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1 **A new efficient method to solve the stream power law**
2 **model taking into account sediment deposition**

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8 **Key Points:**

- 9 • We present an efficient ($O(N)$ and implicit) method to solve a river erosion model
10 taking into account sediment deposition.
11 • We show how the foreland stratigraphy is controlled by the efficiency of river ero-
12 sion and the efficiency of sediment transport by rivers.
13 • We observe autogenic aggradation and incision cycles in the foreland once the sys-
14 tem reaches a dynamic steady state.

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Abstract

The stream power law model has been widely used to represent erosion by rivers, but does not take into account the role played by sediment in modulating erosion and deposition rates. *Davy and Lague* (2009) provide an approach to address this issue, but it is computationally demanding because the local balance between erosion and deposition depends on sediment flux resulting from net upstream erosion. Here, we propose an efficient (i.e., $O(N)$ and implicit) method to solve their equation. This means that, unlike other methods used to study the complete dynamics of fluvial systems (including the transition from detachment-limited to transport-limited behavior, for example), our method is unconditionally stable even when large time steps are used. We demonstrate its applicability by performing a range of simulations based on a simple setup composed of an uplifting region adjacent to a stable foreland basin. As uplift and erosion progress, the mean elevations of the uplifting relief and the foreland increase, together with the average slope in the foreland. Sediments aggrade in the foreland and prograde to reach the base level where sediments are allowed to leave the system. We show how the topography of the uplifting relief and the stratigraphy of the foreland basin are controlled by the efficiency of river erosion and the efficiency of sediment transport by rivers. We observe the formation of a steady-state geometry in the uplifting region, and a dynamic steady state (i.e., autocyclic aggradation and incision) in the foreland, with aggradation and incision thicknesses up to tens of metres.

35

1 Introduction

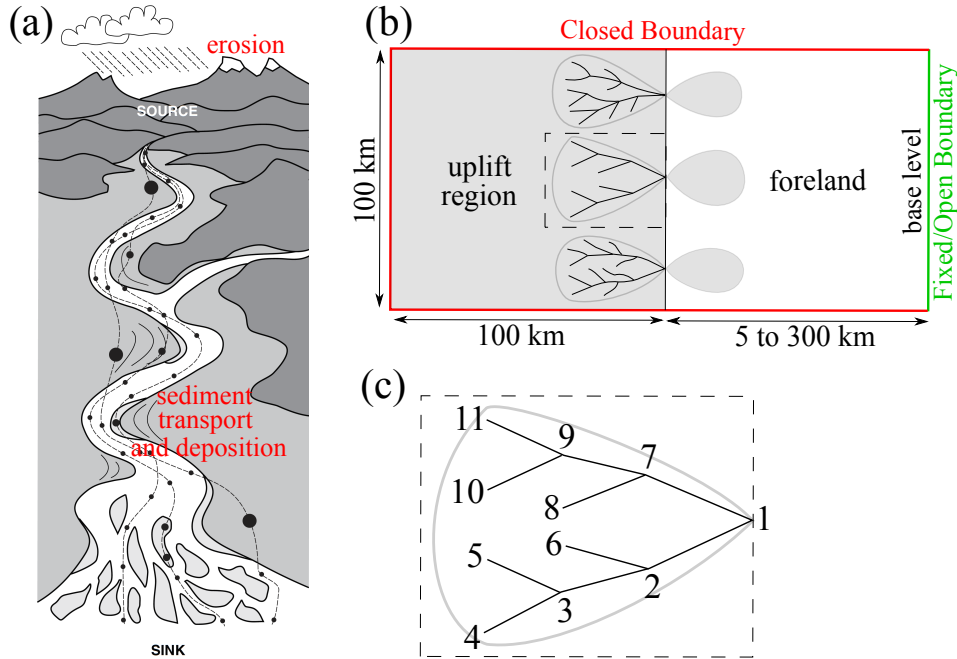


Figure 1. (a) The concept of source to sink with sediment transport and deposition, modified from *Allen and Heller* (2011). (b) Setup for our simulation with uplifting region and foreland basin. (c) Illustration of a simple catchment with normal FastScape stack order (*Braun and Willett*, 2013). In (b), the red lines indicate the closed boundary where sediment flux cannot leave the system, whereas base level (green line) is fixed as an open boundary.

Quantifying the dynamics of river erosion, sediment transport and deposition (Figure 1a) is a fundamental problem in geomorphology that has great relevance for our understanding of landscape evolution in tectonically active areas. Many parameterizations of these processes have been proposed and implemented in numerical landscape evolution models (*Braun and Sambridge, 1997; Chase, 1992; Crave and Davy, 2001; Kooi and Beaumont, 1994; Tucker and Slingerland, 1994*).

The Stream Power Law (SPL) model has been widely used to represent erosion by rivers (*Howard and Kerby, 1983; Whipple and Tucker, 1999*). In its simplest form, it assumes that erosion rate is proportional to the shear stress exerted by the river on its bed which, in turn, is proportional to net precipitation rate, p , drainage area, A , and local slope, S , according to:

$$\frac{\partial h}{\partial t} = U - K_f p^m A^n S^n, \quad (1)$$

where h is topographic elevation, t is time, U is uplift rate, K_f is the fluvial erosion coefficient, and m and n are the SPL exponents. An important assumption of the SPL model is that sediments are efficiently transported by rivers and not deposited in the simulated domain. The SPL model has been shown to describe a number of fluvial landscapes and processes. It is for example commonly used to infer uplift pattern from river profiles or to model topographic evolution at the scale of a catchment (*Braun and Willett, 2013; Campforts et al., 2017; Lavé and Avouac, 2001*). Yet, it is well known that this model might be oversimplified as it does not consider several important processes acting in river channels (*Lague, 2014*). In particular, it is necessary to take into account the role played by sediment in modulating erosion rate and/or deposition (*Whipple and Tucker, 2002*), such as a dependence on bedload transport (*Davy and Lague, 2009; Kooi and Beaumont, 1994*) or a bed-cover effect (*Cowie et al., 2008; Johnson et al., 2009; Sklar and Dietrich, 2001*). In fact, transported sediments provide the tools for abrasion and fracturing of rock but also, if overly abundant, they can protect the bedrock from erosion (*Sklar and Dietrich, 1998*).

Several parameterizations have been proposed to adapt the SPL model to incorporate the effects of transported sediments, including the erosion-deposition formulation proposed by *Davy and Lague (2009)*, which differs from models based on the divergence of sediment flux (e.g., *Paola and Voller, 2005*) in that it conserves mass on the bed and in the water column to treat simultaneous erosion and deposition of a single substrate (*Shobe et al., 2017*). Based on previous erosion-deposition models (e.g., *Beaumont et al., 1992; Kooi and Beaumont, 1994*), *Davy and Lague's (2009)* formulation has a limited number of parameters, while attempting to relate these parameters to physical processes and quantities (e.g., saltation length). In addition, their erosion-deposition framework allows the exploration of both detachment-limited and transport-limited models with simple parameter changes, and displays a smooth transition between the two types of model behavior.

Davy and Lague's (2009) formulation has been used or adapted to obtain simple models (e.g., *Carretier et al., 2016; Ganti et al., 2014; Langston and Tucker, 2018; Mouchéné et al., 2017; Shobe et al., 2017*), which assume that the net rate of topographic change is the sum of the erosion rate (controlled by the SPL model) and of the deposition rate, which is proportional to local suspended sediment flux and to a dimensionless deposition coefficient, and inversely proportional to drainage area, a proxy for water discharge. This parameterization is also receiving growing acceptance due to its ability to reproduce many depositional features of fluvial systems (*Carretier et al., 2016; Mouchéné et al., 2017; Shobe et al., 2017*). However, because the local balance between erosion and deposition depends on sediment flux resulting from net upstream erosion, this parameterization is computationally demanding.

Braun and Willett (2013) have proposed an efficient algorithm for solving the SPL model, which is ideally suited for a large number of model simulations as required for

82 inverting observational constraints in a Bayesian approach. Here we present an equally
 83 efficient (i.e., $O(N)$ and implicit) method to solve the equation proposed by *Davy and*
 84 *Lague* (2009) that takes into account sediment deposition. $O(N)$ means that the com-
 85 putational time increases linearly with the number of points used to discretize the land-
 86 scape. An implicit scheme guarantees unconditionally stable time integration of the land-
 87 scape evolution equation, which means that large time steps can be used without affect-
 88 ing numerical stability. This is potentially an important step as it allows one to use sed-
 89 imentological observations, such as the stratigraphy of foreland basins or the position
 90 and thickness of river terraces, to further constrain landscape evolution models.

91 Nonetheless, the novelty of our study is not limited to the description of the com-
 92 putational efficiency. Based on this algorithm, our model can simulate erosion and de-
 93 position in fluvial landscapes, at large spatial (up to thousands of kilometers) and tem-
 94 poral (up to tens of millions of years) scales. Because these two processes are considered
 95 in one single equation, deposition can occur anywhere in the domain (i.e., not only in
 96 depressed areas but also along channels or in stable continental areas). We therefore use
 97 our new algorithm to explore the impact of coupling erosion and deposition in a fluvial
 98 landscape, and show that this erosion-deposition relationship, which is often ignored, has
 99 a strong impact on relief in the uplifted domain, the concavity of the channels and their
 100 steepness index. We also demonstrate that this relatively simple model leads to the cyclic
 101 formation and destruction of river banks as the stream continuously migrates and some-
 102 times erodes into the sediment it has previously deposited. Such autogenic aggradation
 103 and incision cycles are currently difficult to simulate in landscape evolution models. The
 104 simulations using the new model can thus improve our understanding of the links be-
 105 tween external forcings, internal processes, and depositional features.

106 In the next section, we first present our implementation of *Davy and Lague's* (2009)
 107 model and the $O(N)$ and implicit numerical scheme. Model implications of our new for-
 108 mulation are shown in Section 3. We then explore the model behavior in Section 4 by
 109 performing a range of simulations based on a simple setup composed of an uplifting re-
 110 gion adjacent to a stable continental area on which a foreland basin develops. In Sec-
 111 tion 5, using our model, we observe the formation of autocyclic aggradation and inci-
 112 sion in the foreland once the system reaches a dynamic steady state.

