# INDIAN INSTITUTE OF TECHNOLOGY KANPUR

## DEPARTMENT OF EARTH SCIENCES


# TOPIC: COUPLED MODEL OF SEDIMENTOLOGY AND HYDROLOGY


PRESENTED BY:
GROUP 4

- SHEFALI KURIL
- HARSH VERMA
- YASHWANT
- VANSHIKA MEENA

PROF. RAJIV SINHA

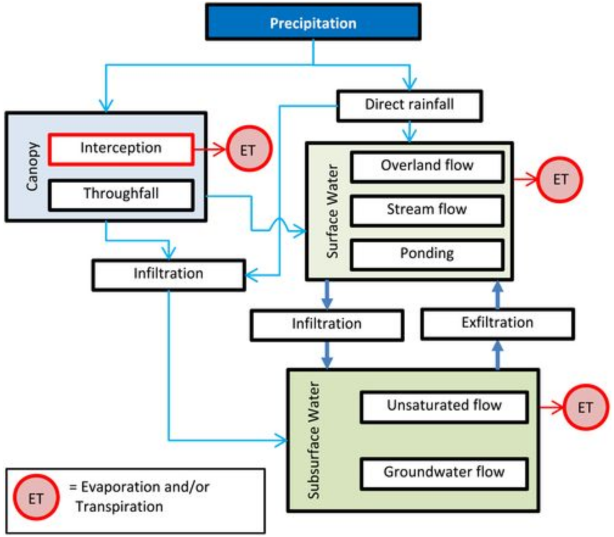
# CONTENTS

# What is a Coupled Model ?

A coupled model of hydrology and sedimentology integrates the processes of water flow and sediment transport, offering valuable insights into various natural and engineered systems.

1. Erosion and Sediment Control in Watersheds
2. Reservoir Management and Sedimentation
3. River Morphodynamics and Channel Evolution

Daniel et.el 2017


# What is a Coupled Model in streambeds

Stream beds are the physical interface between surface water flow and groundwater flow in underlying aquifers.

The topography, permeability and porosity of the streambeds controls water, mass and energy fluxes between the surface water and groundwater

## Hydrology model

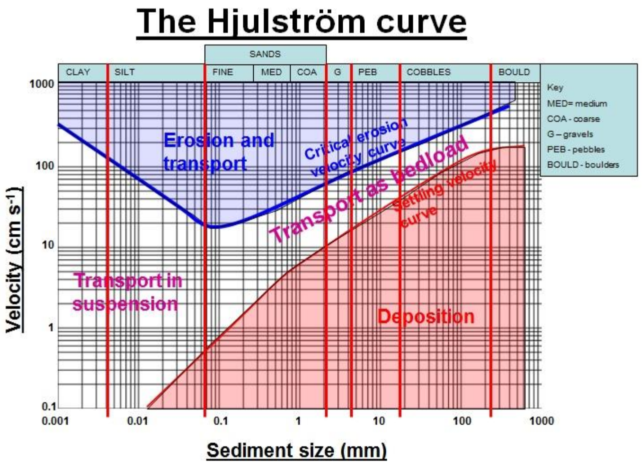
This model aim to simulate flow within catchments by considering both surface and groundwater flow in a physically based way.
Hydraulic head at a given location in a streambed is the sum of hydrostatic and hydrodynamic components.

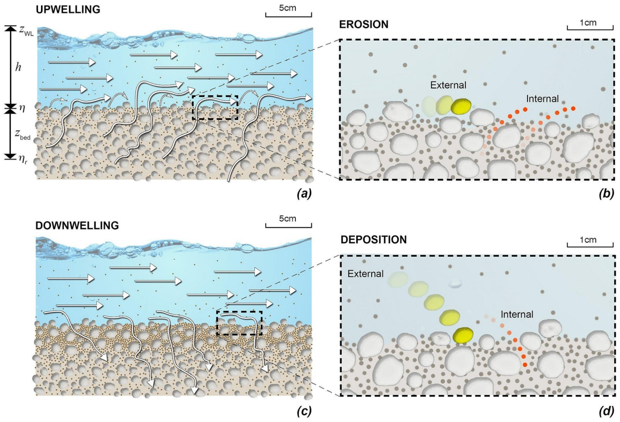
Hydrostatic component consists of sum of elevation and pressure heads whereas the hydrodynamic component is the sum of velocity heads.

Using Darcy's law the exchange fluid flux q is obtained by the product of hydraulic head gradient and the hydraulic conductivity

$$q = K\nabla H$$

**Daniel et.el 2017**




The magnitude of fluid exchange between the surface and subsurface is therefore directly proportional to the hydraulic conductivity of the streambed

$$K = \frac{k\rho g}{\mu}$$

Where k = intrinsic permeability of streambed(L^2), $\rho$ = Density(ML^-3) ,

$\mu$= Dynamic Viscosity(ML^-1T^-1) , g = gravitational constant(LT^-2)

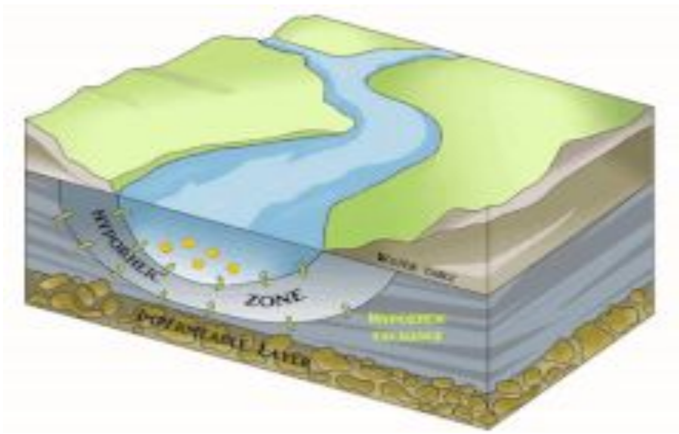
The intrinsic link between the hydraulic properties of sediments and their composition , sedimentological process that modify the composition of a streambed also modify the hydraulic properties at the sediment-water interface.

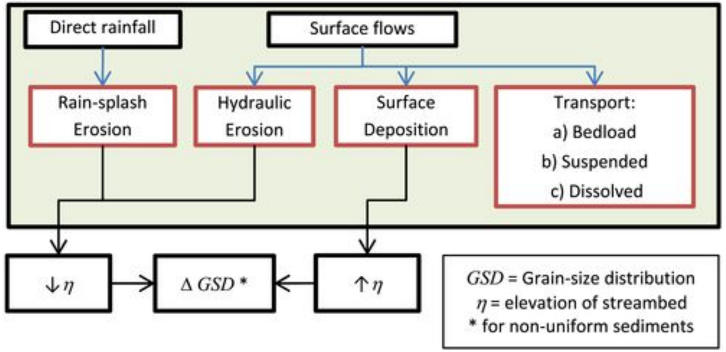
At basic level sediments are deposited during low flow periods and mobilized during high flow events.

During low flow velocity,suspended sediments can settle on the top of the streambed, leading to clogging and reduction of hydraulic conductivity. This can affect streambed topography and the fine sediments that pass through the top layer can also pass deeper into the bed which causes change in streambed composition.

The erosion and deposition is also related to upwelling and downwelling of water. During upwelling Erosion takes place and during downwelling Deposition takes place. Upwelling increases sediment transport and downwelling inhibits sediment transport.

