



Long-term hydro-sediment dynamics of the Ucayali River (Amazon Basin) revealed through combined observations, remote sensing, and SWAT-Amazon modelling

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Abstract. Since the early 1970s, the Amazon basin has experienced growing local and global changes, potentially reaching a climatic tipping point in the coming decades. However, due to cost constraints and limited access, conventional hydrological networks in the basin struggle to provide the spatial resolution and temporal extent required for accurate quantification of water and sediment budgets, which are essential for understanding biogeochemical cycles.

Focusing on the Ucayali River, a major Amazonian foreland tributary, this study provides the first long-term hydro-sediment balances in this region at sub-basin scale, distinguishing fine sediments from sand loads (37 years for water and sands, 20 years for fine sediments). It is achieved by the integration of remote sensing and hydrological-hydraulic modelling using a modified SWAT model, SWAT-Amazon. A new hydraulic module for water routing was implemented in SWAT-Amazon to suit the Amazon diffusive flood wave, representing floodplains as reservoirs. Fine sediment loads were estimated using satellite-derived concentrations and simulated discharges, while suspended sand loads were simulated within SWAT-Amazon. Results indicate that the Andean Ucayali River exports $455 \cdot 10^6 \text{ t yr}^{-1}$ of suspended sediment (40% sand). As the floodplain traps 36% of the Andean sediments (65% sand), mostly by tectonic subsidence, the Ucayali delivers $290 \cdot 10^6 \text{ t yr}^{-1}$ of total suspended sediment to the Amazon River, 26% as sand. Floodplain recycling plays a crucial role as a secondary sediment source (22% of the Ucayali load), with a water storage that peaks at 19.1 km^3 in March (38% of discharge). A previously undocumented sand sedimentation process is identified during the flooding period, capturing 14% of the sand flux at peak discharge and thus decorrelating sediment transport from water discharge. No significant long-term trends in flood duration, discharge, or sediment fluxes were detected, suggesting contrasted evolution patterns of the precipitations in the basin due to its particular position in the Amazon Basin. This study emphasizes the need to rethink hydrological network management with robust and long-term conventional data at ‘super’ stations to support the calibration of remote sensing and modelling at ‘virtual’ stations. Extending this approach to other Amazonian basins could significantly enhance hydro-sediment and biogeochemical cycle research in large river systems. Additionally, it highlights the importance of regionally focused over large-scale assessments, which often carry high uncertainties and may mislead mitigation strategies.



1 Introduction

1.1 Global contribution of the Amazon Basin

35 The Amazon basin is a massive hotspot for water and matter inputs to the Ocean (Syvitski et al., 2005; Martinez et al., 2009; Moquet et al., 2016; Jouanno et al., 2021; Louchard et al., 2021, 2023) and plays a key role in global hydro-biogeochemical cycles (Gaillardet *et al.*, 1999), capable to significantly impact oceanic biogeochemistry (Jouanno et al., 2021; Louchard et al., 2021). Long-term monitoring by the CZO (Critical Zone Observatory) HyBAM (Hydrology of the Amazon Basin) shows that the Amazon River annually discharges 6,500 km³ of freshwater (~20–25% of the global total) (Callède et al., 2010), 1,100 10⁶ t of suspended sediments (~8% of global riverine outputs) (Santini, 2020) and 272 10⁶ t of dissolved matter (Moquet et al., 2016) (~7% of the global flux). It also influences atmospheric circulation, contributing up to 15% of global continental evapotranspiration (Salati, 1979; Soares-Filho et al., 2010; Satyamurty et al., 2013) and It acts as both a carbon sink and a greenhouse gas source, contributing substantially to global cycles (Richey et al., 2002; Melack et al., 2004; Subramaniam et al., 2008; Ward et al., 2016; Pangala et al., 2017; Louchard et al., 2021). Nutrient-rich from the Andean Cordillera, the Amazon 45 hosts 25% of terrestrial species and the Earth's largest rainforest (e.g. Lesack, 1993; Malhi et al., 2008; Fan and Miguez-Macho, 2010).

1.2 Role of the floodplain dynamics

Amazonian floodplains act as dynamic reactors, playing a key role in global water and sediment fluxes. Lateral exchanges between the main channel and alluvial plains are of the same order of magnitude as the fluxes reaching the ocean (Meade et al., 1985; Mertes et al., 1996; Dunne et al., 1998) and dominate the annual floodplain water balance (Rudorff et al., 2014a, b). 50 Around 30% of peak discharge transits through the floodplain (Richey et al., 1989; Lininger and Latrubesse, 2016), with highly variable pathways and residence times (from seconds to months). The flooded area covers 8–10% of the basin (5–6 10⁵ km²) (Fleischmann et al., 2022). Thus, flood dynamics regulate the storage and exchange of sediments, nutrients, organic matter, pollutants, and living organisms (Aufdenkampe et al., 2011; Lewin et al., 2017). Sediment residence time varies from brief 55 periods to tens of thousands of years (Mertes et al., 1996; Allen, 2008), depending on the floodplain's geomorphology, influencing chemical maturation processes essential to global biogeochemical cycles, including CO₂ consumption by silicate weathering (Gaillardet et al., 1999; Guyot et al., 2007; Bouchez et al., 2012). Flexural basins adjacent to the Eastern Cordillera trap 40–50% of Andean sediment exports (Guyot, 1993; Baby and Guyot, 2009; Armijos et al., 2013; Santini et al., 2014; Vauchel et al., 2017; Santini, 2020). Further downstream, sediment balances tend toward equilibrium between deposition and 60 resuspension during flood recession (Espinoza-Villar et al., 2017), though influenced by the basin's structural Arches. Downstream of the Amazon-Madeira confluence, floodplains have remained only partially filled since the last glacio-eustatic lowstand (~125 m below present sea level) and act as fine-sediment sinks, where large, shallow lakes retain overbank floodwaters (Tricart, 1977; Fleming et al., 1998; Park and Latrubesse, 2017).



65 1.3 The impacts of global and local changes

The Amazon Basin is undergoing a dramatic transition (Walling, 2006; Malhi et al., 2008; Davidson et al., 2020), facing pressures from deforestation for agriculture and pasture, resources extraction, and construction of hydroelectric projects (Finer and Jenkins, 2012; Latrubesse et al., 2017; Timpe and Kaplan, 2017; Chaudhari and Pokhrel, 2022). In recent decades, the Amazon Basin has also been affected by global climate changes, experiencing more frequent extreme floods (e.g., in 2009, 2012, 2014, 2015) and severe droughts (e.g. in 2005, 2010, 2023, 2024), with an increase in the amplitude of the annual flood wave (Davidson et al., 2012; Espinoza et al., 2012, 2013; Marengo and Espinoza, 2016; Nobre et al., 2016; Towner et al., 2020). Maximum flooding extent along the central Amazon has expanded by 26% (Fleischmann et al., 2023), mechanically impacting key processes such as CO₂ and CH₄ outgassing. This warming-induced hydrological cycle strengthening is projected to continue in the coming decades (Hirabayashi et al., 2013; Langerwisch et al., 2013; Sorribas et al., 2016; Alfieri et al., 2017) and the rainforest could reach a tipping point by the second half of the century, potentially converting to savanna, particularly in the eastern and southern regions, or persisting in a degraded state (Lovejoy and Nobre, 2019; McKay et al., 2022; Flores et al., 2024).

1.4 Monitoring challenges and integrated approach

The cascading effects of the ongoing transition in the Amazon remain uncertain, as monitoring is limited, particularly regarding sediment fluxes. Sparse measurements, due to high costs and logistical challenges, hinder the establishment of dense, long-term monitoring networks. Such networks are essential to constrain consistent upstream–downstream mass balances and to spatialize them at relevant scales. This is particularly important for identifying key processes, especially those linked to lateral exchange with the floodplain, which are still only roughly estimated. For instance, the CZO HyBAm covers just one gauging station per 160,000 km² on average. Additionally, sediment budgets in lowland sub-basins are challenging to estimate accurately, as their order of magnitude is comparable to the uncertainty in sediment load measurements (e.g. Xiaoqing, 2003; Horowitz et al., 2015; Vauchel et al., 2017; Gitto et al., 2017; Santini et al., 2019; Santini, 2020; Dramais, 2021). A significant portion of the suspended load (up to 70%) consists of very fine sands (Santini et al., 2019; Martinelli, 2022), which are difficult to measure due to their sensitivity to hydrodynamic fluctuations and heterogeneous distribution within the cross section. Furthermore, the buffering effects of such a large basin (Walling, 2006) can mask the impacts on material transfer to the oceans, requiring long-term monitoring.

The scarcity and heterogeneity of observed data directly reduce the robustness and accuracy of hydrological and sediment transport models, limiting their ability to capture key processes and to reliably forecast responses to environmental changes. In response, spatial hydrology has increasingly complemented *in situ* observations in the Amazon, with satellite data, particularly space altimetry and water color imaging, playing a key role in monitoring water levels and sediment concentrations (Calmant et al., 2009; Martinez et al., 2009, 2015; Espinoza-Villar et al., 2012, 2013, 2017; Park and Latrubesse, 2014). On the other hand, hydrological models have also addressed flooding and backwater effects (Yamazaki et al., 2011; Paiva et al.,



2013; Pontes et al., 2017; Siqueira et al., 2018; Santini, 2020; Guilhen et al., 2022). Recently, the question of the sediment routing into semi-distributed models has been explored (Fagundes et al., 2021, 2023; Santini, 2020).

100 However, no study has yet combined remote sensing with modelling for detailed sediment dynamics in the Amazon. This study, building on long-term CZO HyBAm data, proposes an integrated approach using calibration-validation campaigns, remote sensing, and hydraulic-hydrological modelling. The aim is to spatialize long-term mass balances more precisely and differentiate fine sediment fluxes, associated with organic matter and pollutant transfer, from sand loads, related to river dynamics. Finally, due to the importance of sediment dynamics in biogeochemical cycles, this framework holds the potential to deepen the understanding of the Amazon's role in these processes at global scale, as well as the effects of environmental
105 changes on its hydrology and sediment fluxes.