113 2 Model implementation

114 2.1 SPL model taking into account sediment deposition

The effect of upstream sediment flux was first incorporated into the SPL model by
Kooi and Beaumont (1994), by assuming that the rate of topographic change results from
 the imbalance between a sediment “carrying capacity”, q_f^{eqb} , and the upstream sediment
 yield, q_s , according to:

$$\frac{\partial h}{\partial t} = U - \frac{1}{L_f} (q_f^{eqb} - q_s) \quad \text{with} \quad q_f^{eqb} = K q_w S, \quad (2)$$

where L_f is a transport length, K is a dimensionless erosion coefficient, $q_w = Q_w/W$
 is water discharge per unit width (Q_w , water discharge, and W , river width), and q_s is
 the sediment flux per unit width obtained by integrating the net upstream erosion rate:

$$q_s = \frac{1}{W} \int_A \left(U - \frac{\partial h}{\partial t} \right) dA. \quad (3)$$

115 This sediment flux therefore accounts for the whole solid load (bed, suspended, and wash
 116 load). The transport length L_f in (2) can be regarded as the length scale over which the
 117 imbalance between the upstream sediment yield and the river carrying capacity is resolved
 118 either by deposition (in cases where the river is over-capacity) or by erosion (in cases where
 119 the river is under-capacity). Physically, L_f represents the average transport distance of

120 sediment grains within the flow from the location where they are eroded to the location
 121 where they are deposited (*Beaumont et al.*, 1992). L_f thus characterizes the proportion
 122 of incoming sediment flux which is deposited (the larger the value of L_f , the lower the
 123 rate of deposition).

Combining equations (2) and (3) leads to:

$$\frac{\partial h}{\partial t} = U - \frac{1}{L_f} \left[\frac{K Q_w S}{W} - \frac{1}{W} \int_A \left(U - \frac{\partial h}{\partial t} \right) dA \right]. \quad (4)$$

To derive the SPL model, two major assumptions are commonly made. Firstly, it is assumed that river width W varies as the square root of the water discharge (*Lacey*, 1930; *Leopold and Miller*, 1956):

$$W = c Q_w^{0.5}, \quad (5)$$

where c is an empirical constant, typically of the order of $(0.1-1) \times 10^{-2}$ (*Montgomery and Gran*, 2001, and references therein). Secondly, water discharge can be expressed as the product of net precipitation rate p and drainage area A :

$$Q_w = pA = p_0 \tilde{p}A, \quad (6)$$

where p_0 is mean net precipitation rate, and $\tilde{p} = p/p_0$ represents any spatial or temporal variation in precipitation p relative to the mean precipitation p_0 . Combining equations (5) and (6) with equation (4) leads to

$$\frac{\partial h}{\partial t} = U - \frac{1}{L_f} \left[\frac{K p_0^{0.5}}{c} \tilde{p}^{0.5} A^{0.5} S - \frac{1}{c p_0^{0.5} \tilde{p}^{0.5} A^{0.5}} \int_A \left(U - \frac{\partial h}{\partial t} \right) dA \right]. \quad (7)$$

The contribution of *Davy and Lague* (2009) can be regarded as an improvement on *Kooi and Beaumont's* (1994) method in an attempt to relate the transport length L_f to physical parameters (e.g., water discharge and settling velocity of grains within the flow). According to *Davy and Lague's* (2009) formulation, the rate of change of topographic elevation is given by:

$$\frac{\partial h}{\partial t} = U - K' q_w^{m'} S^{n'} + \frac{q_s}{L_f} = U - K' q_w^{m'} S^{n'} + \frac{d^* v_s}{q_w} q_s, \quad (8)$$

124 where K' is an erosion efficiency coefficient, m' and n' are two exponents, v_s is the net
 125 settling velocity of sediment grains, and d^* is the ratio between the sediment concentra-
 126 tion near the riverbed interface and the average concentration over the water column.
 127 The value of d^* (≥ 1) varies as a function of the Rouse number which defines the rela-
 128 tive contribution of bed, suspended, and wash loads (*Davy and Lague's* (2009) Figure
 129 4). *Davy and Lague* (2009) discuss how d^* can be calculated for suspended load and bed
 130 load rivers. For small rivers (or large particles), most of the entrainment mechanisms lie
 131 in the bed load, d^* is much larger than 1 and the transport length L_f is small. Conversely,
 132 for large rivers (or small particles) the Rouse number is small, d^* is close to 1 and L_f
 133 is large. *Davy and Lague* (2009) present such a model based on the relative contribu-
 134 tions of (i) erosion from the bed into the water column (suspended load) and (ii) depo-
 135 sition from the water column onto the bed. Thus, the transport length L_f takes into ac-
 136 count the deposition of the bed load and the suspended load. With this approach, the
 137 deposition term is proportional to the ratio between the sediment flux q_s and the wa-
 138 ter flux q_w . If $q_s \gg q_w$, deposition is high. On the contrary, if $q_s \ll q_w$, the depo-
 139 sition term tends toward 0. Note that both terms are proportional to the drainage area
 140 of the catchment.

The equivalent of *Kooi and Beaumont's* (1994) transport length L_f in the *Davy and Lague's* (2009) approach is therefore:

$$L_f = \frac{q_w}{d^* v_s} = \frac{Q_w}{W d^* v_s} = \frac{pA}{W d^* v_s} = \frac{p_0 \tilde{p}A}{W d^* v_s} = \frac{\tilde{p}A}{WG}, \quad \text{with } G = \frac{d^* v_s}{p_0}, \quad (9)$$

141 where G is a dimensionless deposition coefficient, which is a function of the sediment con-
 142 centration ratio in transport, the settling velocity of sediment, and the mean precipita-
 143 tion rate. G/\tilde{p} is identical to Θ as defined in *Davy and Lague* (2009). They showed that,
 144 for typical values of the various parameters ($p = 10^{-7}$ m/s, $v_s \in [10^{-6} - 10^{-1}]$ m/s),
 145 G/\tilde{p} is of order 1 or greater, in good agreement with estimates from natural sedimentary
 146 systems (*Guerit et al.* (2018), pers. comm.).

Equations (3), (5), (6) and (9) can be combined with equation (8) to yield:

$$\begin{aligned} \frac{\partial h}{\partial t} &= U - K' \left(\frac{Q_w}{W} \right)^{m'} S^{n'} + \frac{1}{WL_f} \int_A \left(U - \frac{\partial h}{\partial t} \right) dA \\ &= U - \frac{K' p_0^{0.5m'}}{c^{m'}} \tilde{p}^{0.5m'} A^{0.5m'} S^{n'} + \frac{G}{\tilde{p}A} \int_A \left(U - \frac{\partial h}{\partial t} \right) dA. \end{aligned} \quad (10)$$

147 In the parametric study of *Davy and Lague* (2009), $m' = n' = 1$ and their erosion term
 148 $\frac{K' p_0^{0.5}}{c} \tilde{p}^{0.5} A^{0.5} S$ is thus similar to that of *Kooi and Beaumont* (1994). The main differ-
 149 ences between their models are: (i) the depositional term is inversely proportional to either
 150 the drainage area (*Davy and Lague*, 2009) or the square root of drainage area (*Kooi*
 151 *and Beaumont*, 1994); and (ii) L_f is applied only to q_s (*Davy and Lague*, 2009), or is ap-
 152 plied to both q_f^{eqb} and q_s (*Kooi and Beaumont*, 1994).

Replacing $K' p_0^{0.5}/c^{m'}$, $0.5m'$ and n' in (10) by K_f , m and n , respectively, we can
 make a more direct connection to the SPL model and write that the rate of topographic
 change $\partial h/\partial t$ in response to tectonic uplift, river erosion and sediment deposition is given
 by:

$$\frac{\partial h}{\partial t} = U - K_f \tilde{p}^m A^m S^n + \frac{G}{\tilde{p}A} \int_A \left(U - \frac{\partial h}{\partial t} \right) dA. \quad (11)$$

153 The modified SPL formulation has only one additional parameter (i.e., the dimension-
 154 less deposition coefficient G) compared to the classic SPL model. The dimensionless con-
 155 stant G multiplying the deposition rate in (11) depends on an assumed mean precipi-
 156 tation rate; any spatial or temporal variation in precipitation rate is introduced through
 157 the variable \tilde{p} . At steady state, the catchment area A in the deposition term vanishes,
 158 and the deposition term is equal to GU/\tilde{p} . Note that this modified SPL formulation is
 159 constructed by considering fluxes and, therefore, it does not specifically consider the ef-
 160 fect of grain size. However, the mathematical definition of G in (9) makes it related to
 161 the size of the sediments in transport through the settling velocity v_s . In this work, we
 162 consider different values of $G = d^* v_s/p_0$ as a whole, rather than studying individually
 163 different values of d^* , v_s , and p_0 .

Fluvial erosion leads to the formation of hillslopes along river channels. Fluvial ero-
 sion and hillslope processes are interdependent, therefore hillslope processes need to be
 included in our model, which are commonly represented by a linear diffusion term (*Ah-*
nerst, 1967):

$$\frac{\partial h}{\partial t} = K_d \nabla^2 h, \quad (12)$$

164 where K_d is a hillslope sediment transport coefficient. In our model, the diffusion equa-
 165 tion (12) is calculated separately, after solving equation (11). Both equations are applied
 166 in every cell of the landscape.

167 Easily detachable materials such as unconsolidated sediments should be character-
 168 ized by a larger erosion coefficient K_f than bedrock (*Davy and Lague*, 2009; *Kooi and*
 169 *Beaumont*, 1994). Therefore, K_f depends on whether topographic elevation h is higher
 170 than basement elevation h_{base} or not. In areas that are in net erosion (i.e., $h \leq h_{base}$),
 171 we assume $K_f = K_{fb}$ (subscript b represents bedrock), whereas in areas covered by sed-
 172 iments (i.e., $h > h_{base}$), we assume $K = K_{fs}$ (subscript s represents sediments). In
 173 most of our simulations, we assume $K_{fb} = K_{fs}$ for the sake of simplicity, and we use
 174 $K_{fb} \neq K_{fs}$ for our sensitivity analysis in Section 4.2.2.

