

Connecting fluvial levee deposition to flood-basin hydrology

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Key Points:

1. Empirical analyses of lidar data show that levee size and shape on the Muscatatuck River, IN are unrelated to the channel planform.
2. Numerical modeling reveals that levees are unrelated to the channel planform because their growth is dictated by inundation dynamics in the floodbasin
3. On tall levees, the levee toe grows faster than the levee crest because it is inundated more frequently, and this causes levees to prograde down-valley over time.

Abstract

Levees are commonly found along every kind of river system, yet there are no widely accepted models for where along the channel they form and what controls their shape. In this study, we investigated whether levee growth is driven by sediment transfer from the channel adjacent to the levee or by inundation dynamics in the flood-basin. To test these ideas, we conducted empirical analyses and numerical modeling of levees on the fine-grained, meandering Muscatatuck River, IN. Using lidar data, we found no statistical relationship between the levee and the adjacent channel planform, which suggests levees are not genetically related to their adjacent channel. Surprisingly, modeling experiments of a simplified Muscatatuck River show that levee initiation can be genetically related to the adjacent channel because bed shear stress on the floodplain is low where channel curvature is high. But after levees initiate, the genetic connection to the adjacent channel is obscured because their shape is modified by inundation dynamics. Tall, mature levees are not inundated regularly and instead obstruct floodplain flow, creating flow shadows on the downstream side. Sediment is preferentially deposited in the flow shadow, which moves the location of maximum

29 **deposition from the levee crest to the toe. This causes levees to prograde down-valley, which in turn**
30 **reshapes the levee and genetically disconnects it from the channel. We propose that this**
31 **morphodynamic mechanism of levee growth is characteristic of fine-grained rivers where flood-**
32 **basins can act as conveyance channels that transport sediment downvalley before deposition.**

33

34 **1. Introduction**

35 Fluvial levees are important geomorphic features located along the banks of alluvial rivers.
36 Levees dictate flooding patterns, and the bi-directional exchange of water, sediment, and pollutants
37 between the channel and floodplain [*Nanson and Croke, 1992; Brierley et al., 1997; Adams et al., 2004*].
38 This, in turn, affects soil development, agricultural activity, and hazards to population centers within the
39 flood basin [*Wolfert et al., 2002; Brakenridge et al., 2017*]. Levees also influence fluvial stratigraphy
40 because as they grow, the potential for avulsion increases as the water surface becomes elevated above
41 the floodplain [*Bryant et al., 1995; Mackey and Bridge, 1995; Mohrig et al., 2000*]. Due to their
42 stratigraphic architecture, levee deposits are often targeted for hydrocarbons [*Fielding and Crane,*
43 *1987*]. Despite the societal and scientific importance of levees we do not know what sets their presence,
44 size, and shape.

45 Our current understanding of levee deposition is surprisingly cursory—empirical studies thus far
46 have not revealed consistent controls on levee presence, size, or shape. Levees are thought to occur
47 along both banks of a river, but on some meandering rivers they are more pronounced along the cut-
48 bank where they are not disturbed by scroll bar formation [*Kesel et al., 1974; Hudson and Heitmuller,*
49 *2003; Brooks, 2005*]. However in other cases, levees show little preference and occur on most river
50 banks [*Smith et al., 1989*], only on straight reaches [*Ferguson and Brierley, 1999*], or on alternating sides
51 [*Iseya and Ikeda, 1989*]. Levee shape should also be related to the sediment load, because all else being
52 equal, channels with fine-grained sediment should produce wide levees, with gentle slopes as sediment

53 is transferred farther into the floodplain, whereas channels with coarser-grains should have narrow,
54 steep levees. Indeed, this occurs in some fluvial systems [Cazanacli and Smith, 1998; Hudson and
55 Heitmuller, 2003], but in many cases, fine-grained levees can be steeper than coarse-grained ones
56 [Ferguson and Brierley, 1999] and the relationship between grain size and levee shape is often
57 ambiguous [Adams et al., 2004]. These empirical studies show no consistent results because we still do
58 not know the basic conditions for levee formation making it hard to isolate controls on levee shape.

59 Levees form when sediment is deposited in a zone of reduced flow competence along the
60 channel margins. A reasonable starting assumption is that the levee is genetically related to the adjacent
61 channel and thus the local channel and sediment characteristics—planform, depth, width, hydrology,
62 and grain settling—govern movement of sediment to the margins where levees form. This notion is
63 encapsulated in early models: assuming a straight channel, it can be shown that suspended sediment is
64 transferred to the channel margin by fluid advection and turbulent eddy diffusion [James, 1985; Pizzuto,
65 1987; Shiono and Knight, 1991; Marriott, 1992; Nezu and Nakagawa, 1993; Mertes, 1997; Adams et al.,
66 2004; Pizzuto et al., 2008]. These studies promote the idea that the levee is genetically related to the
67 adjacent channel segment, even though field data are equivocal and show no consistent relationship
68 between the levee and any measure of the adjacent channel.

69 On the other hand, levees may not be genetically related to the adjacent channel, especially if
70 levee sediment originates from sediment-laden floodwaters moving down-valley. This implies that flood
71 basins are more than just water storage locations—they convey water and sediment long distances,
72 sculpting levees in the process [Filgueira-Rivera et al., 2007; Pierik et al., 2017]. For instance, floodplain
73 sedimentation patterns are sensitive to floodplain microtopography, inundation time, flood wave shape,
74 flood magnitude, and the presence of perirheic zones [Asselman and Middelkoop, 1995; Mertes, 1997;
75 Middelkoop and Van der Perk, 1998; Walling and He, 1998; Nicholas and Mitchell, 2003; David et al.,
76 2017]. One interesting example shows that levee deposition occurs even when water in the adjacent

77 channel does not overtop the levee. *Filgueira-Rivera et al.* [2007] demonstrate that the levee crest and
78 toe can grow quasi-independently because the toe is at a lower elevation and is inundated by sediment-
79 laden water more frequently. This suggests levee deposition may not be related to the adjacent channel,
80 but instead occurs as the flood wave transports fine-grained sediment down-valley to positions behind
81 fluvial levees on the floodplain. The ultimate levee form may be dictated by flood-basin hydrology and
82 sediment transport during flood rather than the detailed mechanisms of sediment transfer from the
83 adjacent channel to the margin.

84 The focus of this paper is to assess whether levee presence and geometry are genetically related to
85 the adjacent channel or to flood-basin hydrology. We assess this by measuring and modeling levee
86 presence, size and shape on the meandering Muscatatuck River, Indiana. We have two hypotheses for
87 how levee deposition should be connected to the adjacent channel in meandering rivers. Our first
88 hypothesis is that levee heights and widths should scale with curvature because curvature drives
89 superelevation of the water surface and promotes advective sediment transport across cutbanks [*Ervine*
90 *et al.*, 1993; *Ervine et al.*, 2000]. Our second hypothesis is that levees should be most prevalent and
91 largest in the crossover region (i.e. where the channel centerline is perpendicular to the floodplain
92 centerline) because this is where advective transport of channel sediment into the floodplain is
93 maximized [*Shiono and Muto*, 1998; *Ervine et al.*, 2000; *Czuba et al.*, accepted]. Our alternative
94 hypothesis is that levee formation is not related to the adjacent channel, and instead dictated by flood-
95 basin hydrology. To test these hypotheses, we conducted an empirical analysis of near channel
96 topography along the Muscatatuck River, IN and its association with the adjacent channel curvature and
97 orientation relative to floodplain centerline. We assess the role of the floodbasin with numerical
98 modeling. Based on our numerical modeling, we show how levee deposition can occur independently of
99 the adjacent channel and how levee form in this scenario is ultimately governed by floodbasin
100 hydrology.

101 **2. Field Study Site**

102 Our field study focuses on the Muscatatuck River (Figure 1), which flows westward from the
103 Muscatatuck Plateau across the Scottsburg Lowlands through southern Indiana eventually meeting the
104 East Fork White River to the west. The Scottsburg Lowlands run approximately north-south between the
105 Muscatatuck Plateau to the east and the Norman Upland to the west. This region lies south of the
106 Wisconsin Glaciation boundary and may have served as a conduit for glacial outwash flowing south to
107 the Ohio River. Underlying this physiographic region is Devonian and Mississippian age limestone, shale
108 and siltstone. Across this area, the Muscatatuck River continues to incise into sediments weathered
109 from the underlying New Albany Shale to the east and the Borden Group shales to the west. The
110 Muscatatuck River incision into this easily eroded, fine-grained siltstone and shale accounts for the fine-
111 grained sediment load.

112 We selected two reaches along the Muscatatuck River for detailed analysis of the near-channel
113 topography (Figure 1A, B). We refer to the topography as near-channel because not all reaches contain
114 levees. These reaches were chosen because both are prone to frequent flood events that fully inundate
115 the surrounding floodplains, and both have levees along the channel margins. Also, neither of the
116 reaches are lined with anthropogenic flood control structures, such as dikes or dams, or elevated road
117 beds which would obscure natural features. Both reaches flow through low gradient, agricultural land in
118 southern Indiana.

