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Interaction between meander dynamics and floodplain heterogeneity in a large tropical sand-bed river: the Rio Beni, Bolivian Amazon

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Keywords:	Meander migration, floodplain heterogeneity, bank erosion, numerical model, planform evolution

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3 1 *Interaction between meander dynamics and floodplain heterogeneity in a large*
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5 2 *tropical sand-bed river: the Rio Beni, Bolivian Amazon*

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25 9 **Abstract**

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27
28 10 The evolution of meandering river floodplains is predominantly controlled by the
29
30 11 interplay between overbank sedimentation and channel migration. The resulting
31
32 12 spatial heterogeneity in floodplain deposits leads to variability in bank erodibility,
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34 13 which in turn influences channel migration and planform development. Despite the
35
36 14 potential significance of these feedbacks, few studies have quantified their impact
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38 15 upon channel evolution and floodplain construction in dynamic settings (e.g.,
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40 16 locations characterized by rapid channel migration and high rates of overbank
41
42 17 sedimentation). This study employs a combination of field observations, GIS analysis
43
44 18 of satellite imagery and numerical modelling to investigate these issues along a 375
45
46 19 km reach of the Rio Beni in the Bolivian Amazon. Results demonstrate that the
47
48 20 occurrence of clay-rich floodplain deposits promotes a significant reduction in
49
50 21 channel migration rates and distinctive styles of channel evolution, including channel
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52 22 straightening and immobilisation of bend apices leading to channel narrowing. Clay
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54 23 bodies act as stable locations limiting the propagation of planform disturbances in
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1
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3 24 both upstream and downstream directions, and operate as 'hinge' points, around
4
5 25 which the channel migrates. Spatial variations in the erodibility of clay-rich floodplain
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7 26 material also promote large-scale (10-50 km) differences in channel sinuosity and
8
9 27 migration, although these variables are also likely to be influenced by channel
10
11 28 gradient and tectonic effects that are difficult to quantify. Numerical model results
12
13 29 suggest that spatial heterogeneity in bank erodibility, driven by variable bank
14
15 30 composition, may force a substantial (c. 30%) reduction in average channel
16
17 31 sinuosity, compared to situations in which bank strength is spatially homogeneous.
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24
25 33 **Keywords**
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27 34 Meander migration; floodplain heterogeneity; bank erosion; numerical model;
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29 35 planform evolution
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36 Introduction

37 Understanding the relationship between meander migration and floodplain evolution
38 is important for a wide range of issues, including river bank erosion and widening,
39 supply of bedload and suspended sediment to the channel, and the associated
40 deposition of sediment on in-channel bars and floodplain surfaces (Nanson and
41 Hickin, 1986; Salo et al., 1986; Lauer and Parker, 2008). Moreover, these processes
42 represent key controls on channel conveyance capacity, flood frequency, long-term
43 floodplain morphodynamics and the ecological functioning of the channel-floodplain
44 environment (Ward et al., 2002; Gueneralp et al., 2012). These issues are thus a
45 primary concern in many areas of river management, including river restoration,
46 floodplain land use and contamination, flood prevention and navigation.

47 The dynamics of meandering rivers has been the subject of intense research
48 over the past four decades, from multiple perspectives. For example, many studies
49 have attempted to classify channel behaviour (e.g., Brice, 1974; Hickin, 1974;
50 Hooke, 1984; Hooke, 2003) and reproduce or explain it using mathematical models
51 (e.g., Ikeda et al., 1981; Ferguson, 1984; Howard and Knutson, 1984; Johannesson
52 and Parker, 1989; Howard, 1992; Zolezzi and Seminara, 2001). Migration of bends
53 has often been explored with respect to freely-meandering rivers in relatively
54 homogeneous floodplains, where characteristic planform patterns have been
55 described (e.g., Hooke, 1995). Numerical models have been shown to be capable of
56 generating realistic planform configurations (Lancaster and Bras, 2002; Camporeale
57 et al., 2005; Frascati and Lanzoni, 2010), including compounding bends, asymmetric
58 up-valley skewing and formation of multi-bend loops. However, understanding of the
59 role of variability in bank strength as a control on meander migration remains
60 incomplete.

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3 61 Variable resistance to erosion of river banks may be due to vegetation
4
5 62 (Perucca et al., 2007), slump blocks (Parker et al., 2011), drift wood, sedimentology
6
7 63 and pedological evolution of bank sediments (Constantine et al., 2009), bedrock
8
9 64 (Limaye and Lamb, 2014) or differences in bank height (van de Wiel and Darby,
10
11 65 2007; Xu et al., 2011). Numerous studies have investigated how such variations in
12
13 66 bank strength impact on meander migration (e.g., Howard, 1996; Sun et al., 1996;
14
15 67 Huang and Nanson, 1998; Hudson and Kesel, 2000; Seminara, 2006, Gueneralp
16
17 68 and Rhoads, 2011; Motta et al., 2012a; Posner and Duan, 2012; Limaye and Lamb,
18
19 69 2014). However, the majority of this work has been based on numerical modelling
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21 70 rather than empirical evidence, due to the relatively short record of high quality
22
23 71 imagery (e.g. c. 40 years in the case of satellite data) available for the study of
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25 72 channel migration at high temporal resolutions. Modelling studies suggest that a
26
27 73 decrease in channel belt width may occur where bank erodibility is spatially
28
29 74 heterogeneous (Sun et al., 1996; Gueneralp and Rhoads, 2011). However, the
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31 75 implications for meander geometry remain to be resolved fully. For example, Sun et
32
33 76 al. (1996) found that floodplain heterogeneity has a limited impact on bend
34
35 77 wavelength, while Gueneralp and Rhoads (2011) show that it may promote
36
37 78 compound bends and downstream-skewing of meanders normally associated with
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39 79 super-resonant conditions (Camporeale and Ridolfi, 2006; Seminara, 2006). Huang
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41 80 and Nanson (1998) found an influence of variable bank strength on channel
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43 81 geometry, in particular width, but indicate that its impact is limited compared to
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45 82 hydraulic factors.

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52
53 83 Significantly, several modelling studies have examined the effects of a
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55 84 random spatial distribution of floodplain erodibility (e.g., Gueneralp and Rhoads,
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57 85 2011) or have employed a stochastic model in which mean erodibility decreases with
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3 86 distance from the channel (e.g., Motta et al., 2012a). These studies demonstrate the
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5 87 scale-dependent influence of floodplain heterogeneity on channel planform
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7 88 complexity. However, floodplain heterogeneity is unlikely to vary in a way that is
8
9 89 random, but is instead likely to be controlled by the spatial scaling of, and
10
11 90 interactions between, floodplain morphology, hydrodynamics, vegetation, and
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13 91 sedimentation processes. Thus the role of sedimentary heterogeneity in floodplain
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15 92 evolution and meander migration requires further research (Güneralp et al., 2012),
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17 93 particularly in the context of the complexity found in natural landscapes.
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21 94 The aim of this paper is to quantify the influence of variations in bank
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23 95 composition on meander migration within large, dynamic sand-bed rivers, using field
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25 96 and remote sensing datasets obtained from the Rio Beni in the Bolivian Amazon.
26
27 97 The specific objectives of the work are threefold: First, to assess the influence of
28
29 98 bank material on rate and style of channel migration at individual meander bends.
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31 99 Second, to examine channel evolution over multiple bends, in order to elucidate the
32
33 100 role of floodplain heterogeneity as a control on large-scale channel belt
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35 101 characteristics and on the propagation of planform irregularities between bends.
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37 102 Third, to explore the potential for simulating and explaining these characteristics
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39 103 using a simple numerical model of meander migration.
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48 105 **Study site**

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50 106 The Rio Beni was chosen for this study due to its extensive floodplain, which is
51
52 107 essentially undisturbed by human influence, and its high rates of meander migration
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54 108 and floodplain sedimentation within a single active channel belt. The reach of the Rio
55
56 109 Beni examined herein is located in the Andean foreland basin in north-eastern
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3 110 Bolivia (Fig. 1), and has been largely unaffected by the effects of Holocene sea level
4
5 111 change. The upstream end of this reach is near Rurrenabaque, where the Beni
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7 112 leaves the piedmont of the Andes (Serrania el Susi) and meanders for approximately
8
9 113 375 km (channel length) through the forested 'Llanos de Mojos', a floodplain built up
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11 114 of late-Miocene and Quaternary sediments (Dumont, 1996; Gautier et al., 2007).
12
13 115 Catchment area at Rurrenabaque is 68,000 km² (Gautier et al., 2010), mean channel
14
15 116 width at low flow is 430 m, mean discharge is 2,300 m³s⁻¹, and annual flood peaks
16
17 117 frequently exceed 20,000 m³s⁻¹ (Environmental Research Observatory (ORE)
18
19 118 HyBAM). The Rio Beni transports a comparatively high sediment load of 219 x 10⁶ t
20
21 119 a⁻¹ (Latrubesse and Restrepo, 2014), which can be characterised as of fresh Andean
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23 120 origin (Guyot et al., 2007) and constitutes 72% of the load of the Rio Madeira (Guyot
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25 121 et al., 1999).
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30 122 Water surface slope decreases dramatically within the upstream section of the
31
32 123 study reach where the Rio Beni leaves the piedmont fan and bed material changes
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34 124 from cobble-gravel to sand. Downstream of this point, channel slope declines from
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36 125 0.0002 to 0.00007 m m⁻¹ over a distance of 300 km. Beyond this (over the final 75
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38 126 km) the river profile steepens, although the paucity of reliable dGPS data make it
39
40 127 difficult to quantify the gradient with confidence (see also Gautier et al., 2007). Mean
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42 128 channel sinuosity within the study reach varies temporally (between 1.8 and 2.0) and
43
44 129 spatially amongst sub-reaches (between 1.3 and 2.7) (see also Dumont, 1996,
45
46 130 Gautier et al., 2007). Downstream of the piedmont fan, median sediment size of bed
47
48 131 and suspended load is relatively constant ranging between 0.09 - 0.15 mm and
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50 132 0.0094 - 0.012 mm respectively (Guyot et al., 1999). The channel and its proximal
51
52 133 floodplain are largely unaffected by anthropogenic modification such as bank
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54 134 protection, dredging or deforestation (Aalto et al., 2003).
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3 135 The Beni channel belt has experienced a counterclockwise shift from a
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5 136 northeast orientation (a position currently occupied by the Rio Yacumu) to a more
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7 137 northerly course during the Holocene (Plafker, 1964). Moreover, the deflection point
8
9 138 migrated northward following a north-striking fault line that separates old upper fluvial
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11 139 terraces and hardened clay sediments in the NW from younger floodplain sediments
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13
14 140 in the southeast (Dumont and Hannagarth, 1993; Dumont, 1996). The migration of
15
16 141 the channel belt also responds to differential subsidence and uplift patterns aligned
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18 142 with southwest – northeast striking lineaments in the Brazilian Craton (Plafker, 1964;
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20
21 143 Allenby, 1988).
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25 26 27 145 **Methods**