2.2 $O(N)$ and implicit algorithm

The most challenging part is to solve equation (11) in an efficient manner (i.e., in $O(N)$ operations) and using an implicit algorithm that allows for large time steps. For this we first discretize equation (11) using a backward Euler implicit finite difference scheme for each of the $n_x \times n_y$ nodes (n_x and n_y are the number of nodes to discretize the landscape in the x - and y -directions, respectively) as follows:

$$\frac{h_i^{t+\Delta t} - h_i^t}{\Delta t} = U - K_f \tilde{p}^m A_i^m \left(\frac{h_i^{t+\Delta t} - h_{rec(i)}^{t+\Delta t}}{\Delta l_i} \right)^n + \frac{G}{\tilde{p} \tilde{A}_i} \sum_{j=ups(i)} \left(U - \frac{h_j^{t+\Delta t} - h_j^t}{\Delta t} \right), \quad 1 \leq i \leq n_x \times n_y, \quad (13)$$

in which, h_i^t and $h_i^{t+\Delta t}$ are the elevations of the i -th node at time t and time $t+\Delta t$, respectively, $h_{rec(i)}^{t+\Delta t}$ is the elevation of the i -th node's receiver (the node in the steepest-descent drainage direction of the i -th node) at time $t+\Delta t$, Δl_i is the distance between the i -th node and its receiver, and $\sum_{j=ups(i)}$ represents the sum of the i -th node's upstream catchment nodes. \tilde{A}_i in equation (13) is a dimensionless catchment area defined as:

$$\tilde{A}_i = A_i / (\Delta x \Delta y) = N_i, \quad (14)$$

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where $\Delta x, \Delta y$ are the horizontal sizes of the cells, and N_i is simply the number of cells upstream of cell i . To compute the catchment areas A_i in $O(N)$ operations, we use the reverse stack order as defined in the FastScape algorithm (*Braun and Willett, 2013*).

To explain the remaining parts of our proposed numerical scheme, we assume that $n = 1$. The general case ($n \neq 1$) is dealt with later. When $n = 1$, equation (13) can be expressed as

$$\begin{aligned} -F_i h_{rec(i)}^{t+\Delta t} + (1 + F_i) h_i^{t+\Delta t} + \frac{G}{\tilde{p} \tilde{A}_i} \sum_{j=ups(i)} h_j^{t+\Delta t} &= b_i^t, \\ \text{with } F_i &= \frac{K_f \tilde{p}^m A_i^m \Delta t}{\Delta l_i}, \quad \text{and } b_i^t = h_i^t + U \Delta t + \frac{G}{\tilde{p} \tilde{A}_i} \sum_{j=ups(i)} (h_j^t + U \Delta t). \end{aligned} \quad (15)$$

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The term b_i^t on the right-hand side of the equation is known from the solution at time t , while elevations on the left-hand side are unknown at time $t + \Delta t$.

For the nodes at base level (open boundary, Figure 1b), we assume that the elevation is constant through time:

$$h_{base\ level}^{t+\Delta t} = h_{base\ level}^t. \quad (16)$$

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We also assume that sediment can leave the system from these base level nodes.

The above finite difference equations can be expressed in the following matrix form:

$$\underline{\underline{B}} \cdot \underline{\underline{h}}^{t+\Delta t} = \underline{\underline{b}}^t. \quad (17)$$

As shown in Appendix A, if we use the FastScape stack order in Figure 1c to solve equation (17), every row i of $\underline{\underline{B}}$ has a single non-zero element before the diagonal element that corresponds to the receiver of node i , and many non-zero elements after the diagonal element that correspond to all upstream nodes of i . Solving (17) by factorizing the matrix $\underline{\underline{B}}$ (e.g., by Gauss-Jordan elimination) is a problem of complexity of $O(n^3)$. To obtain a greater efficiency, we use a Gauss-Seidel iteration scheme to compute $\underline{\underline{h}}^{t+\Delta t}$ in equation (17). This iterative algorithm requires to split the matrix $\underline{\underline{B}}$ into its lower $\underline{\underline{F}}$ and strictly upper triangular matrix $\underline{\underline{E}}$ as follows:

$$\underline{\underline{B}} \cdot \underline{\underline{h}}^{t+\Delta t} = (\underline{\underline{F}} + \underline{\underline{E}}) \cdot \underline{\underline{h}}^{t+\Delta t} = \underline{\underline{b}}^t, \quad (18)$$

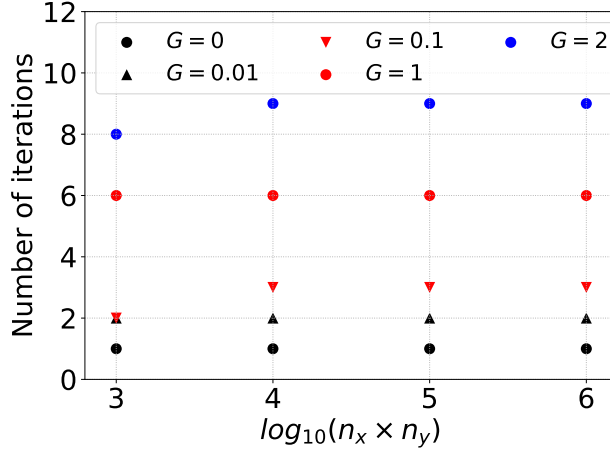


Figure 2. The number of iterations required for the solution to be convergent as a function of resolution $n_x \times n_y$. Solving equation (19) on a square area of resolution $n_x \times n_y = 961, 10000, 99856, \text{ and } 1000000$, using $U = 0.2 \text{ mm/yr}$, $K_f = 2 \times 10^{-5} \text{ m}^{1-2m}/\text{yr}$, $p = 1 \text{ m/yr}$, and varying the value of G .

182 where the matrices $\underline{\underline{F}}$ and $\underline{\underline{E}}$ are shown in Appendix A.

The Gauss-Seidel iterative process starts with an initial guess $h^{t+\Delta t, 0} = \underline{h}^t$ and uses the following recurrence to obtain an improved estimate $\underline{h}^{t+\Delta t, k+1}$:

$$\underline{\underline{F}} \cdot \underline{h}^{t+\Delta t, k+1} = \underline{b}^t - \underline{\underline{E}} \cdot \underline{h}^{t+\Delta t, k}, \quad (19)$$

from the value $\underline{h}^{t+\Delta t, k}$ obtained at the previous iteration. Interestingly, equation (19) can be written in a different form for each node i :

$$h_i^{t+\Delta t, k+1} = \frac{b_i^t - G/(\tilde{p}\tilde{A}_i) \sum_{j=ups(i)} h_j^{t+\Delta t, k} + F_i h_{rec(i)}^{t+\Delta t, k+1}}{1 + F_i}, \quad (20)$$

183 if the nodes are processed in the FastScape stack order.

The procedure is continued until the maximum difference in node elevation between two successive iterations is below a given tolerance ϵ (expressed in meters) as

$$\max |h_i^{t+\Delta t, k+1} - h_i^{t+\Delta t, k}| < \epsilon \quad \text{for all } i \text{ such that } 1 \leq i \leq n_x \times n_y. \quad (21)$$

184 The tolerance is taken as a small fraction (10^{-3}) of the increment in topography, $U\Delta t$.

185 The above procedure based on a Gauss-Seidel iterative scheme is known to con-
 186 verge if the matrix $\underline{\underline{B}}$ is strictly diagonally dominant: $|-F_i| + |\sum_{j=ups(i)} G/(\tilde{p}\tilde{A}_i)| <$
 187 $|1+F_i|$, thus $G/\tilde{p} < \tilde{N}_i/(N_i-1)$ after some derivations. Therefore, the iterative method
 188 is proven to converge unconditionally at least when $G/\tilde{p} \leq 1$, but we show experimen-
 189 tally in section 4 that this method can also converge even if this condition is not satis-
 190 fied.

191 As shown in Figure 2, our new implicit method to solve equation (13) is $O(N)$ as
 192 the number of iterations required in the Gauss-Siedel scheme depends on the value of
 193 G but not on the resolution of the model ($n_x \times n_y$).

194 We note that the left-hand side of equation (19) is the same as in the FastScape
 195 algorithm (*Braun and Willett, 2013*) while the right-hand side only differs by a single

196 term given by: $\underline{E} \cdot \underline{h}^{t+\Delta t, k}$ at time $t + \Delta t$. The implementation of our new algorithm
 197 is therefore a very simple addition to the FastScape algorithm. Note also that, if the value
 198 of the deposition coefficient G is null, the right-hand side term simplifies to \underline{b}^t , and the
 199 new algorithm is identical to the basic FastScape algorithm which does not require the
 200 Gauss-Seidel iteration to obtain the elevation at time $t + \Delta t$.

The above Gauss-Seidel iteration algorithm can be extended to consider different values of K_f and G to differentiate between bedrock and previously deposited sediments during the iteration process. Basement elevation $h_{base}^{t+\Delta t}$ at each step is obtained using:

$$h_{base}^{t+\Delta t} = \min(h^{t+\Delta t}, h_{base}^t + U\Delta t), \quad (22)$$

201 where $h_{base}^t + U\Delta t$ is the basement elevation resulting from uplift without surface processes.
 202

When the value of n is not equal to 1, equation (13) becomes:

$$h_i^{t+\Delta t, k+1} + K_f \tilde{p}^m A_i^m \Delta t \left(\frac{h_i^{t+\Delta t, k+1} - h_{rec(i)}^{t+\Delta t, k+1}}{\Delta l_i} \right)^n = h_i^t + U\Delta t + \frac{G}{\tilde{p}\tilde{A}_i} \sum_{j=ups(i)} (h_j^t + U\Delta t - h_j^{t+\Delta t, k}). \quad (23)$$

203 This non-linear equation can be solved by combining the Gauss-Seidel iterations with
 204 a local Newton scheme. We solve the diffusion equation (12) using an alternating direc-
 205 tion implicit and $O(N)$ scheme (*Peaceman and Rachford, 1955*).

206 3 Model implications on geomorphological relationships

207 Before studying the behaviour of the numerical scheme presented above, we first
 208 wish to derive several basic geomorphological relationships from the evolution equation
 209 (11). These include the steady-state slope-area relationship, the shape of steady-state
 210 river profile, and the expression for the response time (i.e., the time necessary to reach
 211 steady state). In this section, all relationships are for the uplifting region only, and we
 212 neglect the hillslope processes.