119 We analyzed near-channel topography using a 1.5-meter resolution, bare earth digital elevation
120 model (DEM), derived from LiDAR data accessed at opentopography.org. The LiDAR dataset was collected
121 by a nonprofit consortium during the statewide campaign to collect high resolution LiDAR and
122 orthophotography data from 2011 to 2013. Reach 1 has a channel centerline distance of 5,210 meters,
123 the average channel width is 28 meters, and the sinuosity—measured as the channel centerline distance
124 divided by the upstream to downstream straight-line distance—is 1.78 (Figure 1A). Reach 2 is located

125 approximately 8.9 km downstream from Reach 1 and has a channel centerline distance of 7,557 meters,
126 the average channel width is approximately 32 meters, and the sinuosity is 3.04 (Figure 1B). The
127 floodplain slopes for Reach 1 and Reach 2 are 0.00033 and 0.00027, respectively; and the channel bed
128 slopes for Reach 1 and Reach 2 are 0.00021 and 0.00011, respectively. The average bankfull depth
129 through both reaches of the Muscatatuck measured from a single-beam echo sounder is approximately
130 2-meters.

131 To characterize the suspended sediment concentrations in the Muscatatuck River we collected
132 51 suspended sediment samples and augmented these data with 92 suspended sediment concentration
133 samples collected by the USGS at the stream gage in Deputy, Indiana (Figure 1C). We collected
134 suspended sediment samples from four bridges spanning the Muscatatuck River (denoted by stars in
135 Figure 1C) on February 7, 2015; March 4 and 11, 2015; April 12, 2015 and January 21, 2017 during
136 bankfull to flood stage discharge (20 – 198 m³/s) according to USGS standard operating procedures
137 using a US-DH59 sampler [Edwards *et al.*, 1999]. Suspended sediment concentrations were measured
138 with the evaporation method procedures laid out in the ASTM International Test Method, D 3977 –
139 97(B).

140 To characterize the grain-size in the channel and floodplain, we collected four channel-bed-
141 surface sediment samples during bankfull to floodstage flow, and ten hand auger samples from the
142 levee and ten from the floodplain. Channel-bed-surface samples were collected from the Waskom and
143 Wheeler Hollow bridges (Figure 1C) on January 20, 2016 using a Ponar[®] grab sampler according to
144 standard operating procedures laid out by the U.S. Environmental Protection Agency in SESDPROC-200-
145 R3. Levee and floodplain deposits were sampled at Reach 2 using a hand auger (Figure 1D). Two
146 replicate samples were collected at each location, and samples for analysis were taken 0.75 meters
147 below the surface to avoid sampling organic root material or surface soil that may have been modified
148 by agricultural activities. Grain size was measured with a Malvern Mastersizer[®] 3000E.

149 The Muscatatuck River is silt-dominated with moderate to high suspended sediment concentrations.
150 Combining our samples and ones from USGS, the average suspended sediment concentration during
151 bankfull flow ($Q = 35 \text{ m}^3/\text{s}$) or higher for the Muscatatuck River is 0.23 kg/m^3 (Figure 2A). The median
152 grain size for suspended and bed-surface sediment was $15.7 \text{ }\mu\text{m}$ and $27.7 \text{ }\mu\text{m}$, respectively. Median
153 grain size for the levee and floodplain deposits were $9.8 \text{ }\mu\text{m}$ and $10.1 \text{ }\mu\text{m}$, respectively (Figure 2B). There
154 was no systematic difference in the channel-bed grain size across the four measurement locations.

155

156 **3. Methods**

157 *3.1 Measuring Near-channel Topography*

158 Measuring near channel topography is a deceptively challenging task. Idealized levee shapes are
159 conceptually simple—most models depict levees as gull-wing shaped deposits with elevations that
160 exponentially taper from a maximum at the crest out to the levee toe [Hudson and Heitmuller, 2003].
161 But natural levees rarely follow this shape and instead show a wide variability of shape and orientation
162 relative to the channel margin. This variability and complexity is not captured when levee geometry is
163 measured with coarsely-spaced cross-sections perpendicular to the channel centerline.

164 The methodology we have developed captures the full two-dimensional planform shape of the
165 levee. We first detrend the DEM by removing the average floodplain slope from the topography. The
166 average floodplain slope was derived from a triangulated irregular network (TIN). The TIN surface was
167 computed by sampling elevations along evenly spaced cross-sections down the floodplain. The resulting
168 TIN surface maintained all long wavelength topography and removed all short wavelength topography.
169 We found that using this surface to remove down-valley slope minimized distortion of small scale
170 features when detrending the floodplain. Next, we isolate near channel topography in the LiDAR data
171 within three channel widths along both banks using a histogram equalization method. This method is
172 often used by the medical industry to analyze x-rays because the data distribution is reshaped to create

173 a linear cumulative distribution function, which has the distinct advantage of enhancing low contrast
174 areas [Hum et al., 2014]. In the context of a DEM, this enhances the low relief of the levee toe where
175 slopes are small (Figure 3).

176 Using the histogram equalized image, we then pick an elevation value for the levee toe by
177 inspecting cross-sections. Like previous researchers, we identify the levee toe on different cross-sections
178 as the location where the levee slope decreases to approximately 0.01 m/m [Cazanacli and Smith, 1998;
179 *Filgueira-Rivera et al., 2007*]. In areas where levee deposits abut more complex topographic surfaces we
180 determined the levee toe by defining the first major break in topography after the levee crest [Adams et
181 al. [2004]. Otherwise, the levee toe was delineated from changes in land use or vegetation [Hudson and
182 *Heitmuller, 2003*]. The levee toe contour elevation was determined by thorough inspection of individual
183 cross-sections and analysis of satellite and hillshade imagery. We avoided any anthropogenic features,
184 such as elevated road beds or drainage ditches. Additionally, natural features not genetically related to
185 the channel, such as fluvial terraces and continuous low angle hillslopes, were excluded from the levee
186 analysis (inset A, Figure 3).

187 We then define individual levee polygons from closed contours at the elevation of the levee toe
188 (Figure 4). Closed contours define a single levee, and in some cases, these are generated by depositional
189 processes, whereas in other cases channels cut through the levees (inset B, Figure 3). Levee width for
190 each polygon was calculated by calculating channel-perpendicular transects, spaced 2-meters apart, and
191 clipping them at levee polygon boundaries (Figure 4, inset). Average levee width is calculated as an
192 arithmetic mean of all cross-sections within a given polygon on each channel bank. Levee height values
193 were extracted as the highest value for each channel-perpendicular transects and then averaged across
194 all transects. Local channel centerline curvature was measured using RivMAP [Schwenk et al., 2017]. We
195 also calculate the radius of curvature by fitting a circular arc to the channel centerline using satellite

196 imagery available through Google Earth Pro[®]. The channel centerline was defined as the mid-point
197 between the cutbank and the inside point bar.

198 Finally, we calculated the orientation of levee deposits relative to the floodplain centerline. The
199 levee and floodplain centerlines were determined with Thiessen polygons. We grow the Thiessen
200 polygons from the boundaries of the levee and floodplain polygons. The levee polygons were defined by
201 the levee toe contours, and the floodplain polygon was defined by floodplain terrace boundaries. The
202 line where the Thiessen polygons meet is taken as the levee and floodplain centerlines. The angle
203 between the two centerlines was measured as the difference between the azimuth orientation of each
204 centerline measured in degrees.

205

206 *3.2 Numerical Modeling Methods*

207 3.2.1 Flow modeling considerations in Delft3D

208 We conducted numerical modeling experiments using Delft3D, which solves the coupled 3D fluid
209 flow field and bed evolution. For 3D flow, Delft3D solves the shallow-water equations with a user
210 specified number of depth layers. In this paper we used eight layers with thicknesses (as a percentage of
211 total depth) moving from the water surface to the bed as 30, 20, 10, 10, 10, 10, 5, and 5%. The vertical
212 transport of momentum between the flow layers is accounted for using the κ - ϵ turbulence closure
213 model which solves for the kinetic energy (κ) and the turbulent kinetic dissipation (ϵ). These values are
214 used to determine the vertical eddy viscosity and the mixing length of the flow. In addition to the
215 transport of fluid momentum, Delft3D models the transport of suspended sediment by solving the
216 conservation of mass of suspended material within the model domain. The model solves for the 3D
217 transport of sediment by calculating the advection and diffusion of suspended particles using the flow
218 velocity terms and the eddy diffusivities derived using the Navier-Stokes equations.

219 3.2.2 Model Domain Selection

220 To explore controls on levee depositional patterns, we choose a model domain covering a
221 channel reach with several meander bends that allowed for an examination of the hydrodynamic
222 conditions during flood events. We selected Reach 2 for this analysis (Figure 1B) because it contains
223 multiple river meander bends and we know that levees have formed along the channel. The modeling
224 domain covers a floodplain area that is 2,500 meters long by 1,700 meters wide. Our model domain is
225 the same extent as shown in Figure1B. In the model we retain the shape of the river in planform, but
226 synthetically flattened the floodplain topography so we could easily assess how the flood-wave interacts
227 with the channel during flood. The smoothed floodplain has a slope of ~ 0.00027 m/m and channel slope
228 of ~ 0.00011 m/m, consistent with field observations. The channel was 2-meters deep, consistent with
229 field measurements, and with a rectangular cross-sectional shape. A random centimeter-scale
230 topographic perturbation was also added across the model domain to the surface to allow for more
231 realistic flow propagation. We added a straight channel segment at the upstream boundary of the
232 model for flow establishment prior to entering the first meander bend. We also smoothed the channel
233 bankline boundaries to encourage a more natural flow condition and eliminated grid effects.

234 3.2.3 Boundary Conditions and Physical Parameters

235 Flow direction enters the domain along the eastern boundary and exits to the west. The upstream
236 boundary was set across the channel width at the eastern boundary and defined the time-dependent
237 fluid discharge entering the domain. The downstream boundary was a time-varying water level
238 elevation that was chosen to prevent backwater or drawdown effects based on the upstream discharge.
239 The north and south model boundaries were defined as Neumann boundaries, which allows fluid and
240 sediment to smoothly exit or enter the boundary. At the upstream boundary, the incoming sediment
241 flux was set to a constant 0.23 kg/m^3 with a density of $2,650 \text{ kg/m}^3$, and a grain size of 0.016 mm . This
242 size was determined based on the median grain size of suspended sediment sampled during field

243 operations. The settling rate of grains is 0.23 mm/s [Ferguson and Church, 2004]. Additional modeling
244 parameters are summarized in Table 1.