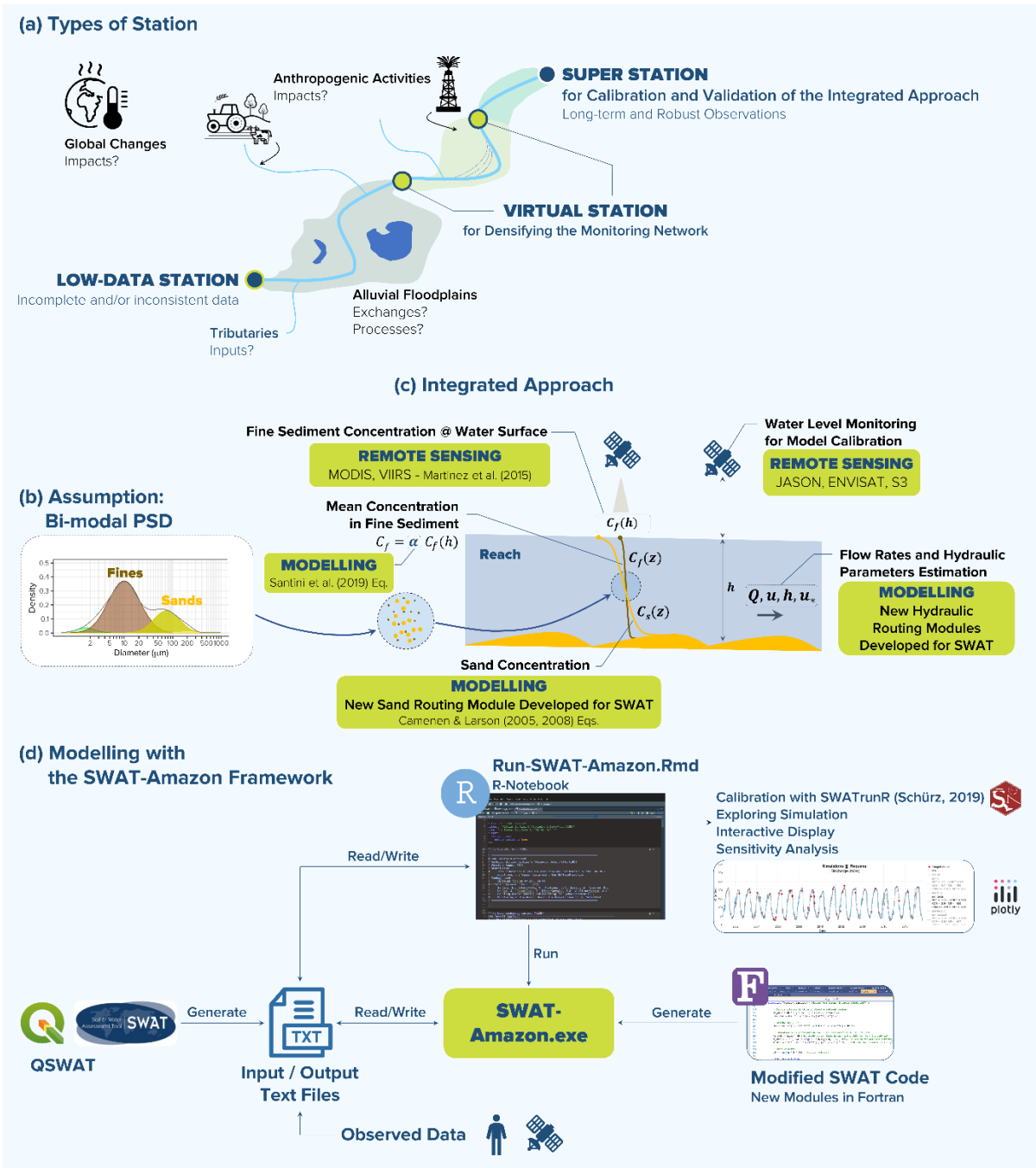
1.5 Case study: the Ucayali Basin

Given the continental scale of the Amazon Basin, this study focuses on the Ucayali River, a major foreland tributary draining 350,000 km² (49% Andes, 51% plains). Only two HyBAm gauging stations monitor hydro-sediment fluxes in this basin, with incomplete records for the study period (1983-2019): one upstream of the lowlands, the other at the basin outlet. Limited water
110 levels, with some unreliable records, along with a few discharge measurements, are also available from other lowland stations. These constraints make the Ucayali a relevant test site for building the proposed integrated approach, before extending it to other Amazonian sub-basins.



2 Integrative strategy

115 The tailored integrative strategy (Fig. 1) for improving water and sediment balances relies on a combination of three station types in the plain: (i) ‘low-data’ (or poorly monitored) conventional stations, characterized by incomplete and/or inconsistent datasets; (ii) ‘virtual’ stations established at locations where satellite altimetry ground tracks intersect the river mainstem, in order to enhance the spatial density of the monitoring network through the integration of remote sensing and modelling; and (iii) ‘super’ stations with long-term, high-quality datasets, which serve as benchmarks for calibrating and validating the integrated approach. This strategy is further supported by dedicated calibration-validation field campaigns. At all stations, water and sediment fluxes are estimated by integrating remote sensing products and hydrological modelling outputs. Water discharges are simulated using a modified version of the Soil and Water Assessment Tool (SWAT) model (Arnold *et al.*, 1998), to account for Amazon flood wave dynamics and attenuation during flooding. It is assumed that the river transports two particle groups (Santini *et al.*, 2019): fine sediments and sands (Fig. 1b). Fine sediments (mean diameter $d_f \cong 10 - 20 \mu\text{m}$), primarily silts with small clay aggregates, behave similarly to passive scalars and their fluxes are not modelled with transport capacity equations. Instead, fine sediment concentrations at the water surface are derived from satellite images (Espinoza-Villar *et al.*, 2012, 2013, 2017; Martinez *et al.*, 2009, 2015), using an inversion model calibrated with *in situ* data. Whereas suspended sands (mean diameter $d_s \cong 80 - 120 \mu\text{m}$) transported in graded suspension, are invisible to satellite spectroradiometers due to Mie scattering (Pinet, 2017). Moreover, according to Santini *et al.* (2019) and Martinelli (2022), observed Rouse numbers (Rouse, 1937) are between 0.2 and 0.8 for this sand fraction, inducing concentrations near the surface. Therefore, sand loads are modelled using sediment transport equations in a new routing module developed in the SWAT model, referred to as SWAT-Amazon (Fig. 1d).



135 **Figure 1: General schematic overview of the proposed methodology. (a) Types of stations. (b) Typical bi-modal particle size**
distribution (PSD) in the large Amazonian rivers, identifying two main size groups: 1- fine sediments that can be monitored by satellite but
not modelled; 2- fine sands in graded suspension, invisible to satellites but whose transport capacity can be modelled. (c) Integrated
approach combining remote sensing, modelling, and calibration-validation campaigns. (d) SWAT-Amazon, a tailored version of the SWAT
140 **model for simulating water and sand fluxes. This modelling framework consists of a Fortran-based executable (SWAT-Amazon.exe), derived**
from the standard SWAT2012 code, and an R notebook (Run-SWAT-Amazon.Rmd) designed to enables model run, simulation analysis,
interactive result visualization, as well as sensitivity analysis and calibration procedures with the SWATrunR package (Schürz et al., 2019).



3 Dataset

3.1 Conventional data

This study relies on long-term hydro-sediment flux data from the CZO HyBAm (Guyot et al., 2007; Santini, 2020). In the
145 Ucayali basin, IRD (*Institut de Recherche pour le Développement*) and SENAMHI (*Servicio Nacional de Meteorología e
Hidrología*) has been operating two HyBAm gauging stations (Lagarto and Requena) since 2001 (Fig. 2), carrying out 82 field
campaigns to establish rating curves. Additional sediment monitoring was carried out at Puerto Inca between 2012 and 2016,
at a conventional SENAMHI station. Water level–discharge relationships were also established at Puerto Inca and Pucallpa,
where water levels are monitored by port authorities. Observed sand fluxes, empirically derived from gauging (see
150 supplementary section S1), carry $\pm 30\%$ uncertainties, affecting simulations statistics.

3.2 Altimetric data and definition of virtual stations

Two virtual stations, JA204-S3B310-R22 (reach 22 on Fig. 2) and S3B331-R5 (reach 5), were defined based on the intersection
of satellite altimetry ground tracks (Jason, Envisat, Sentinel) with the mainstem of the Ucayali River. These stations provided
satellite-derived water level time series used to calibrate the hydrological model. In addition, satellite altimetry was employed
155 to correct water level records at Requena, Pucallpa and Contamana, the latter being a rarely visited SENAMHI station without
any flow measurements. Altimetry data processing was carried out using the open-access VALS (Virtual ALtimetric Stations)
software. By summing the virtual stations, the low-data stations (Contamana and Pucallpa), and the long-term CZO HyBAm
stations (Lagarto and Requena), the Ucayali sedimentary basin was subdivided into five distinct compartments to establish
hydro-sediment budgets with the integrated approach.

160 3.3 Fine sediments monitoring with remote sensing data

3.3.1 Retrieving time series of remote-sensed reflectance data

Given the required revisit frequency and the Ucayali River's width in the plains (500–1000 m), moderate-resolution satellite
imagery from MODIS (MODerate Resolution Imaging Spectroradiometer, 250×250 m, 1999–present, 1–2 days) and VIIRS
(Visible Infrared Imaging Radiometer Suite, 375×375 m, 2012–present, 0.5 days) was used to generate time series of surface
165 water reflectance (Fig. 1c). Reflectance values in the red and near-infrared (NIR) bands were extracted pixel by pixel from
satellite images using the free software GetMODIS and MOD3R, developed by the CZO HyBAm, that was tested and validates
in various previous studies (e.g. Espinoza-Villar et al., 2017; Vauchel et al., 2017). Water masks were applied to the Ucayali
River's main course near virtual stations. To ensure 50-100 pixels per mask, large river stems were covered, with masks
redrawn every 2–3 years due to river mobility. Collected scenes comprise images spanning 8-day periods, selecting pixels with
170 the lowest cloud cover and smallest satellite-viewing nadir angle.