213 3.1 Steady-state slope-area ($S - A$) relationship

At steady state (i.e., when uplift is balanced by channel incision assumed to be governed by the SPL model), the slope and intercept of the relationship between slope and drainage area provide constraints on the concavity (the ratio m/n) and the steepness index ($k_s = (U/K_f)^{1/n} \tilde{p}^{-m/n}$), respectively (e.g., *Wobus et al., 2006*, and references therein). When taking deposition into account, the SPL model must be replaced by equation (11), which, under the assumption of steady state (i.e., $\frac{\partial h}{\partial t} = 0$) leads to the following slope-area ($S - A$) relationship:

$$S = \left[\frac{(1 + G/\tilde{p})U}{K_f} \right]^{1/n} A^{-m/n} \tilde{p}^{-m/n} = (1 + G/\tilde{p})^{1/n} (U/K_f)^{1/n} A^{-m/n} \tilde{p}^{-m/n}. \quad (24)$$

From this equation, a new steepness index can be defined:

$$k'_s = (1 + G/\tilde{p})^{1/n} (U/K_f)^{1/n} \tilde{p}^{-m/n}, \quad (25)$$

214 that only differs by a factor $(1 + G/\tilde{p})^{1/n}$. For typical values of $G/\tilde{p} = 1$ (*Davy and*
 215 *Lague, 2009*) and $n = 1$ (*Whipple and Tucker, 1999*), the effect of sediment deposition
 216 is to increase the steepness index by a factor of 2.

217 3.2 Steady-state river profile

At steady state, a power-law relationship is generally observed between the length of a stream, x , and its upstream drainage area (*Hack, 1957; Lague et al., 2003; Montgomery and Dietrich, 1992; Morisawa, 1962; Walcott and Summerfield, 2009*) that can

be expressed as:

$$A = k x^b, \quad (26)$$

where k and b , usually close to 0.5 and 2, respectively, are called Hack's law coefficients. Combining equation (26) with equation (24) leads to:

$$S = \frac{dh}{dx} = (1 + G/\tilde{p})^{1/n} U^{1/n} K_f^{-1/n} \tilde{p}^{-m/n} k^{-m/n} x^{-\frac{bm}{n}}. \quad (27)$$

By integrating the above equation over the length of the stream, we obtain the steady-state river profile:

$$h(x) = (1 + G/\tilde{p})^{1/n} U^{1/n} K_f^{-1/n} \tilde{p}^{-m/n} k^{-m/n} \left(1 - \frac{bm}{n}\right)^{-1} x^{1-\frac{bm}{n}} + C, \quad \text{for } bm \neq n, \quad (28)$$

218 where $C = 0$ if the elevation at base level is zero (i.e., $h(0) = 0$). We see that the steady-
219 state elevation of every point along the river profile is $(1 + G/\tilde{p})^{1/n}$ times higher than
220 what is expected from the classic SPL model (i.e., without sediment deposition).

221 3.3 Response time

The response time of a fluvial landscape can be defined as the time needed for river profiles to reach steady state. In our model, the response time is

$$\tau' = h(L)/U = (1 + G/\tilde{p})^{1/n} U^{1/n-1} K_f^{-1/n} \tilde{p}^{-m/n} k^{-m/n} \left(1 - \frac{bm}{n}\right)^{-1} L^{1-\frac{bm}{n}}, \quad (29)$$

where L is the length of the uplifting region. In the classic SPL form (without the term of deposition), the response time of a fluvial landscape is (*Whipple, 2001*)

$$\tau = U^{1/n-1} K_f^{-1/n} \tilde{p}^{-m/n} k^{-m/n} \left(1 - \frac{bm}{n}\right)^{-1} L^{1-\frac{bm}{n}}. \quad (30)$$

The ratio between these two response times is once again given by:

$$\frac{\tau'}{\tau} = (1 + G/\tilde{p})^{1/n}. \quad (31)$$

The impact of continental deposition on the evolution of a fluvial landscape is highlighted by the three geomorphological relationships above: steady-state slope is higher, average topography in the uplifting region is higher, and response time is longer than with the classic SPL model (i.e., without deposition). However, neither the shape of the river profile, nor the dependency of the response time or the steepness index to other parameters (such as K_f , m and n , or the length of the channel, precipitation rate or the uplift rate) are affected by sediment deposition. The dimensionless deposition coefficient G alone controls the difference with respect to the classic SPL model and appears only inside a multiplying factor, $(1 + G/\tilde{p})^{1/n}$. In all three relationships we derived above, the factor $(1 + G/\tilde{p})^{1/n}$ multiplies the other poorly constrain factor $K_f^{-1/n}$; this means that the effect of sediment deposition (on most morphometric measures and scales) can be included in the SPL erosion coefficient K_f by simply redefining this constant in the following way:

$$K_f' = K_f / (1 + G/\tilde{p}). \quad (32)$$

222 This also means that the value of the constant G cannot be easily derived from the con-
223 cavity of rivers, the total relief of river channels or the response time of fluvial erosion.

224 4 Model behavior

225 We now demonstrate the behavior and applicability of our numerical implemen-
226 tation of the modified SPL model by performing a range of simulations. These simula-
227 tions are based on a simple setup composed of an uplifting region adjacent to a stable

Table 1. Parameters for the simulations.

Notation	Definition	Values/Range	Unit
h	elevation of topography		m
h_{base}	elevation of basement		m
t	time	10	Myr
Δt	incremental time	1000	yr
x	horizontal dimension	105 to 400	km
y	horizontal dimension	100	km
$\Delta x, \Delta y$	cell size	1	km
U	uplift rate	0.2, 0.5, 1	mm/yr
p	net precipitation rate	0.5, 1, 2	m/yr
p_0	mean net precipitation rate	1	m/yr
$\tilde{p} = p/p_0$	variation ratio of precipitation rate	0.5, 1, 2	-
A	drainage area		m ²
S	surface slope in drainage direction		-
m	SPL exponent	0.4 ^a	-
n	SPL exponent	1 ^b	-
K_{fb}	SPL erosion coefficient of bedrock	2×10^{-5} ^c	m ^{1-2m} /yr
K_{fs}	SPL erosion coefficient of sediments	$(2, 4, 8) \times 10^{-5}$	m ^{1-2m} /yr
G	deposition coefficient	$(0, 0.1, 1, 10)$ ^d	-
K_d	hillslope diffusion coefficient	0.01 ^e	m ² /yr

^aParameters from *Stock and Montgomery (1999)* and *Perron et al. (2009)*; ^bparameters from *Braun and Willett (2013)*; *Stock and Montgomery (1999)*; *Whipple and Tucker (1999)* and *Braun and Willett (2013)*; ^cparameters from *Whipple and Tucker (1999)*; ^dparameters from *Davy and Lague (2009)*; ^eparameters from *Densmore et al. (2007)* and *Armitage et al. (2013)*.

228 continental area on which a foreland basin is allowed to develop. Setup for the simula-
 229 tions is a rectangular area of (105 to 400)×100 km² (Table 1, Figure 1b). The initial to-
 230 pography has random noise elevation of up to 1 m. This domain is discretized into a num-
 231 ber of cells with a cell size of $\Delta x = \Delta y = 1$ km. The left-hand side of the domain (100×
 232 100 km²) is uplifted at a constant rate U while the foreland is fixed, and the foreland
 233 edge (base level) through which the sediments can leave the system is also fixed. We use
 234 a constant diffusion coefficient of $K_d = 0.01$ m²/yr in all simulations. The net mean
 235 precipitation rate p_0 is set to 1 m/yr. We perform a series of model runs for a total time
 236 of 10 Myr (10 000 time steps of duration 1000 yr) (Table 1).

237 4.1 Simulations without foreland

238 We first performed a series of simulations without the foreland to better illustrate
 239 the effect of sediment deposition. We run a simulation with $U = 0.2$ mm/yr, $K_f = 2 \times$
 240 10^{-5} m^{0.2}/yr, $p = 1$ m/yr and $G = 0$ to test the classic SPL model without sediment
 241 deposition. During the experiment, topography rises, and after 4 Myr, reaches steady
 242 state. The final simulated landscape is shown in Figure 3a and the evolution of the mean
 243 elevation is presented in Figure 3b. As observed from the black curve ($G = 0$, Figure
 244 3b), the mean elevation in the uplifting region increases progressively before decreasing,
 245 and then reaches a constant value at steady state. The same behavior is observed in *Davy*
 246 *and Lague's* (2009) detachment-limited simulations (their Figure 1). Note that this non-
 247 monotonous evolution of the mean elevation is probably related to the presence of lo-
 248 cal minima which are important at the beginning because we start with an initial topog-
 249 raphy that has random noise. Local minima are known to artificially reduce the erosional

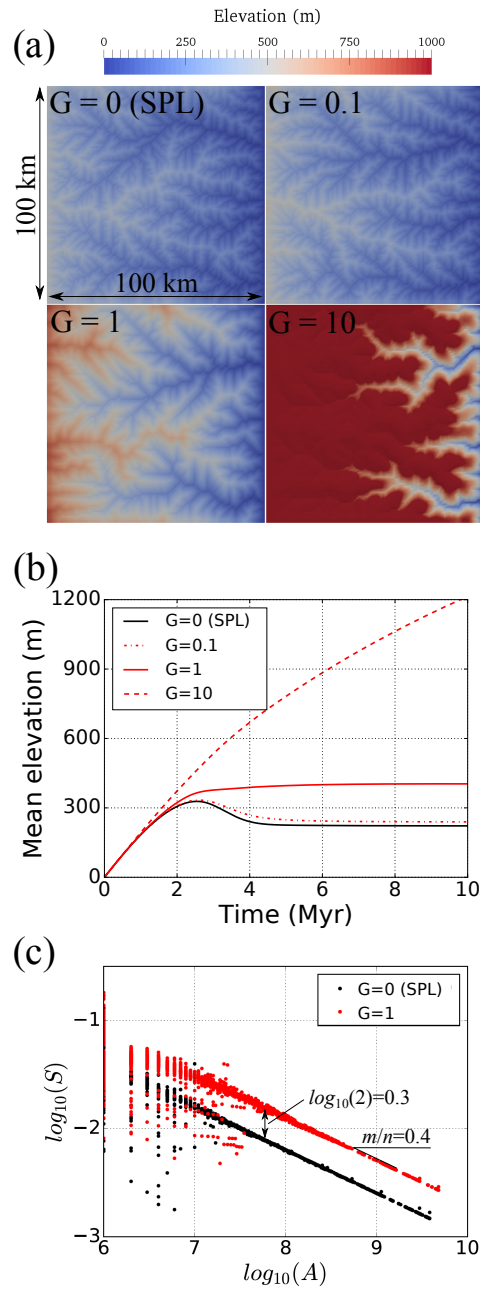


Figure 3. (a) Simulated, plan view topography of the uplifting region for four values of $G = 0, 0.1, 1,$ and $10,$ at 10 Myr. $G = 0$ corresponds with the classic SPL model without taking into account sediment deposition. (b) Mean elevation of uplifting region as a function of time for the four G values. (c) $\log_{10}(S)$ as a function of $\log_{10}(A)$ in steady-state uplifting region for $G = 0$ (SPL model) and $1.$ We observe that the concavity is $m/n = 0.4,$ and the difference in steepness index k_s is by a factor of $2.$

250 efficiency of landscape evolution models by preventing some river discharge to reach down-
 251 stream locations.

