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Modeling Length Adjustment in Alluvial Anabranching Units

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Key Points:

- A dynamic model is formulated to study the length evolution of anabranching units in rivers
- The analysis predicts three possible trajectories depending on reach-scale channel geometry, bank and floodplain attributes, and flow regime
- The theoretical range of lengths that supports persistent anabranching is broadly consistent with satellite observations

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Abstract Anabranching units form when the stream flow divides into two channels that reconnect downstream. Recent observations indicate that these patterns tend to develop fairly regular planforms, characterized by an average length that scales with reach-averaged channel geometry and by weak asymmetry between the anabranches. The question remains as to how such systems evolve toward these regular geometries, and why some of them fail to maintain a multi-thread morphology after their formation. Here, a dynamic model for a solitary anabranching unit is formulated, in which the planform evolution is described in terms of temporal variations in the nodal length separating the upstream bifurcation from the downstream confluence. The governing equation expresses a balance between competing processes acting at different flow stages: elongation during base-flow conditions, assumed to be linked to sediment flux partitioning at the bifurcation, and recession driven by erosion during flood events. The analysis reveals three possible evolutionary trajectories: (i) the attainment of an equilibrium length, in which both branches are capable of carrying sediment over long timescales; (ii) the transition to a single-thread configuration through island removal under dominant erosion; or (iii) the closure of one anabranch following unbounded elongation. The evolutionary pathway depends on reach-averaged hydraulic geometry, bank and floodplain properties, and basin hydrology. The predicted range of lengths for long-term persistence of the multi-thread configuration is broadly consistent with satellite-derived observations.

Plain Language Summary Rivers sometimes split into two branches that flow around an island before merging again downstream, forming anabranching units. These features are commonly observed in both single-thread and multi-thread rivers and often display regular shapes, with lengths on the order of a few hundred times the reach-averaged flow depth. Understanding how they evolve over time is important, not only to advance scientific knowledge, but also to inform river restoration strategies aimed at maintaining or re-establishing multi-channel river forms. In this study, we develop a simplified dynamic model to describe the long-term evolution of anabranching units after their formation. The model represents changes in length as the result of two competing processes depending on the flow regime: one that tends to lengthen the system during low-flow conditions, and another that shortens it through erosion during floods. We show that, depending on the balance between these processes, an anabranching unit may persist with two active channels, or it may transition to a single-channel river as one branch closes or the island is eroded away. These results may help explain why many natural anabranching units exhibit similar lengths and provide insights into the conditions that favor their long-term persistence.

1. Introduction

When a river splits into two branches that flow around an island and reconnect downstream, it locally forms an anabranching system. The islands are typically permanent vegetated landforms that rise approximately the same elevation as the surrounding floodplain (Brice, 1982; Nanson & Knighton, 1996). Locally anabranching systems constitute the basic morphological units of multi-thread rivers, appearing as nested sequences formed by the concatenation of multiple channels. They are also frequently found as isolated features in rivers that display transitional morphologies between meandering and braiding, that is, fluvial patterns that combine attributes of both planform styles (Ferguson, 1987; Zolezzi et al., 2012), where chute channels cut across the necks of meander bends, creating an anabranching structure with sinuous branches flowing around islands, which exhibit varying degrees of lateral mobility (e.g., Constantine et al., 2010; Fisk, 1947; M. Grenfell et al., 2012; Lewis & Lewin, 1983). Our interest will be devoted hereafter to such individual anabranching units (Figure 1).

So far, much information has been accumulated about the physical processes responsible for the generation of river anabranching. Their formation is attributed to two fundamentally distinct mechanisms. The first involves in-

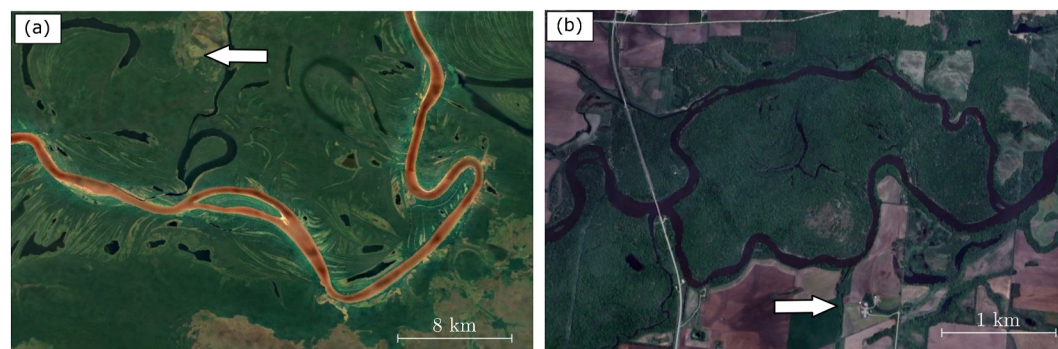


Figure 1. (a) A solitary anabranching unit in the Amazonian Mamoré River along the border between Bolivia and Brazil, 11°38'S 65°14'W. (b) An anabranching segment in the Wapsinicon River (Iowa, USA), 41°44'N 90°38'W. The arrows indicate flow direction. The images were retrieved from Google Earth.

stream processes that promote the initiation of a mid-channel bar. This mechanism is typically associated with variations in channel width, where local widening leads to sediment deposition and, in turn, promotes flow bifurcation (e.g., Ashworth, 1996; Hooke, 1986; Knighton, 1972; Leopold & Wolman, 1957). Under suitable hydraulic and sedimentary conditions, the mid-channel bar progressively accretes and elongates as it traps additional sediment, eventually becoming stabilized through vegetation colonization and developing into an emergent island (e.g., Gurnell et al., 2001; Osterkamp, 1998; Rubey, 1952; Wintenberger et al., 2015). The second mechanism involves channel avulsion, whereby the main flow path shifts by cutting a new channel into the floodplain or reoccupying a pre-existing paleochannel (Nanson & Knighton, 1996; Slingerland & Smith, 2004).

In contrast, the mechanics driving the evolution of anabranching units after their formation is still poorly understood, as well as the key factors needed to describe it, which ultimately determine the resulting morphological shape. The present authors developed a theoretical framework to investigate the equilibrium morphodynamics of individual anabranching units, modeling the mutual feedbacks between the upstream bifurcation and the downstream confluence (Ragno & Tubino, 2025; Ragno et al., 2021). These analyses revealed a close relationship between the length of the anabranches and the reach-averaged hydraulic geometry, which ultimately controls the partitioning of water and sediment fluxes between the anabranches, and thus the conditions of persistence of a multi-thread system in the long term, avoiding either the abandonment or the complete infilling of one of the anabranches. The theoretical model was complemented by remote-sensing analyses of natural anabranching units (Ragno et al., 2022). These observations indicated that these systems typically exhibit a mean length of the anabranches proportional to the reach-averaged bankfull depth, say a few hundreds depths, corresponding approximately to a few tens of reach-averaged widths, as also observed by previous field measurements and satellite observations (Abad et al., 2024; Burge & Lapointe, 2005; Nanson & Knighton, 1996). The scaling appears to be largely independent of the reach slope, and therefore of river type, and is associated with a modest difference in branch lengths. The latter feature, when interpreted in the light of the theoretical model, reflects the apparently great sensitivity of flux partitioning to small slope advantages arising from slight length differences between the two anabranches.

The above studies primarily addressed the physical conditions necessary for an anabranching unit to persist in the equilibrium state. Therefore, these analyses only provided a static picture, neglecting the transient evolution from the formative (young) stage to the mature configuration; they relied on the fundamental assumption that the planform geometry was fixed. Consequently, some major knowledge gaps remain to be addressed: how do locally anabranching systems determine, and possibly adjust, their own length? What is the role played by the mechanisms behind their morphogenesis? Why are chute channels in some cases maintained throughout the history of system development, while in other cases they eventually close off (e.g., Fisk, 1947; M. Grenfell et al., 2012)? More broadly, do anabranching units constitute a possible equilibrium sedimentary pattern, or are they inherently a transient feature (e.g., Rubey, 1952)?

The aim of this paper is to address these questions by developing a mechanistic model for the planform evolution of an anabranching unit, formulated in terms of changes in anabranch length over time. Describing such dynamics necessarily requires the identification of the physical processes that drive these changes. Perhaps, the most

common process is lateral migration, driven by erosion of the outer banks, which progressively increases the sinuosity of the anabranches, as occurs in single-thread meandering rivers (e.g., Seminara, 2006). However, it has been pointed out that in meandering rivers exhibiting a transitional planform, the tendency to chute initiation and the subsequent formation of islands primarily exert a twofold influence on system dynamics. First, bend elongation and migration within the anabranches are substantially reduced or even suppressed (Bridge et al., 1986; M. Grenfell et al., 2012; Luchi et al., 2007). Second, the overall sinuosity of the channel tends to decrease, not only within the anabranches, which may experience internal cutoffs, but also in the adjacent upstream and downstream bends (e.g., Gay et al., 1998; Howard, 1996; Johnson & Paynter, 1967). The development and persistence of internal bends are also strongly influenced by the partitioning of water and sediment at the upstream bifurcation (M. C. Grenfell et al., 2014; Lin et al., 2025; Van Dijk et al., 2012). If an anabranch silts up owing to insufficient flow supply, the system may eventually evolve toward a single-thread configuration. Once an anabranch becomes sealed, the adjustment in channel morphology, including rapid lateral migration, local widening, and steepening of the bed profile, affects both upstream and downstream reaches (Bridge et al., 1986; Schwenk & Foufoula-Georgiou, 2016).

Other processes, rather than being directly associated with the dynamics of the anabranches themselves, act in the vicinity of the island's upstream and downstream tips, and may lead either to an increase or to a decrease in the distance separating the bifurcation and confluence nodes. Among the lengthening mechanisms is that of progressive sediment deposition, often reinforced by vegetation growth (e.g., Gurnell et al., 2006; Leopold & Wolman, 1957; Rubey, 1952; Wintenberger et al., 2015). This process may occur both at the bifurcation head, where field observations by Bridge et al. (1986) and Burge (2006) indicate a relationship between bed material accumulation and flow partitioning, and at the confluence, where slack-water conditions promote sediment deposition (e.g., Best, 1988; Mosley, 1976; Zinger et al., 2013). In contrast, shortening is chiefly driven by scouring of island banks during high flows or overbank events, which may uproot vegetation and destabilize island margins (e.g., Gurnell et al., 2001; Osterkamp, 1998).

The relative importance of all these competing processes is regulated by the flow regime. Observations by Gautier et al. (2021) in an anabranching reach of the Lena River documented alternating phases of island accretion and recession, controlled by the duration and magnitude of flood events. Discharge variability and the associated fluctuations in sediment supply operate as key regulating factors in the temporal evolution of the system. Low or receding flows foster vegetation colonization by exposing bare sediment surfaces to seedling establishment and plant growth (e.g., Gilvear & Willby, 2006; Gurnell et al., 2012; Tal & Paola, 2010), whilst high flows redistribute discharge between the anabranches, sometimes inducing flow switching and thus preventing progressive infilling or abandonment of one branch (e.g., Bertoldi, 2012; Brice, 1982; Sorrells & Royall, 2014).