**Daniel et.el 2017**




# Hyporheic zone

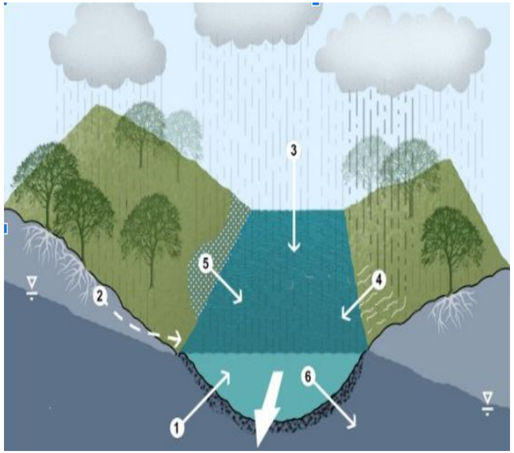
The hyporheic zone is an active ecotone between the surface stream and groundwater.
This interface between surface water and groundwater acts as a key mediator of sediment transport, biogeochemical cycling, and ecological interactions. By integrating the processes of sediment erosion, transport, and deposition with hydrological dynamics, these models capture the intricate exchanges of water and solutes within the hyporheic zone.

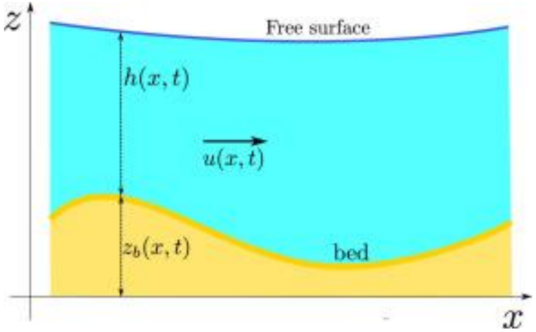
Hyporheic exchange refers to the movement of water between a river or stream and the surrounding sediment or substrate. It occurs in hyporheic zone.
These exchange at meander scale(~10m) can also derive sedimentological change in the subsurface.

# Sedimentological Model

This model focuses on sediment transport process which simulate the mode fluvial geomorphology and sedimentological processes that encompass hydrodynamic flow and sediment transport processes, such as sediment erosion and deposition of the streambed.

**Daniel et.el 2017**



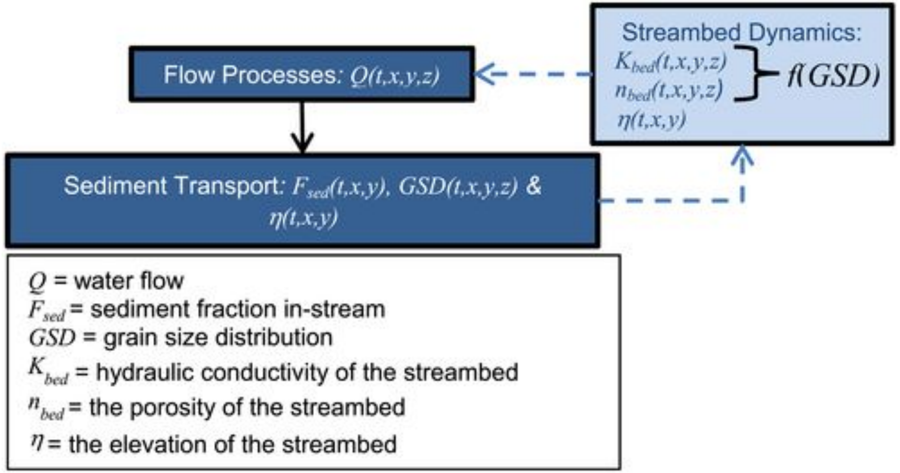

There has been the development of a very wide range of numerical models simulating erosion and deposition processes in and around streambeds. While models differ significantly in terms of processes considered and their conceptualization, the key components of all morphological models of the streambed are a fluid flow model and the sediment continuity equation.

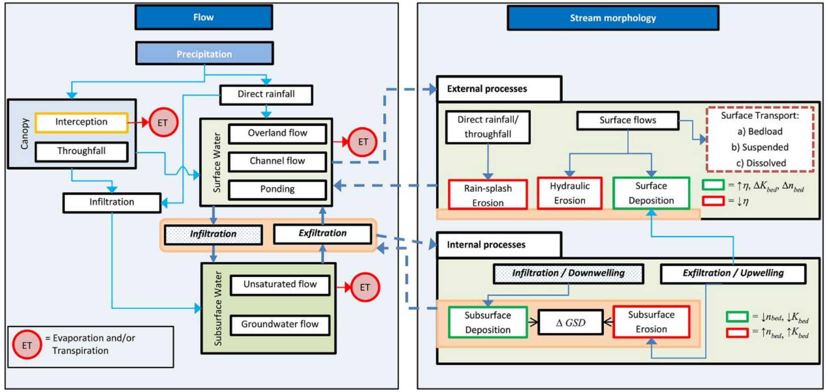
$$\frac{\partial \eta}{\partial t} = -A\frac{\partial U}{\partial x}$$

where η is bed elevation relative to some fixed datum, t is time, A is a coefficient, U is average flow velocity, and x is downstream distance

The majority of sedimentological models simulating reach scale morphodynamic processes use homogeneous sediments." As has been discussed in hydrology model, the hydraulic conductivity of the streambed is related to its grain size distribution. Simulating the transience of hydraulic conductivity, therefore, requires sedimentological models capable of simulating sediment mixtures and classes and their distribution in the streambed. In 2016 a numerical solution of mixed-sediment was introduced which in result give PDEs of hyperbolic solution.

**Daniel et.el 2017**




# Coupled Dynamics

The flow solution accounts for the changing elevation and permeability of the sediment interface, while the sediment transport solution considers how SW-GW interactions control the flow field.

**Flow Model**

The incompressible Navier-Stokes equations are the most common fundamental description of water flow. Solving these equations alone is challenging, but making several assumptions about the nature of surface flows and subsurface flows ,allows description of a surface-subsurface flow model.

Assumptions in the flow model

The model must simulate, on a continuous basis, the important processes and relationships within the system it represents,

The model must be physically relevant to the system that it represents, and the model must be non-unique with respect to both time and space, and applicable over a wide range of hydrologic and geographic conditions.

**Daniel et.el 2017**

# Surface and Subsurface flows

The conceptual model presented here, a suitable flow equation for a stream should account for unsteady flows at the reach or catchment scale, which is achievable with the dynamic wave equations (continuity and momentum eqns); points and times of supercritical flow will exhibit higher velocities and turbulence making them more erosive than subcritical flows and will also affect SW-GW exchange fluxes. Note that flow equations with fewer assumptions (e.g., the Navier-Stokes equations) could be implemented as well, which theoretically would allow simulation of pore-scale processes in the streambed. For flow adjacent to the stream on hillslopes and across plains a much simpler equation may be warranted, such as the diffusion wave or kinematic wave approximation. At the bed form scale, pressure distributions resulting from hydrodynamic pressures induced by undulating bed forms are important for understanding hyporheic flows; hence, the hydrodynamic component of flow cannot be ignored.

$$\frac{\partial h}{\partial t} + \nabla \cdot h\mathbf{q}_{\mathrm{surf}} = h\Gamma_{ex} + \sum Q_{s/s,\mathrm{surf}}$$

$$\frac{Dh\mathbf{q}_{\mathrm{surf}}}{Dt} + gh(\nabla \cdot z_{\mathrm{WL}} - (\mathbf{S} - \mathbf{S}_f)) = 0$$

where h is the height of water above the surface (L), t is time (T), qsurf is the surface flow flux (LT−1), Γex is the exchange between the surface and subsurface (T−1), Qs/s,surf (LT−1) are source/sink terms (which can be positive or negative), zWL is the location of the water surface relative to a fixed datum (L), g is the gravitational constant (LT−2), S is the slope vector (Sx,Sy), where Sx and Sy are the slopes of the surface in the x and y directions respectively (LL−1), and Sf is the friction slope vector (Sfx, Sfy), where Sfx and Sfy are the frictional slopes in the x and y directions respectively (LL−1).