252 On the contrary, a peak in elevation is not observed in experimental models such
 253 as in *Babault et al. (2005)* where they study the development of relief by submitting a
 254 thick layer of silica powder to uplift and runoff-driven erosion. In their experiments, the
 255 mean elevation increases monotonously before stabilizing around a constant value (their
 256 Figure 2A). We suggest that the difference between the two behaviors is due to sediment
 257 deposition. To prove this point, we run simulations with $G = 0.1, 1, \text{ and } 10$. Physically,
 258 the larger G , the more deposition in the river channels. The final landscape for the cases
 259 of $G = 0.1, 1, \text{ and } 10$ are compared in Figure 3a. The two simulations with $G = 0$ and
 260 $G = 0.1$ are very similar to each other (Figure 3a), but the topography changes signif-
 261 icantly if the value of G is ≥ 1 . As demonstrated by *Davy and Lague (2009)*, the value
 262 of $G/\bar{p} = 1$ corresponds to the transition from detachment-limited to transported-limited
 263 behavior.

264 In terms of average topography, the larger G , the higher the mean elevation (Fig-
 265 ure 3b). For $G = 0.1$, only slight differences in mean elevation can be observed with
 266 respect to the $G = 0$ simulation (Figure 3b) and steady-state elevation is reached af-
 267 ter ~ 4 Myr. On the contrary, the mean elevation for $G = 1$ increases continuously
 268 for $\sim 6-7$ Myr before reaching a constant elevation of ~ 400 m, which is almost twice
 269 higher than the classic SPL model simulation ($G = 0$) (Figure 3b). This results agrees
 270 with our estimate of the response time for fluvial systems including the effect of sedi-
 271 ment deposition. When $G \geq 1$, the evolution of the mean elevation is consistent with
 272 the one observed in the experiments by *Babault et al. (2005)*. As expected from equa-
 273 tions (28) and (31), the average elevation at steady state and the response time of the
 274 landscape are $(1 + G/\bar{p})$ times larger when sediment deposition is considered with re-
 275 spect to the classic SPL model. For the case of $G = 10$ (Figure 3b), the mean eleva-
 276 tion increases substantially but it does not reach steady state over 10 Myr of simulation.
 277 If the model run is allowed to last for a longer period, it should reach steady state af-
 278 ter approximately 40 Myr, based on equation (31).

279 The similarities between laboratory experimental landscapes and our simulations
 280 with $G > 0$ confirm that a term for the effect of sediment is required to accurately de-
 281 scribe the evolution of a fluvial landscape: although bedrock erosion is the dominant pro-
 282 cess in the uplifted region (i.e., no net deposition takes place anywhere), the reduction
 283 in erosion rate due to sediment deposition, the so-called “cover effect”, is key to capture
 284 landscape evolution (*Sklar and Dietrich, 1998; Whipple and Tucker, 2002*).

285 4.2 Simulations with foreland sedimentation

286 4.2.1 Reference simulation

287 We run a “reference simulation” that includes the presence of a foreland (i.e., a re-
 288 gion where $U = 0$). In this reference experiment, the erosion coefficients for bedrock
 289 and sediments are first set to the same value for the sake of simplicity (i.e., $K_{fb} = K_{fs} =$
 290 K_f). The value of K_f is set to $2 \times 10^{-5} \text{ m}^{0.2}/\text{yr}$, identical to the one used in the pre-
 291 vious set of experiments without a foreland, and G is set to 1. The uplift rate is set to
 292 $U = 0.2 \text{ mm/yr}$, the precipitation rate to $p = 1 \text{ m/yr}$, and the foreland length to 50
 293 km. The landscape evolution through time and the axial stratigraphy of the resulting
 294 foreland basin, with a synthetic stratigraphic layer generated each Myr, are presented
 295 in Figure 4a. The associated animations Movie S1 and Movie S2 are presented in the
 296 Supporting Information.

297 At the outlets of the catchments that drain the uplifting region, we observe the for-
 298 mation of small sedimentary fans that progressively coalesce into larger ones. This fan-
 299 like deposition is the result of using the SPL model with sediment deposition. As up-

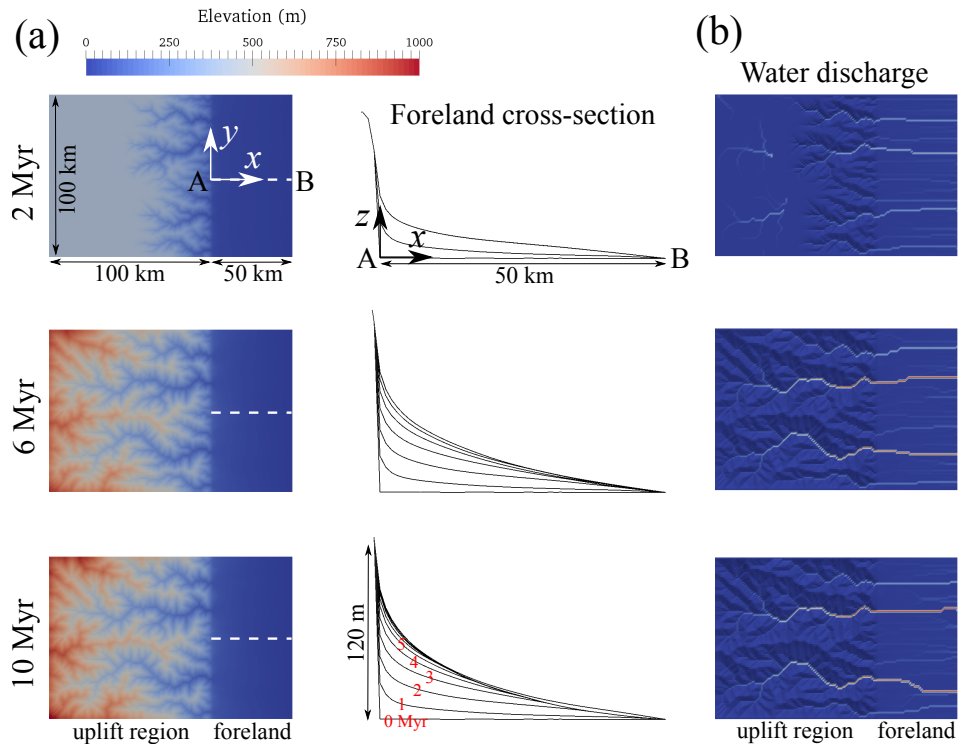


Figure 4. (a) Map view of landscape evolution and longitudinal profile of the central foreland topography (white dashed line in a), and (b) water discharge (the warmer the color, the larger the water discharge) at 2 Myr, 6 Myr, and 10 Myr. See also Movie S1 for landscape evolution, Movie S2 for foreland central cross-section, and Movie S3 for water discharge (lateral variation of river channels in foreland). In (a), the foreland develops a dynamic steady state with the continuous formation and destruction of the riverbeds.

lift and erosion progress, the fans prograde to reach the base level at the far edge of the foreland where the sediments are allowed to leave the system.

The mean elevation of the uplifting region, the maximum thickness of sediments, and the mean surface slope of the foreland basin as function of time are shown by solid curves in Figures 5a-c. Compared to the equivalent simulation without foreland (red curve in Figure 5a), the mean elevation of the uplifting region with a 50-km-long foreland (solid curve in Figure 5a) is about ~ 70 m higher as it reaches a steady-state elevation of ~ 470 m.

Sediment deposition in the foreland basin results in a progressive increase of its slope up to a threshold, over which the sediment flux coming from the uplifting region bypasses the foreland. This steady-state slope varies over the foreland, and the mean slope of the foreland is ~ 0.0032 ($\sum_{i=1}^{N_f} S_i/N_f$, where N_f is the number of foreland nodes, solid curve in Figure 5c), typical of fans built by shallow to deep braided channelized flows (*Stanistreet and McCarthy*, 1993). Many studies of sedimentary fans (*Guerit et al.*, 2014; *Reitz and Jerolmack*, 2012; *Van Dijk et al.*, 2009) have shown that the surface slope stabilizes after some time, which is considered as the steady-state morphology of the fan. In many experimental studies, the fan slope stabilizes quickly because its apex is fixed. Stabilization also occurs if the apex is uplifted through time as the mountain front eventually reaches a steady-state elevation (*Densmore et al.*, 2007). In our simulations, the slope tends to stabilize after several Myr when the mean elevation of the uplifting region has reached steady state.

4.2.2 Sensitivity analysis

In the following, we explore the impact of several parameters such as different foreland lengths, uplift rates, precipitation rates, and different erosion coefficients for bedrock and sediments. The values used for the sensitivity analysis are summarized in Table 1.

In Figure 5, we first assess the effect of foreland length by running four simulations with foreland length of 50 km (the reference simulation, solid curve), 100 km (dashed curve), 200 km (dotted curve), and 300 km (dash-dotted curve), respectively. U is set to 0.2 mm/yr, p to 1 m/yr, G to 1, and K_f to 2×10^{-5} m^{0.2}/yr. The maximum thickness of foreland sediments for larger foreland length is higher than the reference simulation (Figure 5b). The sediment thickness in foreland imposes the local base level of the mountain range. Thus, the mean elevation of the uplifting region for larger foreland length is higher than the reference simulation (Figure 5a). However, in the foreland, the mean slopes are significantly lower than the reference simulation (Figure 5c), as they are related to the length of the foreland basin. In fact, sediments are allowed to deposit everywhere until they reach the edge of the domain. Therefore, for a given amount of sediments, the larger the foreland basin, the lower the slope of the foreland basin. This behavior is observed because in our model, rivers are always able to transport sediments even across a large foreland basin.