3.3.2 Conversion of remotely sensed reflectance to fine sediment concentration

Two radiometric calibration-validation campaigns were conducted in the Ucayali Basin: the first in November 2011 at Requena (Espinoza-Villar et al., 2012) and the second in February 2017 at Lagarto, Puerto Inca, and Pucallpa, spanning three weeks (Santini, 2020). A total of 42 surface water samples were collected to determine total, fine, and sand concentrations. Simultaneously, hyperspectral field radiometers (TriOS) were deployed following the experimental setup of Mobley (1999), as adapted by Martinez et al. (2015) for the Amazon Basin. High-frequency (1 Hz) hyperspectral measurements of surface water reflectance were obtained at sampling locations. Relying on this dataset, a unique model for all the Ucayali Basin was fitted between fine sediment concentration at the water surface and the ratio of radiometer reflectance in the NIR (841–876 nm, according to the satellite sensor bands) and red bands (620–670 nm) (see Section 5.4).

3.3.3 From surface to mean concentration of fine sediments

Due to the considerable depth of Amazonian rivers and the vertical sediment concentration gradient near the surface, the ratio α_f , relating the channel mean concentration to the surface index concentration retrieved by satellite, ranges from 1 to 1.8 according to the CZO HyBAm database (1.1 to 1.2 in the Ucayali). To estimate α_f , the Santini *et al.* (2019) models were applied, parameterized using hydraulic data simulated in SWAT-Amazon.

3.4 Input data for modelling

The study utilizes the Peruvian Interpolated Data of SENAMHI's Climatological Observations (PISCO) (Aybar et al., 2020; Llauca et al., 2021) to support the development of an operational model in collaboration with the SENAMHI. Potential Evapotranspiration (PET) was estimated using the Hargreaves (1985) method, to take advantage of PISCO's temperature data. Land use data was obtained from the Peruvian Ministry of Environment (www.geoservidor.minam.gob.pe), while soil information was sourced from the Harmonized World Soil Database (www.fao.org/soils-portal). The topography layer was derived from the Multi-Error-Removed Improved-Terrain Digital Elevation Model (MERIT DEM) (Yamazaki et al., 2017), resampled from 90 m to 300 m for computational efficiency.

In SWAT-Amazon, water and sand fluxes can be forced at any sub-basin via input files. However, no external forcing was applied in this study. For sand fluxes, we assumed that Andean inflows (basins 3, 8, 9 in Fig. 2) were governed solely by transport capacity within the sand routing module. This assumption is supported by the observed relationship between sand flux and water discharge at Lagarto and Puerto Inca, which indicates, to a first approximation, sediment availability throughout the hydrological cycle in the Andean sub-basins. Lateral contributions from plain tributaries (basins 1, 2, 4, 7) were considered negligible and were likewise represented as transport-capacity limited in the simulations.

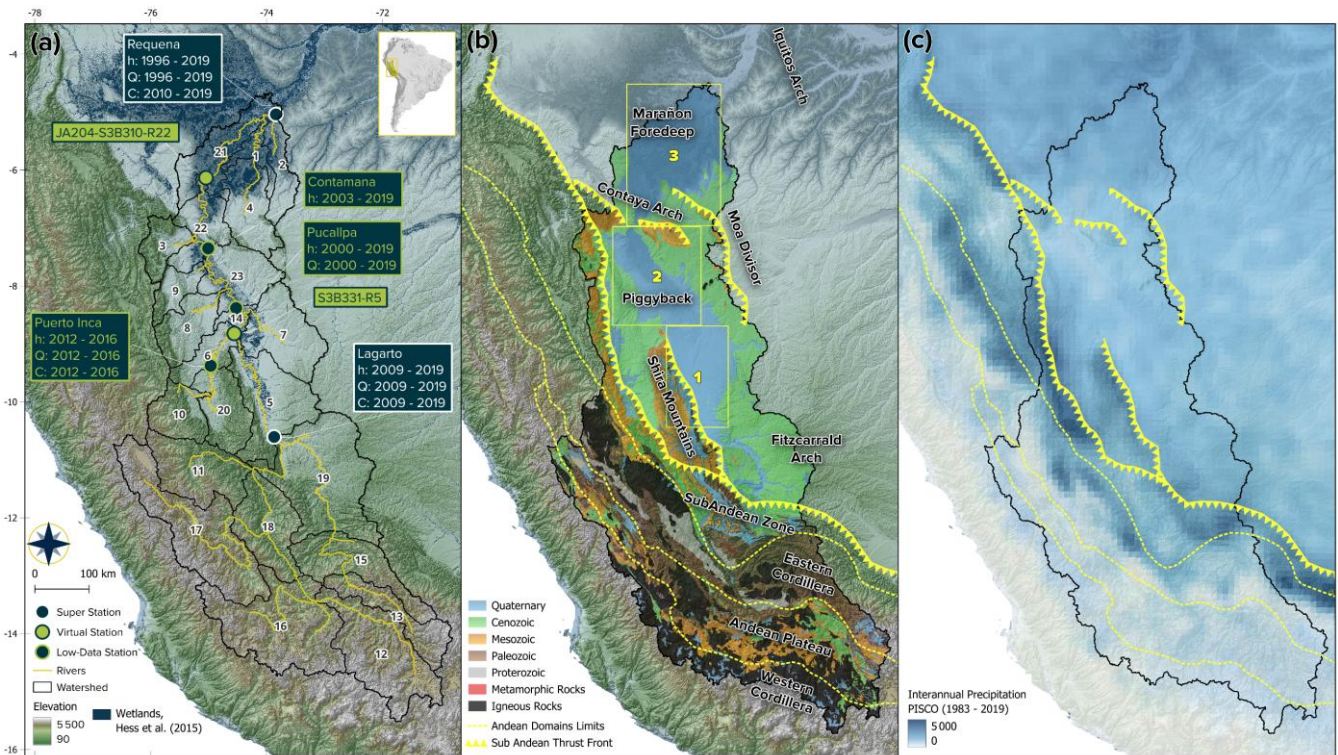


Figure 2: Ucayali's stack (a) Ucayali Basin sub-basins and station locations. Super stations (CZO HyBAm) (blue-filled circles, white outline), virtual stations (green-filled circles, blue outline), and low-data Stations with sparse/inconsistent data (blue-filled circles, green outline). The text box details observation periods for water level (*h*) discharge (*Q*), and suspended sediment concentration (*C*). Main Andean tributaries on the left bank: Cushabatay (3); Pisqui (9); Aguaytia (8); Pachitea (10, 20, 6). **(b) Geomorphological domains and outcrop distribution.** Box 1: Narrow sedimentary basin controlled by Shira Mountains and Fitzcarrald Arch uplift; Box 2: Piggyback Basin backward the Moa Divisor Thrust Fault; Box 3: Marañon Foredeep. **(c) Mean annual precipitation map** (PISCO dataset, 1983–2019).

4 Tailoring SWAT for water and sediment flux simulation

SWAT, a semi-distributed model with physical and conceptual equations, was chosen for its proven robustness in simulating hydrological processes in large basins at daily time step. Its open-source Fortran code and extensive user community provide numerous complementary modules and tools. However, SWAT has limitations in modelling water and sediment routing in large rivers with diffusive flood waves and extensive floodplains. It lacks realistic hydraulic connectivity between floodplains and the main channel, preventing accurate simulation of the relationships between water levels, velocities, and discharges, which are keys for sediment transport. To address this, a major code modification is introduced below.



4.1 New water routing modules

The main channel's trapezoidal cross-section in SWAT was replaced with a rectangular one for consistency with the hydraulic equations used. Floodplains were modeled with either rectangular or triangular cross-sections to ensure smoother flow transitions. Modelled as simple storage units, this streamlined approach was chosen after testing other configurations.

4.1.1 Water level calculation and state variables

The mean water level h (m) is derived from the water volume V (m³) stored in the reach i at time step t (beginning of the simulation day). As long as $h \leq h_f$, where h_f (m) is the floodplain activation threshold, h is computed as:

$$h_i^t = \frac{V_i^t}{\Delta x_i B_i}, \quad (1)$$

where B (m) is the main channel width and Δx (m) is the reach length. When $h \geq h_f$, the floodplain is activated, distributing V between the main channel and floodplain. If the floodplain cross-section is rectangular:

$$h_i^t = \frac{V_i^t - (\Delta x_i B_i h_{f,i})}{\Delta x_i (B_i + W_{fp,i})} + h_{f,i}, \quad (2)$$

where $W_{fp} = k_{fp} B$ (m) is the floodplain width and k_{fp} (–) a coefficient to be calibrated (the SWAT-Amazon parameters are given in Table 1). For a triangular floodplain cross-section, h depends on θ_{fp} (rad), the riverward slope angle (see supplementary section S2). Finally, state variables such as wetted area A_h (m²), wetted perimeter P_h (m), and hydraulic radius R_h (m) are all derived from h .