245 The primary factors that drive suspended sediment deposition and erosion within the modeling
246 domain are the critical shear stress for sedimentation and erosion. There are numerous factors, such as
247 vegetation, microtopographic variations, grain flocculation, for example, that cause deposition of silt
248 and clay sized grains on the floodplain. Our model was not able to account for all of these factors. In
249 order to allow for deposition of sediment during a flood event, we set the critical shear stress for
250 sedimentation at 0.5 N/m². This value permits a reasonable amount of deposition in the domain. The
251 critical shear stress for erosion is set to 2 N/m². This value falls within the reasonable range for erosive
252 shear stresses for fine-grained, cohesive sediment and does not allow for excessive channel erosion
253 during elevated discharge events [Clark and Wynn, 2007; Kimiaghali et al., 2016]. Based on historical
254 satellite imagery, we know that the channels of the Muscatatuck are relatively stable, and any shear
255 stress values for erosion that allow for rapid changes to the channel planform are not representative of
256 the natural system. Deposition of fine-grained silt within the model domain occurs slowly, so we used a
257 morphological scaling factor of 70 for these modeling experiments to speed up change.

258

259 3.2.4 Modeling Experiment and Sensitivity

260 To quantify the evolution of levee morphology, we flooded the model domain from the
261 upstream boundary four times with identical triangular flood waves. The flood waves were symmetrical
262 and peaked at 100 m³/s. Each flood wave lasted two days and started and stopped at a minimum
263 discharge of 10 m³/s. The peak discharge completely inundated the floodplain, and floodplain waters
264 exited the domain before the next flood event. The entire four wave experiment lasted eight days of
265 model time. It is difficult to scale this time precisely, but the discharge magnitude roughly corresponds

266 to 0.5 to 1-year flood and given the morphological scale factor of 70, we argue these simulations could
267 represent 140-280 years of actual time.

268 We assessed the model sensitivity by comparing flow conditions and depositional patterns while
269 varying several physical parameters, the grid resolution, and boundary conditions. The model proved
270 insensitive to implementation of the two-dimensional horizontal large eddy simulation available with
271 the Delft3D modeling package. Model results were also insensitive to finer grid resolutions.
272 Computation time increased dramatically when decreasing grid cell size from 10-meter by 10-meter to
273 5-meter by 5-meter, but the depositional patterns and hydrodynamic conditions were unchanged.
274 Finally, variations in suspended sediment concentrations at the upstream boundary did not affect the
275 planform depositional patterns. The volume of sediment deposited within the domain was affected by
276 reductions in the sediment concentration, but the locations and lateral extent of deposits were
277 unaffected.

278 Model results were sensitive to the horizontal eddy viscosity value, the type of boundary conditions
279 along the northern and southern boundaries, and the threshold water depth. Increasing the eddy
280 viscosity value by one and two orders of magnitude resulted in increasing grid effects to fluid flow across
281 the model domain. High values of eddy viscosity limited diffusion of eddies flowing from one grid cell to
282 another. Changing the northern and southern boundary conditions to closed boundaries resulted in
283 different flood wave propagation and depositional patterns along the distal margins of the domain when
284 compared to the results using neumann boundaries. Closed boundaries forced abrupt down-valley
285 change in flow vector orientation and often modified the shape of the distal edge of sedimentation
286 patterns. Lastly, variations in the water threshold depth, value for determining if a cell is wet or dry,
287 caused unrealistic deposition across the domain. Elevated threshold values often caused grid cells with
288 shallow water depths to fluctuate between wet and dry conditions during the initial inundation of the

289 floodplain and as waters receded during the falling limb of a flood wave. This resulted in unrealistic,
290 gridded depositional patterns along the leading edge of the flood wave.

291 **4. Controls on levee morphology on the Muscatatuck River**

292 Based on our definition of levees and methods for their delineation, we mapped the presence of
293 levees along the two reaches of the Muscatatuck River (Figure 5). On Reach 1 levees are more common
294 and occur along 80% of the banks, whereas in Reach 2 levees occur along 55% and 10% of the left and
295 right banks, respectively (Figure 5). River right and river left series show some degree of correspondence
296 on both reaches. In Reach 1, both sides have a decreasing levee presence until normalized stream
297 distance of 0.1 and then become progressively more leveed downstream. In Reach 2, both river right
298 and left have unleveed portions from normalized stream distances of 0.5 to 0.75. This suggests that
299 levees tend to be paired. For a given reach, paired levees show little similarity between their heights and
300 widths at a given position (Figure 6). We find no statistical evidence that these series are correlated.
301 Median levee heights and widths for both reaches are reported in Table 2.

302 To explore controls on levee shape, we compared the local channel centerline curvature for
303 both reaches to levee geometry. We calculated the channel curvature as the second derivative of the
304 channel centerline (Schwenk et al., 2017) (Figure 6C,F). We calculated the cross-correlation between the
305 curvature series and levee width or height to see if levee geometry is in phase (plus or minus a lag) with
306 channel planform. We found that levee morphology is not in phase with channel curvature (Figure 7). In
307 some cases, there may be a slight correlation between levee width and channel curvature (e.g., river
308 right levees along Reach 1 at the zero-lag position, Figure 7). But, the value does not rise significantly
309 above the background. These results suggest that variations in the height and width of levees along the
310 Muscatatuck River is unrelated to channel curvature.

311 For Reaches 1 and 2, we also explored whether levee morphology was determined by the
312 orientation between levee centerline and floodplain centerline (Figure 8). Angle values near zero

313 degrees indicate a levee centerline runs approximately parallel with a floodplain centerline, and values
314 approaching 90 degrees indicate a levee centerline is perpendicular to the floodplain centerline. There is
315 no obvious relationship between the orientation of the levee relative to the floodplain and all levee
316 variables considered (Figure 8).

317 We finally explored the relationships between channel radius of curvature and levee geometry
318 for 54 meander bends along the Muscatatuck River. For this analysis, we increased our study reach to 44
319 kilometers of channel centerline distance to include more meander bends (Figure 1C). For these bends,
320 we only considered cutbanks where levees are not obscured by scroll bars and easier to identify. We
321 found no relationship between channel radius of curvature and levee morphology (Figure 8). That said,
322 meander bends with the highest radius of curvature values are loosely associated with narrower and
323 taller levees, however, this could be a sample size bias at larger radii of curvature (Figure 9A,B). Aspect
324 ratio, and basal area showed a high degree of variability across the range of channel radius of curvature
325 values and no clear relationships exist (Figure 9C,D).

326 In summary, these empirical data show no clear relationship between the adjacent channel and
327 levee formation. We think this suggests that levee deposition on the Muscatatuck is instead governed by
328 flood-basin hydrology and in the following section we test how a fine-grained river system creates
329 levees during overbank flow.

330

331 **5. Morphodynamic modeling of levee formation**

332 To explore the processes that lead to levee formation and growth on the Muscatatuck River, we
333 model levee morphodynamics in Delft3D along Reach 2 through four consecutive flood waves. Each
334 flood wave is symmetrical and peaks at 100 m³/s discharge and starts and stops at a minimum of 10
335 m³/s discharge. Bankfull discharge is achieved at approximately 35 – 40 m³/s. Peak discharge results in

336 complete inundation of the floodplain while still maintaining the high velocity channelized flow.

337 Floodplain water depth during peak inundation is approximately 0.5 meters.

338 Prior to the morphodynamic simulation, we first simulated the reach with only hydrodynamics. In
339 this case, as the flood wave moves down-valley, there is surprising variability in flood-wave arrival times
340 given the simplicity of the floodplain (Figure 10A). For instance, the southern half of the floodplain
341 inundates later than the northern half. This could be caused by the interaction of the flood with the
342 meandering channel, which serves as a macro roughness element. During peak inundation, there are
343 distinct zones of high and low bed shear stress that emerge as the flood wave interacts with the channel
344 (Figure 10B). Downstream of meander bend apices, the expulsion of high velocity channel waters
345 creates locally high bed shear stress [Sellin *et al.*, 1993; Shiono and Muto, 1998]. However, at the apex of
346 the meander bend channel and floodplain flow are aligned and shear stress in this region low.

347 During the four flood waves, some sediment is deposited on the channel bed, but most of the
348 deposition occurred along the channel margins and across the floodplain (Figure 11). Levee deposition
349 occurs where the bed shear stress is less than the critical value for sedimentation (0.5 Pa, Table 1) in our
350 experiments (compare Fig. 10B and 11). Interestingly, the thickest levee deposits occur at meander bend
351 apices where channel curvature is highest. As expected, total levee deposit thicknesses are small;
352 typically, on the order of 10 to 20 centimeters, but they are consistent with field measurements (Table
353 2; Figure 6).

354 Many areas across the modeling domain show interesting depositional patterns, but we focus on
355 four areas that include three cutbank levee deposits (labeled 1 through 4 on Figure 11). Levees at
356 locations 1 and 2 are oriented approximately 45 and 30 degrees, respectively, relative to the down-
357 valley flow direction. Levees at location 3 and 4 are oriented approximately parallel to the down-valley
358 slope. Temporal evolution of levee shape and size show that all levees grow down valley during
359 subsequent flooding events (Figure 12). Down-valley progradation is especially evident for location 1

360 where levee growth is oblique to the channel bank at the meander bend apex. The levee at location 1
361 also widens as it grows in a down-valley direction. Levee 2 also undergoes widening and down-valley
362 progradation. Progradation of the levee at location 3 is less pronounced; however, this deposit also
363 widens and begins to grow in a down-valley direction after the second flood wave. Down-valley
364 progradation is evident along the left bank at location 4, and both paired levees experience lateral
365 widening.

