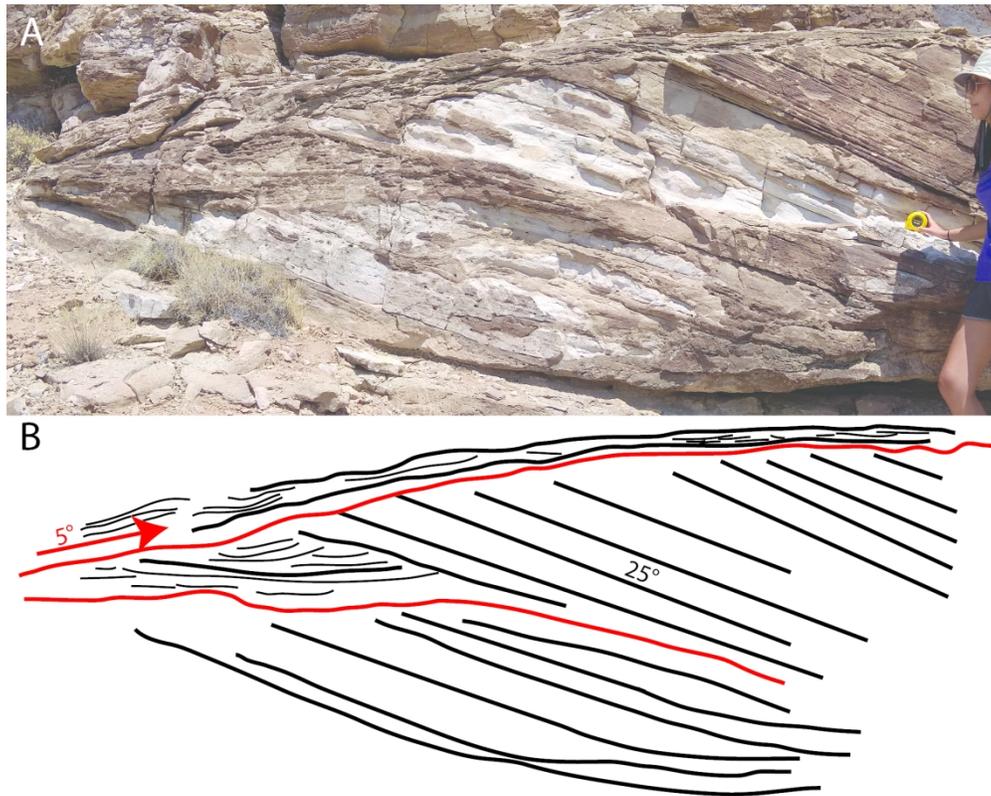
170x230mm (300 x 300 DPI)

Figure 19 – Cross-sectional view showing the internal structure of a barform. Flow was from left to right. In the sketch, the bottom-most red line marks a surface at which lower bar-scale cross strata are truncated. This surface becomes conformable with the bar strata downstream, and features a veneer of smaller-scale sets above it. This arrangement is consistent with a bar stoss-to-lee transition. The upper-most red line defines a 5° upstream-dipping surface truncating 25° downstream dipping bar-scale strata. Above this surface are two dune-scale sets. The architecture arranged around this red surface is consistent with a bar stoss environment. From bottom to top, the transition from stoss-to-lee architecture to lee architecture, all within the same barform, records the forward migration and aggradation of the barform.
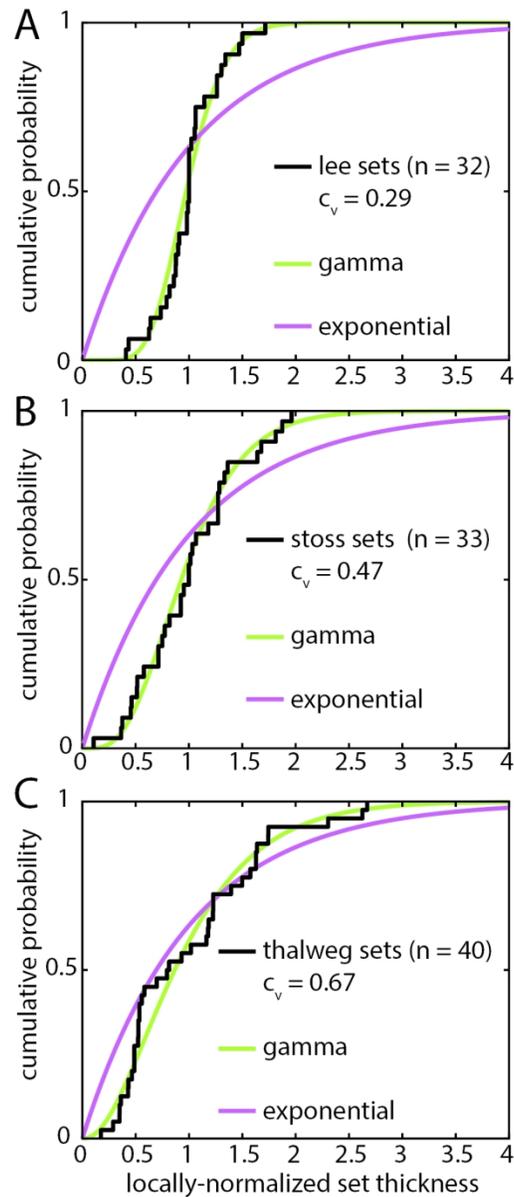
170x136mm (300 x 300 DPI)