Daniel et.el 2017

Three-dimensional saturated and unsaturated flow through anisotropic heterogeneous porous media can be described using the Richards equation as follows:

$$-\nabla\cdot\mathbf{q}_{\mathrm{sub}} + \Gamma_{ex} + \sum Q_{s/s} = \frac{\partial}{\partial t}(\theta_s S_w)$$

where qsub is the subsurface fluid flux (LT-1), Γex is the exchange between the surface and subsurface, Qs/s are source/sink terms (which can be positive or negative) (L3L−3T−1), θs is the saturated water content (L3L−3), and Sw is the water saturation of the porous media related to the water content θ (Sw = θ/θs).The fluid flux qsub is defined as

$$\mathbf{q}_{\mathrm{sub}} = -\mathbf{K}\cdot k_r \nabla(\psi + z)$$

where K is the hydraulic conductivity tensor (LT−1), kr is the relative permeability (−), ψ is the pressure head (L), and z is the elevation above the datum (L). The hydraulic conductivity tensor is defined as

$$\mathbf{K} = \frac{\rho g}{\mu}\mathbf{k}$$

where ρ is fluid density (ML−3), g is the gravitational constant (LT−2), μ is the dynamic viscosity (ML−1T−1), and k is the permeability tensor (L2).

Constitutive relationships are defined that relate pressure head to the saturation (Sw) and relative permeability (kr) of the porous media.

Subsurface flow can also take place in other features such as fractures and macropores as well as through anthropogenic features such as tile drains. While of importance in many catchments, these flow paths are beyond the focus of this particular review.

**Daniel et.el 2017**

# Sediment Transport and Sedimentological Model

$$\text{Stream}: \frac{\partial}{\partial t}\int_{\eta}^{z_{WL}} \alpha_f dz + \vec{\nabla}_H \vec{\phi}_f + \Omega_{out}(\eta) - \Omega_{in}(\eta + z_{WL}) - \int_{\eta}^{z_{WL}} \Gamma_f dz = 0$$

$$\text{Subsurface}: \int_{\eta_r}^{\eta} \frac{\partial \alpha_s}{\partial t} dz + \alpha_s(\eta)\frac{\partial \eta}{\partial t} - \alpha_s(\eta_r)\frac{\partial \eta_r}{\partial t} + \vec{\nabla}_H \vec{\phi}_s + \Omega_{out}(\eta_r) - \Omega_{in}(\eta) - \int_{\eta_r}^{\eta} \Gamma_s dz = 0$$

where zWL is the elevation of the stream surface relative to the datum (L), η is the streambed elevation relative to the datum (L), ηr is the elevation of bottom of the streambed sediments (L), αf and αs are the sediment densities in the stream and subsurface respectively (ML−3), is the horizontal divergence operator , ϕf and ϕs are the horizontal vector mass sediment fluxes per unit width in the stream (i.e., bed load, suspended load and wash load) and subsurface respectively (ML−1T−1), Ωout(η) is the sediment exchange flux per unit area over the interface between the stream and subsurface (ML−2T−1), Ωin(zWL) is the sediment exchange flux per unit area of the water surface boundary (ML−2T−1), Ωout(ηr) is the sediment exchange flux per unit area of the bedrock interface (ML−2T−1), and Γf and Γs are the rate of sediment production or destruction in the stream and subsurface, respectively (ML−3T−1). The elevation of the bottom of the streambed sediments ηr is assumed to correspond to the lowermost elevation where sedimentological processes occur. For example, it could be the interface between the more permeable streambed sediments and a less permeable geological unit, such as bedrock.

**Daniel et.el 2017**

# Assumptions

There is no production or destruction of sediments in the stream or subsurface;

The elevation of the bottom of the streambed $\eta r$ is assumed to be constant because, for the timescales considered, sedimentological processes are assumed to be very slow and negligible below that elevation; therefore, the third and fifth terms in equation 8 can be removed.

Because of the relatively slow sedimentological processes, horizontal sediment flux in the subsurface relative to the stream horizontal sediment flux can be ignored; thus, the fourth term from equation 8 can be removed.

Using the average sediment density in the stream and an average sediment density in the subsurface, the first terms of equations 7 and 8 are simplified.

Because of averaging the sediment density, only a small thickness (i.e., the streambed) is considered.

**Daniel et.el 2017**

# Modified Equations

$$\text{Stream}: \quad \frac{\partial h \bar{\alpha}_f}{\partial t} = -\vec{\nabla}_H \vec{\phi}_f + \Omega_{\text{in}}(z_{\text{WL}}) - \Omega_{\text{exch}}$$

$$\text{Streambed}: \quad \frac{\partial z_{\text{bed}} \bar{\alpha}_s}{\partial t} = -\bar{\alpha}_s \frac{\partial \eta}{\partial t} + \Omega_{\text{exch}}$$

where alpha f are the averaged sediment density in the stream (of thickness h) and streambed (of thickness zbed), respectively.

# Sediment Erosion and Deposition

A key component of the proposed model is in the partitioning of sediment erosion/depositional processes occurring on the bed surface and from within the bed, and there are still challenges ahead therein. This is a very complex process that is dependent on (1) the nature of the SW-GW exchange (upwelling or downwelling) and (2) the composition and structure of the streambed, in particular, the grains size distribution, pore geometry (cross-sectional area along the direction of flow), and pore volume. In the case of downwelling, the pore geometry and volume will control the capacity of the streambed to internally capture sediments, with only stream sediments smaller than the pore openings able to enter the streambed. In the case of upwelling, the flow rate and internal critical shear stress for each of the sediment classes will determine the transport of sediments out of the streambed. As upwelling flow rates are normally very small compared to stream flow, it follows that the shear stress on the sediments generated in the pore spaces is also very low. Hence, transport of sediments out of the streambed due to seepage would only occur for very fine sediments [Martin, 1970]. This upwelling will also be of importance for the deposition of very fine sediments, as upwelling will prevent deposition of the fines when the velocity of upwelling pore water exceeds the fall velocity of the sediment.

**Daniel et.el 2017**

# Erosion and Sediment control in watersheds

A coupled model in sedimentology and hydrology integrates both sediment transport processes and hydrological processes to simulate how water flow influences the movement of sediment particles. These models are crucial for understanding erosion, deposition, and morphological changes in rivers, streams, estuaries, coastal areas, and other water bodies. Here's how such a model can be applied:

1. Hydrological component
2. Sediment Transport Component
3. Coupling Mechanism
4. Numerical Methods
5. Validation and Calibration

Salheddine et.el 2020

# Specificities of the coupled hydro-sedimentary model

1. Coupled Model Components.
2. Software Used.
3. Sediment Class Parameters
4. Numerical Scheme.
5. Sediment Distribution
6. Hydraulic Structures
7. Exchange Terms
8. Stability

Salheddine et.el 2020

# Coupling Principles

1. There is no overlap between the 1-D and the 2-D domains, which means that the boundary between the two domains is constituted of lines and/or areas where the exchange terms are calculated.
2. The exchange terms are the fluxes (water and sediment) through the boundaries between the areas dedicated to each model.
3. The calculation of these fluxes should be performed taking into account the variables of both models.
4. The remaining difficulty lies in defining a calculation method of these fluxes, which should take into account the physical processes at the boundary.