In Figure 4a, the slope of the fan varies over the 50-km-long foreland, and the fan surface is concave up. We performed several simulations with smaller foreland lengths (5 km, 10 km, and 20 km) in Figure 6. The predicted foreland thickness scales directly with the foreland length (Figure 6). The simulations also show that the fan surface is relatively straight when the foreland length is small (e.g., 5 km). The reduction in fan slope could be due to the reduction in the ratio between the sediment flux q_s and the water flux q_w . The larger the foreland length, the larger the reduction in q_s/q_w downstream, and the more curved the fan surface, as observed in km-scale fans (e.g., *Bull*, 1964; *Densmore et al.*, 2007).

We test for the impact of the uplift rate by running three simulations with $U = 0.2, 0.5$ and 1 mm/yr, respectively. Basin length is set to 50 km, p to 1 m/yr, G to 1,

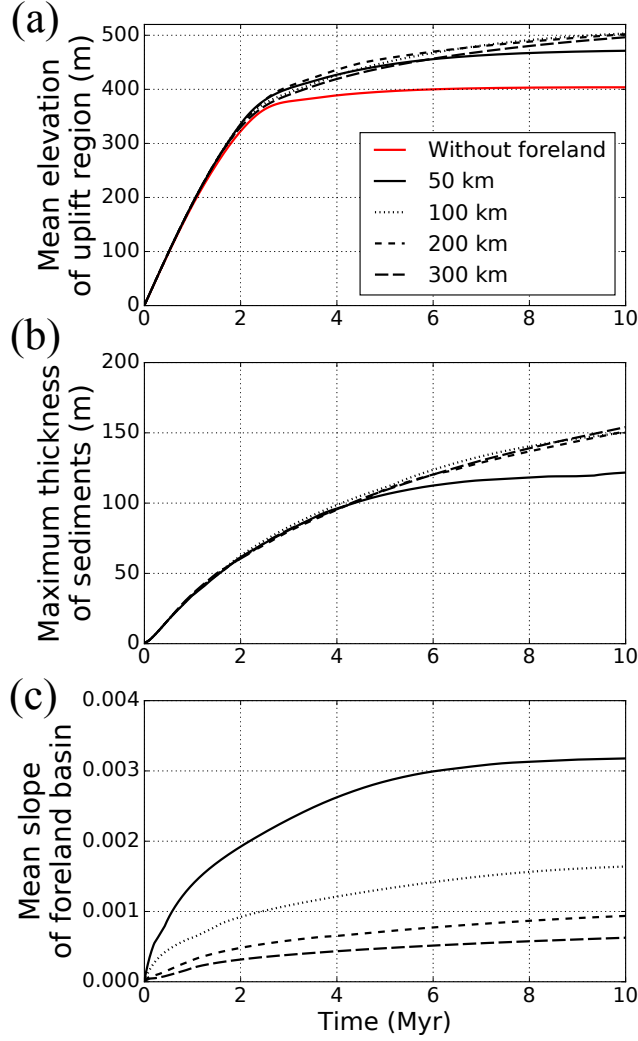


Figure 5. (a) Mean elevation of the uplifting region without foreland, and with foreland length of 50 km, 100 km, 200 km, and 300 km. (b) Maximum thickness of sediments in foreland, and (c) mean slope of the foreland basin as a function of time for foreland length of 50 km, 100 km, 200 km, and 300 km. U is set to 0.2 mm/yr, p to 1 m/yr, G to 1, and K_f to $2 \times 10^{-5} \text{ m}^{0.2}/\text{yr}$.

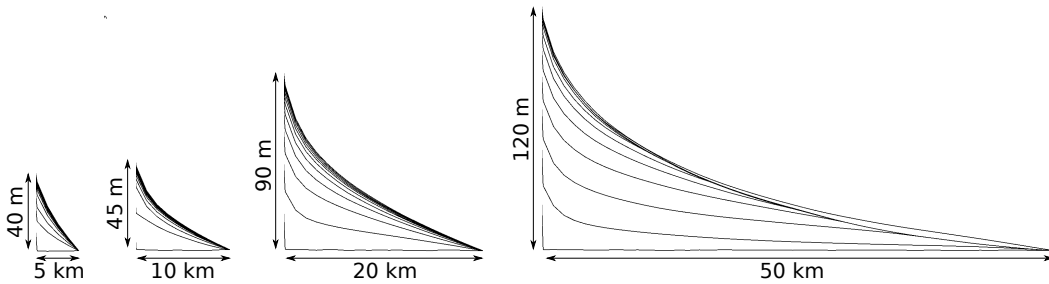


Figure 6. Longitudinal profile of the central foreland topography for various of foreland lengths. Fan surface is relatively straight when the foreland length is small (e.g., 5 km), and is concave up when the foreland length is large (≥ 10 km), consistent with field observations (*Bull*, 1964; *Densmore et al.*, 2007).

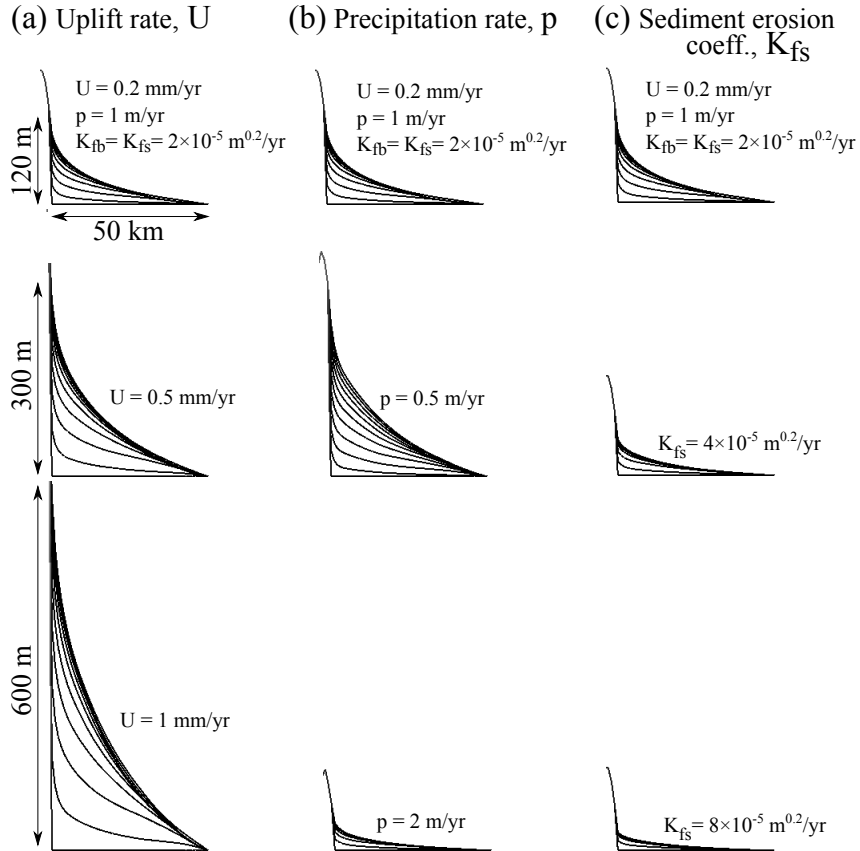


Figure 7. Modelled output for foreland basin showing sensitivity tests on (a) uplift rate U , (b) precipitation rate p , and (c) sediment erosion coefficient K_{fs} . First row shows the standard simulation with the same values throughout. All figures have the same vertical scale.

and K_f to $2 \times 10^{-5} \text{ m}^{0.2}/\text{yr}$. The main impact of an increased uplift rate is a higher steady-state elevation in the uplifting region. In fact, according to equation (28), the steady-state elevation in the uplifting region is proportional to $U^{1/n}$. The increase in sediment accumulation is also a direct result of the increase in sediment flux resulting from a higher uplift rate. In consequence, sediments can accumulate more in the foreland, and the outlets of the drainage basins are significantly higher (up to 600 m for $U = 1 \text{ mm/yr}$ compared to $\sim 120 \text{ m}$ for the reference simulation, Figure 7a).

We then test for the impact of precipitation rate by running three simulations with $p = 0.5, 1, \text{ and } 2 \text{ m/yr}$, respectively. Basin length is set to 50 km, U to 0.2 mm/yr , G to 1, and K_f to $2 \times 10^{-5} \text{ m}^{0.2}/\text{yr}$. We observe that the higher the precipitation rate p , the lower the foreland deposition and the foreland slope (Figure 7b). This is due to the fact that the erosion rate is proportional to \tilde{p}^m in equation (11), but the deposition term is inversely proportional to \tilde{p} , thus, the net deposition rate is proportional to $1/\tilde{p}^{1-m}$. This implies that at higher precipitation rates, the transit of sediments in the foreland is faster (i.e., bypass is more efficient).

Finally, we test for the impact of the erosion coefficient, K_f . In these simulations, we consider different values for the bedrock (K_{fb}) and the sediment (K_{fs}) erosion coefficients. We run three simulations with $K_{fb} = 2 \times 10^{-5} \text{ m}^{0.2}/\text{yr}$ and $K_{fs} = 2 \times 10^{-5}$, 4×10^{-5} and $8 \times 10^{-5} \text{ m}^{0.2}/\text{yr}$, respectively. Basin length is set to 50 km, p to 1 m/yr, and G to 1. As the erosion coefficient increases, sediments in the foreland are more and more remobilized by river erosion, resulting in lower slopes in the foreland basin (Figure 7c). Almost no sediments are preserved in the simulation of $K_{fs} = 8 \times 10^{-5} \text{ m}^{0.2}/\text{yr}$. This means that the reworking (by erosion) of sediments in the foreland is an efficient process that enhances the bypass of sediments from the uplifting region toward the base level.