Building upon this information, in this study we seek a simplified, yet physically consistent, representation of the key processes governing water and sediment dynamics at the bifurcation and confluence nodes, as well as channel-island interactions and morphology-vegetation feedbacks, which may exert a first-order control on the adaptation of the nodal length. Beyond its theoretical relevance for understanding sedimentary patterns so widespread in fluvial environments, the study has relevant practical implications. Because anabranching configurations are often the result of river restoration interventions—such as the reopening of side channels—the proposed framework may be of interest for engineers who may wish to assess the feasibility and long-term sustainability of such measures, potentially reducing the need for repeated and costly maintenance.

The rest of the paper is organized as follows. In Section 2, observed trends of length variations captured through satellite images of several anabranching units are documented. Section 3 is devoted to the formulation of the mathematical model, introducing first the main assumptions adopted for the hydro-morphological characterization of the reach, then proceeding with the derivation of an adaptation equation for the distance between the bifurcation and confluence nodes (Section 3.1). The results are presented in Section 4 and discussed in Section 5, where the model predictions are contextualized on the basis of available data. Finally, Section 6 is devoted to some concluding remarks.

2. Satellite Observations

The starting point of our work is a remotely sensed analysis of several fluvial reaches locally exhibiting anabranching units. The examples presented in the following illustrate different trends in length variations associated with these patterns. Figure 1 shows two reaches: the Mamoré River along the border between Bolivia and Brazil

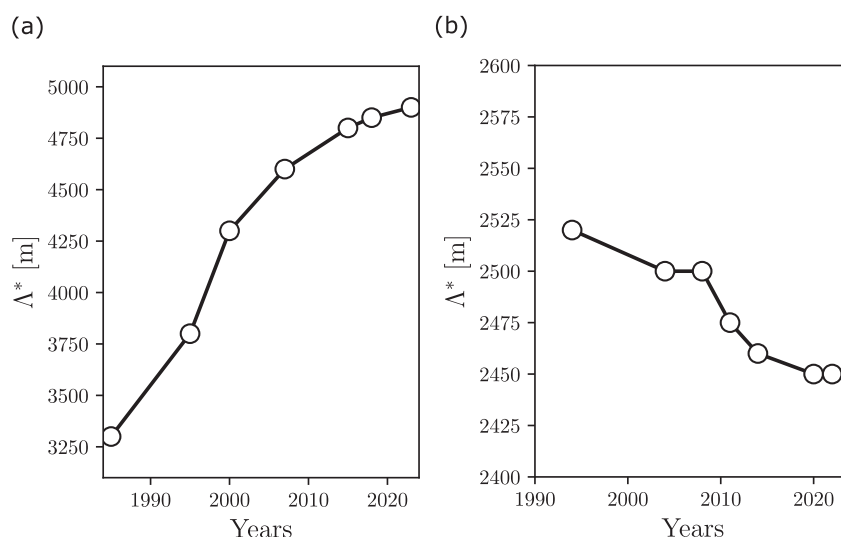


Figure 2. Temporal variation of the nodal length, Λ^* , of (a) the Mamoré River unit reported in Figure 1a, and (b) the Wapsinicon River unit reported in Figure 1b.

and the Wapsinicon River, Iowa, USA. Both anabranching units are characterized by a fairly symmetric planform; the Mamoré River unit comprises a single bend for each anabranch, while the Wapsinicon River unit comprises two or more internal bends. The selected reaches do not show evident signs of anthropogenic disturbance, the floodplains are vegetated and relatively homogeneous, and significant tributary inflows are absent. Our interest is to examine length variations in terms of the nodal length Λ^* (hereafter the superscript * will indicate a dimensional quantity), namely the Euclidean distance between the bifurcation and confluence nodes. This quantity was estimated manually on the basis of Landsat images and aerial photographs retrieved from Google Earth Engine over a temporal window of approximately 40 years, using visible vegetation lines at the head and tail of the island.

Figure 2 shows the temporal variation in nodal length for Mamoré River and Wapsinicon River reaches. The solitary anabranching unit of the Mamoré River exhibited a monotonic increase of Λ^* over the available observation period (Figure 2a). This growth was gradually slowing down, seemingly approaching a plateau around 5,000 m, which corresponds to approximately eight times the width of the reach at bankfull stage (600 m, Latrubesse (2008)). The estimated average rate of elongation was approximately 50 m per year. During the same period, the two anabranches exhibited only minor variations in sinuosity, maintaining a consistently low value around 1.1. Hence, the system was evolving primarily through changes in nodal length, while preserving a fairly symmetric shape with quasi-straight anabranches. The anabranching unit of the Wapsinicon River differs from the previous case in that the nodal length underwent only minor variations and decreased slightly over time, reaching a value in 2020 of about 19 times the reach-averaged bankfull width (122 m, Soar and Thorne (2001)). The two anabranches exhibited limited asymmetry, which remained nearly constant despite minor fluctuations due to an internal cutoff. The average recession rate was about 4 m per year.

These two examples illustrate anabranching units in which adjustments in nodal length, whether increasing or decreasing, occur while preserving the multi-thread structure. The absence of substantial width variations suggests that both anabranches convey comparable fractions of the total flow. A similar behavior for anabranching units in the Renous River, Canada, was reported by Burge and Lapointe (2005). However, this is not the only evolutionary pathway that can be observed.

Figure 3 shows two additional anabranching units from the Guaviare River, Colombia (Figure 3a) and Peace River, Canada (Figure 3b). In both cases, the nodal length exhibited a persistent increase, with no evident slowing trend toward stabilization. The estimated elongation rates ranged from approximately 17 m per year of the Guaviare River to 34 m per year of the Peace River. This lengthening was accompanied by the progressive narrowing of one anabranch, which became gradually colonized by vegetation over time. Although a narrow channel remained still visible in the most recent imagery, the observed evolution would suggest that these systems were transitioning toward a single-thread morphology.

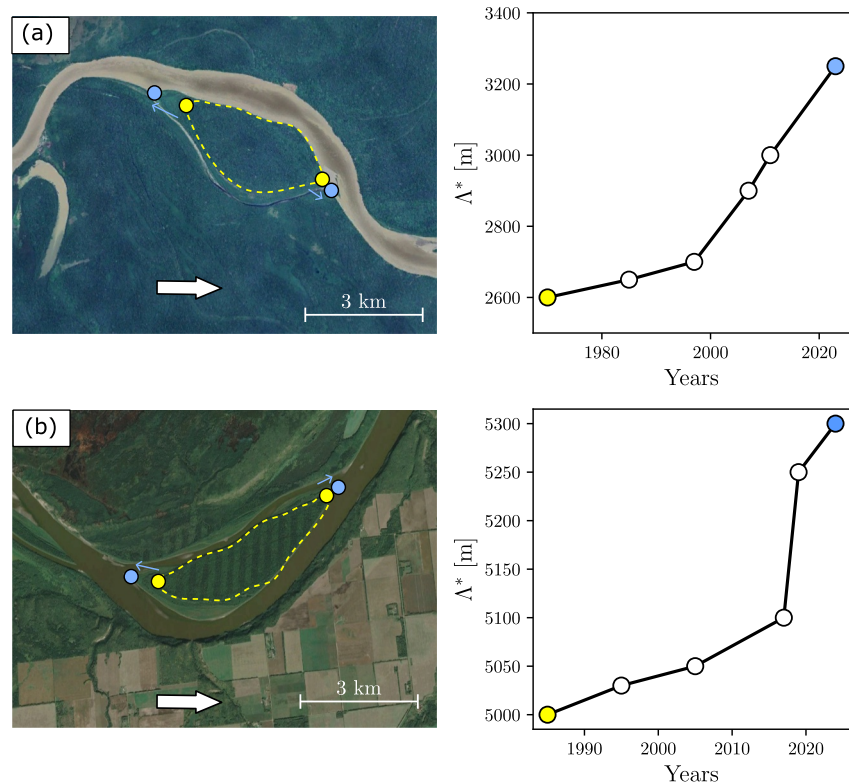


Figure 3. Temporal variation of the nodal length, Δ^* , in closing systems: (a) Guaviare River (Colombia), $2^\circ 52'N$ $71^\circ 29'W$, and (b) Peace River (Canada), $58^\circ 36'N$ $116^\circ 36'W$. The yellow dots and dashed lines indicate the configuration at the first observation, while the blue dots indicate the last observation. The white arrows denote flow direction. The images were retrieved from Google Earth.

Patterns similar to those observed in the Guaviare River and the Peace River were previously documented by Bridge et al. (1986) and Osterkamp (1998). The latter study also identified an additional mechanism leading to the loss of the multi-thread structure: the gradual disintegration of islands during successive flood events. Island removal by flow scouring was likewise documented in several other studies (e.g., Gautier et al., 2021; Guo et al., 2023; Gurnell et al., 2001; Picco et al., 2014).

Despite being sensitive to the duration of the monitoring window, the observations presented here, combined with evidence from earlier studies, suggest that solitary anabranching units may evolve so as to: (a) preserve the multi-thread structure through near-monotonic adjustments in nodal length, with limited variations in width and sinuosity; or (b) transition from a multi-thread to a single-thread morphology, either through the abandonment of one branch, apparently associated with progressive elongation of the system, with vegetation rapidly occupying newly available areas, or by removal of the island as a consequence of intense and frequent floods. This information forms the foundation for the development of the dynamic model that will be presented in the next section.

3. Formulation of the Model

We consider a river section as depicted in Figure 4. The river—denoted with m —is fed by a water discharge Q_m^* and a sediment supply Q_{sm}^* that are time-dependent and flows through a sufficiently wide floodplain so that its movement is not affected by confinement effects. At some point, the channel bifurcates into two anabranches, denoted by subscripts l, r , which reconnect at a downstream confluence, where the river retakes its single-thread shape. The downstream reach is long enough not to be affected by backwater effects. The anabranches, which may comprise a single or multiple bends, are separated by a vegetated island, which rises above the floodplain so that it is not entirely submerged during floods. This is our definition of an anabranching unit.