Salheddine et.el 2020

# Possibilities of Coupling

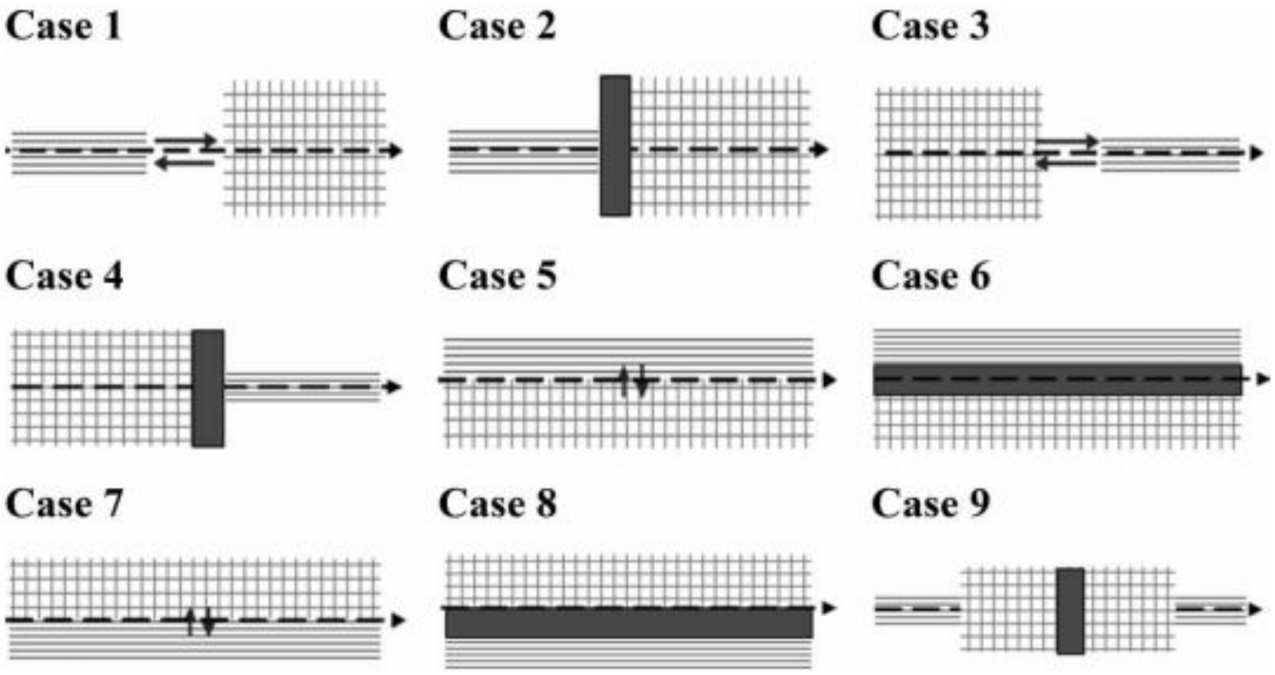



**Figure 1** | Sketch of definition of the coupling cases. ( ) 2-D area, ( ) link with shallow waters, ( ) 1-D reach, ( ) link with hydraulic structure, ( ) flow direction.

Salheddine et.el 2020

# Case Study: Agly River Flood

1. The overtopping of Agly levees during the November 1999 flood in the southern part of France led to the flooding of a wide coastal plain, quite flat, in which flow was constrained by many embankments.
2. Overtopping.

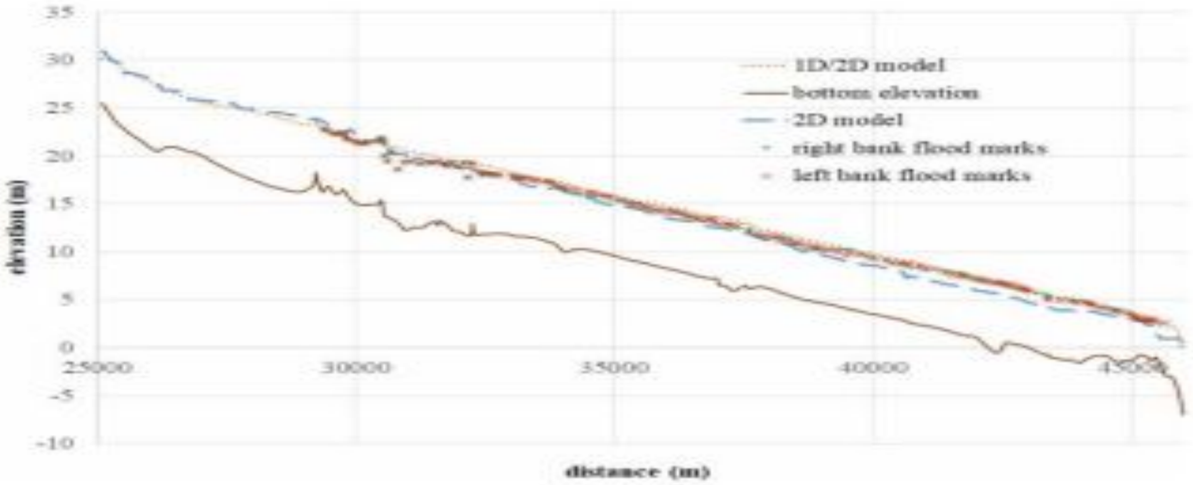



Fig: 1999 Agly flood. Peak water elevation along downstream main ch

Salheddine et.el 2020

# Case Study: Agly River Flood

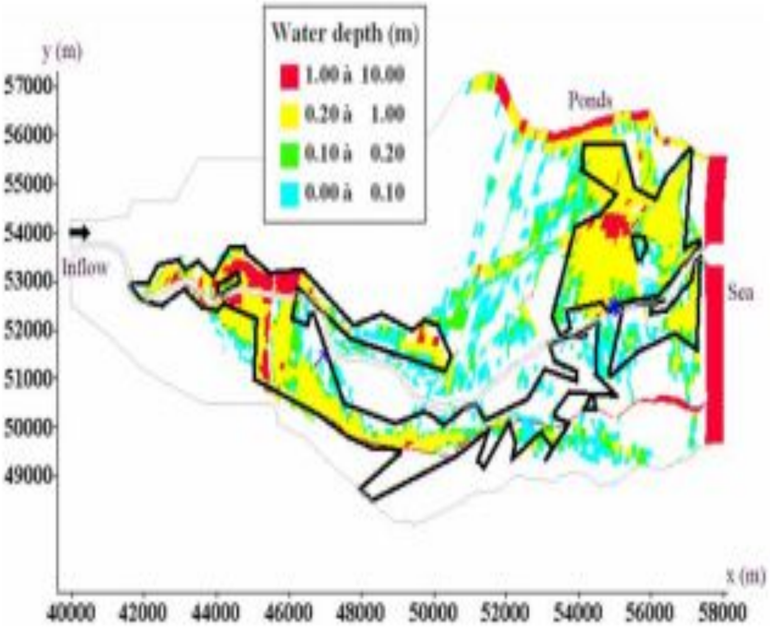



Fig: Agly floodplain. Peak water depths for the 1999 flood (coupled model results). The thick black line (respectively the star) shows the limit of observed flooded area