5 Discussion

5.1 Slope-area ($S - A$) relationship from simulations

By deriving equation (24), we have shown that the inclusion of depositional processes should increase the steepness index, k_s , with respect to the classic SPL model (i.e., with $G = 0$) while the concavity should remain unchanged. This is illustrated by the slope-area analysis of two experiments without foreland with $G = 0$ and $G = 1$ (Figure 3c). With $n = 1$, the concavity and the steepness index can be derived from log-log plots, where the slope of the slope-area relationship is the concavity and the intercept with the origin is the \log_{10} of the steepness index. The slope-area analysis of the simulations $G = 0$ and $G = 1$ reveals a concavity of 0.4 in both experiments, as we expect from our m and n values (0.4 and 1, respectively). On the contrary, despite similar U and K_f , the steepness index k_s is increased by a factor 2 when $G = 1$ (Figure 3c). Through these numerical experiments, we confirm that the steepness index, usually used to derive information about uplift and landscape properties, is also sensitive to the amount of deposition within the fluvial network.

5.2 Autocyclic aggradation and incision in the foreland

During a simulation, the landscape evolves until it reaches an equilibrium topography (when the elevation in the mountain range and the slope in the foreland become stable, Figure 5). In our simulations, river channels continuously migrate laterally in the foreland (Figure 4b, see the animation Movie S3 on water discharge in the Supporting Information), similar to observations made in laboratory experiments (*Van Dijk et al., 2009*) and other numerical simulations (*Croissant et al., 2017; Pepin et al., 2010*). By considering the axial topography (Figure 4a; the corresponding animation Movie S2 is presented in the Supporting Information), we can indeed observe that the foreland steady-

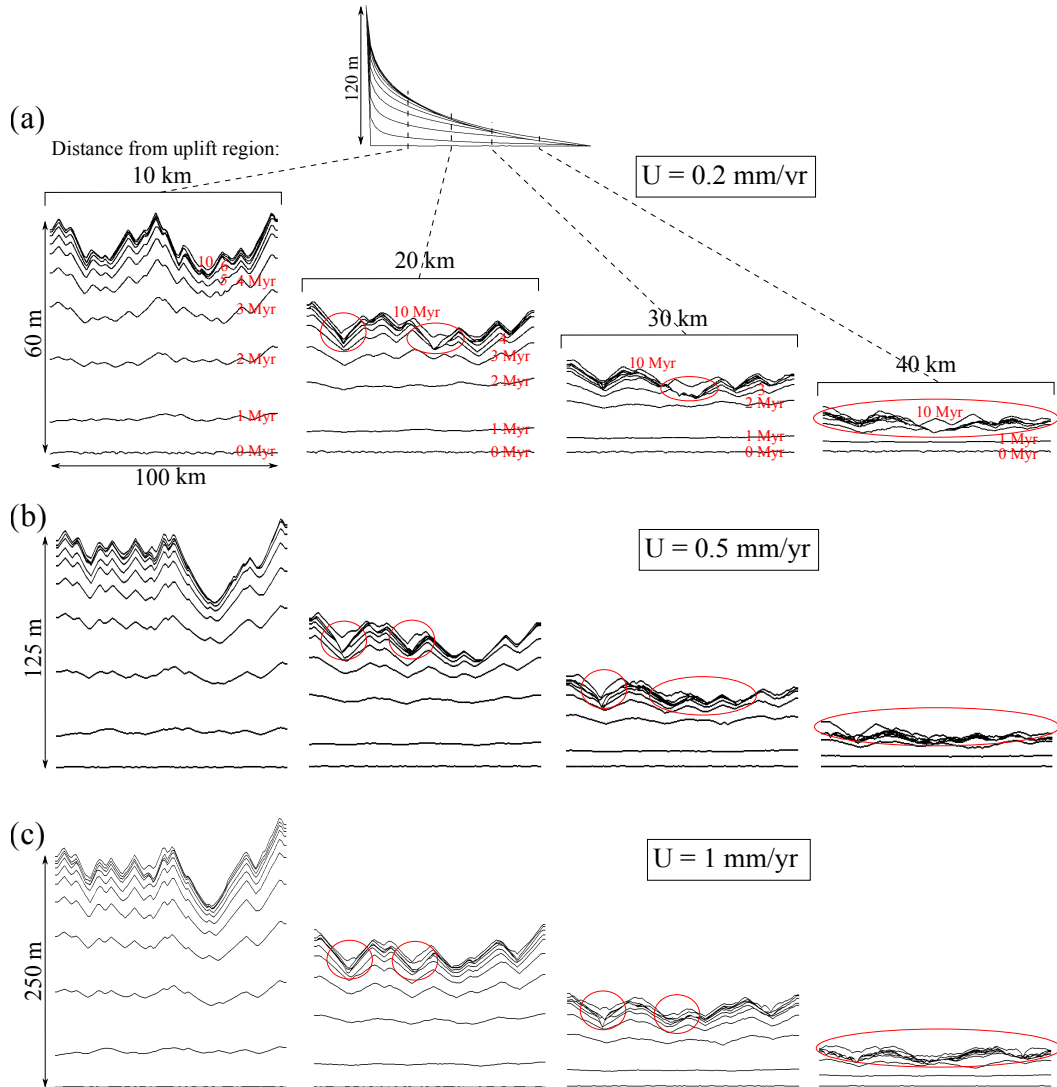


Figure 8. Foreland cross sections at 10 km, 20 km, 30 km, and 40 km from the mountain front for the uplift rate (a) $U = 0.2$ mm/yr, (b) $U = 0.5$ mm/yr, and (c) $U = 1$ mm/yr. Animation of the simulation Movie S4 for aggradation and incision cycles in foreland (a) is presented in the Supporting Information. Red ellipses show the general locations of autocyclic aggradation and incision around a dynamic steady state.

399 state configuration is a dynamic steady state with the continuous formation and destruc-
 400 tion of river banks. The formation and destruction of river banks can be observed from
 401 the foreland sections parallel to the mountain front (Figure 8a; the corresponding an-
 402 imation Movie S4 is presented in the Supporting Information). The successive cycles of
 403 aggradation and incision mainly occur in the lower foreland (30 km and 40 km in Fig-
 404 ure 8a) where the longitudinal slope is low. In the upper foreland, where the slope is higher
 405 (10 km and 20 km in Figure 8a), sediments mainly aggrade until reaching steady state.
 406 This can be seen, for example, in Figure 4a where sediments aggrade up to a maximum
 407 of ~ 120 m at the fan apex (i.e., along the transition between the foreland and the up-
 408 lifting region). Yet, the amplitude of aggradation and incision is higher close to the moun-
 409 tain range than in the lower foreland, in good agreement with *Bull* (1964).

410 The successive aggradation and incision cycles arises from the competition between
 411 the SPL erosion term $K_f \tilde{p}^m A^m S^n$ and the deposition term $\frac{G}{\bar{p}A} \int_A (U - \frac{\partial h}{\partial t}) dA$ in equa-
 412 tion (11) where the uplift rate $U = 0$ mm/yr in the foreland. When the foreland slope
 413 S is low, the erosion term $K_f \tilde{p}^m A^m S^n$ is lower than the deposition term $\frac{G}{\bar{p}A} \int_A (U -$
 414 $\frac{\partial h}{\partial t}) dA$. Then, sediments accumulate in the foreland (because sediments are deposited
 415 more than eroded), resulting in an increase of the foreland slope. However, when the fore-
 416 land slope S is sufficiently large, the erosion term becomes greater than the deposition
 417 term, leading to a decrease in slope. The SPL erosion and deposition terms are never
 418 equal to each other in the foreland, and a dynamic steady state develops characterized
 419 by autocyclic and alternating phases of aggradation and incision.

420 Note that the node spacing in the simulation is 1×1 km² (Figure 8a). To cap-
 421 ture morphological features such as valleys and hillslopes that typically occur over shorter
 422 space scales, we ran two other simulations with a higher spatial resolution of 100 m and
 423 20 m (Figure S1 in the Supporting Information). The model is smaller (15×10 km²)
 424 than the reference simulation, and has an uplifted domain of 10×10 km². The processes
 425 that lead to the autocyclic aggradation and incision are replicated (Figure S1), and the
 426 spacing between valleys and the migration frequency of river channels are not affected
 427 by the spatial resolution of the model.

428 *Langston et al.*'s (2015) simulations show that the repeated aggradation and aban-
 429 donment of high surfaces take place in the upper foreland due to modulation of sediment
 430 supply associated with climate change. Unlike their simulations, ours show that, with-
 431 out changing external forcings, streams migrate and sometimes erode into the sediment
 432 they have previously deposited, mainly in the lower foreland, around a dynamic steady
 433 state. Under a constant forcing, we observe that the upper foreland is quite stable. This
 434 could be related to the relatively limited numbers of outlets at the edges of the moun-
 435 tain range, which limits the possibility of major fluvial reworking close to the range.

436 Riverbed aggradation and incision cycles are often interpreted to reflect changing
 437 external forcings including tectonics (*Bull*, 1991; *Yanites et al.*, 2010), climate (*Bridg-*
 438 *land and Westaway*, 2008; *Hancock and Anderson*, 2002), and base-level change (*Fisk*
 439 *et al.*, 1945). Yet, without varying tectonics (e.g., the value of U) or climate (e.g., the
 440 value of p), our numerical simulations continuously produce lateral migration of river chan-
 441 nels in the foreland, resulting in autocyclic aggradation and incision around a dynamic
 442 steady state. Using different uplift rates (e.g., $U = 0.5$ mm/yr and 1 mm/yr in Fig-
 443 ures 8b and 8c, respectively), we observe larger aggradation and incision thicknesses, be-
 444 cause sediments accumulate more in the foreland for higher uplift rates.

445 5.3 Limitations of the model

446 There are several limitations to this model and how it can be used to represent nat-
 447 ural processes. First of all, our model of course encompasses some limitations of the clas-
 448 sic SPL model from which it is derived and such limitations should be considered to de-
 449 sign future studies. For example, we can not account for river width variations or lat-

450 eral migration in bedrock channel, preventing the simulations of alluvial terraces within
 451 uplifted areas (*Langston and Tucker, 2018*). Channel narrowing and widening also in-
 452 fluence the stream power per unit width, and therefore the transport capacity (e.g., *Finnegan*
 453 *et al., 2005; Whittaker et al., 2007*).