4.1.2 Dynamic and process equations

In large Amazonian rivers, flow variations over time and space are minimal, leading to a subcritical hydraulic regime, that can be modeled using the 1d Saint-Venant equations (Moussa and Bocquillon, 2009). Given the (very) gradual flow variations (Trigg et al., 2009), the convective and local acceleration terms are negligible, making the diffusive flood wave approximation suitable. When water-surface slope effects are also negligible, the pressure gradient is eliminated, allowing the use of the kinematic wave equation. SWAT-Amazon enables reach-specific selection between kinematic wave ($S_f = S_b$), suitable for steep Andean reaches, and diffusive wave approximation ($S_f = S_b + S_w$), preferred for low-slope floodplain reaches, with S_f (m m⁻¹) the energy gradient (or friction slope), S_b (m m⁻¹) the bed slope and S_w (m m⁻¹) the water-surface slope. When the diffusive wave model is used, S_w is assessed as follows:

$$S_{w,i}^t = \frac{(h_{i+1}^t - h_i^t)}{\frac{1}{2}(\Delta x_i + \Delta x_{i+1})}. \quad (3)$$

The reach-averaged velocity u (m s⁻¹) and discharge Q (m³ s⁻¹) are then calculated using the Gauckler-Manning-Strickler (GMS) friction equation (Hager, 2005).



$$240 \quad Q_i^t = A_{h,i}^t u_i^t = A_{h,i}^t \frac{\sqrt{s_{f,i}^t}}{n_{c,i}^t} (R_{h,i}^t)^{2/3} . \quad (4)$$

4.1.3 Continuity equation and water storage in the reach

At the end of the calculation time step $(t + \Delta t)$, V in reach i is updated as:

$$V_i^{t+\Delta t} = V_i^t + (Q_{i-1}^t - Q_i^t) \Delta t + V_{R_i}^t - V_{E_i}^t - V_{T_i}^t , \quad (5)$$

where V_E (m^3) is the volume lost to evaporation, V_T (m^3) is the volume infiltrated into the unsaturated water table and V_R (m^3) is the runoff (surface, subsurface, and baseflow) reaching the river. The computation of V_E , V_T and V_R follows the standard
 245 SWAT model. The updated volume $V_i^{t+\Delta t}$ is then used to determine $h_i^{t+\Delta t}$ for the next simulation step.

4.2 New module for sand sediment routing

4.2.1 Sand load and concentration in the reach

At the beginning of the simulation day, the suspended sand concentration C_s ($t \ m^{-3}$) in the reach i is:

$$250 \quad C_{s,i}^t = \frac{V_{s,i}^t}{\frac{Q_i^t}{u_i^t} \Delta x_i} , \quad (6)$$

where V_s (t) is the sand volume stored in the reach, in the main channel only. The daily suspended sand load Q_s ($t \ d^{-1}$), taking $\Delta t = 86,400 \ s$, is then:

$$Q_{s,i}^t = Q_i^t C_{s,i}^t \Delta t , \quad (7)$$

4.2.2 Transport capacity evaluation

255 Selecting appropriate transport capacity equations for deep, low-gradient rivers is crucial, as most were derived from laboratory studies under opposite conditions (steep slopes, shallow water, uniform flow). The physically based Camenen and Larson (2005, 2008) models for non-cohesive sands were chosen for their calibration with extensive global datasets and proven applicability in large tropical rivers (Camenen et al., 2014). In these models, the transport capacity Q_s^* ($t \ d^{-1}$) for suspended sands is evaluated as a function of the Rouse number P_s ($-$), which defines the concentration profile exponential shape, and a
 260 near-bed reference concentration C_b^* ($m^3 \ m^{-3}$), which determines its magnitude:

$$Q_{s,i}^{*t} = C_{b,i}^{*t} \left[\frac{1}{6 P_{s,i}^t} (1 - \exp(-6 P_{s,i}^t)) \right] Q_i^t s \Delta t , \quad (8)$$

where s is the relative sand density. The Rouse number summarizes the equilibrium between grain settling velocity w_s ($m \ s^{-1}$) and turbulence-induced lift, related to the shear velocity u_* ($m \ s^{-1}$), weighted by the sediment-to-eddy diffusivity ratio β_s ($-$):

$$P_{s,i}^t = \frac{w_{s,i}}{\beta_{s,i}^t \kappa u_{*i}^t} , \quad (9)$$



265 where κ is the Von Kármán constant. The Soulsby (1997) law is used for estimating the sand grain settling velocity, involving the grain size d_s (m) of the suspended sands. The shear velocity is calculated using the depth-slope product:

$$u_{*i}^t = \sqrt{g h_i^t (S_{bi}^t + S_{wi}^t)}, \quad (10)$$

where g (m s^{-2}) is the gravitational acceleration. The diffusivity ratio is either assigned a fixed value for each reach or computed dynamically using the Santini et al. (2019) model:

$$270 \quad \beta_{s,i}^t = 3.1 \exp \left[-0.19 \times 10^{-3} \frac{u_{*i}^t}{w_{s,i}} \left(\frac{h_i^t}{d_{s,i}} \right)^{0.6} \right] + 0.16, \quad (11)$$

The bottom reference concentration C_b^* is given by Camenen and Larson (2005):

$$C_{bi}^{*t} = \frac{0.0015 \theta_i'^t}{\exp \left(0.2 d_{bi} + 4.5 \frac{\theta_{cr,i}^t}{\theta_i'^t} \right)}, \quad (12)$$

275 where θ' is the dimensionless grain-related bed shear stress and θ_{cr} (–) the critical Shields parameter for the inception of transport (Camenen et al., 2014), which can be estimated from the Yalin-Shields curve as a function of the riverbed dimensionless mean diameter d_{b*} :

$$\theta_{cr,i} = \frac{0.25}{d_{b*,i}} + 0.055 (1 - \exp(-0.02 d_{b*,i})). \quad (13)$$

The parameter θ' is calculated as following:

$$\theta_i'^t = \frac{\frac{1}{2} f_i'^t (u_i^t)^2}{g (s-1) d_{bi}}, \quad (14)$$

where f' (–) is the Darcy-Weisbach skin roughness factor, derived from the logarithmic velocity law:

$$280 \quad f_i'^t = 2 \left(\frac{\kappa}{\ln \left(\frac{30}{k'_{s,i}} h_i^t \right) - 1} \right)^2, \quad (15)$$

with κ is the Von Kármán constant and the height k'_s (m) is the hydraulic skin roughness of Nikuradse, which can be expressed as a function of d_b and a coefficient $\sigma = 2.5$ (Engelund and Hansen, 1967; Bartholdy et al., 2010):

$$k'_{s,i} = \sigma_i d_{bi}. \quad (16)$$

4.2.3 Sand load adjustment based on transport capacity

285 The difference ΔQ_s^* (t d^{-1}) between the transport capacity Q_s^* (t d^{-1}) and the sand load Q_s (t d^{-1}) is then calculated at time t :

$$\Delta Q_{si}^{*t} = Q_{si}^{*t} - Q_{si}^t. \quad (17)$$

If $\Delta Q_s^* > 0$, there is excess transport capacity, allowing for riverbed erosion. The eroded mass E_{bed} (t d^{-1}) is defined as:

$$E_{bed,i}^t = K_{bed,i} \Delta Q_{si}^{*t}, \quad (18)$$

where $K_{bed}(-)$ is a coefficient ($0 \leq K_{bed} \leq 1$) representing the riverbed's susceptibility to erosion. The sand flux is then updated:

$$Q_{s_i}^t = Q_{s_i}^t + E_{bed_i}^t. \quad (19)$$

Conversely, if $\Delta Q_s^* \leq 0$, the sand load exceeds transport capacity, and the sand load is set to the transport capacity:

$$Q_{s_i}^t = Q_{s_i}^{*t}. \quad (20)$$

4.2.4 Sand budget at reach scale

295 Drawing on the mass balance proposed by Dunne et al. (1998), an erosion term, E_{bk} ($t d^{-1}$), is introduced to account for both floodplain channel inputs and bank erosion. These processes primarily occur at point bars on the inner bends of meanders, where floodplain inflows, with lower sediment concentrations than the river's transport capacity, enhance erosion and resuspension. Riverbed erosion, E_{bed} (Eq. 18) and two deposition terms (D_{ovbk} , D_{lat}), all expressed in ($t d^{-1}$), are also considered. Sand deposition on bars in low-velocity zones of the main channel is already accounted for in the sand load adjustment (Eq. 20). Deposition in floodplain channels and levee depressions when active (*i.e.* when $h > h_f$) is neglected for sand particles, as the high flow resistance caused by vegetation in these areas is expected to result in complete sedimentation at their inlets. This process is therefore implicitly included in the overbank deposition term, D_{ovbk} .

The term E_{bk} is activated only when the daily water volume ΔV_{fp} ($m^3 d^{-1}$) exchanged between the main channel and the floodplain is negative, meaning floodplain waters contribute to the main channel. Thus, E_{bk} was defined as function of ΔV_{fp} and C_{bk} ($t m^{-3}$), the concentration of these banks and bars inputs, considered as a constant to be calibrated:

$$E_{bk_i}^t = -\Delta V_{fp,i} C_{bk_i}^t. \quad (21)$$

Thus, E_{bk} is neglected when $\Delta V_{fp} \geq 0$ because the volume of water that could flow back from the floodplain during the rising stages is low compared to the water discharge in the main channel, contrary to the flood recession phase. In addition, the term E_{bed} can compensate for this if necessary, when the transport capacity is in excess.