366 Down-valley progradation occurs because when the levee grows it generates a morphodynamic
367 feedback by shifting deposition from the crest to the toe [Figueira-Rivera *et al.*, 2007]. All levees
368 undergo a transition from crest to toe deposition by the fourth flood waves (Figure 12, 13). The change
369 maps for each levee (Figure 12) show that deposition on the crest decreases from flood wave 1 to 4,
370 while deposition on the levee toe increases. For instance, levee deposition at location 2 is initially at the
371 levee crest near the channel margin (plot 2A in Figure 12), but later flood waves deposit more sediment
372 on the lee side of the levee while sedimentation near the channel is reduced (compare plots 2B and 2C
373 in Figure 12). The levee crest at location 1 is subaerial through most of the fourth flood wave. This forms
374 a bed shear stress shadow and low flow velocity zone on the lee side of the levee and leads to more
375 deposition at the levee toe than at the crest. Interestingly, deposition also increases on the up-valley
376 side of all three levee deposits. The combination of these two depositional trends effectively increases
377 levee length.

378 The transition from crest to toe deposition occurs because as the levee grows it obstructs flow
379 and sediment transport over the crest. The transition from front-loading to back-loading can also be
380 seen in the depositional rates measured at the crest, a down-valley position, and an up-valley position
381 (Figure 13). For all four examples, deposition rates at the levee crest are the greatest, relative to the
382 other locations, during the first two flood events but decrease through time (Figure 13, plotted in blue).
383 Up-valley and down-valley depositional rates remain relatively constant at all four levee deposits

384 through time. Down-valley depositional rates generally exceed up-valley depositional rates (Figure 13,
385 plotted in red), indicating levee growth is occurring in a down-valley direction. In all cases, the crest
386 deposition rate is equal to or less than the other positions by the fourth flood wave, indicating the
387 transition from front-loading to back-loading [*Filgueira-Rivera et al., 2007*].

388 Additionally, the arrival time of the floodwave also can influence where and how levees grow. At
389 location 4 the channel is oriented parallel to the floodplain centerline and levees should be built by
390 classic mechanisms of turbulent eddy diffusion [*Pizzuto, 1987; Adams et al., 2004*]. This assumption
391 arises because at peak flood, floodplain flow and channelized flow should be aligned. But this site
392 illustrates how difficult it can be to detect the controls on levee formation based on this kind of
393 reasoning alone. At location 4, the floodwave arrives on the river right side first (Figure 10A). This sets
394 up a water surface gradient from north to south on the river left side, which creates relatively strong
395 flow perpendicular to the channel centerline (Figure 14A). This strong flow advects sediment from the
396 channel into the floodplain on the river left side, and at peak discharge the gradient in sediment
397 concentration on river left side is larger than river right (Figure 14B). In this case, the levee is genetically
398 connected to the channel, insofar that the levee sediment was advected from the channel and
399 immediately deposited on the margin, but the floodbasin hydrology dictates floodwave arrival time and
400 levee formation. This location illustrates that controls on levee formation can be related to channel and
401 floodbasin conditions.

402

403 **6. Discussion: What controls levee size and shape?**

404 It is surprising to us that there are no convincing empirical data or theory on what controls levee size
405 and shape. Because of this we started with the null hypothesis that the levee is genetically related to the
406 adjacent channel. Afterall, the channel is the likely source of levee sediment and the common practice
407 of viewing levees through channel perpendicular cross-sections implies a connection. If levees were

408 genetically related to the adjacent channel, we would expect our data to show some relationship
409 between levee size and local curvature, radii of curvature, or relative angle. But this is not the case:
410 there is no interpretable relationship between river planform and the associated levee (Figures 6-9).
411 This is surprising because if levees form as sediment diffuses and advects from the channel to the
412 margin [Adams *et al.*, 2004], then advection should be enhanced where curvature is high because
413 centripetal acceleration causes overbank flow, and advection should be enhanced in the crossover
414 region because downstream oriented floodwaters would move sediment to the channel margin [Sellin *et*
415 *al.*, 1993]. Even in the case of levee 4, where levee sediment is advected from the channel to the margin
416 (Figure 14), the advection is caused by differences in floodplain arrival time, which in turn is related to
417 inundation dynamics of the floodbasin.

418 Modeling results suggest levee initiation is genetically connected to the channel because levees
419 initially form where curvature is high and shear stress is low (Figure 10B; 11). However, this genetic
420 connection is obscured as levees grow because their inundation dynamics change. Smaller levees are
421 inundated from crest-to-toe as floodwaters spill out of the channel and flow down the levee. But taller
422 levees flood waters flow around the levee perimeter and inundate the levee toe first. These taller levees
423 can exhibit a toe-to-crest inundation pattern.

424 This change in inundation pattern marks a significant transition in deposition. Deposition rate on
425 smaller levees is highest on the crest because that inundates first with sediment-laden water directly
426 from the adjacent channel (Figure 13). But deposition rate on taller levees is highest at the toe because
427 it is inundated longest with sediment-laden floodwater. This depositional transition reshapes the levee
428 as the toe grows faster than the crest [Filgueira-Rivera *et al.*, 2007]. The transition effectively starves the
429 levee crest of new sediment and depending on the hydrodynamic conditions this shifts the center of
430 mass of a levee deposit. This shift moves the levee further from the channel margin and generates levee
431 shapes that are not easily related to the adjacent channel and its orientation.

432 This depositional transition results in levee progradation down-valley. Once levees are tall enough,
433 they obstruct the propagating flood wave. The flow obstruction creates a shear stress shadow on the
434 downstream side, and sediment transported down-valley by floodplain flow is deposited within this
435 zone of reduced flow competence (Figure 15A). Though it is difficult to say conclusively, we propose that
436 down-valley prograding levees explain some of the enigmatic levee topography along the Muscatatuck
437 River (Figure 15B,C) and also make it difficult to connect the levee morphology to the adjacent channel
438 (Figures 6-9). The modeled feedbacks between levee evolution and fluid flow suggest that after the
439 initial levee deposition, levee morphology within fine-grained fluvial environments may be more
440 strongly related to flood wave propagation.

441 We propose that the lack of relationship between the channel planform and levee size, as observed
442 in this study, may be characteristic of fine-grained river systems. In coarser-grained rivers, the
443 suspended sediment quickly falls out of suspension as fluid moves from the channel to the channel
444 margin. If this is the only source of levee sediment, then the levees will be genetically connected to the
445 channel. Indeed, consistent with this notion, *Hudson and Heitmuller* [2003] found that coarser grained
446 levees (84th percentile of grain size ranges from 64 to 240 μm) on the coastal gulf plain rivers of Mexico
447 are largest on the outside of meander bends. This situation may not occur on finer-grained rivers. The
448 levees and floodplain of the Muscatatuck River have much finer sediment (84th percentile is 19-22 μm)
449 with almost no sand-sized sediment (Figure 2B). We see no observable connection between the levee
450 and adjacent river channel, possibly because as the modeling shows, fine-grained suspended sediment is
451 transported long distances during flood before it is deposited. This can decouple the levee from the
452 adjacent channel.

453 **7. Conclusions**

454 Here we tested whether levees grow and form by the transfer and deposition of sediment from the
455 adjacent channel to the margin or by inundation dynamics in the floodbasin. If levees are formed by

456 sediment transfer from the adjacent channel then levee size and shape should be a function of the
457 processes that drive that transfer. Our empirical analyses show no conclusive evidence that levee
458 presence or size is genetically related to the planform morphology of the adjacent channel. We focused
459 on planform attributes, like the curvature of the channel and its orientation relative to the floodplain
460 axis, because, all else being equal, a sinuous channel should transfer more sediment to the channel
461 margins during flood.

462 To see why levees may not be genetically related to their adjacent channel we conducted modeling
463 experiments of a simplified Muscatatuck River that retains the channel planform but has a smooth
464 floodplain. Our modeling results suggest that in fine-grained systems levee initiation is related to the
465 channel planform. But, after levees grow tall enough relative to the floodwave, their crests are starved
466 of sediment and the deposition maximum moves to the levee toe. When this transition occurs, levees
467 prograde down-valley, which in turn reshapes the levee and genetically disconnects it from the
468 channel.

469 From these empirical and modeling results we conclude that in fine-grained rivers, levee size
470 and shape is sensitive to floodbasin hydrology and emerges from the interactions among inundation
471 frequency, flood-wave propagation, and transport pathways of suspended sediment.

472

473 **Acknowledgements**

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475 supporting this work. We would also like to thank the Indiana University Grand Challenges *Preparing for*
476 *Environmental Change* for partial support. GJ would like to thank the American Association of Petroleum
477 Geologists and the Society for Sedimentary Geology for providing additional research grants in support
478 of field and laboratory work. Upon publication all data will be made available through Indiana
479 University's Scholar Works (<https://scholarworks.iu.edu/>).