Figure 20 – (A) Cumulative distribution function (CDF) showing the mean-normalized distribution of set thickness located within bar lee environments (Fig. 14). (B) CDF of set thickness within bar stoss environments (Fig. 13). (C) CDF of thalweg set thickness (Fig. 15). The best-fit gamma and exponential curves are shown for each distribution. In all cases, the exponential fits are rejected using a Kalmogorov-Smirnov test at a significance level of .05, and the gamma fits are not. This suggests all architectures required a significant rate of bed aggradation relative to the rate of dune migration, although the similarity of the two curves for thalweg sets indicates the ratio of bed aggradation to dune migration was the lowest of the three environments (Paola and Borgman, 1991; Jerolmack and Mohrig, 2005).
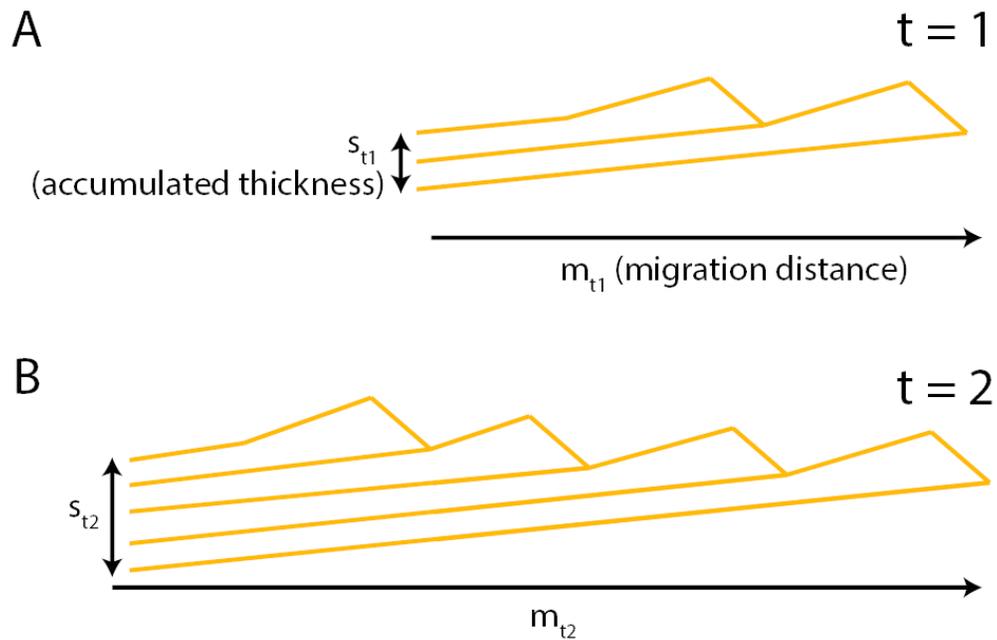
80x188mm (300 x 300 DPI)