310 The daily sand mass D_{ovbk} is defined as a function of ΔV_{fp} :

$$D_{ovbk_i}^t = \Delta V_{fp,i} C_{s_i}^t(z_{surf_i}^t), \quad (22)$$

where $C_s(z_{surf})$ ($t m^{-3}$) is the sand concentration in the upper flow layer, estimated at $z_{surf} \cong (h - h_f)/2$ (m). $C_s(z_{surf})$ is derived from the mean concentration C_s in the reach:

$$C_{s_i}^t(z_{surf_i}^t) = \frac{C_{s_i}^t}{\alpha_s}. \quad (23)$$

315 The ratio $\alpha_s = C_s(z_{surf})/C_s$ is estimated with the Santini et al. (2019) model:

$$\alpha_{s_i}^t = \frac{1}{6 P_{s,cf_i}^t} \exp\left(6 P_{s_i}^t \frac{z_{surf_i}^t}{h_i^t}\right) (1 - \exp(-6 P_{s_i}^t)), \quad (24)$$



4.2.5 Continuity equation

At the end of the calculation time step, the sand volume V_s stored in the main channel of the reach i is updated as:

$$V_{s,i}^{t+\Delta t} = V_{s,i}^t + (Q_{s,i-1}^t - Q_{s,i}^t) \Delta t - D_{ovbk,i}^t + E_{bk,i}^t, \quad (25)$$

4.3 Additional flow resistances

4.3.1 Impact of floodplain activation on flow velocity and transport capacity

When the floodplain becomes active, differences in depth and roughness between the main channel and floodplain develop a shear interface between the two flow zones, associated with Kelvin-Helmholtz instabilities, transferring horizontal momentum from the main channel to the floodplain (e.g. Sellin, 1964; Nicollet et Uan, 1972; Ervin et Baird, 1982; Knight, 1989, 1996; Smart, 1992; Loveless et al., 2000; Yen, 2002; Uijtewaal, 2014; Atabay and Knight, 2018). Sediment-laden water flowing through floodplain channels (Lewin et al., 2017) also transfer large amounts of momentum to the plain and reduces the kinetic energy of the main flow, as does the attenuation of the water surface slope during flooding, which tends toward the valley slope. Moreover, the waters that travel for a short time through the floodplain before returning to the main channel also contribute to reduce the flow velocity. These combined effects significantly reduce flow velocity and, more drastically, transport capacity in the main channel. They change the spatial distribution of velocities and shear stress in the main channel cross-section, especially near the banks and bars, where sediment stocks can be available. To account for this, a flow resistance correction factor ζ_n was defined as:

$$\zeta_n = \frac{u_{cf}}{u_c} = \frac{n_c}{n_{cf}}, \quad (26)$$

where the subscripts “c” and “cf” denote in-bank flow configuration (without floodplain) and flow with an active floodplain, respectively, at the same water level $h > h_f$. To evaluate ζ_n , the Nicollet et Uan (1972) or Smart (1992) equations can be used. However, both formulations only consider the shear layer interface between the main channel and floodplain. Furthermore, the Smart equation is not suitable for large rivers, and the Nicollet et Uan equation requires an estimate of the floodplain's Manning coefficient. Although the latter was implemented in the new water routing module, a simpler approach was preferred. Therefore, a relationship between ζ_n and the relative height $Y = (h - h_f)/h$ (m), from which the water exchanges between the main channel and the floodplain begin to affect the flow velocity, was defined:

$$\zeta_{n,i}^t = \frac{1}{(1 + Y_i^t C_{nfp,i})}, \quad (27)$$

with C_{nfp} (–) a coefficient to be calibrated, superior to zero if the flood impacts the flow resistance. Thus, to account for the floodplain drag when $h > h_f$, the Manning coefficient is reevaluated as following:

$$n_{cf,i}^t = \frac{n_i}{\zeta_{n,i}^t}. \quad (28)$$



345 Following the same reasoning as for velocities the ratio of the dimensionless grain-related bed shear stresses θ'_c and θ'_{cf} should also be a function of ζ_n . Indeed, according to Eq. 14:

$$\frac{\theta'_{cf}}{\theta'_c} = \left(\frac{f'_{cf} u_{cf}}{f'_c u_c} \right)^2 = \zeta_n^2 . \quad (29)$$

Here, $f'_{cf} = f'_c$ is assumed, as f' is a grain-related friction factor, not a flow resistance factor (Yen, 2002): the floodplain drag is already accounted for in u_{cf} , through n_{cf} . The shear velocity term used for calculating the Rouse number (Eq. 10) should
 350 be also affected by the floodplain drag:

$$u_{*cf} = \zeta_n u_{*c} . \quad (30)$$

However, when shifting from a 1d to a 2d framework, the transverse profiles of θ' , and consequently of C_b^* and $Q_{s_{cf}}^*$, are likely to be more strongly affected than the lateral profile of the depth-averaged velocity (see supplementary section S3), in particular near the banks. To account for the complex 2d effects on sediment transport capacity, effects not considered in the initial
 355 computation of the transport capacity $Q_{s_{cf0}}^*$ which was initially calculated using the corrections for θ' and u_* corrections in Eqs. 29 and 30, the following formulation is applied when the floodplain is active (i.e. when $h > h_f$):

$$Q_{s_{cf_i}}^{*t} = \zeta_{n_i}^{\eta t} Q_{s_{cf0_i}}^{*t} , \quad (31)$$

with η an exponent to calibrate which accounts for these complex 2d effects.

4.3.2 Bed roughness influence for low waters

360 In the large Amazonian rivers, a decrease in bed roughness influence with increasing water levels has been observed (see example in supplementary section S4). To model this in the SWAT-Amazon version, the Manning coefficient is modified using the factor $\zeta_{n_{ch}}$, defined as:

$$\zeta_{n_{ch_i}}^t = 1 + C_{n_{ch_i}} \left(\frac{h_{ch_i} - h_i^t}{h_i^t} \right) , \quad (32)$$

where $C_{n_{ch}}$ (–) is a coefficient and h_{ch} (m) the water height that ends the additional bed roughness influence. Both needs to be
 365 calibrated.

4.4 Calibration and sensitivity analysis

The model calibration was performed using the SWATrunR package (Schürz, 2019), which enables parallel processing. To run the SWAT-Amazon executable and calibrate parameters, including the newly introduced ones (Table 1), an R-Notebook was written (Fig. 1d). It allows users to export interactive figures and perform sensitivity analyses. Both SWAT-Amazon and
 370 its R-Notebook for calibration are available for download at: <https://github.com/william-santini/SWAT-Amazon>.



Table 1: Main parameters in the new routing modules of SWAT-Amazon

Variable name	Unit	Routing module	Definition	Input file
h_f	(m)	Water	Water height that triggers the floodplain activation	.rte
B	(m)		Width of the rectangular main channel	.rte
S_b	(-)		Channel bed slope, calculated from the MERIT DEM with QWAT	.rte
k_{fp}	(-)		Coefficient to determine the floodplain width: $W_{fp} = k_{fp} B$ (case of floodplain with rectangular cross-section)	.rte
θ_{fp}	(rad)		Angle of the floodplain riverward slope (case of floodplain with triangular cross-section)	.rte
n	(s m ^{-1/3})		Manning coefficient	.rte
C_{nfp}	(-)		Coefficient for increasing the flow resistance in the main channel when the floodplain is active	.rte
h_{ch}	(m)	Sand	Water height that ends the additional bed roughness influence	.rte
C_{nch}	(-)		Coefficient for increasing the flow resistance in the main channel when low waters	.rte
d_s	(m)		Arithmetic mean diameter of suspended sands	.rte
d_b	(m)		Arithmetic mean diameter of riverbed sands	.rte
s	(-)		Relative sand density. $s = 2.65$ was taken in this study	.rte
β_s	(-)		Ratio of suspended sand to eddy diffusivity, imposed or calculated with the Santini <i>et al.</i> (2019) model (Eq. 12)	.rte
ν	(m ² s ⁻¹)		Kinematic water viscosity. The value corresponding to a temperature of 28°C was considered here	.rte
σ	(-)		Coefficient to determine k'_s : $k'_s = \sigma d_b$	.rte
K_{bed}	(-)		Main channel susceptibility to erosion (riverbed only), between 0 and 1	.rte
C_{bk}	(t m ⁻³)		Concentration of bank and bar inputs (constant)	.rte
η	(-)		Correction exponent for transport capacity when the floodplain is active (cf. Eq. 31)	.rte