480

481 **Table 1.**

Table 1 Summarized model parameters for Delft3D		
<i>Hydrodynamic Parameters</i>		
Water Density	1000	kg/m ³
Horizontal Eddy Viscosity	0.001	m ² /s
Horizontal Eddy Diffusivity	1	m ² /s
3D Turbulence Model ^a	κ-ε	--
Manning Roughness	0.035	--
<i>Sediment Parameters</i>		
Settling Velocity	0.23	mm/s
Specific Density	2650	kg/m ³
Dry Bed Density	1350	kg/m ³
Density for Hindered Settling	1600	kg/m ³
τ-crit for Sedimentation ^b	0.5	N/m ²
τ-crit for Erosion ^c	2	N/m ²
Erosion Parameter	0.0001	kg/m ² /s
Initial Sediment Layer	0.5	m
Morphological Scaling Factor	70	--

482

483 ^a Kappa-Epsilon (κ-ε) 3-dimensional turbulence closure model

484 ^b Critical shear stress for sedimentation

485 ^c Critical shear stress for erosion

486

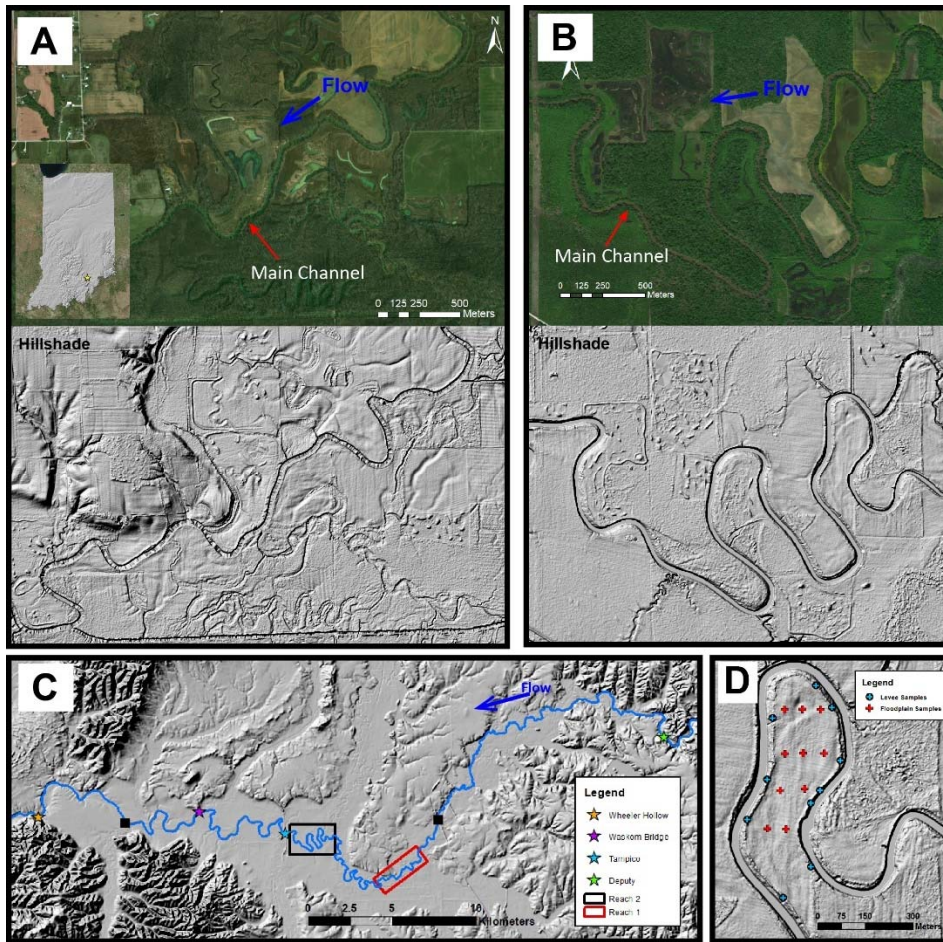
487 **Table 2:** Levee statistics on Reaches 1 and 2 of the Muscatatuck River

	Median Height (m)	Range in levee height (m)	Median width (m)	Range in levee Width (m)
Reach 1 – river left	0.21	0.05 – 0.84	16.9	0.6 – 30.7
Reach 1 – river right	0.19	0.08 – 1.1	13.8	0.6 – 36.2
Reach 2 – river left	0.20	0.04 - 0.9	21.9	4.2 – 16.1
Reach 2– river right	0.17	0.07 - 0.7	16.1	6.3 – 38.8

488

489

490 **Figures and Captions**



491

492 **Figure 1.** (A, B) Satellite and a hillshade image, generated from 1.5-meter resolution elevation data, for
493 Reaches 1 and 2 along the Muscatatuck River, Indiana. Flow direction in the main channel is indicated on
494 the satellite imagery. Channel bathymetry is from hydro-flattening. (C) Longer reach showing the
495 locations of Reaches 1 and 2 (boxes), and locations of river-spanning bridges where suspended sediment
496 samples were collected (stars). These are also locations of USGS gauges used for data in figures 2. Black
497 squares mark the beginning and end of the reach of the Muscatatuck River analyzed in radius of
498 curvature analysis in figure 6. (D) Sediment sample locations along Reach 2. Inset map in (A) shows
499 location of study reaches in Indiana.

500

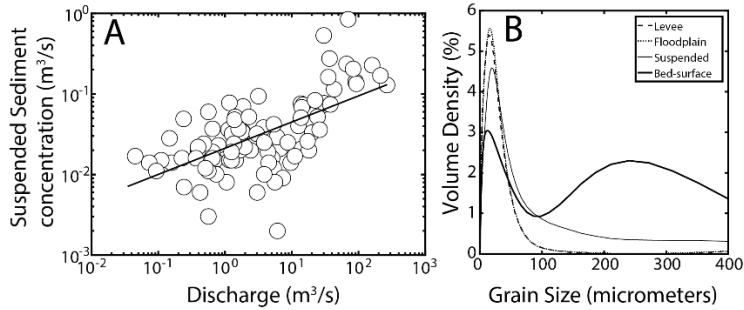
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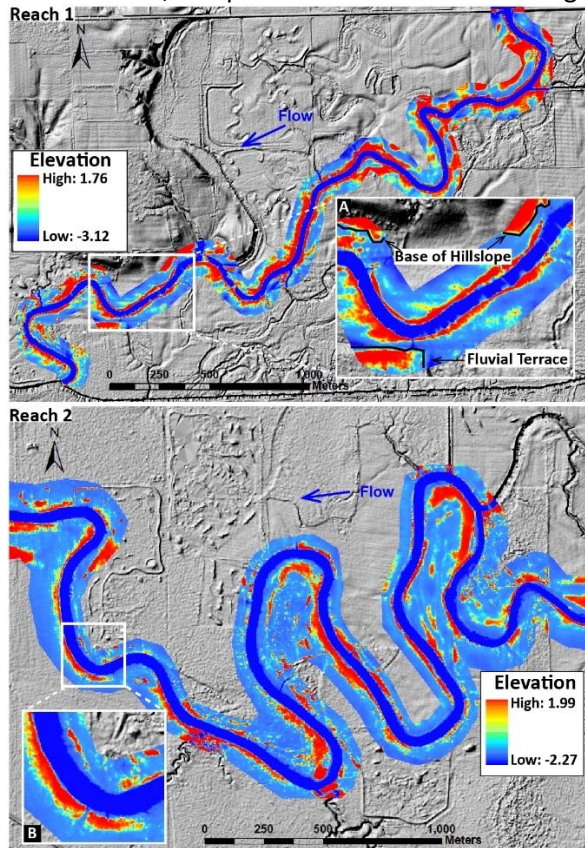
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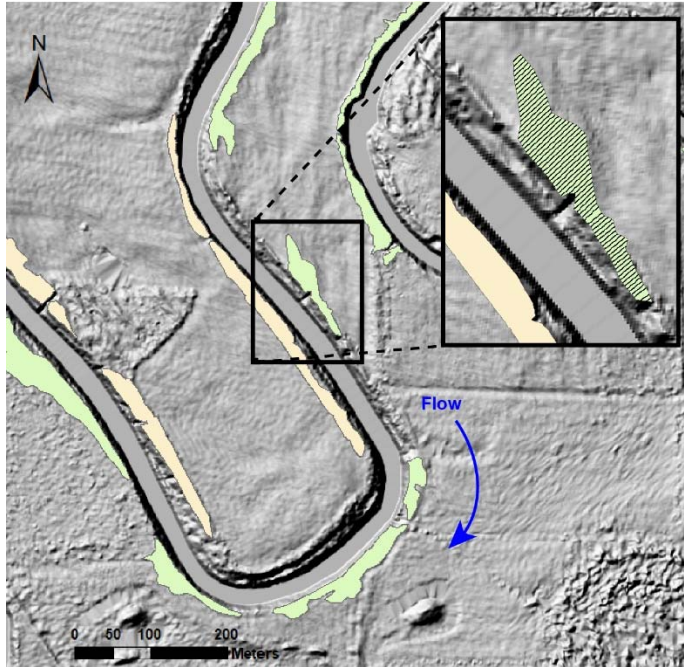
506

507 Figure 2. (A) Suspended sediment rating curve for the Muscatatuck River from USGS suspended
 508 sediment concentration samples collected at the Deputy, Indiana gauge station. (B) Channel-bed-surface
 509 distribution is the average of four samples collected near Waskom and Wheeler Hollow bridges and
 510 suspended sediment distribution is the average of 51 samples collected at Deputy and Wheeler Hollow.
 511 Floodplain and levee crest grain size distributions are the averages of ten samples collected along Reach
 512 2. In all cases, sample locations are shown in Figure 1C.