Before proceeding to the details of the formulation, it may be useful to introduce the main assumptions adopted to simulate the planimetric evolution. Planform variations are the result of competing processes that may elongate or

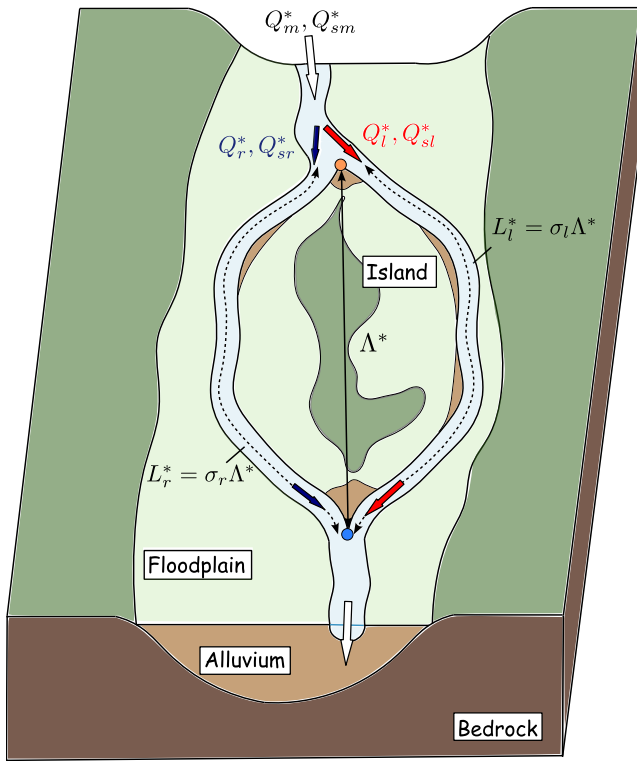


Figure 4. Schematic of the modeling planform with relevant variables. The green layer indicates vegetation cover on the floodplain. Light and dark green layers denote young and mature vegetation, respectively. The light brown layer indicates bare sediment. The arrows denote the water fluxes within the channels.

shorten the system. The lengthening mechanisms include the accumulation of bed material at the bifurcation head associated with vegetation establishment, the deposition of sediments in the recirculation zone just upstream of the confluence during the falling limb of a flood, as well as the lateral shift of each anabranch due to bank erosion that increases their sinuosity. In contrast, shortening may be caused by recession of the bifurcation head due to excess sediment removal during floods, or a decrease of the branch sinuosity due to internal bend cutoffs. All of these mechanisms should, in principle, be investigated locally, accounting for their spatial variability. This approach would introduce great complexity since one should account, for instance, for the spatial variations of bank erosion rate due to the local structure of flow and bed topography, or for the vegetation recruitment and growth onto deposited surfaces during consecutive months or even years of low-flow conditions. However, the problem is greatly simplified by adopting a reach-averaged description of elongation and recession, which allows these processes to be treated as spatially continuous. Therefore, it does not matter whether the lengthening or shortening occurs at a specific node (bifurcation or confluence). Likewise, a single representative value of sinuosity and migration rate is used to characterize each branch. Consequently, denoting with σ_i the anabranch sinuosities, the intrinsic length, L_i^* , which is defined as:

$$L_i^* = \Lambda^* \sigma_i, \quad i = l, r, \quad (1)$$

is a function of a single independent variable, the time t^* .

In this study, our attention is restricted to the particular case in which the planimetric evolution is driven exclusively by variations in nodal length Λ^* . The sinuosities σ_i are therefore assumed to be constant over time. Moreover, the planform geometry is assumed to be symmetric, such that $\sigma_l = \sigma_r = \bar{\sigma}$, where $\bar{\sigma}$ denotes the mean sinuosity. This assumption is equivalent to pre-

scribing a self-similar evolution in which the overall shape of the system is preserved throughout its temporal development. It follows that the rate of change of L_i^* is given by:

$$\frac{dL_i^*}{dt^*} = \bar{\sigma} \frac{d\Lambda^*}{dt^*}. \quad (2)$$

A further simplification of the problem is achieved by considering the estimates of the rate of change in nodal length displayed in Figures 2 and 3, which suggest that it is plausible to assume that, on the one hand, the role of tectonic uplift or subsidence can be reasonably neglected, and, on the other hand, the time scale of the planimetric variations is slower than that of in-stream morphodynamical processes. Moreover, we assume that the latter processes occur on a much larger time scale than that of flow adaptation, allowing the well-known quasi-steady approximation to be introduced (de Vries, 1965).

The steady state in each thread corresponds to a uniform turbulent flow. Considering a rectangular cross-section, let W_m^* and S_m denote the main channel width and bed slope, respectively, which are assumed to remain constant over the long-term timescale considered here. These quantities are set by the hydrological regime together with the sedimentological characteristics of the catchment, which are described below. Hence, the main channel flow depth D_m^* is computed using a normal flow relation of Chézy-type as follows:

$$D_m^*(t^*) = \left(\frac{Q_m^*(t^*)}{W_m^* C_m(t^*) \sqrt{g S_m}} \right)^{2/3}, \quad (3)$$

where g is the acceleration due to gravity, and C_m is the Chézy coefficient, which is estimated using a power-law formula:

$$C_m = \alpha_c \left(\frac{D_m^*}{d_s^*} \right)^{n_c}, \quad (4)$$

where d_s^* is the median grain size of the bed material. The coefficient α_c and the exponent n_c are set based on the regression values reported by Parker et al. (2007) and Wilkerson and Parker (2011) for gravel-bed and sand-bed rivers, respectively. Specifically, $\alpha_c = 3.71$ and $n_c = 0.263$ for gravel-bed rivers, while $\alpha_c = 1.44R_p^{0.197}$ and $n_c = 0.413$ for sand-bed rivers, where R_p is the particle Reynolds number.

Attention is now turned to the flow and sediment transport conditions within the anabranches. Let the widths of the left and right channels be equal to half the width of the main channel, that is, $W_l^* = W_r^* = W_m^*/2$. As for the main channel, these widths are assumed to be preserved throughout the planform evolution. Together with the assumption of constant sinuosities, this represents one of the main simplifications of the present formulation. The two assumptions are likely not independent, as both the width and the length of the individual branches may vary substantially depending on the morphogenesis of the system and on the temporal evolution of water and sediment fluxes distribution. Therefore, their validity is expected to be mainly restricted to systems generated by the growth of a mid-channel bar, which is likely to produce relatively symmetric planforms, or for systems whose intermediate evolutionary stages are characterized by limited asymmetry between the two branches, as displayed by the Mamoré and Wapiniccon rivers shown in Figure 1. In contrast, when an anabranching unit originates from chute-cutoff or avulsion, substantial differences in branch widths and lengths may arise from the early stages of development. Similarly, during the final stages of evolution, when one branch progressively approaches closure, its width may differ significantly from that observed during earlier stages, as illustrated by the examples shown in Figure 2.

The flow depths $D_{l,r}^*$ are expressed by Equation 3. Unlike in the case of the main channel, neither the discharges nor the bed slopes are known a priori, since both depend on the flow and sediment partitioning at the bifurcation. To determine the flux distribution between the anabranches, we rely on the model developed by Ragno et al. (2021), and subsequently extended in Ragno and Tubino (2025), who introduced a one-dimensional framework for computing the equilibrium states of an anabranching unit.

Hydraulic variables are assumed to be spatially uniform within each branch, except in regions adjacent to bifurcation and confluence nodes, where a two-cell scheme is employed to account for lateral exchanges of mass and momentum. At the bifurcation, five internal conditions are specified: the conservation of water and sediment mass, the same water level in all branches at the node, and a relation governing the sediment transport partition. Following Bolla Pittaluga et al. (2003), the latter results from the combined effect of water flux exchange and gravity, which deflects sediment particles moving over a laterally sloping bed. Similarly, four nodal conditions are imposed at the confluence; in addition to the conservation of water and sediment mass, two further conditions are derived from the streamwise momentum balance applied to two distinct control volumes. Further features and outcomes of the model will be summarized in the next section.

We shall now discuss how the flow regime is treated. Following previous studies (e.g., Carlin et al., 2021; Lauer & Parker, 2008; Naito & Parker, 2019), the full hydrograph is not modeled; instead, flow variability is taken into account using the flow duration curve, which is assumed to be stationary. Alterations in the hydrological and sediment supply regimes, such as those, for instance, associated with climate change or dam construction, are not investigated henceforth.

Let $p_Q(Q_m^*)$ denote the probability density of the water discharge supplied to the river, defined as the derivative of the flow duration curve. To simplify the formulation, $p_Q(Q_m^*)$ is approximated as a series of N discrete steps corresponding to discharges Q_{mj}^* ($j = 1, \dots, N$). This approach is analogous to the method used in morphodynamic studies dealing with grain sorting, where the distribution of sediment grain sizes in a mixture is approximated by discrete size fractions (e.g., Seminara et al., 1996). Accordingly, the density function p_Q is expressed as follows:

$$p_Q = \sum_{j=1}^N p_j \delta(Q_m^* - Q_{mj}^*), \quad (5)$$

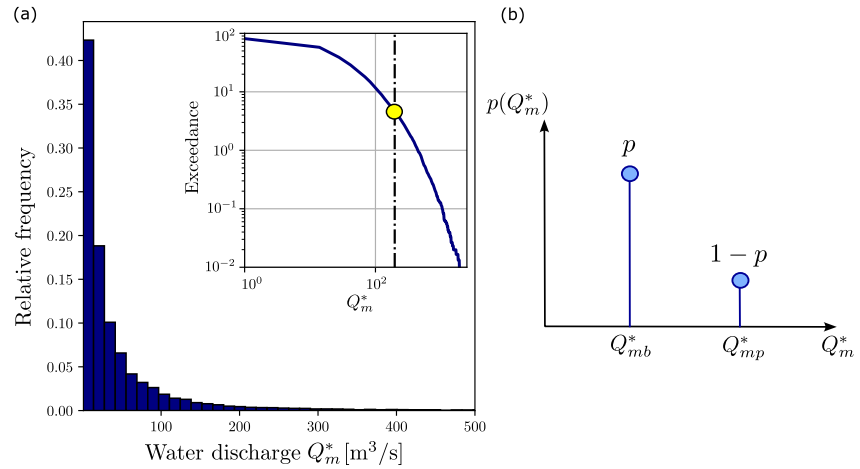


Figure 5. (a) Relative frequency distribution of flow stages and duration curve (inset) of the Big Sioux River near Akron, IA, USA. The plots are constructed on the basis of approximately 80 years of discharge record at the U.S. Geological Survey (USGS) gauging station at Akron (station 06485500). The yellow circle on the duration curve indicates the bankfull discharge, which is about 197 m³/s (Soar & Thorne, 2001). (b) Schematic of the flow hydrograph by means of two-modes flow discharges.

where $\delta(Q_m^* - Q_{mj}^*)$ is the Dirac distribution centered at Q_{mj}^* . To further simplify the analysis, it is assumed that $N = 2$, and then the hydrological regime is reduced to two modes: a base flow Q_{mb}^* that occurs for a fraction of time p , and a peak flow Q_{mp}^* that occurs for the remaining fraction $(1 - p)$ (Figure 5). This simplified description of the flow regime, as discussed in Blom et al. (2017), is quite useful because it allows the analytical computation of a representative *formative discharge* characterizing the long-term average response of the system. The concept of formative discharge embraces a variety of definitions that have been proposed over the years (e.g., see for a review Blom et al. (2017) and Bolla Pittaluga et al. (2014)). Here we suppose that the formative discharge Q_{mf}^* —the suffix f denotes formative conditions—is the steady water discharge that, for a given slope, provides the same sediment load as the long-term hydrograph (e.g., Doyle & Shields, 2008). The mean sediment yield of the reach $\bar{Q}_{sm}^*$ is given by:

$$\bar{Q}_{sm}^* = W_m^* \sqrt{(s_g - 1)gd_s^{*3}} [p \Phi_m(Q_{mb}^*) + (1 - p)\Phi_m(Q_{mp}^*)], \quad (6)$$

where s_g is the ratio between sediment and water density, and Φ_m is the equilibrium transport intensity, which is a function of the Chézy coefficient and of the Shields number θ_{mj} :

$$\theta_{mj} = \frac{D_{mj}^* S_m}{(s_g - 1)d_s^*}, \quad j = b, f, p. \quad (7)$$

Once the mean value of the sediment load is known from Equation 6, the formative flow depth of the main channel D_{mf}^* is obtained by reversing the transport function Φ_m , which depends on the type of river considered. The formative discharge Q_{mf}^* is then computed from Equation 3. The above formulation sets out the basis for the nodal-length adaptation model described in the following section.