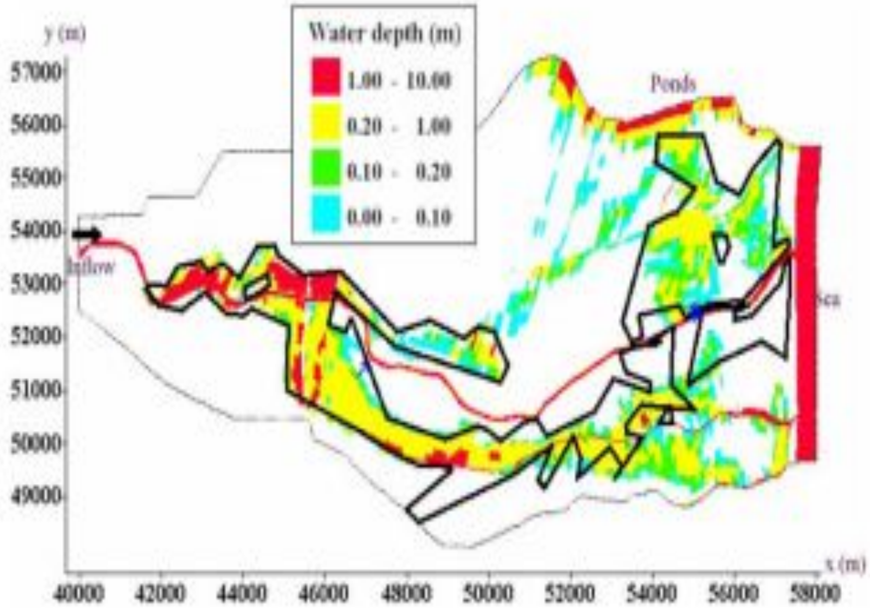



Fig: Agly floodplain. Peak water depths for the 1999 flood (2-D model results). The thick black line (respectively the star) shows the limit of observed flooded area

Salheddine et.el 2020

# Case Study: Agly River Flood

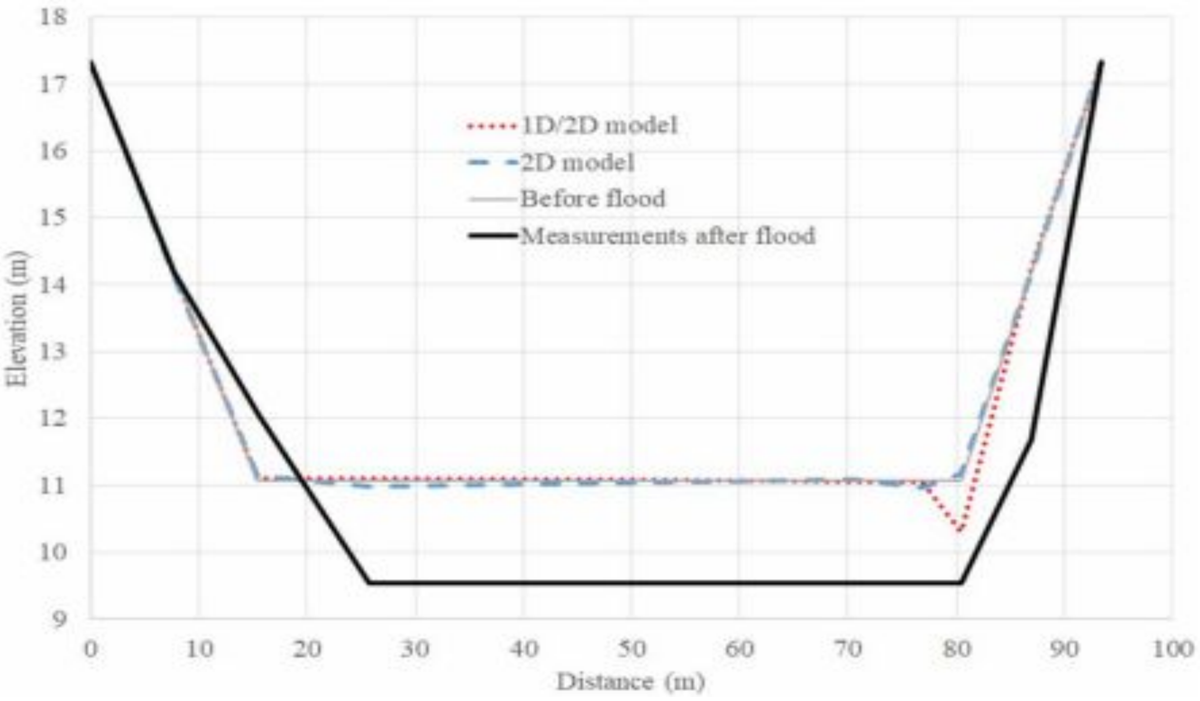
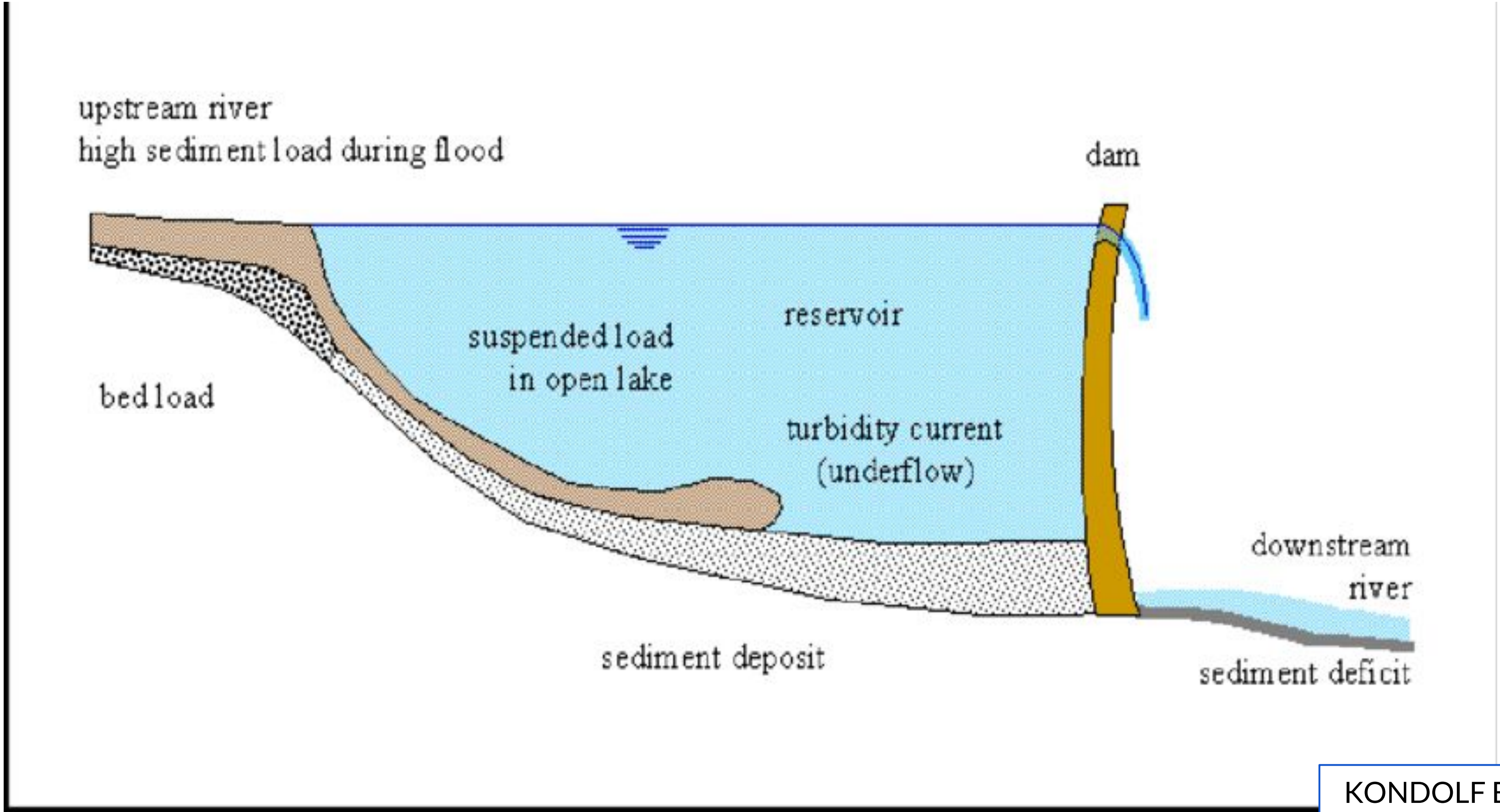



Fig: Evolution of the cross section of Agly main channel at distance 33,750 m during 1999 flood.