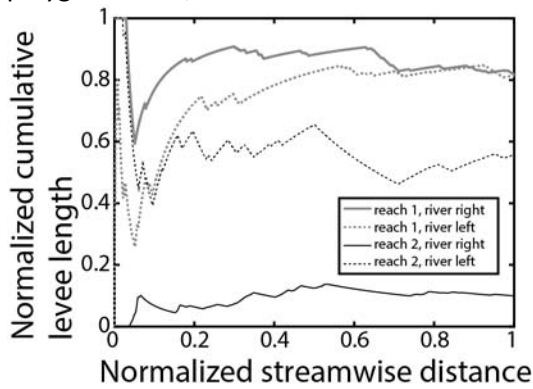
513

514 Figure 3. Histogram equalization and detrending visualization methods for near-channel topography
 515 along Reach 1 and Reach 2 of the Muscatatuck River. Elevation is measured relative to the detrended
 516 floodplain surface.



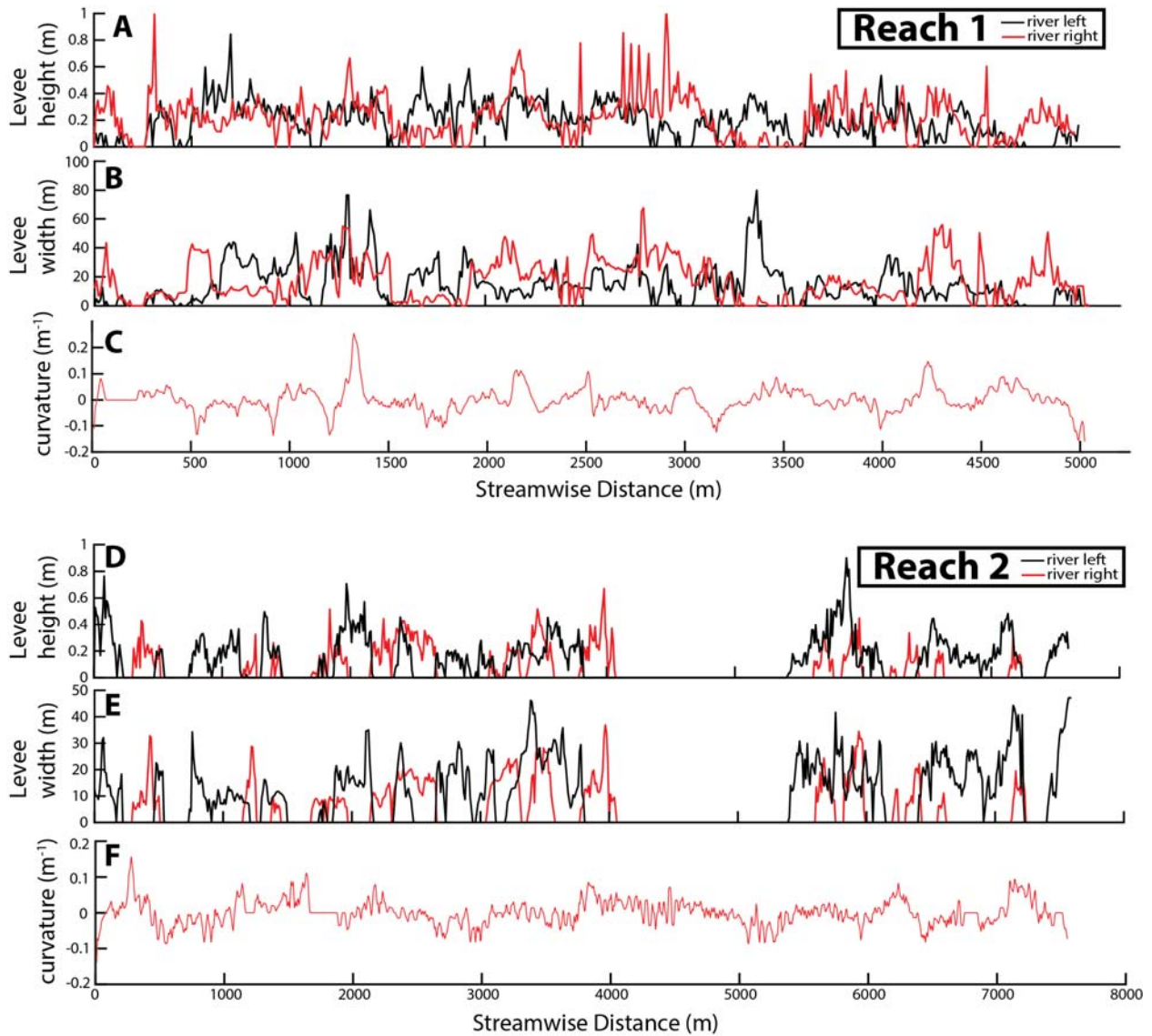
517

518 Figure 4. Examples of closed-contour polygons showing the spatial extent of the levee deposits
519 determined from levee toe contour elevation. The levee widths and height values were extracted for
520 each levee polygon. Levee width is measure along channel-perpendicular transects clipped to the levee
521 polygon extent, as shown in inset box. Colors indicate river right or river left side.



522

523 Figure 5. Levee presence along Reaches 1 and 2. Levee length is measured cumulatively and
524 nondimensionalized by cumulative bank length. Increases and decreases in normalized levee length,
525 respectively, represent leveed and unleveed parts of the reach.



526

527 Figure 6. For both reaches, levee height (A, D) and width (B, E) are shown for every 2-meter channel-
528 perpendicular cross-section. (see example in Figure 4). Curvature (C, F) is calculated from the channel
529 centerline following Schwenk et al., (2017).

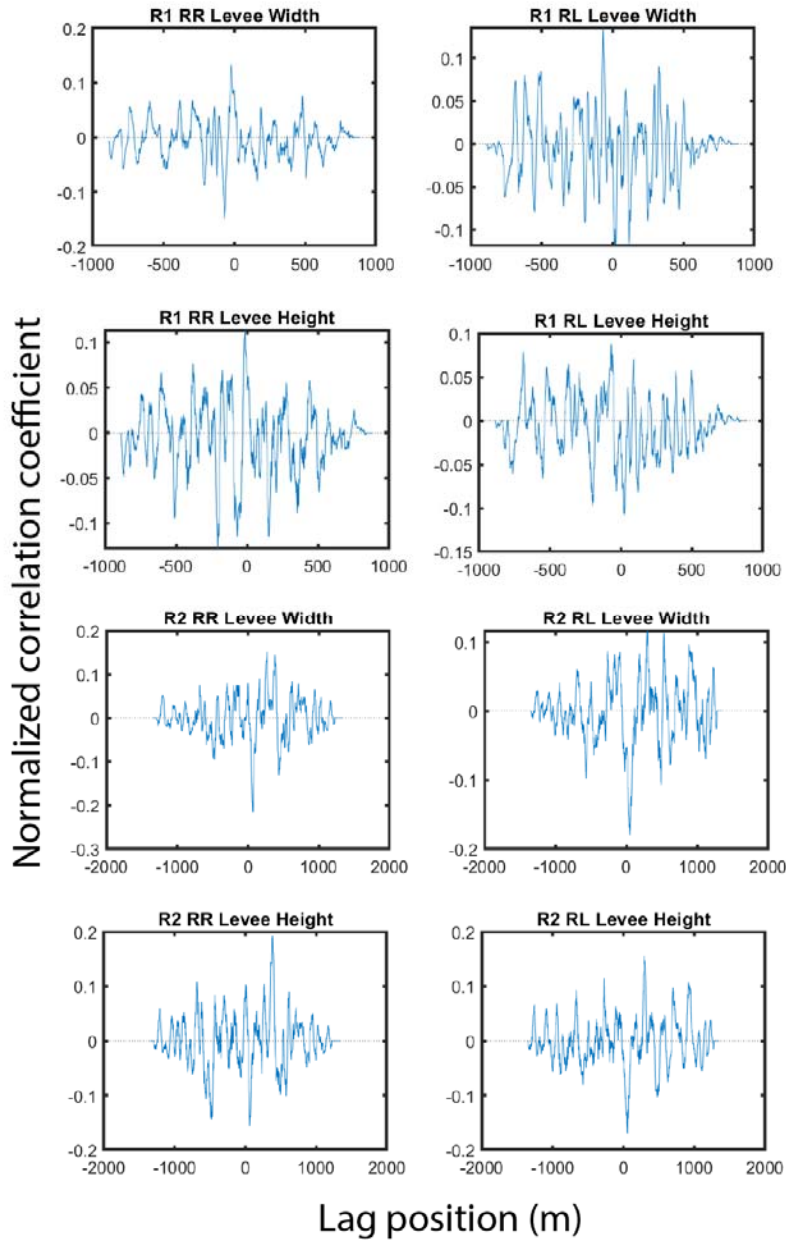
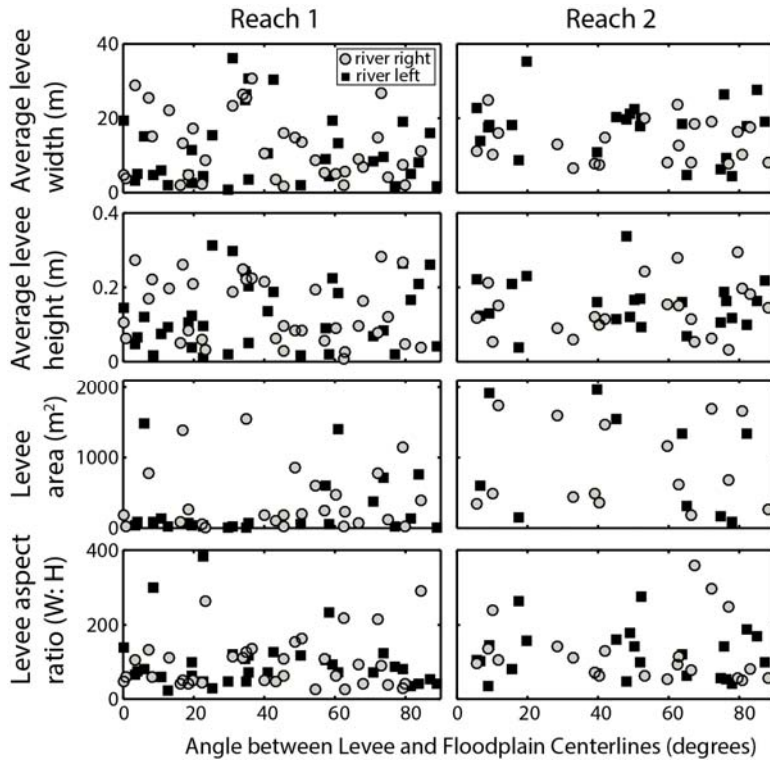
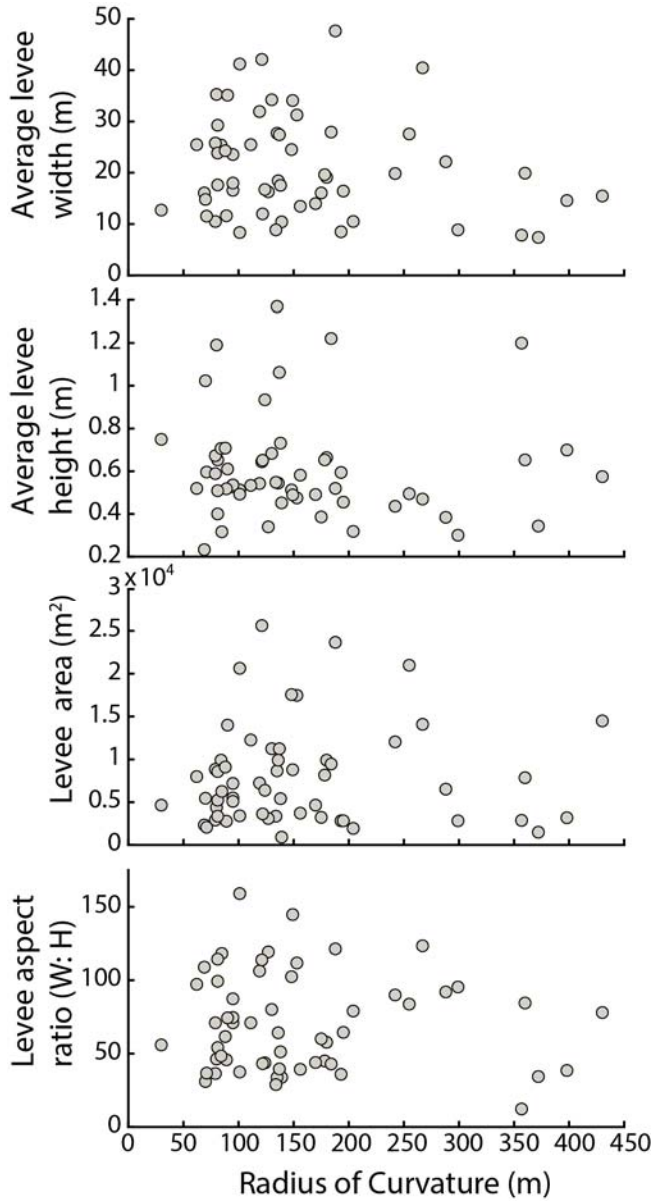