3.1. An Adaptation Equation for the Nodal Distance

The equation that models the evolution of the length Λ^* must express the balance between the processes promoting the elongation of the island and those contributing to its shortening, depending on whether the flow stage corresponds to base flow or peak flow. We therefore assume the following adaptation equation:

$$\frac{d\Lambda^*}{dt^*} = p \mathcal{F}_E(\Lambda^*) - (1 - p) \mathcal{F}_S(\Lambda^*), \quad (8)$$

where F_E and F_S are the two functions describing elongation and shortening rates, respectively.

In modeling the elongation term, observations (e.g., Gran & Paola, 2001; Gurnell et al., 2006; Leopold & Wolman, 1957; Osterkamp, 1998; Rubey, 1952; Wintenberger et al., 2015) suggest that the role of vegetation in trapping and stabilizing sediment deposits should be taken into account. The mathematical formulation adopted to simulate plant growth is not intended to provide a detailed treatment of the vegetation-morphology interaction. Rather, it serves as a minimal representation of what we consider to be the main effect of vegetation growth relevant to island elongation.

Vegetation may occupy and stabilize emerging sediment during low-flow periods. After a flood event is extinguished, the water level declines, resulting in the exposure of bare soil that provides a substrate for vegetation establishment, particularly near the head and tail of the island. Plants that survive flooding continue to grow, while successive seeding increases the vegetation density. Over time, this process merges the initially unvegetated sediment deposit with the island. A similar mechanism typically supports the accretion of point bars in meandering rivers (e.g., Parker et al., 2011; Pizzuto, 2008; Zen et al., 2016).

In an anabranching system, it is plausible that the encroachment rate is further influenced by the sediment flux partitioning at the bifurcation. Field studies by Bridge et al. (1986) and Burge (2006) provided some evidence that island elongation was promoted when anabranches carried a similar proportion of flow. Under conditions of nearly balanced flow partitioning, a gradual wedge-shaped bar was observed to form upstream of the bifurcation. This bar increased in elevation over time, supported by vegetation growth. In contrast, no sediment accreting ramp was observed when the flow partition was strongly unbalanced. In this case, a bed step developed at the inlet of the two anabranches, the channel receiving the smaller fraction of water and sediment being distinguished by a higher bed elevation due to the formation of a depositional bar at the entrance.

The presence of an inlet step is a common feature of channel bifurcations both in gravel-bed and sand-bed rivers, as widely documented in previous experimental (e.g., Bertoldi & Tubino, 2007) and field studies (e.g., Abad et al., 2024; Leopold & Wolman, 1957; Zolezzi et al., 2006). Furthermore, if only one of the anabranches transports sediments under base-flow conditions, and if these conditions persist for a sufficiently long period, the other branch may become completely colonized by vegetation, ultimately leading to island attachment to the floodplain (Nadler & Schumm, 1981; Osterkamp, 1998).

The degree of discharge asymmetry between the two anabranches also influences the dynamics of the downstream confluence. When the two anabranches convey a similar amount of flow, a backwater effect is produced by the colliding streams, creating a zone of slack water that promotes sediment deposition at the tip of the island (e.g., Best, 1987; Biron et al., 2002; Mosley, 1976). In contrast, when the incoming fluxes are unequal, transverse gradients of the water surface and bed topography develop. As shown theoretically (Ragno et al., 2021) and documented by field data (Sukhodolov & Sukhodolova, 2019), the most-carrying branch experiences higher water surface elevation and bed degradation, and vice versa for the non-dominant branch, at least for confluences that do not display marked asymmetric geometry. When nearly all flow is conveyed by one branch, a drawdown effect is induced at the confluence. This leads to aggradation of the less-carrying branch, up to the point that a sediment plug may form, causing the branch to close off, with the subsequent formation of an oxbow channel (Le Coz et al., 2010; Zinger et al., 2013). In other words, the lower-carrying branch may be unable to impinge on flow across the confluence zone due to the dominant influence exerted by the flow of the higher-carrying branch. It follows that under strongly unbalanced flow conditions, erosion and deposition may be such that the downstream shift of the confluence would be hampered.

Based on these arguments, the elongation term F_E is written as:

$$F_E(\Lambda^*) = \zeta_{veg}^* \mathcal{E}_1 \cdot \mathcal{E}_2(\Lambda^*), \quad (9)$$

where ζ_{veg}^* is a characteristic speed of vegetational encroachment, and $\mathcal{E}_1 \cdot \mathcal{E}_2(\Lambda^*)$ is the product of two non-negative functions representing two distinct processes, of which only $\mathcal{E}_2$ is assumed to depend on the nodal length.

The function $\mathcal{E}_1$ represents a *shoaling function*, which describes the dependence of vegetation growth on the extent of exposed dry areas. Lower discharges result in larger areas potentially available for vegetation establishment (e.g., Garcia Lugo et al., 2015; Tal & Paola, 2010). As the wetted width decreases and the sediment

surfaces emerge, the expanded area available for colonization can support a higher density and diversity of vegetation (e.g., Camporeale & Ridolfi, 2006; Corenblit et al., 2009; Gurnell et al., 2012). This enhanced vegetative colonization is likely to contribute to accelerated island growth and elongation.

We express this behavior by introducing the functional $\mathcal{E}_1 = f(\beta_{mb})$, where $\beta_{mb} = W_m^*/2D_{mb}^*$ denotes the half-width-to-depth (or aspect) ratio of the main channel under base flow. Neglecting the fraction of width occupied by vegetation, a simple way to take into account the shoaling effect is to write $\mathcal{E}_1$ as:

$$\mathcal{E}_1 = \begin{cases} 1, & \delta \leq 0, \\ 1 + \delta, & \delta > 0, \end{cases} \quad (10)$$

with the parameter δ measuring the relative distance from the emersion conditions:

$$\delta = \frac{\beta_{mb} - \beta_{emr}}{\beta_{emr}}, \quad (11)$$

where β_{emr} is an *emersion-limited* value of the channel aspect ratio. An estimate for β_{emr} is obtained on the basis of the non-linear bar theory of Colombini et al. (1987), as shown by Redolfi et al. (2020), generating values of β_{emr} that range from 20 to 30, depending on the relative grain roughness. In this work, a constant value of $\beta_{emr} = 20$ is set.

The function $\mathcal{E}_2(\Lambda^*)$ is a *flux partition function*, representing the dependence of elongation on the sediment discharge asymmetry under base-flow conditions, ΔQ_{sb} :

$$\Delta Q_{sb} = \frac{Q_{slb}^* - Q_{srb}^*}{Q_{smb}^*} = f(\Lambda^*). \quad (12)$$

The flux partition function is defined to satisfy the following criteria: $\mathcal{E}_2$ is maximized when $\Delta Q_{sb} = 0$, that is, when sediment discharge is evenly distributed between branches, while $\mathcal{E}_2 \rightarrow 0$ when $|\Delta Q_{sb}| \rightarrow 1$, that is, when the entire flow is routed through a single branch. To meet these requirements, a possible functional form for $\mathcal{E}_2(\Lambda^*)$ is as follows:

$$\mathcal{E}_2(\Lambda^*) = 1 - \alpha_\Delta |\Delta Q_{sb}(\Lambda^*)|, \quad (13)$$

where α_Δ is a coefficient lower than 1, and the modulus is included because it is assumed that elongation is independent of whether the left or right branch carries the majority of the flow. The choice of taking $\alpha_\Delta < 1$ ensures that the rate of elongation gradually decreases as $|\Delta Q_{sb}|$ increases, while preventing it from being entirely suppressed even in cases of complete diversion of sediment flux to one branch.

As anticipated, the sediment discharge asymmetry is computed employing the equilibrium solution of an anabranching unit derived by Ragno et al. (2021). If the lengths and widths of the anabranches are the same, the response of the system is governed by two dimensionless parameters: the *bifurcation number* $\mathcal{R}$, which controls the dynamics of the bifurcation, and the dimensionless mean length of the anabranches $\bar{L}$, which measures the magnitude of coupling between the bifurcation and confluence nodes. Temporarily omitting the subscript j indicating the flow stage, these parameters are defined as:

$$\mathcal{R} = \frac{r\alpha}{\beta_m \sqrt{\theta_m}}, \quad \bar{L} = \frac{\Lambda^* \bar{\sigma}}{C_m^2 D_m^*}, \quad (14)$$

where r is an empirical coefficient that spans between 0.3 and 0.6 and measures the gravity effect on transverse sediment transport (e.g., Ikeda, 1982; Talmon et al., 1995), and α is a parameter of the order of one (Bertoldi & Tubino, 2007; Redolfi et al., 2016), physically representing the upstream portion of the main channel affected by two-dimensional effects on flow and bed topography induced by the unequal hydro-morphodynamical response

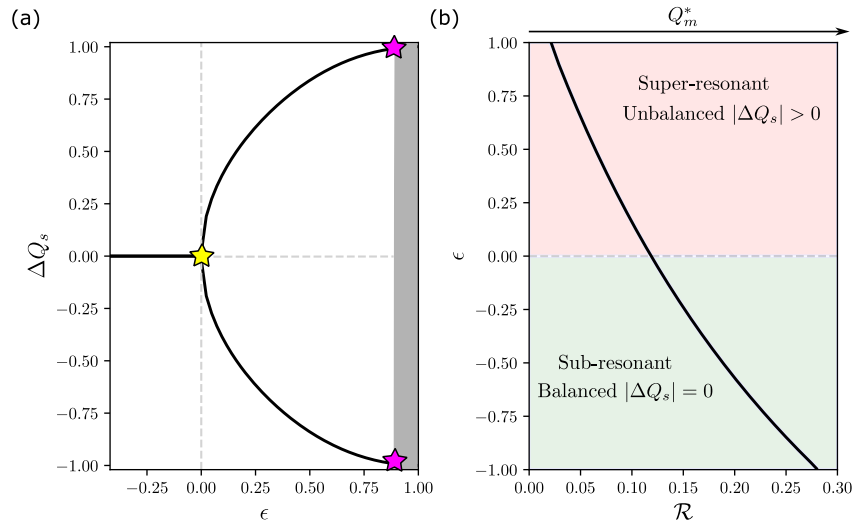


Figure 6. (a) Example of equilibrium diagram for a “free” anabranching unit: the stable solutions of the sediment discharge asymmetry ΔQ_s are plotted as a function of the deviation from criticality, ϵ . The yellow and magenta stars denote the criticality and the point of vanishing transport in one of the two branches, respectively. The gray box highlights conditions where just one anabranch is morphodynamically active. (b) The control parameter ϵ is plotted against the bifurcation number, R , for a given mean length $\bar{L} = 5$. Higher (lower) values of R correspond to higher (lower) discharges, Q_m^* . Without loss of generality, both plots refer to a generic sand-bed river for which the Engelund and Hansen (1967) transport formula is adopted.

of the bifurcates. The bifurcation number computed with reference to bankfull conditions is typically of $O(10^{-1})$ and $O(10^{-2})$ for gravel-bed and sand-bed rivers, respectively (Ragno & Tubino, 2025).