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30 146 Field data were acquired during visits to the study area between 2011 and 2013.
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32 147 Data include measurements of water surface slope and ground elevation obtained
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34 148 using dGPS (XRT, Trimble Navigation Ltd, Sunnyvale, USA) in conjunction with real-
35
36 149 time OmniSTAR HP correction or post-processing (CSRS-PrecisePointPositioning,
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38 150 Natural Resources Canada, Ottawa, Canada). Relative bank height (the difference
39
40 151 between the bank top and low flow water level) was measured from a boat with a
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42 152 dGPS supported laser range finder (Impulse 200 LR, Laser Technology Inc.,
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44 153 Centennial, USA). These measurements were taken on cut bank and point bar sides
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46 154 of bends and along straight sections in intervals of approximately 100 m. Differences
47
48 155 in bank composition were mapped using digital photographs of bank sections along
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50 156 approximately 350 km of channel combined with sampling of bank sediments (n =
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52 157 67) for laboratory grain size analysis, carried out using a Sedigraph 5100
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56 158 (Micromeritics Instrument Corp., Norcross, USA). Bank material was sampled at a
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3 159 number of heights above the water level at representative locations for each bank
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5 160 material class.
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8 161 Rates and styles of river migration were quantified by digitizing channel bank
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10 162 lines in ArcGIS (ESRI, Redlands, USA) from geo-referenced multispectral Landsat
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12 163 imagery (Table 1), taken during dry season (May to September), for 18 years
13
14 164 between 1975 and 2011 (Gautier et al., 2007). Additional bank lines were acquired
15
16 165 from aerial photography collected in 1960 (provided by the Bolivian Navy, see also
17
18 166 Plafker, 1964) that covered the upper 280 km of the study reach (Table 1). Individual
19
20 167 bends were numbered from the upstream end of the reach (117 bends in total). It
21
22 168 should be noted that not all bends are present over the entire period of study (due to
23
24 169 periodic bend initiation and abandonment). The study reach was divided into 19 sub-
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26 170 reaches (mean length ~ 20 km) at locations where the channel has experienced only
27
28 171 minor lateral migration over the past 50 years (see Fig. 1).
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33 172 Bank lines were converted to centrelines, which were then resampled to a
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35 173 node spacing of 100 m (approximately a quarter of one mean channel width) for use
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37 174 in subsequent analysis. Curvature was calculated following Motta et al. (2012b;
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39 175 equation 15), while migration rate at bends was measured as the area between two
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41 176 centrelines divided by the bend length, whereby bends are bounded in up- and
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43 177 downstream directions by points of inflection of the centreline. Apparent migration
44
45 178 associated with centerline movement following bend cutoff has been excluded from
46
47 179 all calculations. It should be noted that migration rates calculated from image pairs
48
49 180 may be sensitive to the time period between images. For example, where the
50
51 181 channel does not move in a consistent direction over time the migration rate from a
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53 182 single pair of images may be under-estimated (i.e. where the migration direction
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55 183 reverses during the period covered by the image pair). Between 2003 and 2011,
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3 184 errors introduced by image rectification were found to be negligible, due to precise
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5 185 pre-rectification of the Landsat images. For older images, migration distances of less
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7 186 than 16 m can be affected by image rectification errors (RMSE < 16 m relative to
8
9 187 2000 image; Table 1). The mean random error induced by pixel resolution (30 m) is
10
11 188 expected to be close to zero over an entire bank line.

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15 189 Bend evolution was investigated at 117 bends over a period of 51 years. This
16
17 190 involved visual assessment of individual bends in ArcGIS and classification
18
19 191 according to channel migration style (see Fig. 2; see also Hooke, 1984). Styles of
20
21 192 migration include: a) longitudinal expansion or contraction of the bend (i.e., changes
22
23 193 in bend wave length); b) lateral extension or contraction (i.e., changes in bend
24
25 194 amplitude); c) confined or unconfined translation longitudinally up or down the valley;
26
27 195 d) lateral displacement (bend migration without alteration of planform shape between
28
29 196 points of inflection); e) bend rotation; and f) no change (stable). Complex, irregular
30
31 197 and compounding patterns were separated from simple bend evolution styles. Some
32
33 198 bends are characterised by mixed styles of migration involving several elements of
34
35 199 the behaviour outlined above. Thus this classification results in numerous
36
37 200 combinations of the basic types of channel change, which were then generalised and
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39 201 grouped into 7 common migration styles in order to remove some of the subjectivity
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41 202 introduced by the visual assessment. The focus of this classification is on dominant
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43 203 style of migration and not the quantification of the magnitude of displacement.

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49 204 In order to explore the relationship between channel geometry, migration rate
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51 205 and bank erodibility further, numerical simulations were carried out using a simple
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53 206 model of meander migration and floodplain sedimentation. The approach adopted
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55 207 herein follows that of Howard (1992), differing only in the detail of the model
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57 208 formulation (see below). Specifically, we simulate overbank sedimentation using a
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3 209 form of exponential decay law, as is common in models of long-term floodplain
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5 210 evolution (e.g., Howard, 1992; Mackey and Bridge, 1995):
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9 211
$$D_i = C_i H^{1.5} e^{-\alpha_i x} \quad (1)$$

10

11 212 where D_i is the deposition rate for size fraction i , C_i and α_i are a grain size dependent
12
13 213 constant and decay coefficient, H is the height difference between the floodplain
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15 214 surface and an assumed maximum flood water level, and x is distance to the nearest
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17 215 channel. The non-linear dependence of D upon H reflects the increased frequency of
18
19 216 inundation of low-lying floodplain areas. In the current model application, two grain
20
21 217 size fractions have been used (one fine fraction and one coarse fraction). They are
22
23 218 not intended to represent specific sediment sizes because the model should be
24
25 219 considered phenomenological rather than physically-based. The parameters C_i and
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27 220 α_i were assigned values of 0.035 and 0.0015, respectively, for the coarse size
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29 221 fraction, and 0.005 and 0.00015, respectively, for the fine size fraction. These values
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31 222 were selected to approximate the decline in sedimentation rates and deposit grain
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33 223 size observed for the Beni by Aalto *et al.* (2003). This equation is applied over a grid
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35 224 of cells to update the floodplain grain size composition (and topography) during each
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37 225 model iteration.
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43 226 Meander migration is simulated herein using the model of Howard and
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45 227 Knutson (1984), implemented using the parameterization that is equivalent to the
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47 228 approach of Ikeda *et al.* (1981). In this model, migration rates are a product of the
48
49 229 weighted sum of local and upstream channel curvatures, and a local bank erodibility
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51 230 coefficient. Thus the detail of flow and sediment transport are not represented and
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53 231 channel migration is a function of planform geometry and bank strength only.
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56 232 Channel curvature is calculated from the coordinates of nodes spaced at one half
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3 233 mean channel width along the centerline. Migration leads to movement of these
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5 234 nodes and, ultimately, to neck cutoff (when two sections of channel migrate close to
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7 235 one another). Chute cutoffs are not modelled herein, and are rare along the Beni.
8
9
10 236 Erodibility is defined as a function of the floodplain grain size composition in the grid
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12 237 cell into which the channel is migrating. The relationship between bank erodibility (E)
13
14 238 and the fraction of the floodplain composed of fines (F) is represented by:

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16
17 239
$$E = \beta(0.05+0.95(1-F)^k) \quad (2)$$

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19

20
21 240 where β and k are constants. The value of β controls the average rate of channel
22
23 241 migration, but not the dependence of migration on floodplain heterogeneity. The
24
25 242 value of k determines the strength of the relationship between erodibility and bank
26
27 243 composition (a higher value of k yields a stronger grain size dependence). The form
28
29 244 of equation (2) was chosen by combining relationships between bank silt-clay
30
31 245 content, critical shear stress, and bank erosion rate presented by Julian and Torres
32
33 246 (2006; their Figures 4 and 6). They report an inverse relationship between critical
34
35 247 shear stress and bank erosion rates, which would lead to infinite erosion rates where
36
37 248 F tends to 0. Consequently, equation (2) is adopted herein, which overcomes this
38
39 249 problem and provides a good fit to the relationships shown by Julian and Torres
40
41 250 (2006) where $k = 6$. The combined bank erosion and floodplain sedimentation model
42
43 251 was implemented herein using $\beta = 4$ and $k = 6$, $\beta = 2$ and $k = 3$ (i.e. reduced
44
45 252 dependence of E on F), and $\beta = 1$ and $k = 0$ (i.e. bank erodibility independent of
46
47 253 floodplain grain size composition). Simulations used a floodplain domain with
48
49 254 dimensions of 125 km (downstream) by 50 km (cross-stream) and a grid resolution
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51 255 of 500 m. The model is initialized using a flat floodplain and straight channel with
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53 256 very small random perturbations in channel centerline coordinates.
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3 257 **Results**
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5
6 258 *Rates and styles of migration at individual bends*
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8
9 259 Cutbanks characterised by clay-rich sediments (hereafter also termed clay banks)
10
11 260 were mapped at 19 bends during several field visits between 2003 and 2011. Such
12
13 261 banks were identified by their grain size assemblage and their distinctive colour and
14
15 262 geometry. For example, clay-rich banks often contain ferrous concretions, giving
16
17 263 them a speckled appearance, while on the lower bank the matrix material is often of
18
19 264 a greyish colour, indicating oxygen depletion over prolonged periods (see Fig. 3).
20
21 265 The coherent nature of these banks leads to a stable, often slightly less than vertical
22
23 266 upper bank geometry, below which banks can have a scalloped shape with regular
24
25 267 protrusions into the channel. The average grain size composition of banks identified
26
27 268 visually as clay-rich was found to be 39-84% clay and 15-61% silt, compared with
28
29 269 15-25% clay and 75-83% silt in other bank sections. Sand constitutes a small
30
31 270 fraction in most banks with a range of 0-11%. The D_{50} of bank material classified as
32
33 271 clay-rich ranges between <0.6 and $3.42 \mu\text{m}$ but is typically smaller than $2 \mu\text{m}$.
34
35 272 Although clay banks are unusually high in some places (e.g., due to tectonic uplift at
36
37 273 the upstream end of the study site, in sub-reach 1) their mean height above low flow
38
39 274 water level over the entire study reach (6.47 m) is not significantly higher than for
40
41 275 banks composed of any other substrate (6.40 m) (two-sample t test, $\alpha = 0.05$).
42
43 276 Moreover, there is no significant relationship between migration rate and bank height
44
45 277 evident within the study reach as a whole (Pearson's $r = -0.24$, $p = 0.067$, $n = 61$).
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52 278 Individual clay-rich banks vary in grain size, extent of reddish concretions,
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54 279 height and erodibility, which may reflect differences in their age and post-
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56 280 depositional development. Despite this variability, locations occupied by clay banks
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3 281 experience significantly lower mean rates of channel migration at the scale of whole
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5 282 bends (two-sample t test, $p < 0.05$) compared to bends with cutbanks made of any
6
7 283 other alluvial deposit (Fig. 4). Mean annual bend migration rate of individual bends
8
9
10 284 ranges between 3.4 ma^{-1} and 530.7 ma^{-1} . It is lowest at bends with clay banks (23.8
11
12 285 ma^{-1}), followed by bends migrating into mixed substrates (37.3 ma^{-1}), point bar
13
14 286 deposits (43.3 ma^{-1}) and infilled channels (58.8 ma^{-1}). Mean migration rate is highest
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16 287 at bends migrating into oxbow lakes (98.8 ma^{-1}), although migration rates are highly
17
18 288 variable in such cases and depend on cutoff age (degree of fill), substrate
19
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21 289 (erodibility) and angle of approach by the migrating bend. Mean annual bend
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23 290 migration rates show a significant correlation ($\alpha = 0.05$) with a number of discharge
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25 291 metrics such as accumulated discharge during wet seasons (December to April),
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27 292 maximum annual discharge and the number of days with discharge in excess of
28
29 293 bankfull ($6000 \text{ m}^3 \text{ s}^{-1}$) with Pearson's r of 0.66, 0.71 and 0.82, respectively.
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32 294 Correlations based on metrics that summarise the previous three wet seasons are
33
34 295 stronger than those based on the previous wet season only.

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37 296 The channel in the region of clay banks is characterised by distinct narrowing
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39 297 just downstream of the apex, often due to a resistant notch of clay protruding into the
40
41 298 channel (average channel width of 290 m in 2011, for $n = 22$ bends). Channel width
42
43 299 at the equivalent position for bends without clay banks is 534 m ($n = 63$). Despite the
44
45 300 particular planform configuration at these bends, no specific pattern in water surface
46
47 301 slope could be established. Bend migration rate is not correlated to mean slope at
48
49 302 various distances upstream of, downstream of, or around bend apices.
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53 303 Classification of bend migration style (over the period from 1960 to 2011)
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55 304 indicates that 28.1% of the bends are relatively immobile, as indicated by their low
56
57 305 mean rate of migration (26.0 ma^{-1}). Sub-reaches dominated by clay-rich banks
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3 306 (especially sub-reaches 1 and 12, Fig. 5) are characterised by a high proportion of
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5 307 immobile bends. Wavelength expansion is apparent at 13.3% of bends, often in
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7 308 combination with bend rotation and/or changes in amplitude. This occurs mainly in
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10 309 the more mobile sub-reaches located in the central part of the study reach. In
11
12 310 contrast 16.8% of bends experienced a reduction in wavelength, often accompanied
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14 311 by lateral extension (e.g., a common style in the early phase of bend development,
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16 312 just after bend inception) and/or rotation, but, in a few cases, also by a reduction in
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18 313 bend amplitude.