454 Second, the thickness and location of deposited sediments above bedrock elevation
 455 is not recorded, as erosion and deposition are calculated within the same equation and
 456 at the same time step. A valuable addition to this work would be to track sediment de-
 457 position within uplifting regions, either within channels or alongside channels in the form
 458 of terraces. The transient sediment storage in inter-montane basins or within uplifting
 459 regions could be a potential application to better understand landscape evolution. For
 460 this, the thickness of sediments above bedrock elevation should be tracked during land-
 461 scape evolution (e.g., *Shobe et al., 2017*).

462 In addition, the model is built in fluxes and it does not account for grain-size dis-
 463 tribution (i.e., the eroded material behaves as a single grain-size material). Accordingly,
 464 we neither distinguish between bedload and suspended load, nor simulate processes such
 465 as sediment downstream fining. The size of the sediments is an important parameter that
 466 partly controls the shape and the geometry of sedimentary bodies such as alluvial fans
 467 (*Armitage et al., 2011; Delorme et al., 2016; Paola et al., 1992*). Multiple grain sizes can
 468 be incorporated into the model framework by splitting the flux into different loads as-
 469 sociated with different deposition coefficients. This would also require to track grain size,
 470 which is not trivial to perform in *Davy and Lague's* (2009) flux-based formulation.

471 Last, the model did not account for flexural isostasy either in the uplifted domain
 472 or in the foreland, mostly for the sake of simplicity, to explain the behavior of the sur-
 473 face process model, independent of any assumption regarding isostatic response. Flex-
 474 ural subsidence of foreland basins creates accommodation space and allows sediments
 475 supplied from the uplifted relief to be preserved through time. To simulate sediment stratig-
 476 raphy with the current model, a first-order approximation would be to impose a nega-
 477 tive uplift in the foreland basin to simulate subsidence and enhance sediment deposition
 478 and preservation.

479 6 Conclusion

480 We have developed a new efficient (i.e., $O(N)$ and implicit) method to solve the
 481 equations arising from *Davy and Lague's* (2009) formulation to investigate the role of
 482 sediment deposition on fluvial erosion of landscapes. In our implementation, the depo-
 483 sition rate is assumed to be inversely proportional to drainage area, and proportional
 484 to the local suspended sediment flux (estimated from the integrated upstream net ero-
 485 sion rate) and to a dimensionless deposition coefficient, which we term G . This depo-
 486 sition coefficient G is a function of the sediment concentration ratio in transport, the set-
 487 tling velocity of sediment, and the mean precipitation rate. From a macro point of view,
 488 G/\tilde{p} in equation (11) is a direct measure of the local deposition rate relative to the mean
 489 net erosion rate of upstream drainage area.

490 To study the behavior of this new scheme and its numerical implementation, we
 491 performed a series of simple experiments in which a region subject to uplift at a constant
 492 rate is adjacent to a stable area. Based on multiple simulations, we demonstrate that
 493 the relief in the uplifting area and the stratigraphy in the foreland basin are controlled
 494 by the river erosion coefficient K_f and by the value of G . The deposition coefficient G
 495 controls the average elevation in the uplifting region and the surface slope in the asso-
 496 ciated foreland basin, especially when the value of G is ≥ 1 . At steady state, the slope,
 497 the average elevation along river profiles and the response time of the uplifted relief are
 498 a factor $(1 + G/\tilde{p})^{1/n}$ higher than those derived from the SPL model without a depo-
 499 sition term.

500 In our simulations, as uplift and erosion progress, sediments aggrade in the fore-
 501 land basin and prograde toward the foreland edge (the base level). The foreland slope
 502 increases until it reaches a threshold over which sediments bypass the foreland. The fore-
 503 land then develops a dynamic steady state, in which the stream continuously migrates
 504 and sometimes cuts into the sediment it has previously deposited. This process leads to
 505 the continuous formation and destruction of river banks. This behavior mainly takes place
 506 in the lower foreland where the slope is low, whereas in the upper foreland where the slope
 507 is relatively high and rivers are less free to migrate, the sediments mainly aggrade un-
 508 til reaching steady state. This observed dynamic steady state suggests that aggradation
 509 and incision cycles can occur in the absence of variations in external forcings such as tec-
 510 tonics, climate or change of base level.

511 As our method is optimally efficient (i.e., $O(N)$ and implicit), it is therefore highly
 512 suitable for performing large numbers of simulations and it can thus be used in optimiza-
 513 tion procedures that require many simulations to infer the amplitude of variations in tec-
 514 tonic forcings, climatic events or base-level changes in natural examples. In the future,
 515 we plan on developing this model to better constrain the nature and timing of erosional
 516 events on continents, through an inversion of the stratigraphy of the adjacent foreland
 517 basins.

518 A Matrix of finite difference equations in FastScape stack order

To illustrate our algorithm, we now explicitly write the coefficients of matrix \underline{B} and
 vectors $\underline{h}^{t+\Delta t}$ and \underline{b}^t in equation (17) for a single catchment made of 11 nodes (Figure
 1c) using the FastScape stack order. This gives:

$$\underline{B} = \begin{pmatrix} 1 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ -F_2 & 1 + F_2 & G/(\tilde{p}\tilde{A}_2) & G/(\tilde{p}\tilde{A}_2) & G/(\tilde{p}\tilde{A}_2) & G/(\tilde{p}\tilde{A}_2) & 0 & 0 & 0 & 0 & 0 \\ 0 & -F_3 & 1 + F_3 & G/(\tilde{p}\tilde{A}_3) & G/(\tilde{p}\tilde{A}_3) & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & -F_4 & 1 + F_4 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & -F_5 & 0 & 1 + F_5 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & -F_6 & 0 & 0 & 0 & 1 + F_6 & 0 & 0 & 0 & 0 & 0 \\ -F_7 & 0 & 0 & 0 & 0 & 0 & 1 + F_7 & G/(\tilde{p}\tilde{A}_7) & G/(\tilde{p}\tilde{A}_7) & G/(\tilde{p}\tilde{A}_7) & G/(\tilde{p}\tilde{A}_7) \\ 0 & 0 & 0 & 0 & 0 & 0 & -F_8 & 1 + F_8 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & -F_9 & 0 & 1 + F_9 & G/(\tilde{p}\tilde{A}_9) & G/(\tilde{p}\tilde{A}_9) \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & -F_{10} & 1 + F_{10} & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & -F_{11} & 0 & 1 + F_{11} \end{pmatrix}, \quad (\text{A.1})$$

$$\underline{h}^{t+\Delta t} = \begin{pmatrix} h_1^{t+\Delta t} \\ h_2^{t+\Delta t} \\ h_3^{t+\Delta t} \\ h_4^{t+\Delta t} \\ h_5^{t+\Delta t} \\ h_6^{t+\Delta t} \\ h_7^{t+\Delta t} \\ h_8^{t+\Delta t} \\ h_9^{t+\Delta t} \\ h_{10}^{t+\Delta t} \\ h_{11}^{t+\Delta t} \end{pmatrix}, \quad \text{and} \quad \underline{b}^t = \begin{pmatrix} h_1^t \\ h_2^t + U\Delta t + G(h_3^t + h_4^t + h_5^t + h_6^t + 4U\Delta t)/(\tilde{p}\tilde{A}_2) \\ h_3^t + U\Delta t + G(h_4^t + h_5^t + 2U\Delta t)/(\tilde{p}\tilde{A}_3) \\ h_4^t + U\Delta t \\ h_5^t + U\Delta t \\ h_6^t + U\Delta t \\ h_7^t + U\Delta t + G(h_8^t + h_9^t + h_{10}^t + h_{11}^t + 4U\Delta t)/(\tilde{p}\tilde{A}_7) \\ h_8^t + U\Delta t \\ h_9^t + U\Delta t + G(h_{10}^t + h_{11}^t + 2U\Delta t)/(\tilde{p}\tilde{A}_9) \\ h_{10}^t + U\Delta t \\ h_{11}^t + U\Delta t \end{pmatrix}. \quad (\text{A.2})$$

519 Let's now take node 2 (the second row in equation A.1) as an example (Figure 1c). Its
 520 receiver is node 1 thus there is the term $-F_2 = -K_f \tilde{p}^m A_2^m \Delta t / \Delta l_2$ in the second row
 521 of the first column of the matrix \underline{B} . The node 2 has several upstream nodes (node 3, node

522 4, node 5, and node 6) which contribute the drainage water, thus, there are dimension-
 523 less terms $G/(\tilde{p}\tilde{A}_2)$ in the the second row of the third to the sixth column of the matrix
 524 $\underline{\underline{B}}$.

The matrix $\underline{\underline{B}}$ is split into its lower $\underline{\underline{F}}$ and strictly upper triangular matrix $\underline{\underline{E}}$ as:

$$\underline{\underline{F}} = \begin{pmatrix} 1 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ -F_2 & 1 + F_2 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & -F_3 & 1 + F_3 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & -F_4 & 1 + F_4 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & -F_5 & 0 & 1 + F_5 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & -F_6 & 0 & 0 & 0 & 1 + F_6 & 0 & 0 & 0 & 0 & 0 \\ -F_7 & 0 & 0 & 0 & 0 & 0 & 1 + F_7 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & -F_8 & 1 + F_8 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & -F_9 & 0 & 1 + F_9 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & -F_{10} & 1 + F_{10} & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & -F_{11} & 0 & 1 + F_{11} \end{pmatrix}, \quad (\text{A.3})$$

and

$$\underline{\underline{E}} = \begin{pmatrix} 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & G/(\tilde{p}\tilde{A}_2) & G/(\tilde{p}\tilde{A}_2) & G/(\tilde{p}\tilde{A}_2) & G/(\tilde{p}\tilde{A}_2) & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & G/(\tilde{p}\tilde{A}_3) & G/(\tilde{p}\tilde{A}_3) & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & G/(\tilde{p}\tilde{A}_7) & G/(\tilde{p}\tilde{A}_7) & G/(\tilde{p}\tilde{A}_7) & G/(\tilde{p}\tilde{A}_7) \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & G/(\tilde{p}\tilde{A}_9) & G/(\tilde{p}\tilde{A}_9) \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \\ 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 & 0 \end{pmatrix}. \quad (\text{A.4})$$

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 529 github.com/fastscap-lem/fastscapelib-fortran.
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