The two quantities R and $\bar{L}$ control the “free” response of the anabranching unit, which corresponds to a symmetric geometrical configuration without “forcing” effects that may determine a preferential path for water and sediment distribution in the downstream branches (e.g., a branch that is shorter and thus steeper than the other). In this free configuration, the system exhibits a critical behavior, where the shift from balanced to unbalanced flux partitioning represents a *continuous phase transition* (Ragno, 2023). For a given value of R , if the dimensionless mean length is shorter than a critical length $\bar{L}_{cr}$, the only stable state is $\Delta Q_s = 0$; differently, if $\bar{L} > \bar{L}_{cr}$ the system attains a multi-stable state, where either the left or right branch captures the largest fraction of the flow. A similar reasoning can be applied by keeping $\bar{L}$ fixed and changing R . The effect of R and $\bar{L}$ is combined into a single variable, ϵ :

$$\epsilon = \frac{\bar{L} - \bar{L}_{cr}(R)}{\bar{L}_{cr}(R)}, \quad (15)$$

which is a dimensionless measure of the deviation from criticality.

Figure 6a illustrates a typical plot of spontaneous breaking of the balanced state as ϵ exceeds the critical value. The free solution displays a pitchfork bifurcation, with the sediment discharge asymmetry behaving as $\Delta Q_s \sim \epsilon^{1/2}$ near the critical point (Ragno, 2023). For $\epsilon > 0$, the system has a stable equilibrium in which one of the two branches captures the majority of the incoming discharge, while the other is still able to carry sediment. Increasing the value of ϵ , hence the mean length, at some point one of the branches becomes morphologically inactive ($|\Delta Q_s| = 1$), thereby establishing the conditions for its closure.

The critical threshold is associated with the concept of *morphodynamic resonance* (Redolfi et al., 2016), originally defined by Blondeaux and Seminara (1985) in the theory of regular meanders. It delineates two distinct regimes: sub-resonant ($\epsilon < 0$), characterized by a balanced flux partition, and super-resonant ($\epsilon > 0$), where the flux distribution is unbalanced. The analysis carried out in Ragno and Tubino (2025) showed the quasi-universal tendency of anabranching units to exhibit a super-resonant behavior.

The analytical expression for $\bar{L}_{cr}$ can be found in Ragno and Tubino (2025). A simple expression is obtained if the sediment transport function is computed using a power-law formula. For example, using the Engelund and Hansen (1967) formula and assuming for simplicity that $\bar{\sigma} = 1$, it can be shown that:

$$\bar{L}_{cr} = \Lambda_{cr} = -\frac{1}{4} \left[\frac{23 + 12\mathcal{R}}{12\mathcal{R} - 5} \right], \quad (16)$$

whence for $\mathcal{R} = 0$ ($\beta_m \rightarrow \infty$) the lower limit is $\Lambda_{cr}|_{\mathcal{R}=0} = 23/20$, whereas for $\mathcal{R} \geq 5/12$, the system is sub-resonant independently of its length.

The value of ϵ not only changes during the planform evolution, according to the variation in nodal length, but also varies with the flow stage, as the critical length depends on the bifurcation parameter, which is inversely proportional to both the aspect ratio and the Shields number (Equation 14). In particular, $\bar{L}_{cr}$ is an increasing function of $\mathcal{R}$. As displayed in Figure 6b, ϵ decreases with increasing discharge, which corresponds to higher values of the aspect ratio, implying a reduction in flow asymmetry. This trend implies that more unbalanced configurations are associated with low-flow stages, as commonly reported in field studies (e.g., Bertoldi, 2012; Sorrells & Royall, 2014), where an anabranch may be inactive during base flow but becomes active during peak-flow events.

It should be noted that, according to the simplified representation of the elongation dynamics adopted here, the two functions $\mathcal{E}_1$ and $\mathcal{E}_2$ are not strictly independent, because they both depend on the aspect ratio through the relationship $\mathcal{R} \propto 1/\beta$. This reflects the implicit assumption that sediment partitioning and vegetation growth in bare areas are linked processes. A strong imbalance in sediment partitioning promotes aggradation in the lower-carrying branch, leading to enhanced emergence, and consequently to suitable conditions for vegetation colonization. In contrast, more balanced sediment fluxes may still promote vegetation growth through the gradual development of gently sloping depositional surfaces, similar to the ramp-like features described by Burge (2006). From a mathematical standpoint, for a given length, large values of $\mathcal{R}$, such that $\delta = 0$, correspond to $|\Delta Q_s| \rightarrow 0$, which maximizes $\mathcal{E}_2$. In contrast, as $\mathcal{R}$ decreases and $|\Delta Q_s|$ progressively approaches unity, $\mathcal{E}_2$ decreases toward the lower bound $(1 - \alpha_\Delta)$, while $\mathcal{E}_1$ increases significantly.

Consider now the shortening term. Assuming that the island recession is mainly caused by erosion of bank material, a Partheniades-type relationship is employed (e.g., Eke et al., 2014; Howard, 1992), according to which the channel shortening rate is linearly related to the excess shear stress between the dimensionless shear stress of the reach at peak flow (θ_{mp}) and that at formative flow (θ_{mf}):

$$\mathcal{F}_S(\Lambda^*) = \zeta_{ers}^* S(\Lambda^*) \frac{\theta_{mp}}{\theta_{mf}}, \quad (17)$$

where ζ_{ers}^* is a reference erosion rate, which should depend upon a number of factors, including the characteristics of bank material (e.g., Hasegawa, 1989) and the sediment load (e.g., Constantine et al., 2014). The function $S(\Lambda^*)$ is a *backwater function*, which ranges between 0 and 1, representing the reduction in transport intensity caused by the backwater effect associated with the super-elevation of the water surface at the confluence. Thus, $S(\Lambda^*)$ is expressed as:

$$S(\Lambda^*) = \begin{cases} 1, & L_B \geq 1, \\ (1 - \alpha_S)L_B + \alpha_S, & L_B < 1, \end{cases} \quad (18)$$

where α_S is a positive constant lower than 1, and L_B is the ratio between the mean length of the anabranches and the backwater length, L_B^* , which is given by:

$$L_B = \frac{\bar{L}^*}{L_B^*} = \Lambda^* \bar{\sigma} \frac{S_m}{D_{mf}^* (1 - F_{mf}^2)}, \quad (19)$$

where $F_{mf} = C_{mf}\sqrt{S_m}$ is the formative Froude number of the main channel. The backwater function therefore is the only term of $\mathcal{F}_S$ that depends (linearly) on the length of the anabranches, at least for values of $\bar{L}^*$ shorter than L_B^* .

Based on Equation 17, it is assumed that island recession is driven by the flow and occurs mainly at flood stage, at a rate ζ_{ers}^* , with the formative Shields number θ_{mf} acting as the lower limit for erosion (Eke et al., 2014; Howard, 1992). The key difference from standard bank erosion models lies in the function $S(\Lambda^*)$, which provides a simplified closure to account for the flow deceleration in the vicinity of the confluence, and the associated reduction in shear stress, an effect that may create a potential sediment sink (Hackney et al., 2018).

Substituting Equations 9 and 17 into Equation 8 gives the dimensional adaptation equation:

$$\frac{d\Lambda^*}{dt^*} = p \zeta_{veg}^* \mathcal{E}_1 \cdot \mathcal{E}_2(\Lambda^*) - (1-p) \zeta_{ers}^* S(\Lambda^*) \frac{\theta_{mp}}{\theta_{mf}}. \quad (20)$$

It may be noticed that Equation 20 is reminiscent of the adaptation equations commonly employed to simulate biological transport networks containing closed loops, such as blood vessels, leaf venation, and slime molds (e.g., Hu & Cai, 2013; Ronellenfitch & Katifori, 2016; Tero et al., 2010), topologically similar to the anabranching units investigated here.

It is now convenient to make Equation 20 dimensionless. The timescale of planform evolution, T_p^* , should be dictated by the vegetation encroachment rate ζ_{veg}^* , or alternatively by the erosion rate ζ_{ers}^* . Thus, we set:

$$T_p^* = \frac{C_{mf}^2 D_{mf}^*}{\zeta_{veg}^*}, \quad (21)$$

where the numerator represents the characteristic scale of length variations (Ragno et al., 2021, 2022). Consequently, T_p^* can be interpreted as the time it takes for vegetation to colonize a distance of the order of $C_{mf}^2 D_{mf}^*$.

Hence, the dimensionless adaptation equation becomes the following:

$$\frac{d\Lambda}{dt} = p \mathcal{E}_1 \cdot \mathcal{E}_2(\Lambda) - (1-p) \Upsilon S(\Lambda) \frac{\theta_{mp}}{\theta_{mf}}, \quad (22)$$

where $\Upsilon = \zeta_{ers}^*/\zeta_{veg}^*$ is the ratio between the reference erosion rate and the encroachment rate. The dimensionless representation is particularly convenient because it shows clearly how the length evolution is driven by terms of different physical nature: in-stream processes ($\Delta Q_s, S$), bank and floodplain processes (Υ), and the characteristics of the hydrological forcing (p).

The solution of Equation 22 is computed numerically adopting an explicit Adam-Bashforth scheme. The initial condition is imposed by prescribing an initial nodal length along with the values of flow parameters for the main channel and of the parameters $\bar{\sigma}, p, \alpha_\Delta, \alpha_S$, and Υ . In particular, since the median grain size, the reach slope, the width and the Chézy coefficient of the main channel are fixed, the flow conditions corresponding to different flow stages are uniquely determined by assigning values to β_{mb}, θ_{mb} , and θ_{mp} . Specifying these parameters allows for the computation of the formative Shields number, θ_{mf} , using Equation 6. For example, when the formula of Engelund and Hansen (1967) is used to compute the sediment transport function Φ_m , θ_{mf} reduces to:

$$\theta_{mf} = [p\theta_{mb}^{5/2} + (1-p)\theta_{mp}^{5/2}]^{2/5}. \quad (23)$$

Once θ_{mf} is known, the formative aspect ratio β_{mf} is computed from the following relation:

$$\beta_{mf} = \beta_{mb} \frac{\theta_{mb}}{\theta_{mf}}. \quad (24)$$

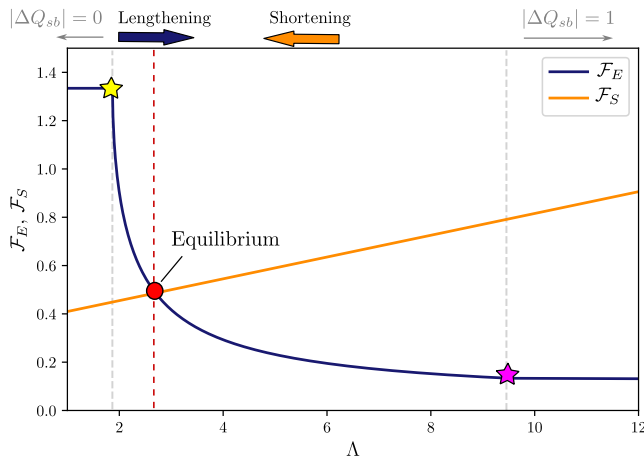


Figure 7. The dimensionless elongation (F_E) and shortening (F_S) functions against the dimensionless nodal length Λ , for values of $R_b = 0.08$, $\Upsilon = 1$, $\alpha_S = 0.5$ and $\alpha_\Delta = 0.95$. The yellow and magenta stars correspond, respectively, to the critical point $\epsilon = 0$, and the point of vanishing transport in one of the two branches ($|\Delta Q_{sb}| = 1$). The red dot indicates the equilibrium nodal length.