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21 314 Channel belt widening, as signified by the bend extension and lateral displacement
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23 315 without a change in bend shape, occurs at 30.1% of bends, and is typical of mobile
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25 316 sub-reaches with few clay-rich banks (e.g., sub-reaches 6, 9, 10, 13, 16). Bend
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27 317 rotation is also prominent in these sub-reaches (11.1% of bends) and takes place
28
29 318 more frequently in a downstream direction. Translation is experienced by 14.8% of
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31 319 bends, is most significant in sub-reaches 2, 4, 8 and 14, and is not preferentially
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33 320 associated with particular bank composition. Up-valley bend translation is rare and
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35 321 occurs mainly at a slow rate and is often associated with migration along a clay-rich
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37 322 bank. A small proportion (2.1%) of all bends showed signs of compounding and
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39 323 increased complexity. Compound bend development occurs in slowly expanding
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41 324 meanders (e.g., in sub-reach 17), as a result of flow diversion at bifurcations (e.g., in
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43 325 sub-reach 4) or partial contact with a clay body (e.g., in sub-reach 8).

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45 326 Bends with clay-rich banks may share common planform and evolutionary
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47 327 characteristics, depending on the precise location of clay bodies. For example,
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49 328 migration of the apex of a bend into a clay body can lead to wavelength expansion,
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51 329 eventually followed by compounding (Fig. 2a). In the upstream part of the study
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53 330 reach the active channel belt appears to be confined by elongated clay bodies
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3 331 parallel to the valley axis. In this region, the entire active cutbank of some bends
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5 332 consists of clay-rich deposits, which usually render the bend immobile with limited
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7 333 up- or down-valley translation (e.g., bend 13 up-valley, Fig. 6). In such cases, some
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9 334 upstream rotation of the bend around the apex may occur. The downstream limb of
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11 335 bends with clay-rich banks tends to elongate and straighten with time. Since
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13 336 meanders not constricted by clay-rich banks typically have shorter life spans (i.e.
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15 337 they evolve to cutoff more quickly), and higher rates of migration and down-valley
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17 338 translation, bends located upstream of less mobile clay-rich meanders can appear to
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19 339 move down-valley past the relative fixed apices of clay banks. As a result, bends
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21 340 with clay-rich banks often appear in planform to be skewed up-valley, with the
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23 341 downstream limb being laterally immobilised by contact with the clay-rich bank (e.g.,
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25 342 bend 14 in Fig. 6). Consequently, over time, bends with clay-rich banks can
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27 343 experience a reduction in wave length due to faster migration of the more mobile
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29 344 upstream limb than of the downstream limb, which eventually leads to a neck cutoff.
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346 **Planform evolution at multi-bend scales**

347 Downstream of clay banks the channel planform often exhibits straightening unless
348 influenced by contact with other clay bodies (Fig. 7). This process involves a gradual
349 reduction in sinuosity over several years, in contrast to rapid channel shortening due
350 to bend cutoff. Channel straightening usually involves an increase in wavelength and
351 contraction in amplitude of bends, for example confined downstream translation,
352 when the downstream limb of a bend translates faster down-valley than the
353 upstream limb (as visible in sub-reaches 1, 8, 10, 12). Such straightening
354 downstream of bends with clay-rich banks may be terminated when the bend in

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3 355 question is cutoff. The steady lengthening of straight reaches can be accelerated by
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5 356 cutoffs, or slowed down or reversed where the channel that is straightening migrates
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7 357 into another clay body (dashed lines in Fig. 7), which introduces new bends and thus
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9 358 increases sinuosity. Moreover, in locations characterised by many closely spaced
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11 359 clay bodies, straights may be unable to form, and cutoffs may be frequent (e.g. sub-
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13 360 reaches 4 and 6). Sustained channel straightening over distances greater than 10
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15 361 km and periods longer than 10 years is rare and only found in sub-reaches 7 and 15
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17 362 in the last 50 years. For example, in the latter sub-reach, sinuosity declined from
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19 363 1.31 to 1.24 over a 15 year period follow a cutoff in 1960. This process appears to
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21 364 have been induced by contact with the clay body on the western edge of the channel
22
23 365 belt. More recently, the channel in sub-reach 7 (see Fig. 8) was characterised by
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25 366 straightening between 1998 and 2005 with a reduction in sinuosity from 1.82 to 1.32.
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27 367 Although this involved several cutoffs, these events were not followed by a gradual
28
29 368 increase in sinuosity, which is a typical response to channel shortening (Hooke,
30
31 369 2003). In this case, straightening may have been initiated by contact between the
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33 370 channel and a clay body in bend 35 at the upstream end of the sub-reach (Fig. 8 a,
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35 371 b) as early as 1987. This triggered a series of neck cutoffs at bends 36 and 39 (Fig.
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37 372 8 b and 8 c), accelerated bend migration down-valley, rapid planform adjustment via
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39 373 chute cutoffs at bends 41 and 43 (Fig. 8 d) and thus significant lengthening of the
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41 374 straight reach below bend 35 until 2005 (Fig. 8 e; see also Fig. 7). Although new
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43 375 planform perturbations developed upstream of former bend 39, the downstream
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45 376 translation of these perturbations has been impeded by a clay-rich bank at bend 40,
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47 377 which acts as a hinge with a fixed apex, thus limiting channel adjustment
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49 378 downstream (Fig. 8 e and 8 f). Consequently, the newly formed bend 38 has
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51 379 increased in amplitude and become gradually more asymmetric due to stabilisation of
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3 380 the apex of bend 40, leading to a steady increase in sinuosity since 2005. The
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5 381 overall tendency for clay-rich banks to induce channel straightening leads to an
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7 382 abundance of channel sections with low planform curvature. Moreover, the
8
9 383 frequency distribution of channel curvature for sub-reaches with numerous clay bank
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11 384 sections exhibits a distinct exponential form, compared to a more linear distribution
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13 385 in other sub-reaches (Fig. 9).

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17 386 The previous sections have highlighted the influence of clay banks on channel
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19 387 form at the scale of individual bends and sub-reaches (lengths of c. 20 km). Spatial
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21 388 variations in the frequency of clay bodies along the 375 km study reach also appear
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23 389 to promote changes in channel dynamics at larger spatial scales, both in terms of
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25 390 rates of bend migration and cutoff (Fig. 10) and channel sinuosity (Fig. 11). For
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27 391 example, in the upper part of the study reach (sub-reaches 1 to 5 and the upstream
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29 392 part of sub-reach 6) the active channel belt is laterally confined by clay bodies
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31 393 (elements of the Chore and Caupolican complexes in the northwest, and the Ichilo
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33 394 complex in the southeast; GEOBOL, 1979). Consequently, channel sinuosity is low
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35 395 in general (< 2 in most sub-reaches) and the upper end of this zone in particular has
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37 396 experienced low migration rates ($< 5 \text{ ma}^{-1}$ in large areas). Further north, the western
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39 397 side of the active channel belt is also bounded by clay bodies of the Ixiamas
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41 398 complex, while in the east relatively few clay contacts were found. In this area, sub-
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43 399 reaches 12 and 15 are strongly influenced by clay banks and exhibit a combination
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45 400 of low migration rates ($< 20 \text{ ma}^{-1}$) and low sinuosity (< 2). In contrast, in sub-reaches
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47 401 where meandering is relatively unconstrained (e.g., sub-reach 6, 9, 10, 14, 16)
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49 402 migration rates are higher (27.0 ma^{-1} on average), cutoffs are frequent (Fig. 10) and
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51 403 sinuosity is high (2.24 on average). Indeed, the transition between reaches in which
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53 404 confining clay bodies are common and absent can be associated with a marked
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3 405 change in channel planform character. For example, in sub-reach 6 downstream of
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5 406 bend 25 the channel leaves a zone where clay banks provide a strong stabilising
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7 407 influence. As a result, sinuosity increases and the channel belt widens, promoting
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9 408 rapid bend migration, increased cutoff frequency, rotation of bends, and the
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11 409 formation and cutoff of a multi-loop bend (Fig. 12). However, along the central
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13 410 section of the study reach, downstream propagation of this dynamic channel
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15 411 behaviour is prevented by the contact with the clay banks located further
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17 412 downstream (e.g. bend 35, Fig. 8).
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23 24 414 **Numerical modelling**

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27 415 The numerical model described above was applied to investigate the relationship
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29 416 between heterogeneity in bank erodibility and planform channel characteristics
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31 417 further. Model simulations were run for a period sufficient for the channel to develop
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33 418 a statistically steady form (i.e. without a long-term trend in channel sinuosity) and to
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35 419 rework the floodplain multiple times (8000 model iterations). Figure 13 shows the
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37 420 time series of channel sinuosity for model runs with contrasting k values. All
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39 421 simulations show a rapid initial increase in sinuosity as a meandering channel
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41 422 develops, as described by others (e.g., Howard and Knutson, 1984; Howard, 1992).
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43 423 Subsequently, sinuosity declines (following the first cutoff events) and then oscillates
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45 424 (reflecting the balance between channel lengthening, due to migration, and
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47 425 shortening, due to cutoffs). Figure 13 illustrates that spatial variability in erodibility
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49 426 (due to floodplain heterogeneity) promotes a substantial reduction in channel
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51 427 sinuosity. Moreover, simulation results are relatively insensitive to the precise value
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53 428 of k (for $k \geq 3$). For example, in the latter half of the simulation following the model
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3 429 spin-up phase, mean sinuosity is 2.87 ($k = 0$), 2.05 ($k = 3$), and 2.07 ($k = 6$). Figure
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5 430 14, shows that frequency distributions of channel curvature differ between
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7 431 simulations that neglect ($k = 0$) and account for floodplain heterogeneity ($k = 3$; note
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9 432 that the distribution for $k = 6$ is very similar). More specifically, spatial variability in
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11 433 heterogeneity promotes a curvature distribution similar to that observed on the Beni
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13 434 in sub-reaches where clay-rich banks are common, while neglecting spatial
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15 435 variability in erodibility leads to a distribution more similar to reaches along the Beni
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17 436 where clay banks are uncommon (Fig. 9).

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21 437 Figure 15 shows the simulated planform pattern of floodplain sediment
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23 438 heterogeneity after 3900 iterations and the meandering channel position (at two
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25 439 instants in time) during model runs with constant bank erodibility ($k = 0$) and bank
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27 440 erodibility dependent on grain size assemblage ($k = 3$). It is evident from these plots
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29 441 that spatial variability in bank erodibility leads to a reduction in active channel belt
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31 442 width (by c. 35% on average), meander migration rates (by c. 40%) and cutoff
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33 443 frequency (by c. 50%), where the latter is indicated in Figure 15 by the number of
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35 444 abandoned channels that are characterised by a high proportion of fine sediment.
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37 445 Moreover, simulations in which bank erosion is dependent on floodplain grain size
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39 446 composition show several of the features of channel migration observed along the
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41 447 Rio Beni. These include the development of angular bend configurations,
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43 448 compounding, straightening downstream of contacts with fine sediment deposits and
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45 449 a tendency for the channel to become relatively immobile at locations where bank
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47 450 sediment is fine.
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56 452 **Discussion**
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3 453 The results presented above demonstrate that the spatial distribution of clay-rich
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5 454 sediments within the floodplain of the Rio Beni exerts a significant influence on rates
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7 455 of bank erosion, styles of meander migration, and channel morphology at bend and
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9 456 channel-belt scales. The influence of floodplain heterogeneity on channel migration
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11 457 has been recognised in many previous studies (e.g. Howard, 1996; Sun et al., 1996;
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13 458 Huang and Nanson, 1998; Hudson and Kesel, 2000; Seminara, 2006, Gueneralp
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15 459 and Rhoads, 2011; Motta et al., 2012a; Posner and Duan, 2012; Limaye and Lamb,
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17 460 2014). However, the controls on and causes of heterogenities in bank erodibility
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19 461 remain to be understood fully. Several studies have focused on the influence on
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21 462 channel migration of resistant clay plugs formed in oxbow lakes (e.g. Howard, 1996;
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23 463 Sun et al., 1996; Hudson and Kesel, 2000). Similarly, it is in such locations that
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25 464 resistance to erosion is highest in the simple numerical model utilised here. In
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27 465 contrast, rates of bank erosion into channel cutoff sites are some of the most rapid
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29 466 observed in this study (c. 140 ma^{-1} on average). This probably reflects the young age
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31 467 of these cutoffs, which means that they have yet to be filled completely and the
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33 468 sediment within them has not been consolidated or altered by pedogenic processes
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35 469 (see also Gautier, 2007). The grain size of clay-rich banks is similar to that of
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37 470 deposits found near the outlet of an oxbow lake (D_{50} : $1.5\text{-}3.0 \mu\text{m}$; Gautier et al.,
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39 471 2010), which is a typical environment for the formation of clay plugs, although oxbow
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41 472 lake deposits can be highly heterogeneous.