Salheddine et.el 2020

# Reservoir Management and Sedimentation




KONDOLF ET AL 2014

- Sedimentation is a crucial problem.
- It causes reduced service life of the reservoir due to the large sediment yield inside the reservoir.
- Extreme floods brought a massive amount of sediments.

Based on the existing data on reservoir sedimentation trapping rates-

Global annual loss of storage capacity of the reservoirs was around 1%.
The global average sedimentation rate - 0.8% per year.
In India, reported the annual average sedimentation rate - 1.41%.
In China, The annual average loss rate of reservoir capacity - 2.3% .

The mean sedimentation rate of reservoirs worldwide was 0.94%

The sedimentation is so large that it would decrease the capacity storage of the reservoir and shorten the length of operation.

KONDOLF ET AL 2014

# Analysis of Sedimentation in Wonogiri Reservoir

**Location** - on the Upper Solo River basin in Central Java Province, Indonesia.

Sediments inlets-

1. Keduang
2. Tirtomoyo
3. Temon
4. Alang
5. Wuryantoro
6. Solo

- Total storage = 730 million cubic meters
- Catchment area = 1,260 square kilometers
- Established in 1981

Santosa et.el 2016

The suspended load discharge can be calculated as-

$$Q_s = a \times Q_w{}^b$$

Qs - suspended sediment discharge (ton/days)

Qw - stream discharge (m3 /s)

a , b - constants, which its value is depend on the measurement data on field.

Bed load was computed as the percentage of the suspended load

$$\theta_c = \frac{\frac{1}{2} f_c U_c^2}{(s-1) g d_{50}}$$

$f_c$ - friction factor (dimensionless)
$U_c$ - current velocity
g - acceleration due to gravity
s- ratio of densities of sand and water
$d_{50}$ - median diameter

Total sediment = Sum of both suspended load and bed load

Santosa et.el 2016

| No | Watershed Name | Area (km$^2$) | Suspended sediment load discharge 85% (ton/year) | Bed Load Sediment Discharge (ton/year) | Sediment Load Total Discharge (ton/year) |
|---|---|---|---|---|---|
| 1 | Keduang | 426 | 615,028.90 | 180,890.24 | 795,917.17 |
| 2 | Tirtomoyo | 206 | 2,458,694.28 | 723,145.25 | 3,181,839.65 |
| 3 | Temon | 69 | 34,746.60 | 10,219.54 | 44,966.19 |
| 4 | Upstream reach of B. Solo | 200 | 98,527.55 | 28,978.74 | 127,506.24 |
| 5 | Alang | 235 | 198,202.56 | 58,294.97 | 256,497.43 |
| 6 | Wuryantoro | 73 | 259,299.81 | 76,264.60 | 335,564.46 |
| 7 | Others | 51 | 148,324.89 | 43,624.97 | 191,949.86 |
| | Total | 1,260 | 3,812,822.58 | 1,121,418.29 | 4,934,240.99 |

- The annual volume of the sediment that entered the reservoir is 6.68 million m3/year.

➢ Analyse annual sedimentation

reservoir inflow
sediment transport
sediment deposition

Key aspects of coupled sedimentology and hydrology

Santosa et.el 2016

# River Morphodynamics and Channel Evolution

- River morphodynamics is all about studying how rivers change over time because of water and sediment interaction.
- It looks at how water moves, carries sediment, and shapes the riverbed.
- Traditionally, people used Newton's laws of motion to understand rivers, focusing on when things are stable and the flow is steady. But this might miss out on how rivers change over time and behave differently in different parts.
- River morphodynamics is a complicated field because it involves lots of different processes, like how sediment moves during big events, how sandbars form and change, and how plants on the riverbanks affect things.
- These processes happen over different time periods, from quick changes to slow ones.
- Understanding river morphodynamics helps us predict what rivers might do in the future and take care of them better, like reducing flood risks and keeping river habitats healthy for people and wildlife.

Church et.el 2015

# Numerical Model

Nicholas et.el 2013

➢ **<u>HSTAR (Hydrodynamics and Sediment Transport in Alluvial Rivers)</u>** is a computer program that can mimic both braided and meandering river types.

The new morphodynamic model for <u>simulating river and floodplain co-evolution</u> has several key features:

1. Capable of simulating a wide range of river styles, including braided, anabranching, and high sinuosity meandering channels.
2. Utilizes a simple grid structure to show how the river channel moves and changes over time. It can simulate things like the river changing its course, forming new channels, or breaking off from the main river without needing to constantly adjust the grid.
3. Includes a bank erosion scheme i.e When the river erodes the banks of the channel, this model limits how much the shape of the riverbank changes over time. It helps to keep the river's shape somewhat stable even as it moves.
4. Incorporates a momentum conserving hydrodynamic model i.e how water moves in the river, considering things like the swirling motion of water (called secondary circulation) and how that affects the movement of sediment (like sand or silt) along the riverbed.
5. Accounts for at least two sediment size fractions associated with channel and overbank deposits.
6. Features a simple treatment of vegetation that represents active channel conversion to floodplain and the tendency for plants to enhance surface stability.

➢ This numerical modeling framework integrates hydrodynamics, sediment transport, bank erosion, floodplain dynamics, and morphodynamic scaling to simulate river and floodplain co-evolution over long timescales.

1)NUMERICAL MODELLING APPROACH:

HSTAR model solves the depth-averaged shallow water form of the Navier–Stokes equations.
It includes equations for hydrodynamics and sediment transport.
The hydrodynamic equations are solved using an explicit time integration method.
The model has been validated using data from the Rio Paraná, Argentina.

$$\frac{\partial h}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = 0$$

$$q_T = \frac{0.05 f \chi |U|^5}{(1-\delta)\sqrt{g} C^3 \psi^2 D}$$

|U| = depth-averaged velocity magnitude,
Ψ = sediment relative density,
D = sand diameter
f = sand fraction in the surface bed layer.
Delta = bed porosity,
Xi is a function of the local bedslope

3)FLOODPLAIN DYNAMICS:

Grid cells = active channel bed or vegetated floodplain surfaces.
Vegetation colonization converts active channel cells to floodplain cells
Floodplain reworking occurs mainly by lateral bank erosion.
changes in bed elevation and grain size composition are determined using the Exner mass balance relation.

2)SEDIMENT TRANSPORT MODELLING:

Sediment transport is modeled for sand and silt fractions.
sand transport rates are calculated using the Engelund–Hansen relation
The direction of sand transport deviates from the mean flow direction due to secondary circulation effects.
Silt transport is represented by solving a two-dimensional advection-diffusion equation.

$$\frac{\partial(h\varphi)}{\partial t} + \frac{\partial(q_x\varphi)}{\partial x} + \frac{\partial(q_y\varphi)}{\partial y} - \frac{\partial}{\partial x}\left(D_H h \frac{\partial\varphi}{\partial x}\right) - \frac{\partial}{\partial y}\left(D_H h \frac{\partial\phi}{\partial y}\right) + D_R - E_B = 0$$