Alternatively, a value for the base flow bifurcation number, R_b , can be assigned instead of β_{mb} . It follows that the formative bifurcation number is given by:

$$R_f = R_b \sqrt{\frac{\theta_{mf}}{\theta_{mb}}} \quad (25)$$

The parameter R_f plays a central role in the analysis, as it controls the distance from the criticality, ϵ , and synthesizes the reach-scale characteristics upon which the sediment flux distribution at formative stage ΔQ_{sf} depends (Figure 6). The variable ΔQ_{sf} serves as a fundamental indicator for assessing whether the system can sustain a multi-thread shape in the long term. Specifically, we assume that the transition from multi-thread to single-thread, resulting from the closure of one anabranch, reflects morphological inactivity under formative conditions.

Equation 22 admits an equilibrium nodal length when the two competing processes are balanced, that is, $F_E = F_S$, as illustrated in the example of Figure 7. When the bifurcation and confluence nodes are in close proximity, specifically for $\tilde{L} < \tilde{L}_{cr}$ (implying $\epsilon < 0$), the sediment partition remains balanced and the elongation function is maximized. In this case $F_E > F_S$. As

Λ increases, the critical threshold is eventually exceeded ($\epsilon > 0$), so that a growing fraction of sediment is directed to a single branch. Consequently, F_E decreases in accordance with Equation 13. Conversely, when Λ is sufficiently large, the shortening term dominates ($F_S > F_E$). As Λ decreases, F_S gradually declines, following the behavior of the functional S , as described by Equation 18, while F_E increases due to a lower degree of transport asymmetry.

4. Results

4.1. Example of Model Implementation: Mamoré River

We first consider the reach of the Mamoré River shown in Figure 1a, which represents a typical sand-bed river, as a test case for the proposed model. The dynamics of this system is of particular interest not only because it provides a nice example of an anabranching unit, but also because it may be representative for those units formed as a consequence of river restoration works, where embankments restrict the lateral migration of the anabranches, a condition frequently observed in regulated fluvial systems.

Basic hydro-morphological data were retrieved from Latrubesse (2008) and the Global Runoff Data Centre (GRDC, <https://grdc.bafg.de/>). By employing the Engelund and Hansen (1967) formula to compute the sediment load, we obtain the following estimates of the model parameters: $\beta_{mb} = 35$, $\beta_{mp} = 19$, $\theta_{mp} = 2.9$, and $p = 0.8$. The corresponding values of the formative parameters are: $\beta_{mf} = 28$, $\theta_{mf} = 1.9$, and $C_{mf} = 18$. In addition, we set $\bar{\sigma} = 1.1$, $\alpha_\Delta = 0.95$ and $\alpha_S = 0.5$. In the absence of direct measurements, the erosion-to-encroachment rate ratio Υ is treated as a calibration parameter, and the sensitivity to this parameter is examined. To compute the characteristic time scale, we assign the value of $\zeta_{veg}^* = 50$ m per year, based on the average variation of Λ^* , as reported in Figure 2a.

Figure 8 shows the temporal evolution of the nodal length Λ^* and the modulus of sediment discharge asymmetry evaluated under formative conditions $|\Delta Q_{sf}|$, for different values of the parameter Υ . For $\Upsilon = 1$ (solid line), that is, $\zeta_{veg}^* = \zeta_{ers}^*$, the system undergoes rapid elongation over the first 100 years. This phase is associated with balanced flux partitioning (i.e., sub-resonant state, with $|\Delta Q_{sf}| = 0$). As the nodal length exceeds the critical value (yellow star), approximately around 5,000 m, the elongation rate decreases and the system gradually approaches an equilibrium state, with a length around 6,000 m. This equilibrium lies in the super-resonant regime and is characterized by an unbalanced sediment flux partitioning under formative flow in which both anabranches remain morphologically active.

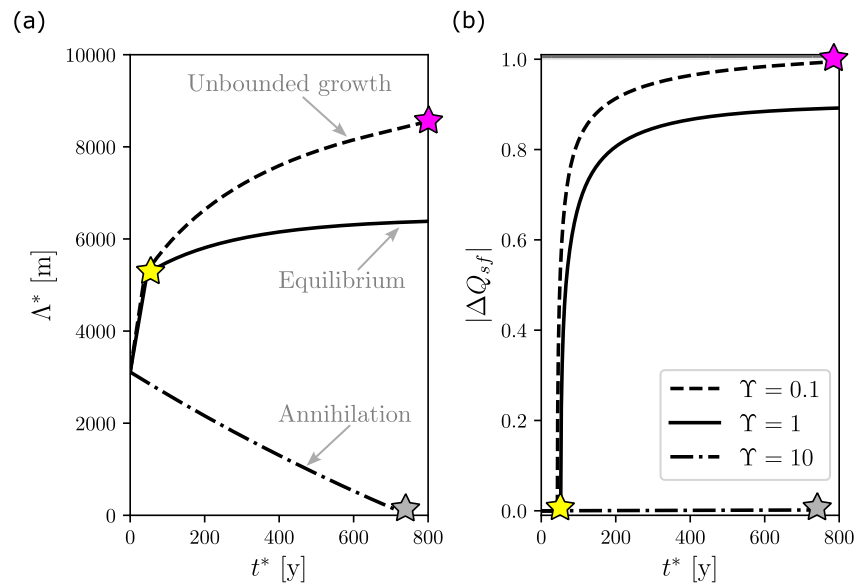


Figure 8. Model simulation based on data from Mamoré River at the border between Bolivia and Brazil (Figure 1a). Temporal variation of (a) nodal length, Λ^* , and (b) modulus of the discharge asymmetry index in formative conditions, $|\Delta Q_{sf}|$, for different values of the erosion-to-encroachment rate ratio, Υ . The yellow star indicates the time at which the system outreaches the critical point, then changing its morphodynamic regime. The gray and magenta stars indicate the time when the system is expected to turn into a single-thread configuration due to dominant erosion (gray star) or closure of one branch due to excessive elongation (magenta star).

In contrast, the additional curves displayed in the figure indicate two markedly different scenarios. For $\Upsilon = 0.1$ (dashed line), that is, $\zeta_{veg}^* = 10 \zeta_{ers}^*$, the solution shows unbounded growth. After exceeding the critical nodal length, the system continues to elongate until $|\Delta Q_{sf}| \rightarrow 1$ (magenta star), implying that one of the two anabranches is no longer active. A transition from a multi-thread to a single-thread morphology also occurs for $\Upsilon = 10$ (dash-dotted line), that is, $\zeta_{ers}^* = 10 \zeta_{veg}^*$. In this case, the transition is not driven by sediment flux partitioning but by the progressive annihilation of the island, which ultimately leads to the disappearance of the anabranching unit after about 750 years (gray star).

The different evolutionary pathways predicted by the model are summarized in the phase space $\mathcal{F}_E - \mathcal{F}_S$ of Figure 9, where the asymptotic states are visualized as convergence lines. Depending on the value of Υ , $\mathcal{F}_E$ and $\mathcal{F}_S$ may take arbitrary values. The three lines plotted in the figure, each starting from a circle corresponding to initial elongation and recession rates, show how the system evolves toward the asymptotic state. Over time, the values of $\mathcal{F}_E$ and $\mathcal{F}_S$ decrease, but the condition $\mathcal{F}_E = \mathcal{F}_S$ is not satisfied in all cases, so the system can continue to elongate (blue arrows) or shorten (orange arrows).

Figure 9 shows the three possible regimes. Two of them correspond to parameter regions where the system is likely to become single-threaded. Specifically, this happens either (a) under dominant erosion ($\mathcal{F}_S > \mathcal{F}_E$), which progressively shortens the nodal length until the anabranching unit disappears (orange box), or (b) under dominant elongation ($\mathcal{F}_S < \mathcal{F}_E$), which drives unbounded growth of the nodal length until one of the two anabranches eventually closes (blue box). In the latter regime, prolonged morphological inactivity of one anabranch eventually results in its closure under formative flow conditions. In the third regime, the system converges to an equilibrium state in which the anabranching structure is preserved ($\mathcal{F}_S = \mathcal{F}_E$). In this state, both branches remain active and convey sediment under formative conditions, despite an unbalanced flow partition.

Although the purpose of this section is to illustrate the fundamental behavior of the model rather than to quantitatively predict the evolution and ultimate fate of the system, a qualitative comparison with the observations reported in Figure 2a can still be made. In particular, for $\Upsilon = 1$ the model reproduces successfully the trend of nodal-length adjustment and the tendency of the system to keep open both branches while approaching a length between roughly 5,000 m and 6,000 m. By contrast, simulations characterized by values of Υ far from unity predict evolutionary trajectories that are inconsistent with satellite-derived observations, suggesting that such

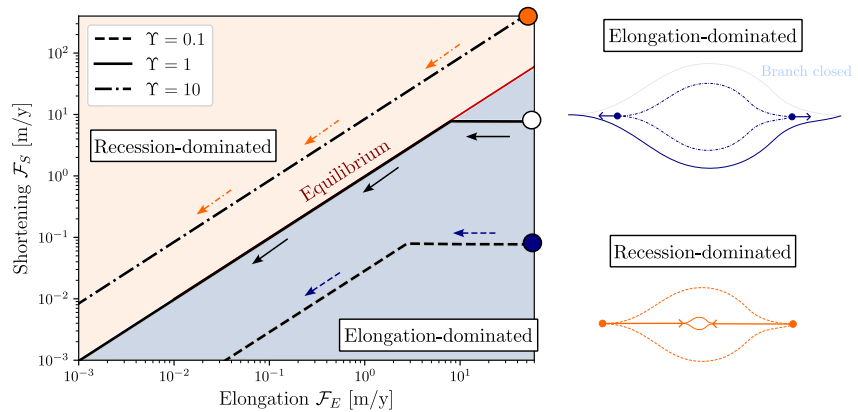


Figure 9. Plot in phase space $F_E - F_S$ showing the sensitivity to the parameter Υ . Circles denote starting point for different values of Υ . The arrows indicate the temporal trajectories. Input parameters and initial conditions coincide with that of Figure 8.

parameter ranges are unlikely to represent the system. One might surmise that a more careful calibration of Υ , combined with different choices of parameters α_Δ and α_S , could likely improve the agreement between modeled and observed nodal lengths. Nevertheless, the analysis indicates that the characteristic vegetation and erosion rates should be of comparable magnitude.