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48 Overall, we find no consistent association between clay-rich bank material
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50 474 and former channel locations along the Rio Beni, hence the origin of some of the
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52 475 resistant bank sediments observed here remains uncertain. Moreover, previous
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54 476 studies of channel migration along the Rio Beni that have investigated oxbow fill
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56 477 processes provide no insight into the question of the evolution of resistant clay
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3 478 bodies (e.g. Gautier et al., 2007; 2010). The clay-rich bank sediments observed
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5 479 along the Beni may have a range of origins, including fine-grained sedimentation
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7 480 within floodplain depressions, distal floodbasins, bar-top chute channels or remnants
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9 481 of mainstem cutoff infills (Fisk, 1947; Kolb, 1975; Schumm and Spitz, 1996). Image
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11 482 analysis confirms that clay bodies identified in bank sections have not been
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13 483 reworked since 1960, and that in many places the time since deposition of these
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15 484 sediments is likely to be much longer, although they may include a surface drape of
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17 485 more recent deposition. Clay-rich material with a similar mottled appearance (see
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19 486 Fig. 3a) was identified in floodplain sediments sampled in back swamp areas to the
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21 487 west of the Rio Beni and in older channel belts that are likely to have experienced
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23 488 slow sedimentation (perhaps $<1 \text{ mm a}^{-1}$, based on ^{210}Pb measurements from
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25 489 floodplain cores) over a prolonged time period (Aalto et al., 2003; Dumont, 1996).
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27 490 Based on these observations, we speculate that deposit age may be an important
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29 491 control on the properties and erodibility of these sediments, and that the resistant
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31 492 bank sediments observed here are likely hundreds to thousands of years old, and
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33 493 may thus be the product of prolonged sedimentation in distal floodbasins.
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39 494 Our analysis suggests that the presence or absence of clay-rich bank
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41 495 sediment is the dominant spatial control on bank erosion rates along the Rio Beni.
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43 496 The resulting erosion rates are 60% and 48 % of those of coarser pointbar and
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45 497 channel deposits, while rates of bend migration into oxbow lakes are much greater
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47 498 and highly variable. The measured mean migration rates over many bends agree
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49 499 well with those of Gautier et al. (2007) for the same reach of the Beni although sub-
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51 500 division of bends into different sub-reaches prohibits direct comparison. Bank
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53 501 erosion rates were found to be unrelated to cutbank height or slope in the vicinity of
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55 502 the bend. Moreover, the relationship between mean bend curvature and migration
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3 503 rate differs markedly between bends with and without clay-rich banks. For example,
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5 504 at the former sites, migration rates are generally low over a wide range of curvatures
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7 505 and thus largely independent of curvature. In the absence of clay rich banks,
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9 506 migration rate is maximized at intermediate bend curvatures, as observed previously
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11 507 for many freely meandering rivers (Nanson and Hickin, 1983; Hooke, 1997; Crosato,
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13 508 2009). Temporally, mean annual bend migration rates are related to flood
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15 509 magnitude, intensity and duration, as previously found by Gautier et al. (2007) on the
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17 510 same river reach. However, we find that discharge metrics based on the previous
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19 511 three wet seasons are better predictors of bank erosion rates than discharge metrics
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21 512 based on a single wet season.
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26 513 Channel morphology and styles of bend migration in the presence of clay-rich
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28 514 deposits along the Rio Beni differ from the predictions of existing theory and simple
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30 515 conceptual models, but are consistent with some past observations of channel
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32 516 behaviour. For example, where the apex and the downstream limb of a bend is in
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34 517 contact with a clay-rich bank, faster down-valley translation of the more mobile
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36 518 upstream limb leads to a reduction in wave length, up-valley skewing (e.g. bend 24
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38 519 in Fig. 12), and eventually to neck cutoff. Similar processes have been described
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40 520 previously for the Mississippi River (Fisk, 1947). The observed decrease in channel
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42 521 width downstream of the bend apex, where the channel is immobilised by resistant
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44 522 banks, contrasts with the peak in width at this location that has been associated with
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46 523 'free-meandering bends' (Luchi et al., 2011; Zolezzi et al., 2012). However, this
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48 524 difference in channel morphology is consistent with the difference in boundary
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50 525 conditions, and serves to emphasize the significant control exerted by variable bank
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52 526 strength. Overall, fewer than 10% of bends are characterised by styles of
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54 527 development that fit simple conceptual models of bend evolution (Hickin, 1978;
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3 528 Hooke, 1987) and associated changes in migration rate between inception and cutoff
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5 529 (Hooke and Yorke, 2010). This can be attributed, in part, to the influence of
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7 530 heterogeneity in floodplain composition on bend migration and planform shape. For
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9 531 example, during a period of straightening in sub-reach 7 downstream of the clay
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11 532 contact in bend 35 a number of cutoffs occurred (Fig. 8) but instead of a gradual
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13 533 increase in sinuosity, as predicted by conceptual models (Hooke, 2003), the reach
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15 534 continued to straighten until 2005. Such cutoff avalanches are typical for meandering
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17 535 rivers and have been observed along the Beni (Gautier et al., 2007) and elsewhere
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19 536 (Stolum, 1996; Hooke, 2004), but the concurrent medium-term straightening is less
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21 537 common.
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26 538 In general, there appears to be an inverse relationship between clay body
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28 539 frequency and mean annual migration rate and channel sinuosity, which is related to
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30 540 the role of clay banks in reducing bank erosion and promoting channel straightening.
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32 541 However, this tendency over-simplifies the relationship between clay body frequency
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34 542 and channel dynamics. For example, sub-reach 8 is characterised by very low
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36 543 sinuosity and migration rates despite a lack of contact with clay-rich bank sections.
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38 544 More generally, although migration rates at clay banks tend to be low, average rates
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40 545 within sub-reaches where clay-banks are common need not necessarily be low if
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42 546 migration rates between these immobile locations are enhanced. Figure 16 shows
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44 547 that in sub-reaches 1 to 8 the apices of bends where clay-rich banks have been
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46 548 found in 2011 have not moved significantly over up to five decades, and that channel
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48 549 migration occurs only between these bends. Thus, bends with resistant clay-rich
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50 550 banks can act as fixed hinges around which the channel migrates. Even where
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52 551 cutoffs occur immediately upstream of these fixed points this has relatively little
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54 552 impact on the position of these bends, and usually results in only short-term and
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3 553 limited bend translation (e.g., for 3 years at bend 18, Fig. 12a). Since planform
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5 554 adjustment downstream of cutoffs is suppressed, new bends formed at the cutoff
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7 555 tend to lengthen and extend rapidly (e.g., bends 16 and 20 in Figs. 6c and 12c).
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10 556 When bends with clay-rich banks are cut off, this can result in a short-term increase
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12 557 in migration rate and can lead to a change in migration direction into substrate of
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14 558 different erodibility, thereby further increasing mobility (e.g., downstream of bend 43
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16 559 after 2003, see Fig. 8e, f). Consequently, floodplain heterogeneity may also modify
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18 560 the characteristic dynamics of channel evolution by promoting long periods of
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20 561 stability punctuated by rapid channel adjustment when the influence of clay-rich
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22 562 banks is temporarily removed. Moreover, the straight channels found downstream of
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24 563 clay-rich bends may decouple these bends from planform evolution downstream if
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26 564 the spacing between clay-rich bends is sufficient (e.g. between bend 20 and 24, but
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28 565 not below bend 24 (Fig. 12d)).
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33 566 Channel behaviour in sub-reaches 11 and 17 to 19 deviates from the
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35 567 relationships identified above. For example, there are few clay bodies exposed along
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37 568 channel bank lines in these areas, they are relatively stable (migration rates $< 20 \text{ m a}^{-1}$
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39 569 ¹), and yet are characterised by relatively high sinuosity (Fig. 11). Sub-reaches 17 to
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41 570 19 lie towards the downstream end of the Beni foredeep, hence channel evolution in
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43 571 this area is likely controlled in part by tectonics. For example, downstream of the
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45 572 study reach the channel is bounded by higher terraces that are indicative of vertical
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47 573 displacement of the channel bed relative to the surrounding floodplain. Gautier et al.
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49 574 (2007) emphasize the role of tectonics in promoting low slopes and channel incision,
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51 575 which in turn leads to lateral channel stability downstream of the junction between
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53 576 the Rio Beni and the Rio Madidi. It is possible that these effects extend further
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55 577 upstream beyond the junction with the Madidi into the study site considered here.
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3 578 However, the limited dGPS data that we obtained in this area indicate a steepening
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5 579 of the channel water surface upstream of the Madidi. Moreover, channel stability in
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7 580 sub-reaches 17 to 19 over the past few decades is not necessarily representative of
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9 581 longer-term channel evolution, and the floodplain in this area contains ample
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11 582 evidence of previous channel migration, abandonment and bifurcation.
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15 583 The model results presented here illustrate that the simulation model captures
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17 584 many of the effects of spatial variability in bank erodibility, despite the simplicity of
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19 585 the process parameterisations used. Moreover, they support the conclusion that
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21 586 floodplain heterogeneity may have a substantial influence on channel geometry and
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23 587 styles of bend evolution. These results are consistent with those of previous
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25 588 modelling studies that have highlighted the effects of heterogeneity in bank strength
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27 589 on channel belt width (Sun et al., 1996) and meander dynamics (Gueneralp and
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29 590 Rhoads, 2011). However, the current simulations also suggest that heterogeneity in
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31 591 floodplain alluvium can promote a substantial reduction in channel sinuosity, which
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33 592 would in turn have significant implications for channel gradient, flow conveyance,
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35 593 sediment transport and, potentially, basin accommodation space. The differences in
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37 594 channel behaviour simulated by these models likely reflects several factors, including
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39 595 differences in the underlying meander migration models used, their parameterisation
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41 596 of bank erodibility, and their ability to simulate spatial structure in floodplain
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43 597 heterogeneity. As a consequence of this, the true sensitivity of channel behaviour to
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45 598 heterogeneity in bank composition is difficult to quantify accurately using these
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47 599 models, due to their phenomenological nature. Resolving this issue likely requires
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49 600 the use of models with a stronger physical basis. Specifically, such models should:
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51 601 (i) incorporate a physically-based (and perhaps fully three-dimensional) treatment of
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53 602 hydrodynamics that is suitable for representing the controls on boundary shear
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3 603 stress in channels with complex planform geometries; (ii) be capable of representing
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5 604 streamwise variations in channel width and depth linked to pinning of bend apices by
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7 605 resistant floodplain sediments; (iii) include an improved treatment of floodplain
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9 606 sediment conveyance and sedimentation, and the physical and chemical processes
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11 607 that control the long-term evolution of floodplain sediment properties and erodibility.
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13 608 Development of such a model represents a significant challenge because the nature
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15 609 of and controls on long-term post-depositional changes in floodplain sediment
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17 610 properties are poorly understood. Moreover, requirements (i) and (iii) are, to some
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19 611 extent, mutually exclusive in that they may require the application of computationally
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21 612 expensive hydrodynamic models over periods of millennia.
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27 28 29 614 **Conclusions**