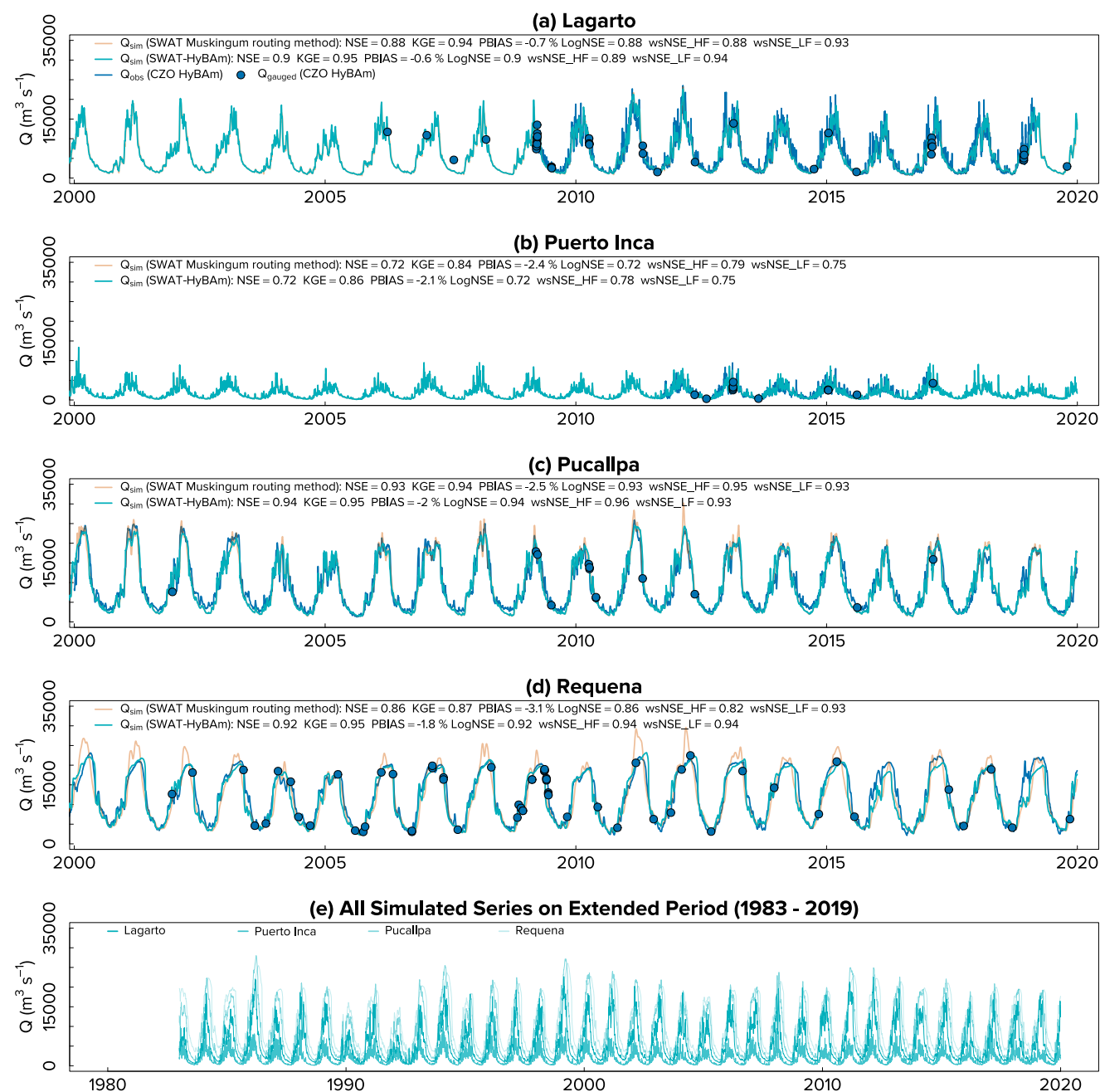
5 Results

5.1 Water discharge simulations

At a daily time step, SWAT-Amazon simulations at the watershed outlet show excellent performance (Fig. 3d): NSE (Nash-Sutcliffe Efficiency) = 0.92, KGE (Kling–Gupta efficiency) = 0.95, PBIAS (Percent Bias) = -1.8%, LogNSE (NSE on the logarithms of the series) = 0.92 on the 2000–2016 calibration period (see Moriasi et al. (2007) for details on these metrics). SWAT-Amazon significantly improves over the standard SWAT model (NSE = 0.86), using Muskingum routing with maximum flood attenuation. Moreover, the standard SWAT simulation predicts flood peaks 1–2 months earlier than observed, whereas SWAT-Amazon correctly synchronizes them. The model accurately captures hydrological dynamics and interannual variability. As highlighted by Yamazaki et al. (2011), the difference between kinematic and diffusive wave simulations was minimal (not shown), confirming that Ucayali flood attenuation mainly results from floodplain buffering.

At Pucallpa (Fig. 3c), SWAT-Amazon shows only slight improvements over standard SWAT due to the less developed floodplain. At the Andean outlet (Lagarto and Puerto Inca), where floodplain influence is minimal, both models perform similarly, though SWAT-Amazon slightly outperforms the default version. Despite a good daily NSE (0.72) at Puerto Inca, the model struggles to reproduce rapid flood oscillations typical of piedmont hydrographs. This issue, independent of the routing model, stems from uncertainties in rainfall estimation. Before final calibration, systematic biases (-20% to +20%) were observed, with underestimation in piedmont stations and overestimation in plains, primarily due to the precipitation dataset. These biases were corrected using interannual adjustment factors in SWAT .sub files. Additional errors in the PISCO dataset were identified and corrected by standardizing precipitation time series across station subgroups. However, PISCO still underestimated precipitation between Contamana and Requena for 2016, 2017, and 2019, leading to their exclusion from efficiency calculations.

Despite these limitations, the bias-corrected PISCO dataset demonstrated a high degree of homogeneity and robustness, allowing extension of observations across all stations (virtual and conventional) for simulations covering 1983–2019 (Fig. 4e), adding 13 years at Requena and 26 years at Lagarto. This extension is particularly valuable for future studies in this poorly monitored region, especially given the high accuracy of the weighted seasonal Nash-Sutcliffe Efficiency for low (wsNSE_LF) and high (wsNSE_HF) flows (see Zambrano-Bigiarini and Bellin, 2012, for details).

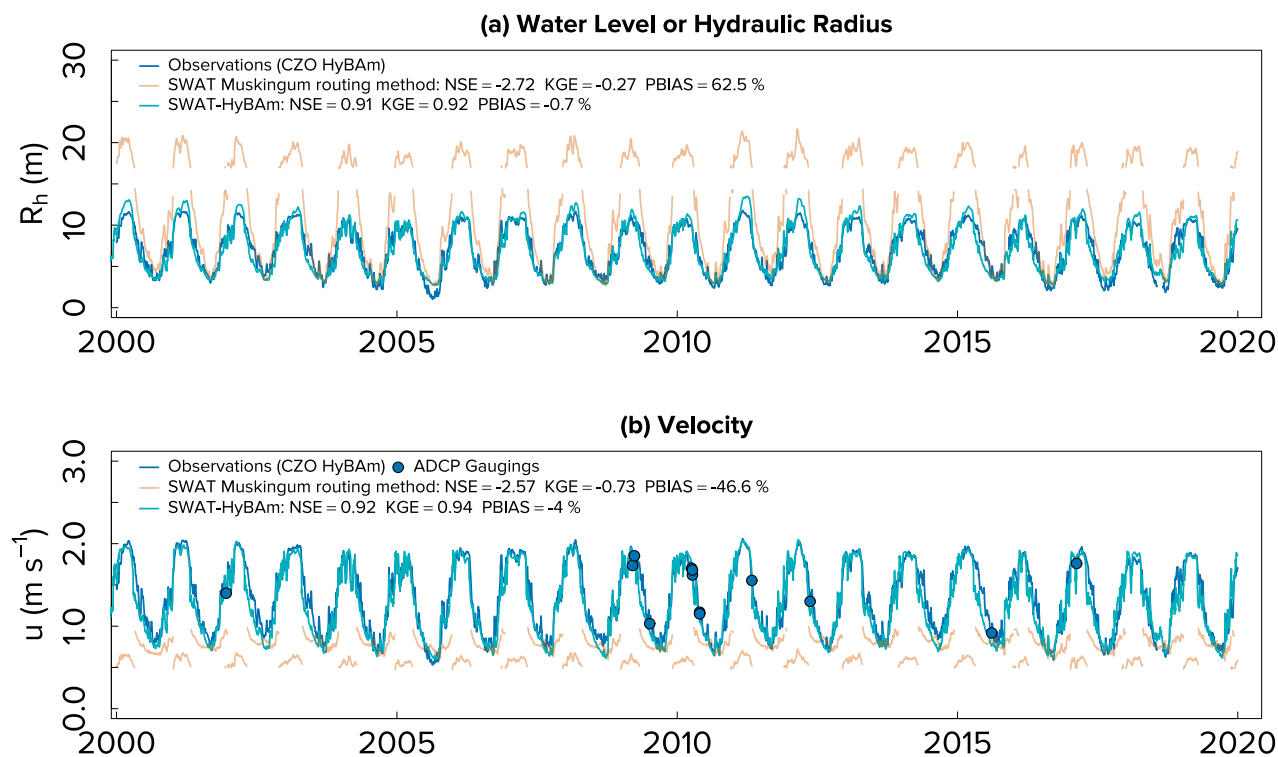


410 **Figure 3: Water discharge simulations for the stations with gauging data (a) – (d) Observed discharge (marine blue) vs. simulated discharge using the default SWAT model (orange) and SWAT-Amazon (cyan) at a daily time step, with punctual ADCP gauging values (blue circles). (e) Full 37-year simulation (1983–2019) with SWAT-Amazon for the same station group.**



5.2 Water levels, velocities, and rating curves simulations

The hydraulic radius R_h was used to compare simulations with observations (Figs. 4 & 5). Indeed, for large cross-sections, the mean depth h_m (m) approximates R_h , and for the modelled rectangular cross-sections, $R_h \approx h = h_m$. Moreover, observed water levels were not directly comparable due to offset differences in staff gauge zero-values relative to assumed river bottom elevation. In standard SWAT, once the bankfull height is exceeded, flow instantly spreads into the floodplain, forming a single cross-section instead of the usual approach in hydraulics of separating channel and floodplain flows (e.g. Einstein, 1950; Einstein and Barbarossa, 1952; Yen, 2002). This sudden change in cross-section geometry thus causes a discontinuity in flow velocity and hydraulic radius (Fig. 4), since P_h increases sharply while A_h grows more moderately ($R_h = A_h/P_h$). Below bankfull height, standard Muskingum simulations overestimate hydraulic radius and underestimate velocities (Fig. 4), and it is not possible to calibrate $u(h)$ rating curves. Therefore, the standard SWAT model is not able to simulate realistic hydraulic radius (water levels) and velocities and even less sand loads with transport capacity laws, for which these variables are required. Conversely, SWAT-Amazon generates robust daily water level and velocity time series, closely matching observations (Fig. 4), with NSE values between 0.77 and 0.93 for water levels and 0.79 to 0.92 for velocities, the lowest at Puerto Inca, while all others exceed 0.89. It produces consistent $Q(R_h)$ and $u(R_h)$ rating curves (Fig. 5), accurately capturing slope-controlled hysteresis and ‘duckbill’ damping when h_f is exceeded, as Manning’s coefficient increases with relative water height due to floodplain effects.



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Figure 4: Example of (a) water levels and (b) velocities simulation at Pucallpa. Marine blue line: observations, orange line: best default SWAT simulation with Muskingum, cyan line: SWAT-Amazon simulation, blue filled circle: ADCP gauging values. Error bars, which were less than 3% for ADCP measurements, are not shown for clarity.