Furthermore, it should be observed that the three runs shown in Figures 8 and 9 share a common feature, in that the final solution is independent of the initial condition. In each run, the nodal length increases or decreases monotonically until an equilibrium or disequilibrium state, which coincides with the disappearance of the anabranching unit, is eventually reached. The only influence of the initial condition reverberates on the relaxation timescale. This is expected from the form of the adaptation Equation 22. The trend displayed by the nodal length is largely determined by the variation of ΔQ_{sf} , provided the value of Υ is not too large. As time progresses, the gradual reduction or increase in ΔQ_{sf} depends on whether the system moves closer to, or farther from, the critical conditions and consequently on whether the nodal length shortens or lengthens. Since the sinuosities are fixed and planform asymmetry is absent, the system lacks any additional mechanism capable of reversing the trend of variation of flow division, for instance by enabling a switch in the most-carrying anabranch.

4.2. General Trends

The results presented so far are limited to a single river system and to prescribed flow conditions and parameter values. To generalize these findings, the dimensionless form of the adaptation Equation 22 is examined for $t \rightarrow \infty$. As detailed in Appendix A, an inspection of this equation indicates that the asymptotic solution of the nodal length is primarily controlled by two parameters: (a) the bifurcation number $\mathcal{R}$, which depends on the reach-averaged characteristics and affects both the distance from critical conditions ϵ and the shoaling term δ ; and (b) the parameter Γ , which encapsulates the combined effects of the erosion-to-encroachment rate ratio Υ , the fraction of time associated with base flow with respect to peak flow, and the ratio between the peak and formative Shields numbers. The latter parameter is defined as:

$$\Gamma = \frac{\Upsilon \Theta}{r_p}, \quad (26)$$

where:

$$r_p = \frac{p}{1-p}, \quad \Theta = \frac{\theta_{mp}}{\theta_{mf}}. \quad (27)$$

By definition, $\Gamma \rightarrow \infty$ when $\zeta_{ers}^* \gg \zeta_{veg}^*$ or $p \rightarrow 1$, while $\Gamma \rightarrow 0$ when $\zeta_{ers}^* \ll \zeta_{veg}^*$ or when the duration of base-flow stages approaches that of flood stages. Since Θ ranges approximately between 1 and 2, typical values of the

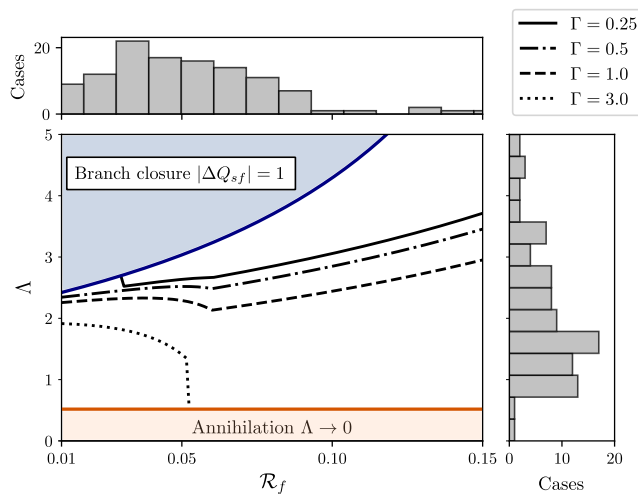


Figure 10. Asymptotic value of Λ plotted in terms of the formative bifurcation number R_f for different values of Γ , with $\theta_{mb} = 0.5$, $\alpha_s = 0.5$, and $\alpha_\Delta = 0.95$. The red and teal boxes indicate the regions of the parameter space where the system is predicted to become single-threaded due to the closure of one anabranch (red box) or excessive recession (teal box). The histograms represent the frequency distribution of the sand-bed data set gathered by Ragno et al. (2022).

5. Discussion

The analysis presented in this paper indicates that the preservation of the anabranching structure requires the system to adjust the nodal length so that elongation during base flow balances recession during peak flow. An equilibrium nodal length is eventually reached, provided that neither the vegetation encroachment rate nor the erosion rate dominates the other. This equilibrium state is generally characterized by unbalanced flow partitioning under formative stages, corresponding to a super-resonant regime, in spite of the fact that under base-flow conditions only one of the anabranches may actively convey sediment.

To assess whether theoretical predictions can reproduce, at least qualitatively, the range of nodal lengths observed in rivers, the model results are compared with the data set previously compiled by Ragno et al. (2022). This is shown in Figure 10. The comparison focuses on the sand-bed subset, for which R_f was computed using measured values at bankfull flow. The predicted range of Λ -values is consistent with remotely-sensed data, which include 115 cases. As shown in the figure, the observed nodal lengths are generally of the order $O(1)$, rarely exceeding values of 3. Overall, the theoretical range of the nodal length under which the anabranching structure should be preserved is fairly close to field estimates. It is seen that, since the Chézy coefficient is typically $C_{mf} \approx 10$ in both sand-bed and gravel-bed rivers (Parker et al., 2007; Wilkerson & Parker, 2011), the corresponding values of the (dimensional) nodal length Λ^* are of the order of one hundred reach-averaged water depths.

To our knowledge, this work represents the first attempt to describe the length evolution of anabranching units through a mathematical model. Overall, the proposed model provides a possible explanation of why the observed lengths of many natural anabranching units fall within a specific range of values and why some of them fail to maintain a multi-thread morphology in the long term.

It seems now appropriate to discuss the main limitations of the analysis. A primary limitation is that temporal variations in anabranch sinuosity are not considered. This assumption is associated with the absence of length asymmetries between the anabranches. Although the field examples presented in Section 2 are fairly consistent with this idealization, it is quite common to observe sinuosity variations in anabranching systems, especially when they originate from chute cutoffs. Changes in anabranch sinuosity are closely linked to flow partitioning (e.g., Caballero et al., 2025; Lin et al., 2025; Van Dijk et al., 2012), as the branch receiving a larger fraction of the discharge is expected to adjust its length faster, thereby reducing its slope. This process may eventually lead to a flow-switching event if the slope of the most-carrying branch becomes much lower than that of the other branch. In this case, where planform changes are driven by a combination of nodal length adjustment and sinuosity

parameter Γ fall in the range $(10^{-1}, 1)$, as shown in Figure A1. For instance, with $\Theta = 1.5$, assuming that the base flow occurs 80% of the time ($p = 0.8$) and that the vegetation encroachment rate is equal to the erosion rate ($\Upsilon = 1$), one obtains $\Gamma = 0.38$.

Figure 10 shows the asymptotic solution for the dimensionless length Λ as a function of the formative bifurcation number R_f , for different values of the parameter Γ . The plot is partitioned into three regions corresponding to the three evolutionary trajectories predicted by the model. Although the equilibrium nodal length increases with the bifurcation number, it remains confined within a narrow range between 1 and 4, unless the values of gamma are relatively large. The trend of the curves changes when the emersion-threshold is exceeded, and thus the shoaling effect becomes significant, which occurs around $R_f \approx 0.06$. The disappearance of the island due to excessive recession is governed mainly by Γ , with larger values increasing the likelihood of rapid erosion and eventual disappearance of the unit. This is the case for the dotted curve that coincides with $\Gamma = 3$. In contrast, the closure of an anabranch resulting from excessive elongation is governed by the threshold value of the nodal length. This threshold increases with the bifurcation number R_f (Figure 6b). The formative sediment discharge asymmetry reaches unity when $\epsilon \approx 1$, and thus $\Lambda \approx 2\Lambda_{cr}$ (Figure 6a).

variations, the sediment flux asymmetry index ΔQ_s would no longer be controlled solely by the internal parameter ϵ , but would instead reflect the interaction between internal and external dynamics, the latter being governed by the degree of length asymmetry. It should be noted that, according to Ragno and Tubino (2025), most of the observed anabranching units exhibit a small degree of length asymmetry. This suggests that anabranching systems may be unable to support a multi-thread configuration over extended periods if the length asymmetry becomes too pronounced.

Second, the spatial and temporal variations of the channel width are neglected and the widths of the anabranches are assumed to be equal. It is evident that the latter condition is never exactly met in nature. Since differences in anabranches widths depend on the discharge distribution, which in turn is influenced by the width differences itself, this feedback mechanism introduces an additional layer of complexity in discharge computations. Hence, accounting for the effect of width asymmetries represents a key priority for future research. Preliminary analyses by Pirlot (2024) revealed a much broader spectrum of variability in the width differences between anabranches compared to their length differences. This greater variability may indicate that anabranching systems better accommodate disparities in cross-sectional geometry than in anabranch length to mitigate imbalances in discharge distribution.

Regarding the variation in width over time, the timescale of width adjustment may be comparable to that of planform changes, as is typically observed in meandering rivers (e.g., Monegaglia & Tubino, 2019). Preliminary analyses indicate that the rates of length variations appear to be of similar magnitude to those measured in single-threaded meanders, calling into question the validity of the fixed-width assumption. In this regard, for multi-thread systems it would be necessary to further investigate how the evolution of the bifurcates affects the morphological properties of the entire unit. Several studies suggested that anabranch dynamics may influence, among other factors, channel sinuosity at the reach scale and bed slope, and consequently sediment transport capacity, both upstream and downstream of the anabranching unit (e.g., Bridge et al., 1986; M. C. Grenfell et al., 2014; Howard, 1996; Schwenk & Foufoula-Georgiou, 2016).

The third limitation concerns the fact that, although physically consistent, the model adopts a simplified representation of the first-order processes governing elongation and recession, channel-island interactions, and morphology-vegetation feedbacks. In particular, the coefficients ζ_{ers}^* and ζ_{veg}^* roughly synthesize the underlying physics. Their estimation requires calibration against field data. These empirical coefficients, therefore, would benefit from a more physically-based description that necessarily should take into account for in-stream-island interactions. One of the main interactions to be considered is arguably the feedback between vegetation dynamics and island topography. In this respect, a direct quantitative comparison between model predictions and field observations would require not only a comprehensive data set documenting the temporal evolution of developing anabranching units, which is currently lacking, but also information on the initial configuration of the system immediately after its formation.

More generally, the simplified formulation adopted for the elongation and recession functions, $\mathcal{F}_E$ and $\mathcal{F}_S$, is not derived directly from physical principles. Rather, these functions, and the resulting adaptation equation, should be regarded as minimal mathematical representations of the dominant processes governing the studied dynamics, in the spirit of the so-called reduced-complexity (or explanatory) approach (e.g., Murray, 2003). At the same time, however, the model does not rely exclusively on empirical or conceptual parametrizations. Some of its key components, notably the sediment partitioning index, are derived from conservation laws. The model should therefore be viewed as a hybrid formulation, combining a reduced-complexity representation of reach-scale evolution with a reductionist description of the local processes governing flux partitioning within the anabranches.