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32 615 Many previous studies have hypothesized that spatial heterogeneity in floodplain
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34 616 sediments and associated bank erodibility may represent an important control on the
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36 617 evolution of alluvial meanders. This study has quantified these effects for the case of
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38 618 the Rio Beni, Bolivia, which is a large dynamic sand-bed river, characterised by high
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40 619 average rates of channel migration and overbank sedimentation. Field data and GIS
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42 620 analysis demonstrate that floodplain clay bodies along the Rio Beni are a key control
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44 621 on meander form and evolution at spatial scales ranging from individual bends up to
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46 622 distances of several tens of kilometers. While the origin of these clay deposits is not
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48 623 certain, we find no simple association between recent cutoff channels and the
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50 624 location of clay bodies. Rather, we speculate that clay rich bodies may be a product
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52 625 of slow sedimentation in distal flood basins, and that pedogenic processes may be
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54 626 an important control on the long-term evolution of floodplain erodibility.
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3 627 Bends that interact with such clay bodies are associated with a significant
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5 628 reduction in migration rate and channel width, and a change in the frequency
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7 629 distribution of local channel curvature that is linked to the formation of more angular
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10 630 geometries characterised by sharp bends and extended straight channel segments.
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12 631 At larger spatial scales, clay bodies act as hinge points that can limit the upstream
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14 632 and downstream propagation of morphodynamic perturbations (e.g., bend translation
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16 633 and response to cutoff). As a consequence, the spatial distribution of clay bodies can
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18 634 promote marked streamwise changes in meander sinuosity and dynamics (over
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20 635 distances of 10-50 km).
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24 636 Our numerical model results are consistent with field observations, and imply
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26 637 that spatial heterogeneity in bank erodibility driven by variable bank composition may
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28 638 drive a substantial (c. 30%) reduction in average channel sinuosity, compared to
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30 639 situations in which bank strength is spatially homogeneous. Moreover, this effect is
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32 640 not simply a consequence of a reduction in average bank erodibility. Rather, reduced
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34 641 channel migration rates and sinuosity are driven, in part, by the change in the
35
36 642 curvature distribution of the channel. Thus, the increase in the length of straight
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38 643 channel segments and the creation of sharp bends near clay bodies may both
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40 644 influence the streamwise development of secondary flows and reduce rates of bank
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42 645 erosion. Although the simple modelling framework employed does not represent flow
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44 646 hydrodynamics or sediment transport processes, it does appear to provide a first
45
46 647 order representation of the link between floodplain erodibility and meander geometry
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48 648 that supports this conclusion.
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53 649 These results illustrate that heterogeneity in floodplain sedimentology is a first
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55 650 order control on channel planform and dynamics. Moreover, by controlling channel
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57 651 belt width, lateral channel migration and levee reworking, floodplain heterogeneity is
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3 652 likely an important influence on the geometry and rate of aggradation of alluvial
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5 653 ridges, the frequency of avulsions and the resulting basin alluvial architecture. Since
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7 654 the spatial-scaling of floodplain heterogeneity is linked to the processes (and
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9 655 associated scales) that control floodplain construction, these effects cannot be
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11 656 accounted for fully in numerical models using stochastically-generated patterns of
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13 657 floodplain erodibility. Rather, attempts to model and thus understand meander
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15 658 dynamics, even over relatively short time scales (e.g., 10-100 years), may need to
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17 659 simulate time periods that are long enough to represent the evolution of floodplain
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19 660 heterogeneity and the bi-directional feedbacks with channel morphodynamics.
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25 662 **Acknowledgements**

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27
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29
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32 665 improvements in the manuscript.
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825 Table 1: Aerial and satellite imagery used for GIS analysis of migration patterns. All
 826 Landsat scenes are from Path 001, Row 069 and 070. Rectification errors (relative to
 827 the year 2000 Landsat images or SRTM in 2000) are given for both tiles or combined
 828 tiles where applicable.

Date	Type of image	Resolution (pixel size)	Rectification RMSE
1960	aerial		>30 m
October 1975	Landsat LM2	83 m	>30 m
August 1987	Landsat LT5	30 m	>30 m
August 1993	Landsat LT5	30 m	>30 m
July 1996	Landsat LT5	30 m	76.7 m
August 1997	Landsat LT5	30 m	11.9 - 14.7 m
July 1998	Landsat LT5	30 m	18.0
July/ August 1999	Landsat LE7	30 m	7.5 - 11.7 m
July/ August 2000	Landsat LE7	30 m	8.8 - 9.3 m (relative to SRTM)
June/ August 2001	Landsat LE7	30 m	9.2 - 18.4 m
August 2003	Landsat LT5	30 m	8.2 - 15.6 m
September 2004	Landsat LT5	30 m	8.2 - 15.6 m
September 2005	Landsat LT5	30 m	8.2 - 15.6 m
July 2006	Landsat LT5	30 m	8.2 - 15.6 m
August 2007	Landsat LT5	30 m	8.2 - 15.6 m
August 2008	Landsat LT5	30 m	8.2 - 15.6 m
August 2009	Landsat LT5	30 m	8.2 - 15.6 m
May 2010	Landsat LT5	30 m	6.8 - 7.3 m
May/ June 2011	Landsat LT5	30 m	8.2 - 15.6 m

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3 830 **Figure Captions**
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6 831 Figure 1: a) Location of the Rio Beni, northeastern Bolivia; b) Rio Beni study reach
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8 832 showing sub-reach boundaries and areas used to illustrate channel dynamics in
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10 833 figures 6, 8 and 12. The background Landsat TM imagery was taken during the dry
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12 834 season 2011.
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15 835 Figure 2: Dominant styles of meander bend migration along the study reach: a)
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17 836 Compounding; b) Expansion; c) Extension; d) Lateral displacement; e) Rotation; and
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19 837 f) Longitudinal translation. Adapted from Hooke (1984). The red outline shows the
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21 838 bank lines some time before the Landsat image was taken.
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25 839 Figure 3: Typical bank sections along the study reach: a) Stable cutbank consisting
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27 840 of clay-rich material at Bend 72; b) Bank composed of silty, highly erodible
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29 841 sediments, with evidence of vertical layering.
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33 842 Figure 4: Mean migration rates of bends of the Rio Beni between 1960 and 2011 (95
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35 843 % confidence intervals shown as error bars). Bends are classified by the substrate or
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37 844 the morphological feature into which they are migrating: banks composed of mixed-
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39 845 sized sediments, clay-rich banks, silty and sandy point bar (and counter point bar)
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41 846 deposits, active oxbow lakes, and former channels infilled with mixed-sized
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43 847 sediments.
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47 848 Figure 5: Proportion of bends within the 19 sub-reaches experiencing different styles
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49 849 of migration between 1960 and 2011. Sub-reaches 1, 3-5, 7, 12 and 15 are
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51 850 dominated by clay-rich banks.
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54 851 Figure 6: Typical sequence of channel migration influenced by clay-rich banks
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56 852 between bends 13 and 18 (in sub-reach 4) over the periods: a) 1975-1987; b) 1987-
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3 853 1993; and c) 1993-1996. Landsat image dates correspond to the end of each time
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5 854 period while banklines shown in red represent the start of the time period in each
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7 855 panel. The presence of resistant clay-rich cutbanks in bend 13, 14 and 18 (marked
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9 856 orange in c) limit rates of bend migration and lead to up-valley skewing of mature
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11 857 bends, such as bend 14. Flow direction is indicated by the arrow.

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15 858 Figure 7: Changes over time in the length of straight channel segments (identified in
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17 859 the key by the sub-reach number in which they are located), illustrating the gradual
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19 860 down-valley lengthening that occurs over multiple decades downstream of clay-rich
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21 861 banks. Typically these segments lengthen due to down-valley translation, increase in
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23 862 bend wavelength and cutoff of bends downstream (solid lines), but contact with other
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25 863 clay-rich banks can lead to the development of perturbations and thus termination of
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27 864 the lengthening process (dashed lines). This process is also illustrated in Fig. 8e and
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29 865 8f. Each line represents a single channel segment.

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33 866 Figure 8: Migration of the Rio Beni between bends 34 and 44 over the periods: a)
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35 867 1987-1993; b) 1993-1998; c) 1998-2001; d) 2001-2005; e) 2005-2011, and f) 2011-
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37 868 2013. Landsat image dates correspond to the end of each time period while
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39 869 banklines shown in red represent the start of the time period in each panel. The
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41 870 channel in this location (sub-reach 7) straightens between 1998 and 2005, following
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43 871 a series of cutoffs. Flow direction is indicated by an arrow and relevant bend
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45 872 numbers are given. Exposed clay-rich banks were found at bend 35 and 40 (marked
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47 873 orange in e and f).

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52 874 Figure 9: Frequency distribution of channel curvature at sub-reaches that are
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54 875 dominated by clay-rich banks (sub-reaches 1, 3-5, 7, 12 and 15) compared with the
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56 876 remaining sub-reaches with few clay-rich banks.

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3 877 Figure 10: Mean centerline migration rates for the 19 sub-reaches along the Rio Beni
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5 878 over four periods between 1960 and 2011. Symbols indicate the proportion of the
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7 879 channel in each sub-reach experiencing mean migration rates above and below
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9 880 threshold values (5 ma^{-1} and 50 ma^{-1} , respectively). Arrows at the top of the image
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11 881 indicate the occurrence of chute (grey) and neck (black) cutoffs.

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15 882 Figure 11: Mean channel sinuosity for the 19 sub-reaches along the Rio Beni at five
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17 883 points in time between 1960 and 2011.

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20 884 Figure 12: Migration of the Rio Beni between bends 18 and 29 over the periods: a)
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22 885 1993-1996; b) 1996-2001; c) 2001-2003; d) 2003-2009; and e) 2009-2011. Landsat
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24 886 image dates correspond to the end of each time period while banklines shown in red
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26 887 represent the start of the time period in each panel. Flow direction is indicated by an
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28 888 arrow and relevant bend numbers are given. Exposed clay-rich banks (marked
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30 889 orange in e) were found at bend 18, 20, 22, 24, 25, 26 and 29.

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34 890 Figure 13: Time series of channel sinuosity for three numerical model simulations in
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36 891 which the dependence of bank erodibility on grain size composition is altered by
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38 892 changing the value of k in equation (2).

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42 893 Figure 14: Frequency distributions of channel curvature for simulations in which bank
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44 894 erodibility is constant ($k = 0$) and bank erodibility varies as a function of grain size
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46 895 composition ($k = 3$).

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49 896 Figure 15: Spatial patterns of the fraction of fine sediment in the floodplain and the
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51 897 meandering channel position (at two instants in time) during model simulations in
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53 898 which bank erodibility is constant (upper panel; $k = 0$) and bank erodibility varies as a
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55 899 function of grain size composition (lower panel; $k = 3$).