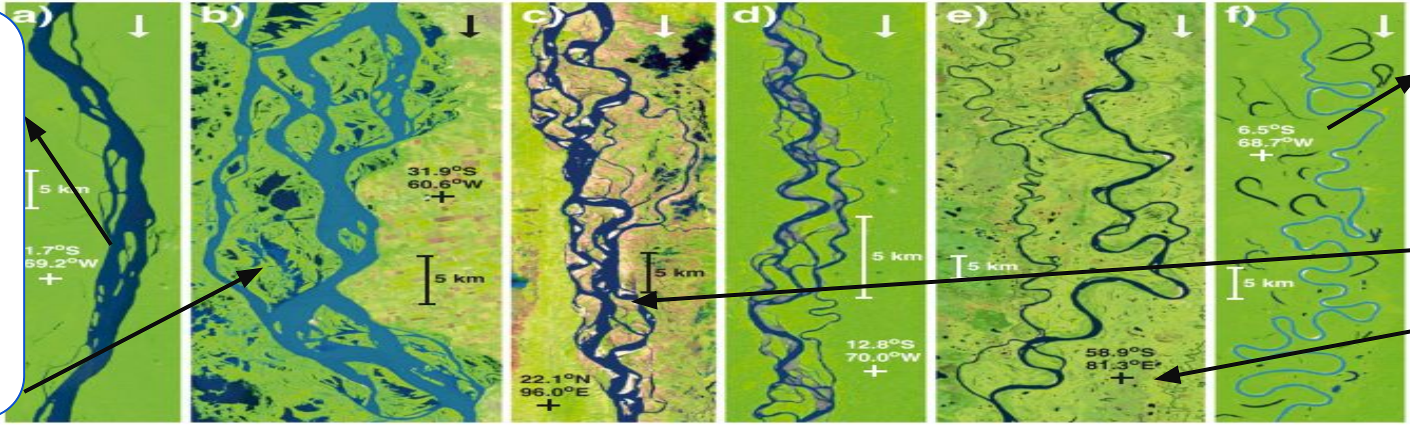
Phi = depth-averaged silt concentration,
$D_H$ = diffusion coefficient,
$D_R$ = source term representing the net deposition rate, which includes the effects of particle setting and remobilization,
$E_B$ = sediment supply by bank erosion

Nicholas et.el 2013

4)DECOUPLING OF HYDRODYNAMIC AND MORPHODYNAMIC MODEL TIME STEPS:

Sediment fluxes are multiplied by a constant morphological scaling factor to allow simulation of river evolution over long timescales.
Initial conditions include a straight channel with small elevation perturbations and sinewave hydrographs for inflow conditions.

**Low Sinuosity Anabranching Channels** dominated by relatively stable vegetated islands




**High Sinuosity Single-thread Meandering Channels**

Variety of **multi-thread rivers**

Fig:Large river channel patterns found in environments in which vegetation is more abundant.
Examples of large river channel patterns: (a) Rio Japurá, Brazil; (b) Rio Paraná, Argentina; (c) Irrawaddy River, Burma; (d) Rio Inambari,Peru; (e) River Ob, Russia; (f) Rio Juruá, Brazil

Nicholas et.el 2013

**Table I.** Key parameter values used in model simulations shown in Figures 1, 3 to 5, and 8.

| Figure | $\Delta x$ (m), $\Delta y$ (m) | $Q_{min}$ ($m^3 s^{-1}$) | $Q_{max}$ ($m^3 s^{-1}$) | $S$ | $D$ (mm) | $C$ ($m^{0.5} s^{-1}$) | $T_{veg}$ (yr) | $H_{cr}$ (m) | $E$ | $L$ (mg $l^{-1}$) |
|---|---|---|---|---|---|---|---|---|---|---|
| 1 | 60, 30 | 10,000 | 30,000 | 0·00005 | 0·4 | 55 | 10 | 0·1 | 10 | 150 |
| 3a | 80, 40 | 10,000 | 30,000 | 0·00005 | 0·4 | 55 | 6 | 0·3 | 1 | 450 |
| 3b | 80, 40 | 10,000 | 30,000 | 0·00005 | 0·4 | 55 | 6 | 0·3 | 10 | 450 |
| 3c | 80, 40 | 10,000 | 30,000 | 0·0001 | 0·2 | 40 | 6 | 0·3 | 10 | 450 |
| 3d | 80, 40 | 10,000 | 30,000 | 0·0001 | 0·2 | 40 | 10 | 0·1 | 10 | 150 |
| 4 | 80, 40 | 10,000 | 30,000 | 0·0001 | 0·3 | 40 | 6 | 0·3 | 10 | 450 |
| 5 | 60, 60 | 10,000 | 30,000 | 0·0001 | 0·3 | 47·5 | 1 | 0·3 | 2·5 | 450 |
| 8 | 10, 5 | 100 | 3,000 | 0·005 | 35 | — | — | — | 0·5 | — |

S = initial channel slope; D = bed sediment particle diameter; C = chezy roughness coefficient; E = dimensionless bank erodibility ;
L = inlet silt concentration;

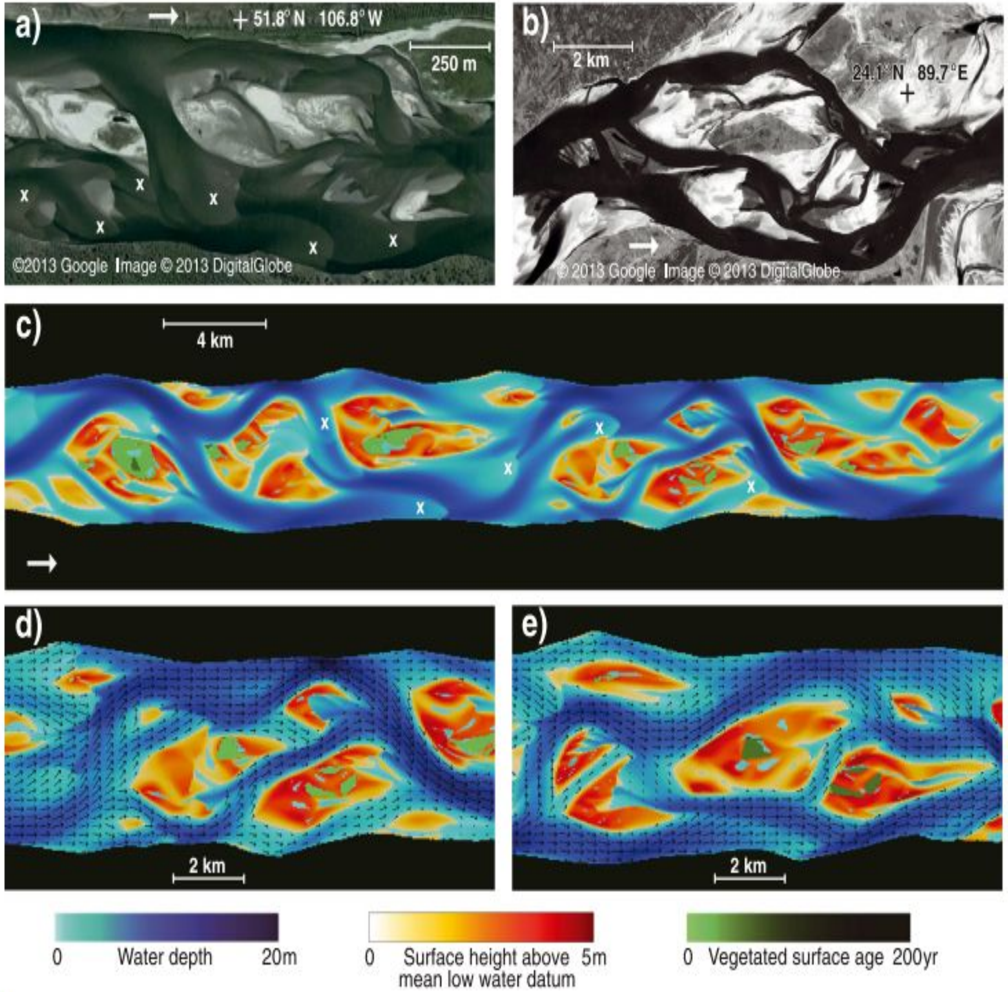



**Fig:<u>Natural and modelled Braided sand-bed rivers</u>:**

(a) South Saskatchewan River, Canada;

(b) Jamuna River, Bangladesh;

(c) simulated morphology after 150 years of channel evolution

- **Vegetation Growth**: Slow plant growth in the simulation makes the river <u>more dynamic</u>, forming lobate bedforms (unit bars) that move downstream and create mid-channel bars.
- **Compound Bars**: These are bars made up of smaller bars. They have a V-shape and limbs that extend downstream. Older compound bars include abandoned channels and quiet water areas.
- **Flow Patterns**: Bars cause the flow of water to <u>split (diverge)</u> at the front of the bar and come together <u>(converge)</u> downstream. This creates variations in how much sediment the water can carry, affecting where new bars form.
- **Constant Changes**: Because of these interactions, the river's shape is always changing, which makes it hard for plants to grow and stabilize the riverbanks.
- **Model challenges**: they need to carefully control how they simulate the river's beginning (inlet) to match what happens naturally.