Finally, in the analysis an approximate description of the hydrological regime is adopted. A straightforward way to overcome this limitation would be to implement the full flow duration curve following, for example, the approach adopted by Naito and Parker (2019) and Carlin et al. (2021) to investigate the bankfull geometry of meanders and the long-term response of river bars, respectively. Alternatively, to account for the flow variability in environmental systems characterized by alternating growth and decay phases depending on flow conditions, as is the case of anabranching units, one could adopt the strategy proposed by Camporeale and Ridolfi (2006), who employed a stochastic model of riparian vegetation dynamics, in which river discharge was treated as a random variable. In this work, vegetation was assumed to decay or grow depending on whether the flow discharge

exceeded a threshold value. Under certain assumptions, an analytical solution corresponding to the (steady) probability density function of the overall vegetation mass was obtained. This methodology can be translated into our theoretical model with relatively little effort, replacing the vegetation mass with lengths. A preliminary version of the stochastic model was recently presented in Ragno et al. (2025).

6. Conclusions

A theoretical framework is introduced to study the long-term evolution of a solitary anabranching unit. The analysis is restricted to a symmetric configuration, in which planform evolution is described in terms of nodal length adjustments between the upstream bifurcation and the downstream confluence. Based on the observation of several anabranching units derived from remotely sensed imagery, an adaptation equation for the nodal length evolution is formulated under the assumption that the two anabranches have identical and time-invariant sinuosity. The adaptation equation is based on a balance between elongation during base-flow conditions and the recession driven by flood-induced erosion. The evolution emerges from a competition between these processes, the relative importance of which is controlled by a limited set of dimensionless parameters representing reach-averaged hydraulic geometry, bank and floodplain properties, and the catchment-scale hydrology. Depending on this balance, anabranching units may persist in a stable multi-thread configuration, collapse through island erosion, or transition to a single-thread morphology following excessive elongation, which ultimately results in the closure of one anabranch.

A key result of the analysis is that the long-term persistence of the anabranching unit is predicted to correspond to values of the nodal length confined to a relatively narrow range, on the order of a few hundred reach-averaged flow depths. This theoretical range is consistent with the values inferred from the observational data. Finally, the analysis provides the basic framework for future extensions of the model that should relax the assumptions of symmetry and fixed sinuosity, as well as the assumption of constant widths.

Appendix A: Derivation of the Asymptotic Length

In this Appendix, we derive an approximate expression of the asymptotic length Λ to elucidate the main governing parameters.

Let us consider the dimensionless form of the adaptation Equation 22 and let $t \rightarrow \infty$. This operation reduces the governing equation to a balance between elongation and recession, thus:

$$p \mathcal{E}_1 \cdot \mathcal{E}_2(\Lambda) = (1 - p) \Upsilon S(\Lambda) \frac{\theta_{mp}}{\theta_{mf}}. \quad (\text{A1})$$

It is convenient to define the lumped parameter $\Gamma = \Upsilon \Theta / r_p$, where r_p and Θ are defined by (27). As shown in Figure A1a, for a given value of Υ , Γ spans two orders of magnitude as p is varied.

Equation A1 then becomes:

$$\mathcal{E}_1 \cdot \mathcal{E}_2(\Lambda) = \Gamma S(\Lambda). \quad (\text{A2})$$

Expanding each function yields:

$$(1 + \delta)(1 - \alpha_\Delta |\Delta Q_s|) = \Gamma \left[(1 - \alpha_s) \frac{F_{mf}^2}{(1 - F_{mf}^2)} \bar{\sigma} \Lambda + \alpha_s \right]. \quad (\text{A3})$$

At first sight, Equation A3 depends on several parameters: $\mathcal{R}$, controlling δ and ΔQ_s , Γ , F_{mf} , $\bar{\sigma}$, α_Δ , and α_s . However, it will be shown that $\mathcal{R}$ and Γ play the major role, while the latter parameters are only of secondary importance.

The solution of Equation A3 requires a numerical approach because ΔQ_s is an implicit function of Λ . However, the procedure can be simplified by noting, from Figure 6a, that the sediment flux asymmetry can be expressed as a power law in ϵ as follows:

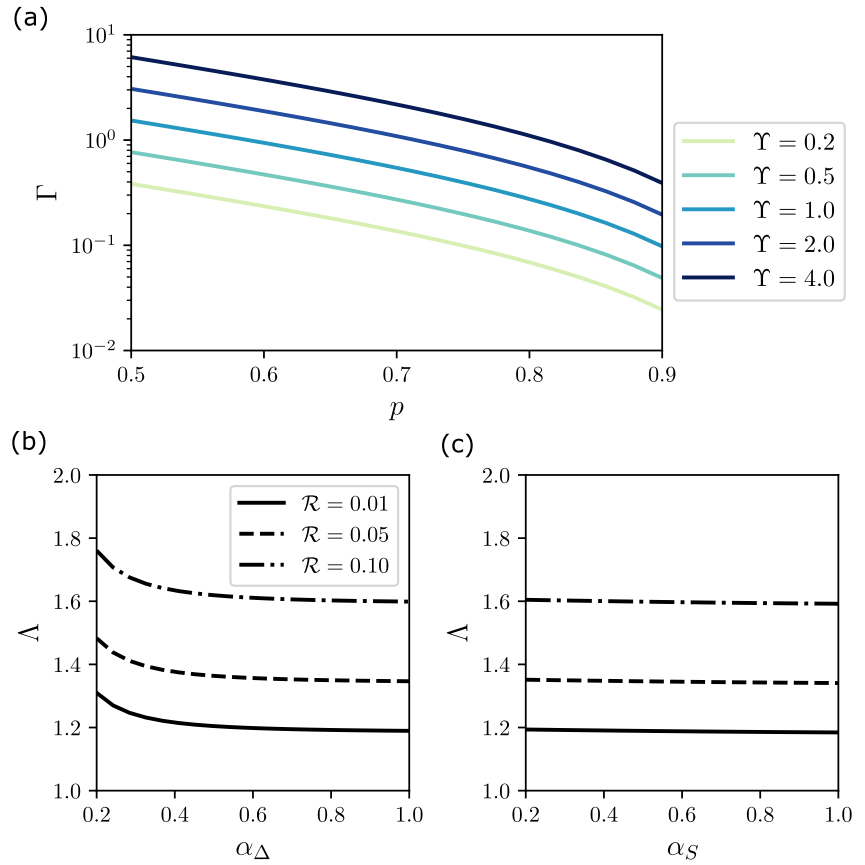


Figure A1. In the upper panel (a), the dependence of Γ on the base-flow fraction p is plotted for different values of Υ and $\Theta = 1.5$. In the lower panels, the nodal length Λ , computed from Equation A7, is plotted as a function of (b) the coefficient α_Δ , and (c) the coefficient α_S , for different values of the bifurcation number $\mathcal{R}$. Fixed parameters: $\Gamma = 0.38$, $\bar{\sigma} = 1$ and $F_{mf} = 0.1$.

$$\Delta Q_s = \alpha_Q \epsilon^{n_Q}, \quad (\text{A4})$$

where the coefficient α_Q is, in general, a function of the Chézy coefficient and of the Shields number. A formal derivation of Equation A4 by means of a perturbation expansion was provided by Ragno (2023), who showed that $n_Q = 1/2$, at least for $\epsilon \rightarrow 0$.

The power-law relation (Equation A4) is now substituted into Equation A3. Considering sand-bed streams, a comparison between Equation A4 and the nonlinear solution of the equilibrium model of Ragno et al. (2021) shows that $\alpha_Q \simeq 1$. Moreover, for lowland rivers $1 - F_{mf}^2 \approx 1$, with $F_{mf}^2 \sim O(10^{-2})$. Equation A3 then becomes:

$$(1 + \delta)(1 - \alpha_\Delta \epsilon^{1/2}) = \Gamma[(1 - \alpha_S)F_{mf}^2 \bar{\sigma} \Lambda + \alpha_S]. \quad (\text{A5})$$

Using Equation 15 and the expression for the critical length Equation 16, the balance reduces to:

$$\alpha_\Delta \sqrt{\frac{4\Lambda(5 - 12\mathcal{R}) - 23 - 12\mathcal{R}}{23 + 12\mathcal{R}}} = 1 - \frac{\Gamma}{(1 + \delta)} [(1 - \alpha_S)F_{mf}^2 \bar{\sigma} \Lambda + \alpha_S]. \quad (\text{A6})$$

The parameter $\bar{\sigma}$, which can be assumed to lie between 1 and 2, affects only slightly the value of Λ because it multiplies a small quantity, F_{mf}^2 . We then assume for simplicity $\bar{\sigma} = 1$.

Squaring both sides of Equation A6, neglecting terms proportional to F_{mf}^4 , eventually yields the following analytical expression for Λ :

$$\Lambda = \frac{\Omega_2(\mathcal{R}; \Gamma, \alpha_\Delta, \alpha_S)}{\Omega_1(\mathcal{R}; \Gamma, \alpha_\Delta, \alpha_S)}, \quad (\text{A7})$$

where:

$$\Omega_1(\mathcal{R}; \Gamma, \alpha_\Delta, \alpha_S) = 4\psi(5 - 12\mathcal{R}) + \frac{2\Gamma F_{mf}^2}{(1 + \delta)}(1 - \alpha_S) \left[1 - \frac{\Gamma \alpha_S}{(1 + \delta)} \right], \quad (\text{A8a})$$

$$\Omega_2(\mathcal{R}; \Gamma, \alpha_\Delta, \alpha_S) = \psi(12\mathcal{R} + 23) + \left[1 - \left(\frac{\Gamma \alpha_S}{(1 + \delta)} \right) \right]^2, \quad (\text{A8b})$$

$$\psi = \frac{\alpha_\Delta^2}{(23 + 12\mathcal{R})}. \quad (\text{A8c})$$

An analysis of the order of magnitude of the different terms indicates that the second term of Ω_1 is two orders of magnitude smaller than the first term, so that:

$$\Omega_1 \approx 4\psi(5 - 12\mathcal{R}) = f(\mathcal{R}; \alpha_\Delta). \quad (\text{A9})$$

Figure A1b and A1c illustrate the dependence of Λ , computed from Equation A7, on the coefficients α_Δ and α_S , for different values of the bifurcation number $\mathcal{R}$ and a given Γ . It appears that Λ exhibits a slight increase at low values of the partition coefficient α_Δ and remains relatively unchanged as α_Δ increases, while it is largely insensitive to the backwater coefficient α_S . A common feature of both plots is the strong sensitivity of Λ to the bifurcation number, with Λ increasing as $\mathcal{R}$ increases. Taken together, the results shown in Figures A1 and 10 indicate that the asymptotic value of Λ as $t \rightarrow \infty$ is primarily controlled by $\mathcal{R}$ and Γ .

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

The Python code used to solve the ordinary differential Equation 22 is freely available at Ragno (2026, <https://zenodo.org/records/20084530>).

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