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3 900 Figure 16: Channel centre lines between 1960 and 2011 in the upper study reach
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5 901 (sub-reaches 1-8) with arrows marking bends with clay-rich banks (orange lines).
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7 902 These bends are relatively immobile over >50 years and act as fixed 'hinges'
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9 903 between more mobile reaches (see text). Flow is from left to right (in order A-B-C).
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For Peer Review

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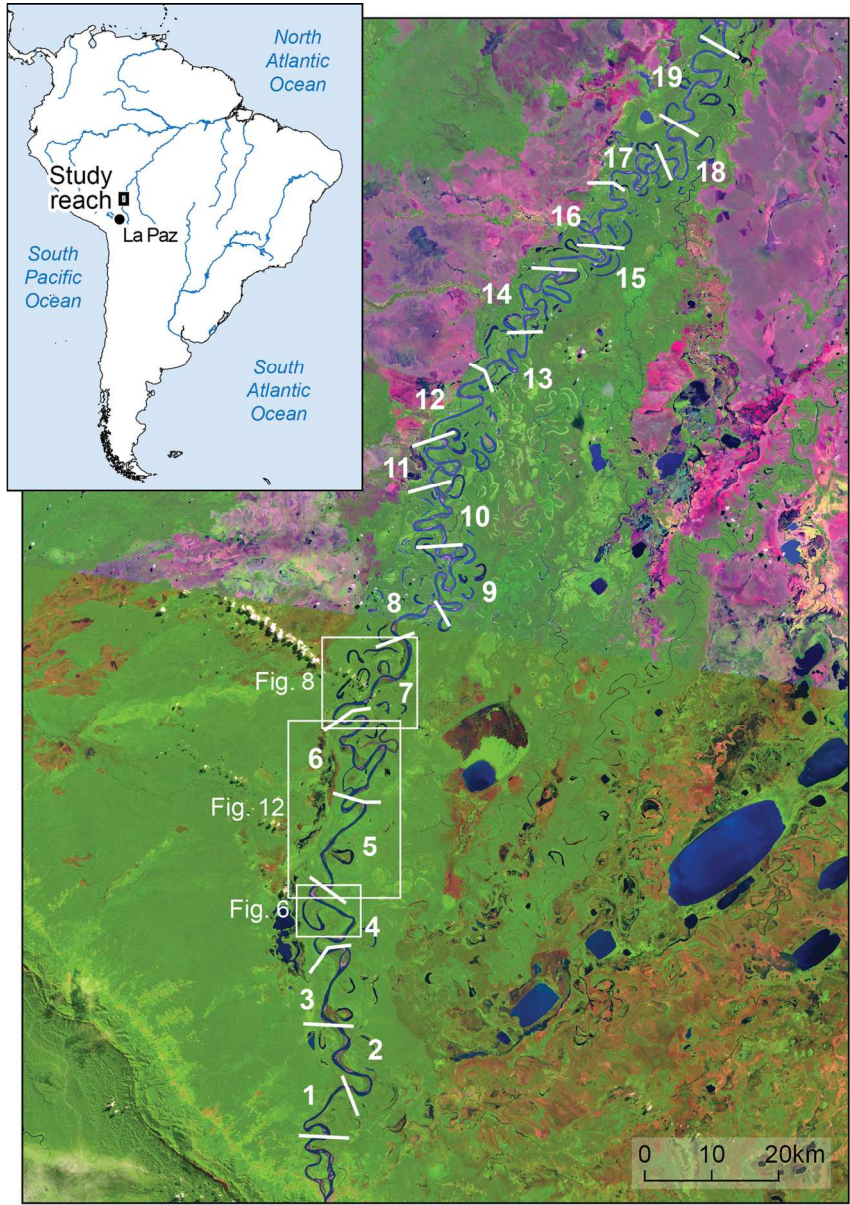


Figure 1: a) Location of the Rio Beni, northeastern Bolivia; b) Rio Beni study reach showing sub-reach boundaries and areas used to illustrate channel dynamics in figures 6, 8 and 12. The background Landsat TM imagery was taken during the dry season 2011. 119x170mm (300 x 300 DPI)

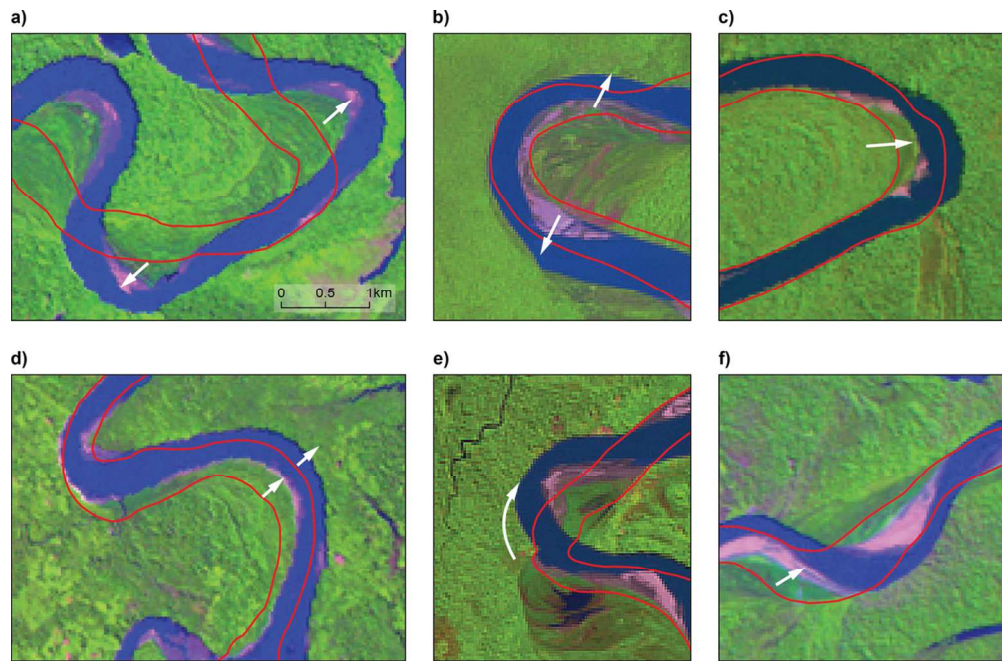


Figure 2: Dominant styles of meander bend migration along the study reach: a) Compounding; b) Expansion; c) Extension; d) Lateral displacement; e) Rotation; and f) Longitudinal translation. Adapted from Hooke (1984). The red outline shows the bank lines some time before the Landsat image was taken.
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Figure 3: Typical bank sections along the study reach: a) Stable cutbank consisting of clay-rich material at Bend 72; b) Bank composed of silty, highly erodible sediments, with evidence of vertical layering.
141x237mm (300 x 300 DPI)

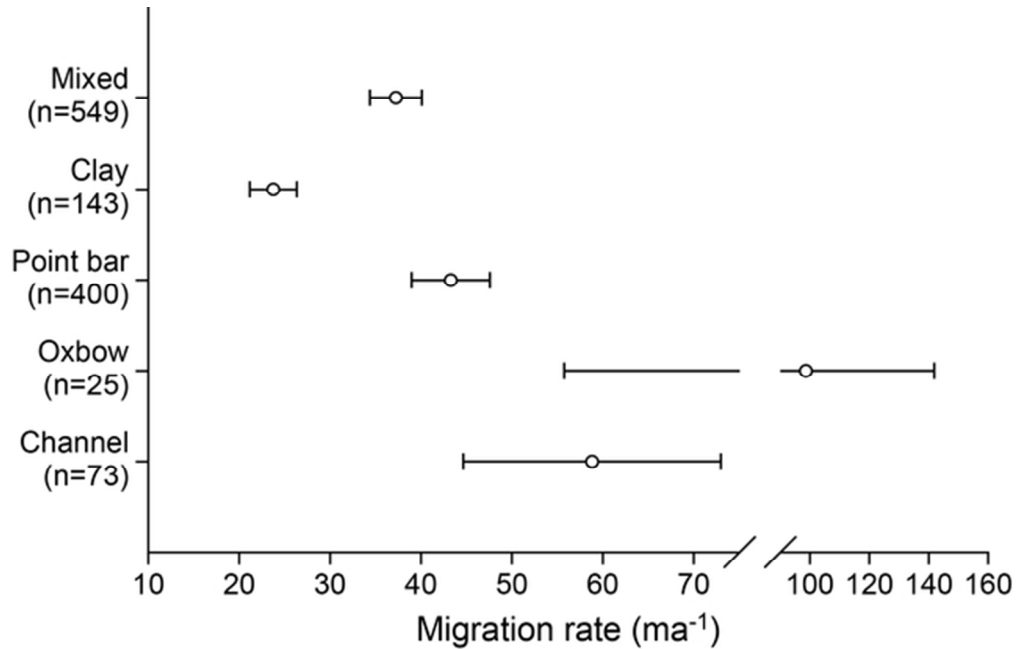


Figure 4: Mean migration rates of bends of the Rio Beni between 1960 and 2011 (95 % confidence intervals shown as error bars). Bends are classified by the substrate or the morphological feature into which they are migrating: banks composed of mixed-sized sediments, clay-rich banks, silty and sandy point bar (and counter point bar) deposits, active oxbow lakes, and former channels infilled with mixed sediments.
54x34mm (300 x 300 DPI)

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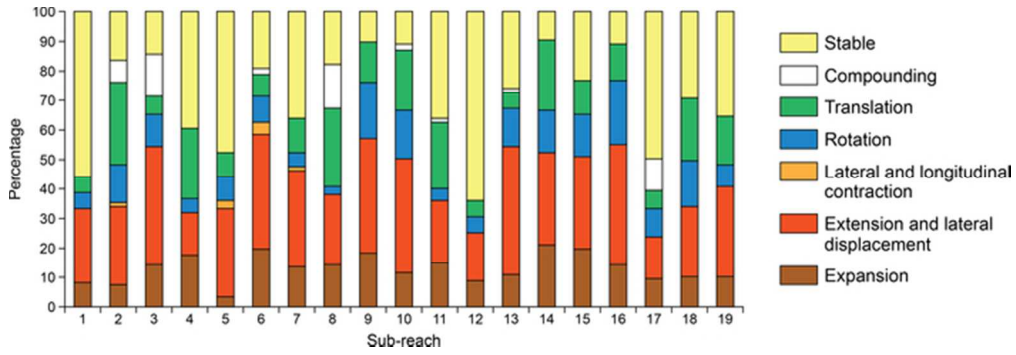


Figure 5: Proportion of bends within the 19 sub-reaches experiencing different styles of migration between 1960 and 2011. Sub-reaches 1, 3-5, 7, 12 and 15 are dominated by clay-rich banks.
59x20mm (300 x 300 DPI)

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Figure 6: Typical sequence of channel migration influenced by clay-rich banks between bends 13 and 18 (in sub-reach 4) over the periods: a) 1975-1987; b) 1987-1993; and c) 1993-1996. Landsat image dates correspond to the end of each time period while banklines shown in red represent the start of the time period in each panel. The presence of resistant clay-rich cutbanks in bend 13, 14 and 18 (marked orange in c) limit rates of bend migration and lead to up-valley skewing of mature bends, such as bend 14. Flow direction is indicated by the arrow.
59x19mm (300 x 300 DPI)

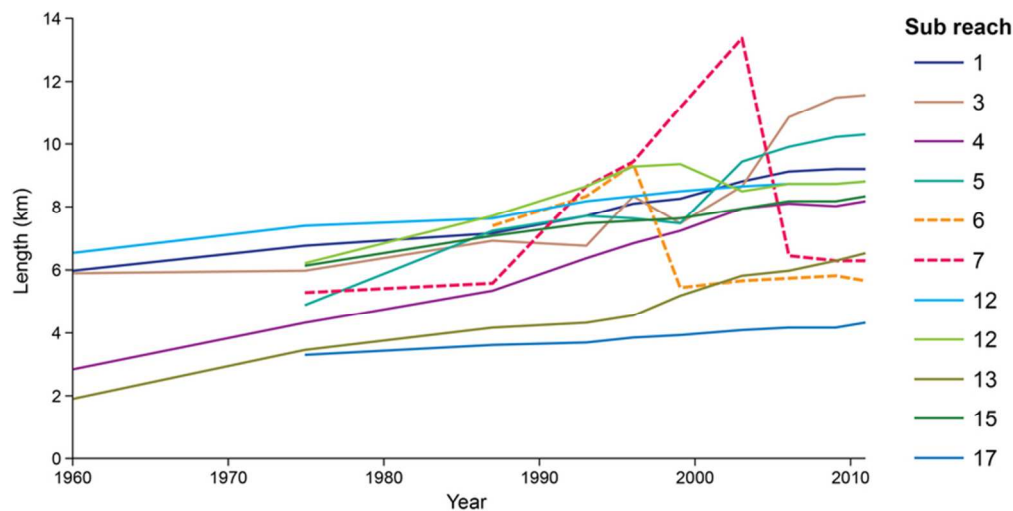


Figure 7: Changes over time in the length of straight channel segments (identified in the key by the sub-reach number in which they are located), illustrating the gradual down-valley lengthening that occurs over multiple decades downstream of clay-rich banks. Typically these segments lengthen due to down-valley translation, increase in bend wavelength and cutoff of bends downstream (solid lines), but contact with other clay-rich banks can lead to the development of perturbations and thus termination of the lengthening process (dashed lines). This process is also illustrated in Fig. 8e and 8f. Each line represents a single channel segment.