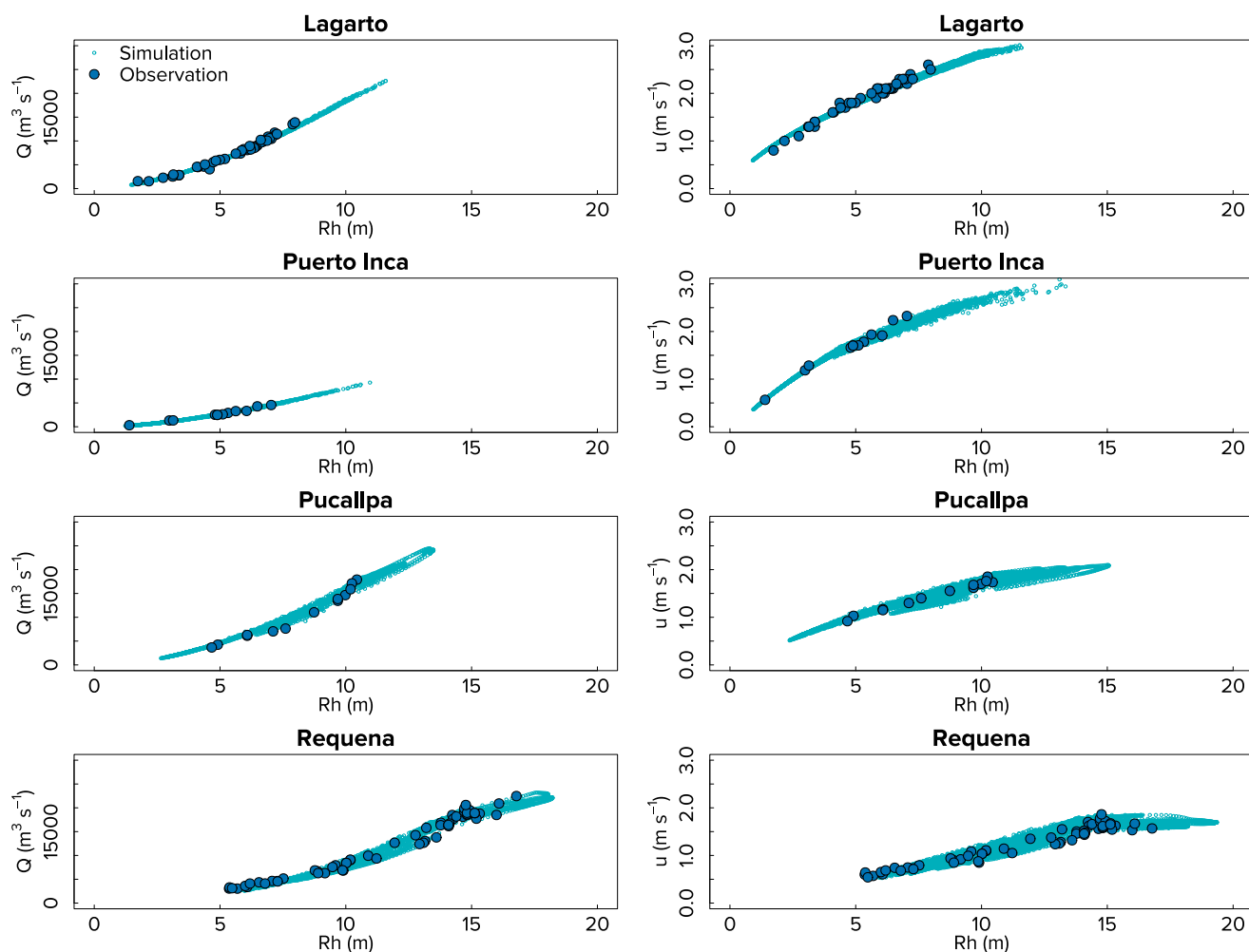


Figure 5: Rating curves between hydraulic radius (R_h) and discharge (left) or velocity (right). Cyan circles: SWAT-Amazon simulation, blue filled circles: ADCP measurements.



5.3 Sand routing

Calibration focused on the 09/2009–09/2015 period, when sediment monitoring protocols were enhanced, including higher sampling frequency at Requena between 11/2012 and 06/2013, where one sample was collected each two days plus three
450 sampling repetitions each ten days. Beyond, sampling was conducted at five-day intervals during the wet period between July 2013 and September 2015. Additionally, the concentration gaugings were performed in all sites with a higher number of samples collected throughout the cross-section, particularly in the first half of the water column, to ensure more accurate sand concentration calculations.

At Lagarto, the sand routing model accurately reproduces sand fluxes (Fig. 6a, daily NSE = 0.8), validating the capacity-
455 limited flux assumption at the Andean outlet (cf. section 2.8.2). At Puerto Inca (Fig. 6b, NSE = 0.44), the model struggles due to rainfall data and sharp flux peaks. Nevertheless, at the mainstem level of the Ucayali, the influence of this discrepancy is limited. At Requena, the model closely matches observed sand fluxes (Fig. 6c, NSE = 0.86) with peaks coinciding with maximum rainfall in January-February. From March, sand flux decreases while flow increases, indicating no correlation between sand flux and discharge. This decline is concomitant with the crossing of the threshold h_f from which the floodplain
460 watering impacts the transport capacity. In the 2010 drought year, the river briefly reached this threshold, with minimal impact on sand flux: Q and Q_{ss} are well correlated. In the remaining years, a second sand peak occurred in May-June. This is concurrent with the recovery of river transport capacity, which is enabled by a reduction in flow resistance due to the dewatering of the floodplain and an increase in energy availability in the main channel, resulting from the influx of floodplain and black waters supplies, which have low sediment concentrations. The 2012 extreme flood event, intensively monitored,
465 highlights this key process for sediment routing dynamics.

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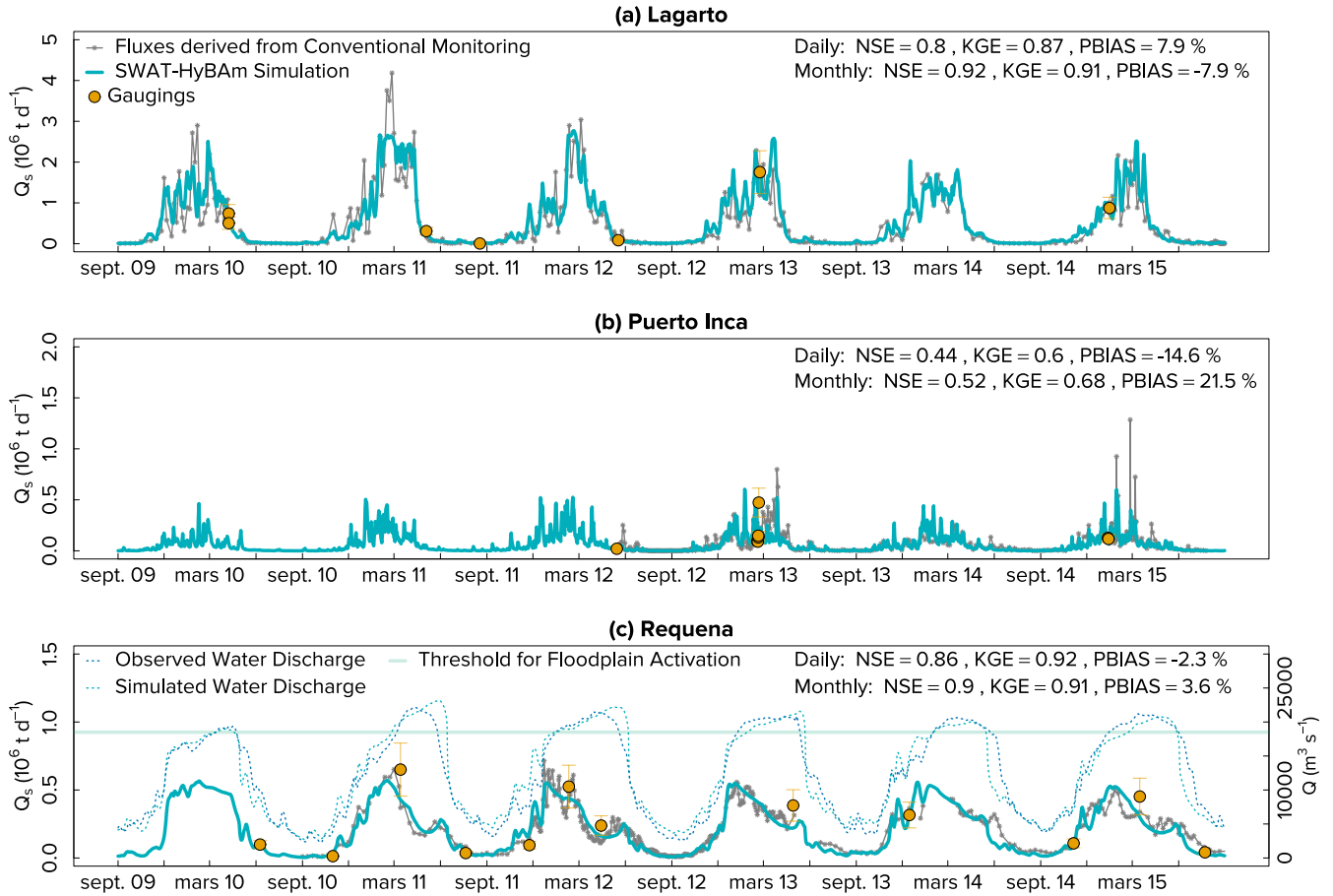


Figure 6: Observed and simulated sand fluxes for the gauging stations with sediment monitoring: (a) Lagarto, (b) Puerto Inca, (c) Requena. Gray line with stars: observations, cyan line: SWAT-Amazon simulations, orange filled circles with error bars: gauged sand flux values. Observed (blue dashed line) and simulated (cyan dashed line) water discharge at Requena are plotted in (c). The green horizontal line represents the discharge triggering floodplain activation (corresponding to h_f), approximated due to the non-bijective stage-discharge relationship.

5.4 Remote-sensed fine sediments fluxes

Calibration-validation campaigns established a relationship between fine sediment concentration C_f (mg l^{-1}) (at $z = h$) and the NIR-to-red reflectance ratio R , supporting a single model for the entire basin (Fig. 7) with a high coefficient of determination ($R^2 = 0.94$) and a low Mean Absolute Error (MAE = 59 mg l^{-1}):

$$C_f(h) = 822.8 \times R^{2.86} + 0.001 \times \exp(10.4 \times R), \quad (33)$$

This model accounts for reflectance saturation in the red band at high concentrations, providing a better fit across the full concentration range than a simple power-law equation. It was validated across all hydrological conditions from 2000 to 2019 using matchups between time series of *in situ* fine sediment concentrations monitored at Requena and Lagarto and co-located satellite reflectance ratios at a monthly time step ($R^2 = 0.78$, MAE = 132 mg l^{-1}) (Fig. 7). Note that Eq. 33 is already corrected



for adjacency effects, through a simple offset of +0.2 applied to the reflectance ratio to account for water pixel contamination by riverbanks.