Nicholas et. el 2013

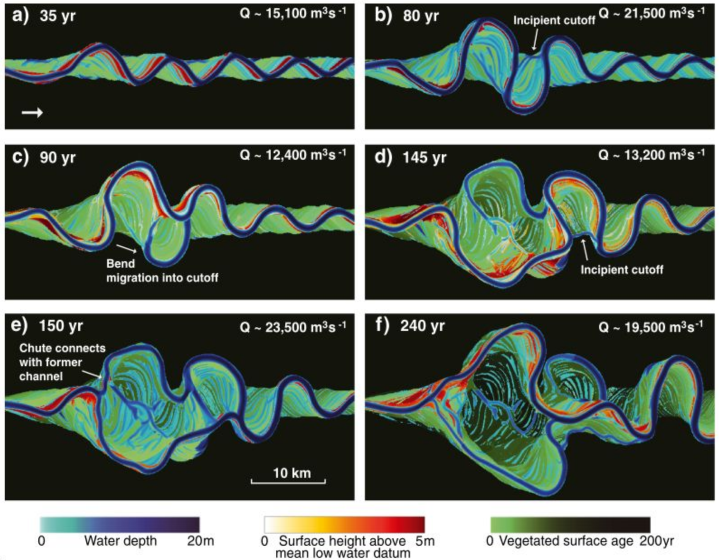


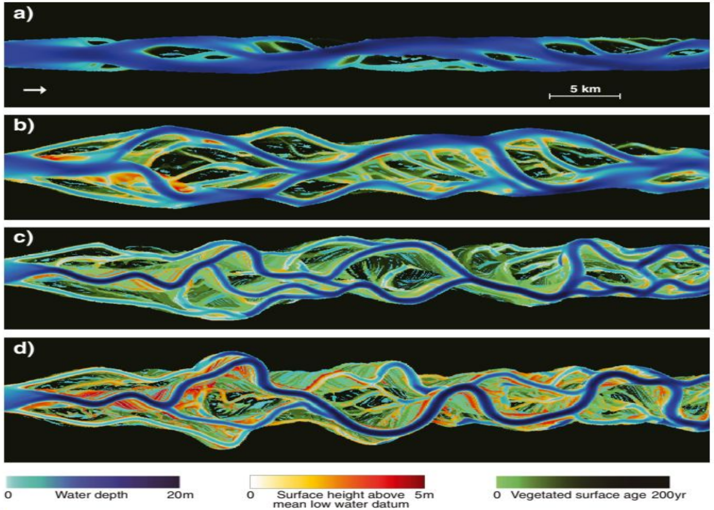

**Fig:Modelled channel morphology at six points in time over a period of 240 years during a simulation of Meandering river evolution**

- Bifurcation and Meander Migration: Bends shifting in rivers alter the alignment of split channels.
- Changing Channel Orientation: Bend movements affect how water and sediment distribute at channel splits.
- Abandonment of Bifurcated Channels: Channels may be deserted if they deviate too much from the main flow.
- Bend Migration and Point Bars: Bends erode outer banks, reducing their height and potentially increasing overbank flow, leading to bar breakdown.

**Fig:Simulated channel morphology for contrasting Anabranching rivers after > 300 years of modelled channel evolution**

- Rapid vegetation growth and strong banks result in straight channels with minimal changes (Figure a).
- Erodible banks lead to more branching in channels, forming elongated islands (Figure b).
- Fine grains and high downstream gradients create rivers with many bends and splits (Figures c and d).

Nicholas et.el 2013



# Results and Discussion:

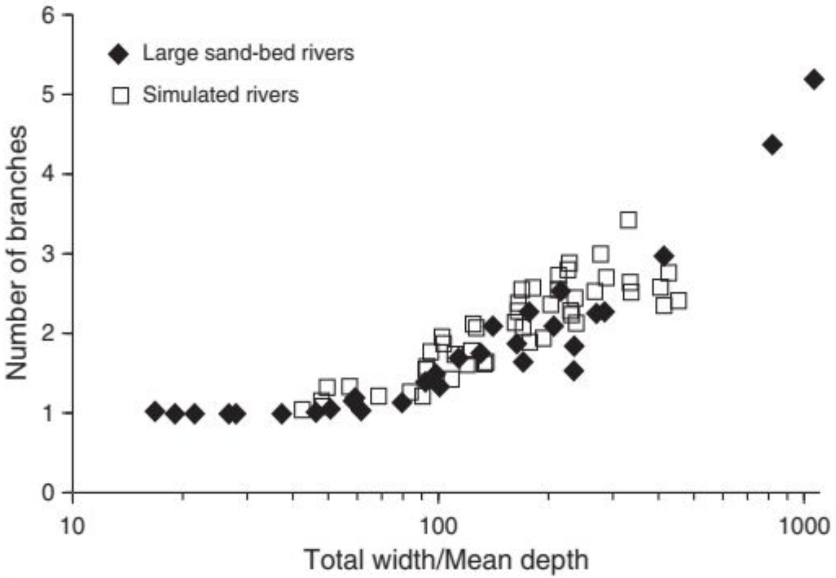



Fig:Relationship between mean number of channel branches at a cross-section and the ratio of total water surface width to mean flow depth for simulated and natural sand-bed rivers

- Preliminary evaluations indicate that the model accurately captures many features of natural rivers, such as bar geometry and meander characteristic.
- Using a momentum-conserving approach helps the model deal with size problems and makes the river shape look realistic.
- Channel pattern dynamics are strongly influenced by coupling mechanisms between the channel and floodplain, affecting channel abandonment, reactivation, and overbank sedimentation.
- Vegetation plays a significant role in stabilizing bar surfaces, limiting vertical erosion, and promoting floodplain construction, shaping river behavior despite simplified modeling approaches.
- There are still difficulties in making the model show all the complicated ways rivers change, like how the banks erode and how sediment moves, so more work is needed to make it better.

Nicholas et.el 2013

# REFERENCES


- Coupled model of surface water flow, sediment transport and morphological evolution
- A coupled 1-D/2-D model for simulating river sediment transport and bed evolution
- Coupled modeling of rainfall-induced floods and sediment transport at the catchment scale
- Integrated sediment transport process modeling by coupling Soil and Water Assessment Tool and Environmental Fluid Dynamics Code
- A generalized Exner equation for sediment mass balance
- Blueprint for a coupled model of sedimentology, hydrology, and hydrogeology in streambeds
- Modelling the continuum of river channel patterns
- https://agupubs.onlinelibrary.wiley.com/doi/full/10.1002/2014WR016862

# THANK YOU