72x36mm (300 x 300 DPI)

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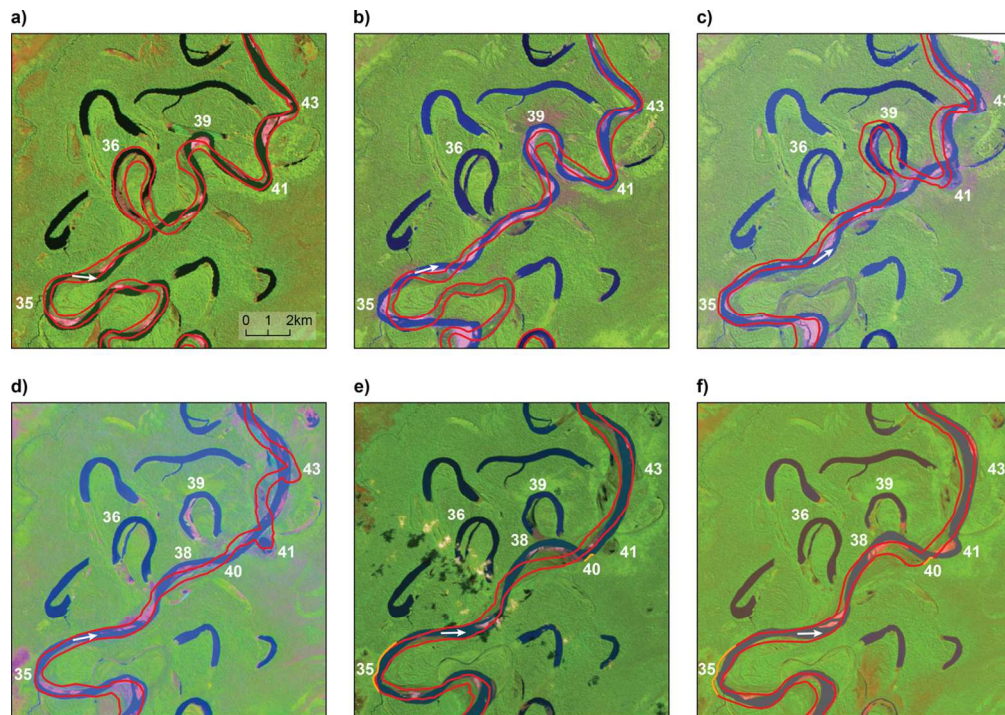


Figure 8: Migration of the Rio Beni between bends 34 and 44 over the periods: a) 1987-1993; b) 1993-1998; c) 1998-2001; d) 2001-2005; e) 2005-2011, and f) 2011-2013. Landsat image dates correspond to the end of each time period while banklines shown in red represent the start of the time period in each panel. The channel in this location (sub-reach 7) straightens between 1998 and 2005, following a series of cutoffs. Flow direction is indicated by an arrow and relevant bend numbers are given. Exposed clay-rich banks were found at bend 35 and 40 (marked orange in e and f).

123x87mm (300 x 300 DPI)

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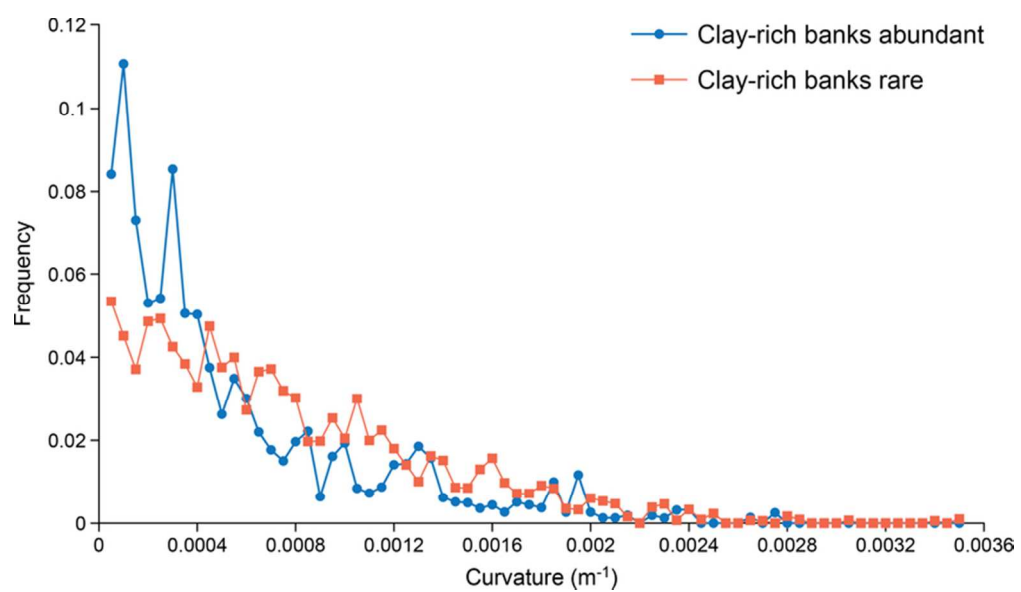


Figure 9: Frequency distribution of channel curvature at sub-reaches that are dominated by clay-rich banks (sub-reaches 1, 3-5, 7, 12 and 15) compared with the remaining sub-reaches with few clay-rich banks. 71x41mm (300 x 300 DPI)

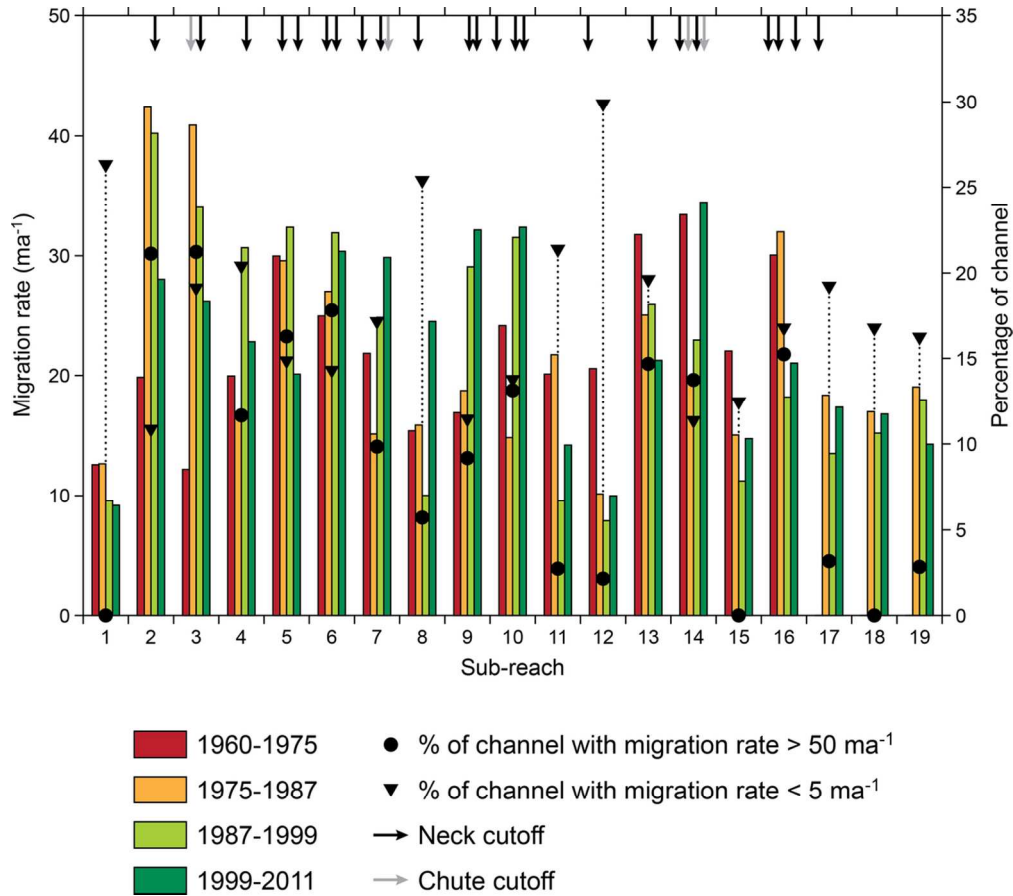


Figure 10: Mean centerline migration rates for the 19 sub-reaches along the Rio Beni over four periods between 1960 and 2011. Symbols indicate the proportion of the channel in each sub-reach experiencing mean migration rates above and below threshold values (5 ma⁻¹ and 50 ma⁻¹, respectively). Arrows at the top of the image indicate the occurrence of chute (grey) and neck (black) cutoffs.

110x98mm (300 x 300 DPI)

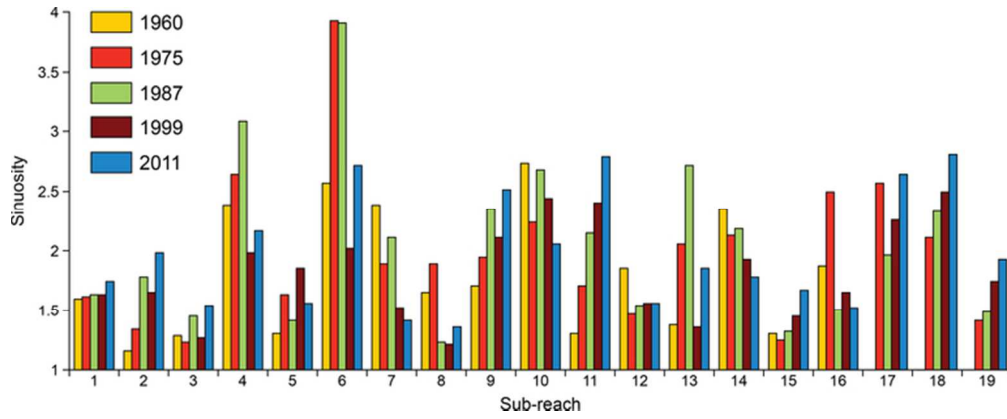


Figure 11: Mean channel sinuosity for the 19 sub-reaches along the Rio Beni at five points in time between 1960 and 2011.
65x26mm (300 x 300 DPI)