Figure 7. Cross-correlation function between channel curvature and levee height and width data at different lag positions. R1 – Reach 1; R2 – Reach 2; RR – River Right; RL – River Left.



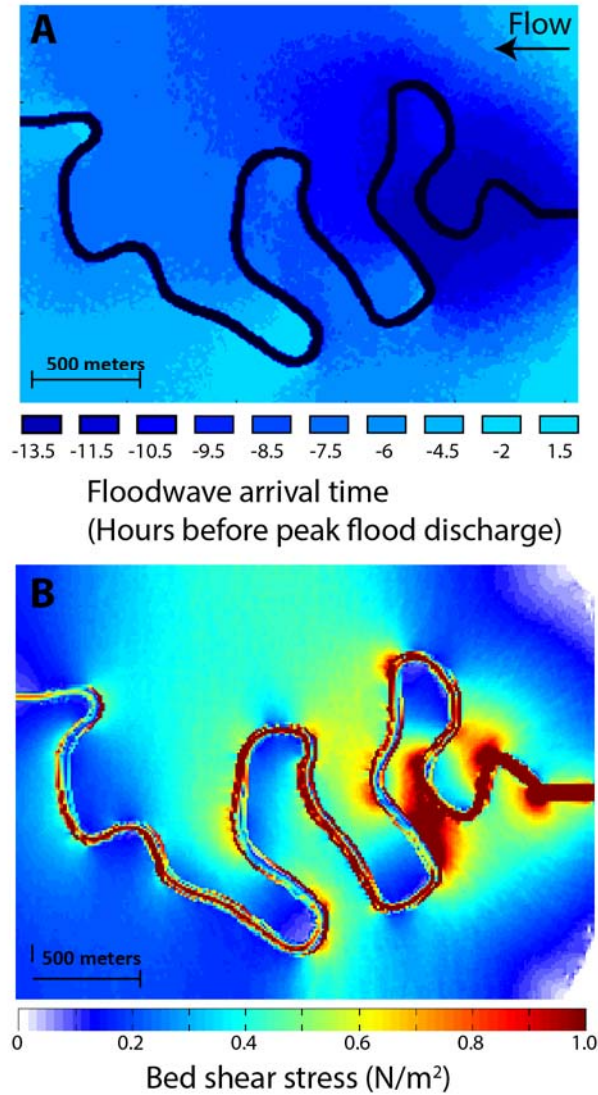
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534 Figure 8. Levee morphometrics plotted against relative orientation for Reaches 1 and 2 along the
535 Muscatatuck River. Relative orientation (in degrees) is the angle between the centerline of the levee and
536 floodplain. Zero indicates the long-axis of the levee is parallel to the floodplain centerline. Data are
537 average values for each closed-contour levee polygon (see example in Figure 4). See Figure 1 A,B for
538 location of these reaches.



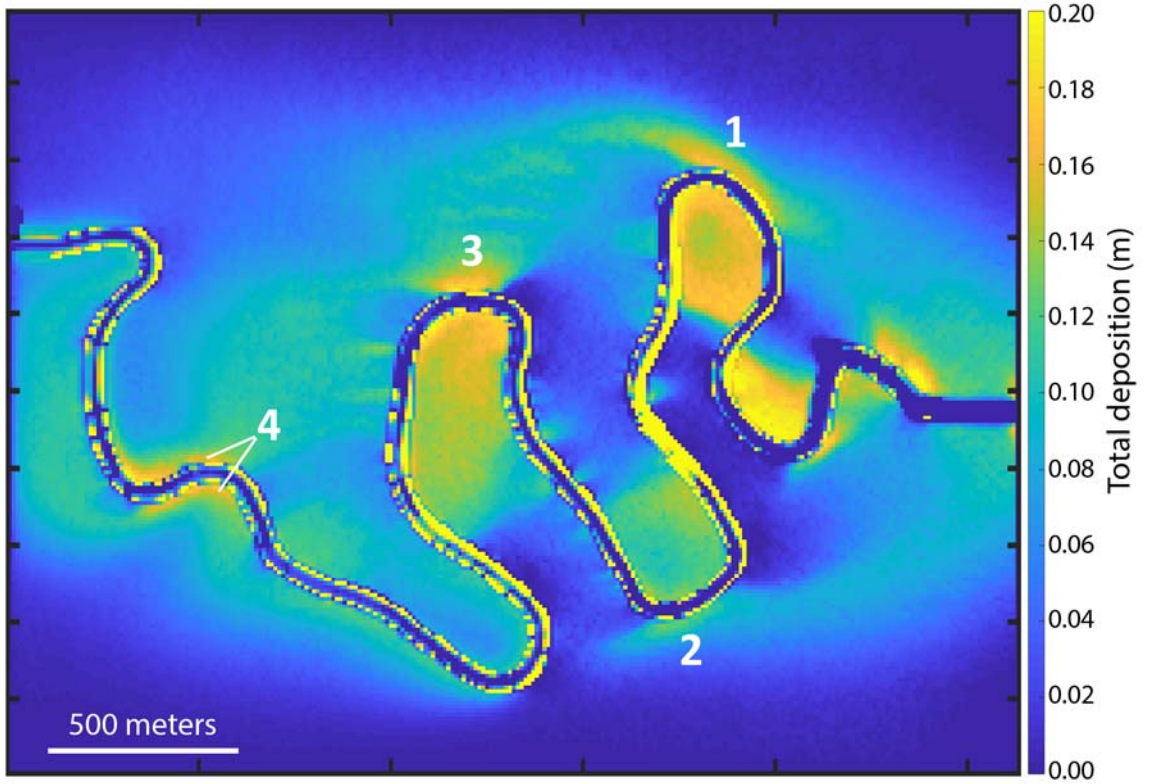
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540 Figure 9. Levee morphometrics plotted against channel radius of curvature for 54 meander bends along
541 the Muscatatuck River. See Figure 1C for location of this reach.