Figure 21 – A diagram explaining the calculation of accumulation time of dune cosets with thickness s, assuming continuous transport (Eqs. 2-5). Dune cosets may be located on bars or the thalweg. Dune migration is towards the right. (A) At t = 1, dunes with a given ratio of aggradation rate (r) to migration rate (c) migrate a distance of mt1, leaving a deposit of thickness st1. (B) At t = 2, this process has continued with the same r and c. Dunes have migrated a farther distance, mt2, accumulating a thicker coset package, st2.
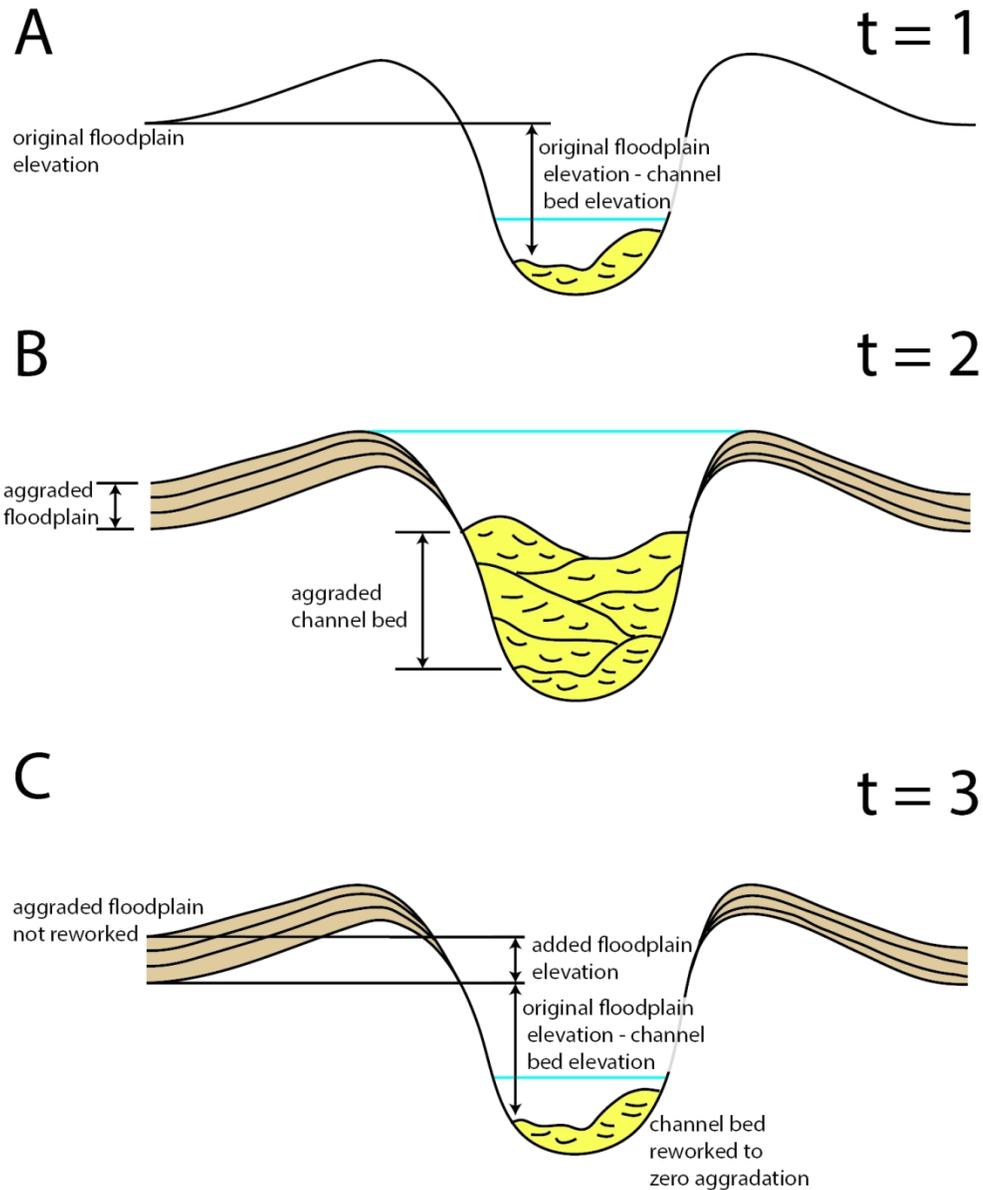
80x52mm (300 x 300 DPI)

Figure 22 – A diagram merging the observations of the usually bypass-state channel bed and the relatively high amount of channel-bed aggradation preceding avulsion, when compared to other fluvial strata (e.g., Mohrig et al., 2000). (A) at t = 1, the distance between the lowest part of the floodplain and the channel bed during low flow is defined. (B) At t = 2, the channel is in flood stage. The floodplain has aggraded less than the channel bed. Avulsion does not occur. (C) At t = 3, the channel bed has been reworked to the configuration of t = 1, maintaing its net-sediment bypass state. The floodplain has not been reworked. Thus, the vertical distance between the channel bed and the lowest part of the floodplain has increased. On average, more channel-bed aggradation will be required to set up an avulsion.

122x149mm (300 x 300 DPI)