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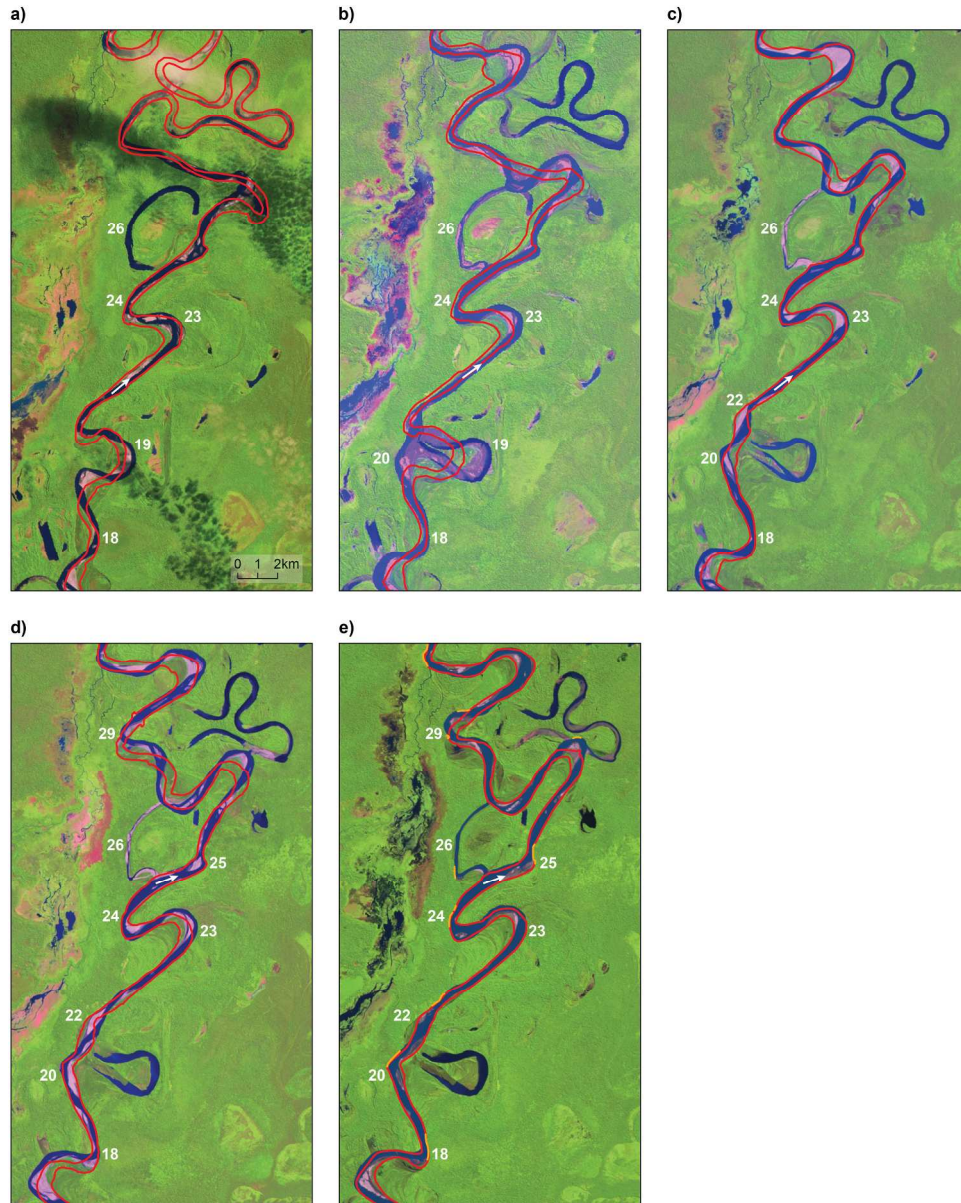


Figure 12: Migration of the Rio Beni between bends 18 and 29 over the periods: a) 1993-1996; b) 1996-2001; c) 2001-2003; d) 2003-2009; and e) 2009-2011. Landsat image dates correspond to the end of each time period while banklines shown in red represent the start of the time period in each panel. Flow direction is indicated by an arrow and relevant bend numbers are given. Exposed clay-rich banks (marked orange in e) were found at bend 18, 20, 22, 24, 25, 26 and 29.
218x272mm (300 x 300 DPI)

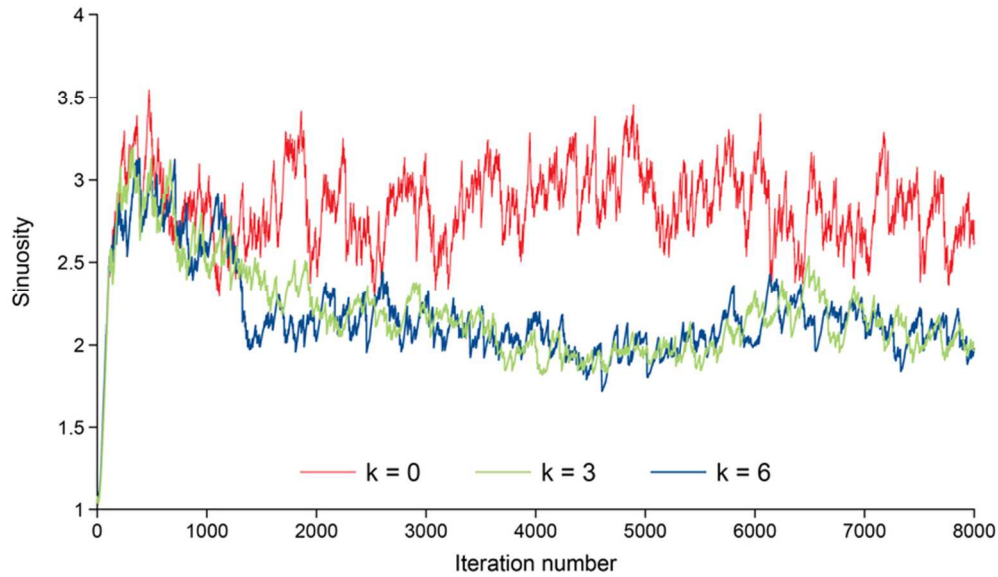


Figure 13: Time series of channel sinuosity for three numerical model simulations in which the dependence of bank erodibility on grain size composition is altered by changing the value of k in equation (2).
73x42mm (300 x 300 DPI)

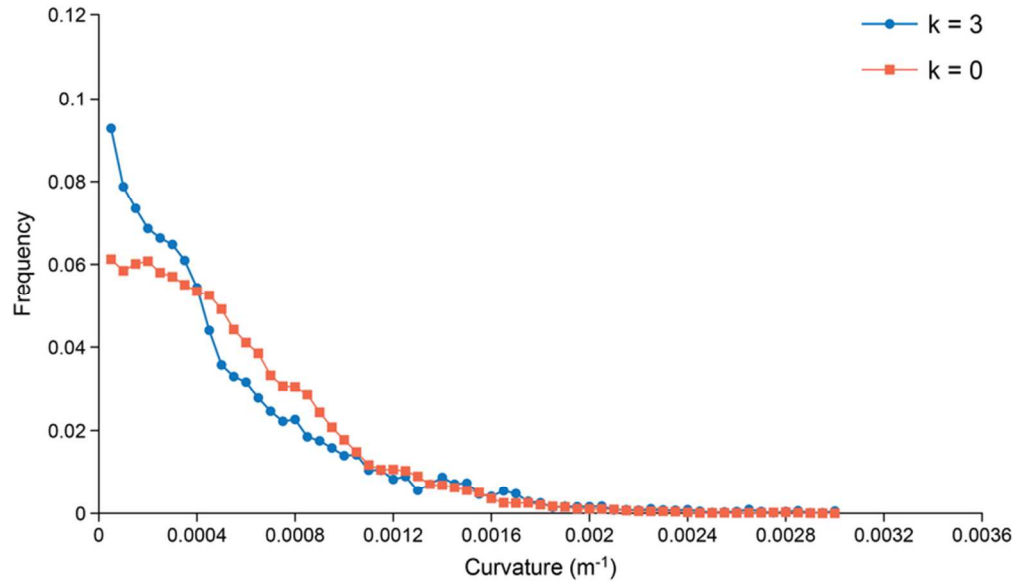


Figure 14: Frequency distributions of channel curvature for simulations in which bank erodibility is constant ($k = 0$) and bank erodibility varies as a function of grain size composition ($k = 3$).
72x41mm (300 x 300 DPI)

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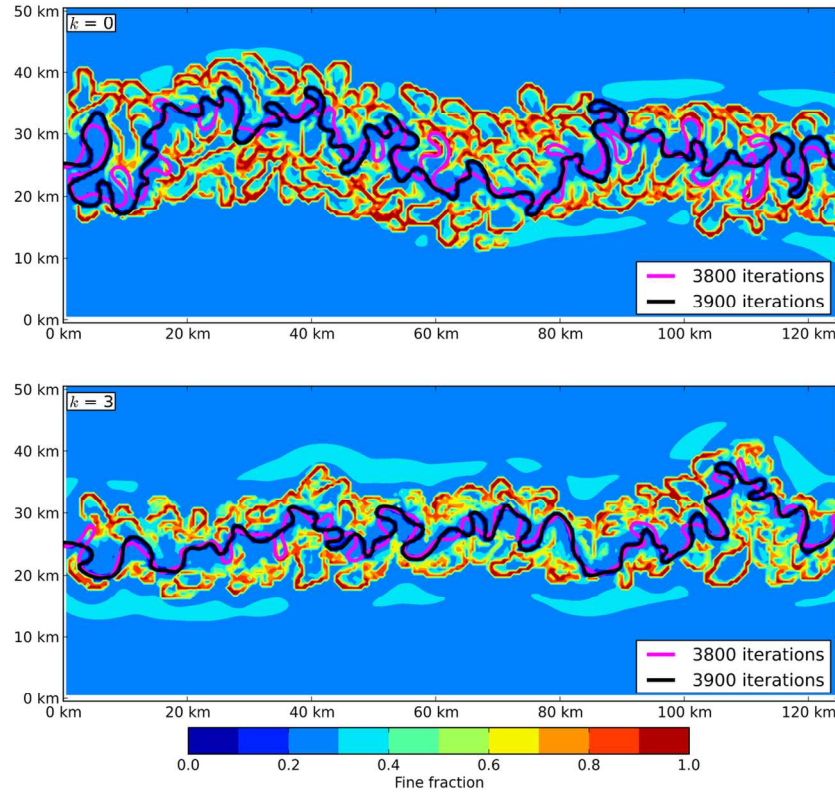


Figure 15: Spatial patterns of the fraction of fine sediment in the floodplain and the meandering channel position (at two instants in time) during model simulations in which bank erodibility is constant (upper panel; $k = 0$) and bank erodibility varies as a function of grain size composition (lower panel; $k = 3$).
127x110mm (300 x 300 DPI)

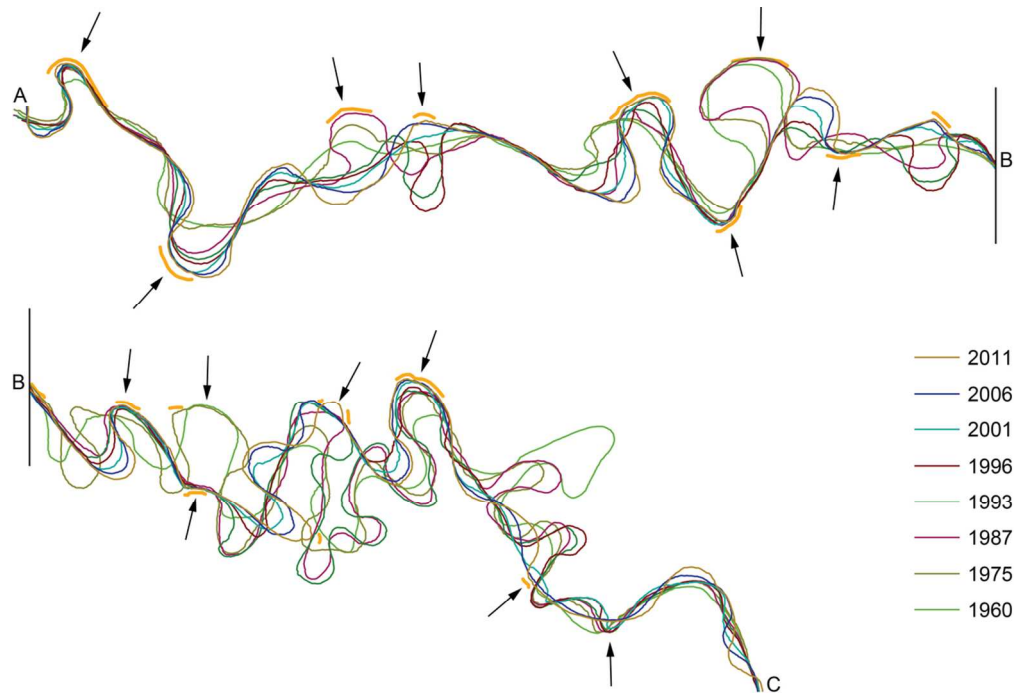


Figure 16: Channel centre lines between 1960 and 2011 in the upper study reach (sub-reaches 1-8) with arrows marking bends with clay-rich banks (orange lines). These bends are relatively immobile over >50 years and act as fixed 'hinges' between more mobile reaches (see text). Flow is from left to right (in order A-B-C).

107x74mm (300 x 300 DPI)