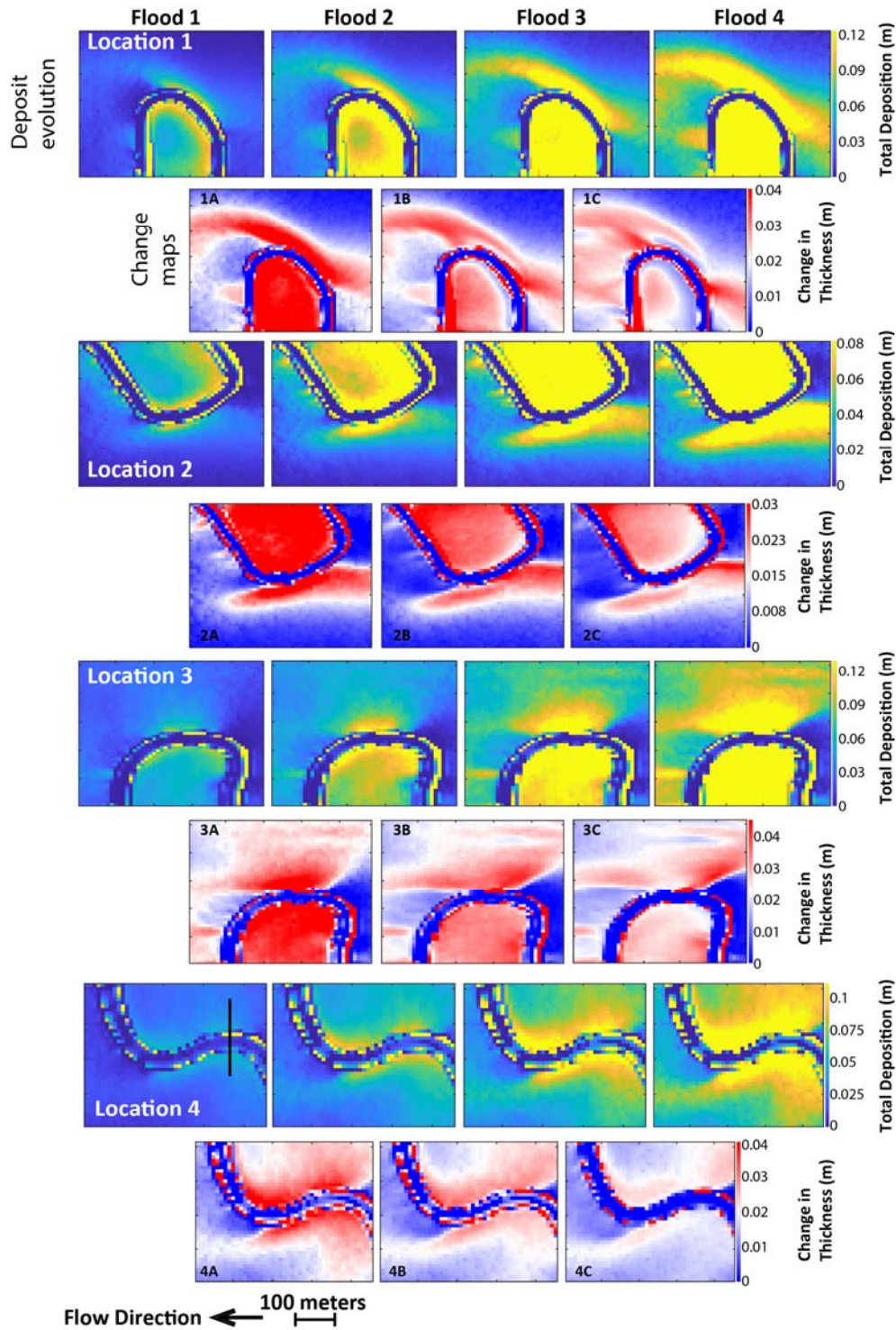
542

543 Figure 10: (A) Map of floodwave arrival time prior to any morphodynamic change. Floodwave arrival is
544 defined as inundation greater than 10 cm. The inundation front is noisy because of the random
545 topographic roughness. Negative numbers indicate floodwave arrival before peak flood. (B) Bed shear
546 stress at peak discharge prior to any morphodynamic change.



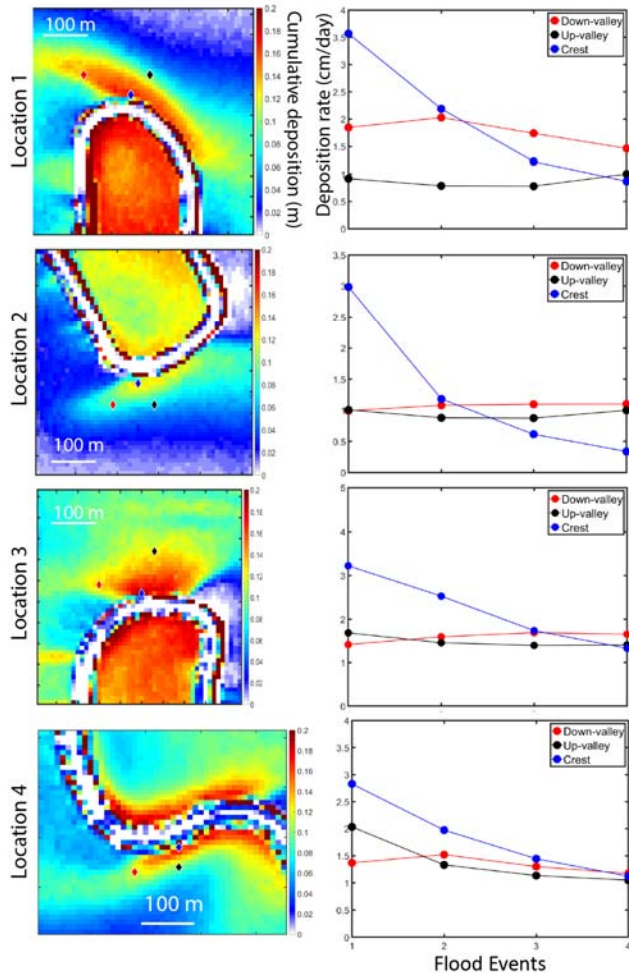
547

548 **Figure 11:** Total deposition in the model domain over the four flood-wave experiment. Numbers 1-4
549 refer to levees analyzed in Figures 12-14.



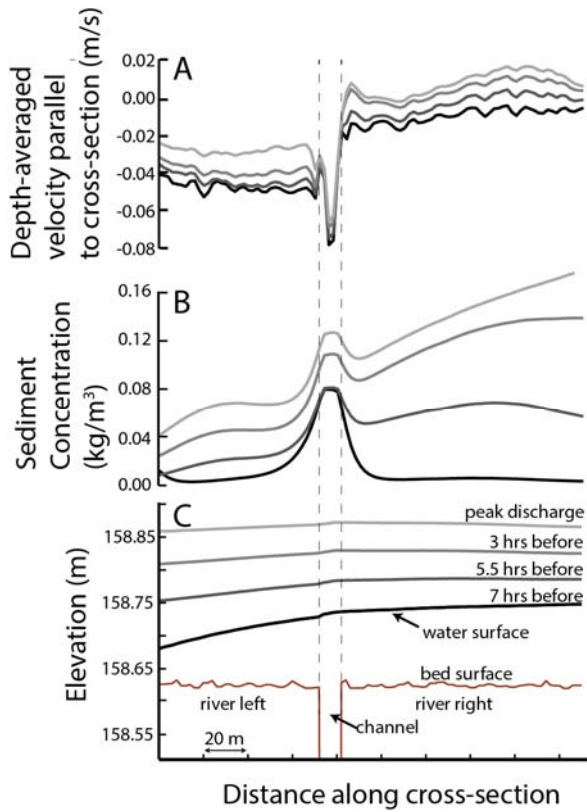
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551 Figure 12: Time evolution of levee thicknesses for locations 1-4. For each levee, the top row shows
552 development of the total deposit after each floodwave is complete, and bottom row shows the change
553 between each flood.



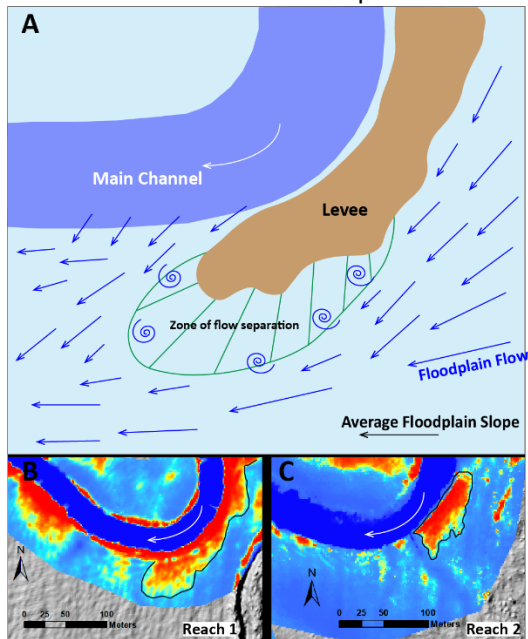
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555 Figure 13: Deposition rate on the levee crest decreases with time and all levees show a transition, by the
556 fourth flood wave, where the highest deposition rate is on the levee toe.



557

558 Figure 14: Time evolution of cross-section parallel flow velocity (A), sediment concentration (B), and
 559 water surface (C) at location 4 during flood 1. Flow is into the page, and negative velocities in C
 560 correspond to a southerly flow direction, while positive velocities mean northerly flow. Cross-section
 561 location is shown on the first panel for location 4 on Figure 12.



562

563 Figure 15: A) Cartoon showing how tall levees create flow shadows and deposition leading to down-
 564 valley progradation. B,C) Possible examples of progradational levee morphologies on Reaches 1 and 2.

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