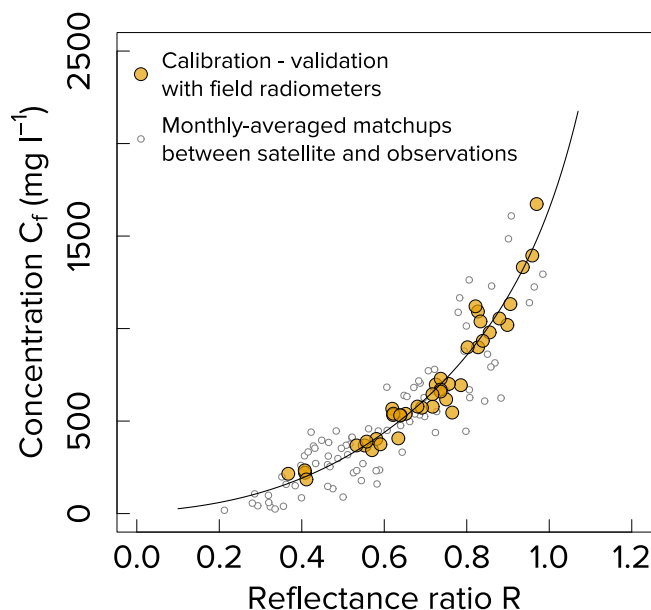


Figure 7: Relationship between fine sediment concentration (C_f) at the water surface and the ratio R of NIR to red reflectance. Orange dots represent calibration and validation points, based on 42 field measurements in the Ucayali Basin, where reflectance was measured using a hyperspectral radiometer and fine sediment concentrations were measured at the water surface. Gray dots correspond to matchups between satellite-derived reflectance ratio and fine sediment concentrations monitored at gauging stations, averaged at a monthly time step.

5.5 Validation of the integrated approach at super stations

The integration of remote sensing and hydrological modelling was validated at two super stations in the basin: Requena (Fig. 8) and Lagarto (not shown). Total sediment fluxes (Fig. 8c) were calculated by summing: (1) sand flux (Fig. 8a) from SWAT-Amazon simulations and (2) fine sediment flux (Fig. 8b) from satellite data combined with SWAT-Amazon flows. The results align well with *in situ* flux measurements (daily NSE: 0.87 at Requena, 0.79 at Lagarto, monthly NSE: 0.87 at Requena, 0.86 at Lagarto), and suggests that both stations could be monitored in this way with a few calibration – validation campaigns. The relevance and validity of the method developed here for an integrated monitoring of hydro-sediment fluxes has thus been demonstrated. This is particularly supported by the increase of satellite data availability over time, as it can be noticed on Figure 8b where missing data were mainly observed at the beginning of the time series, due to the increasing number of operating satellite sensors (Terra since 1999, Aqua since 2002, VIIRS since 2012). It is now possible to envisage a densification of the monitoring network in Amazonia to better understand sediment dynamics and biogeochemical cycles. The hydro-sediment part of this exercise was applied to the Ucayali basin in what follows.

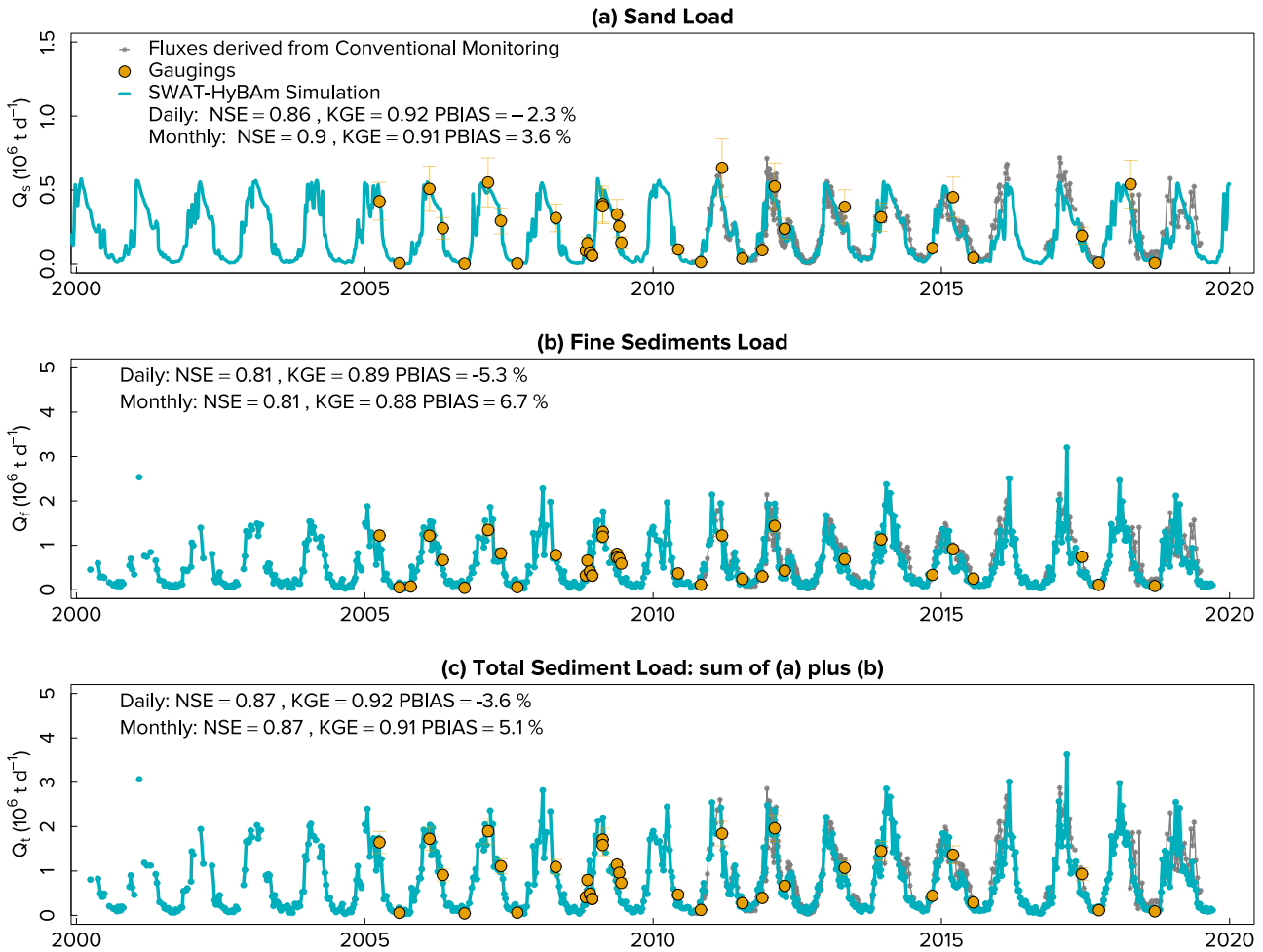


Figure 8: Integrated monitoring of sediment fluxes at the basin outlet. The gray line with stars shows observed sediment fluxes, the cyan line represents SWAT-Amazon simulations, and orange filled circles indicate gauged flux values.



6 Discussion

This study presents the first integrated approach for monitoring hydro-sediment fluxes in the Amazon basin, providing the most extensive and robust daily time series of hydro-sediment fluxes for the Upper Amazon (20 years for fine sediments to 37 years for water discharge and sand flux) with exceptionally high NSE values. It introduces a physically based methodology for modelling transport capacity, using realistic hydraulic parameters derived from the calibration of $u(h)$ and $Q(h)$ rating curves.

Previous large-scale efforts with the MGB (*Modelo de Grandes Bacias*) model (Fagundes et al., 2021, 2023), contributed significantly to understanding sediment transport across South America. However, challenges persist in representing sand transport, especially its suspension dynamics. The MGB model assumes sand transport is mainly bedload, while field observations show sand can make up to 70% of the suspended sediment load in large Amazonian rivers. The Rouse numbers for suspended sand (0.2–0.8) indicate graded suspension rather than intermittent transport (Santini et al., 2019; Martinelli, 2022), leading to underestimates of sediment load in the Ucayali Basin by the MGB model, nearly three times lower than observed values (Santini, 2020). These findings underscore the necessity of detailed, regionally focused studies based on long-term measurements and rigorous data consistency analyses, rather than broad continental-scale assessments, which often involve considerable uncertainties and may lead to misinformed sustainable development strategies and mitigation policies. For the first time, satellite-based sediment monitoring is applied exclusively to the fine fraction. A relationship (Eq. 33) is proposed that is independent of hydrodynamic fluctuations affecting the surface concentration, since sand contributes from ~5% to 50% to the surface concentration in the Ucayali. This contrasts with previous studies (Espinoza-Villar et al., 2012, 2013, 2017; Park and Latrubesse, 2014; Martinez et al., 2015) where satellite reflectance was solely compared with satellite reflectance or where remote-sensing was only used to calibrate the model (Fagundes et al., 2020). However, Pinet (2017) noted hysteresis in Madeira River relationships due to variations in grain diameter at the water surface, and Santini (2020) suggested that the coarser fraction of fine sediments might be also sensitive to turbulence-induced lift variations. This could affect the relationship established for the Ucayali Basin, particularly during resuspension events and low-water conditions. Further investigation is needed, and this aspect will not be discussed in the present study.

Before drawing conclusions and interpreting the mass balances (Section 4.4), it is crucial to assess the robustness and limitations of the method to ensure that the necessary nuances are applied, particularly at the virtual stations. To this end, a sensitivity analysis of the SWAT-Amazon model was conducted.

6.1 Model sensitivity and uncertainties

The Sobol method was applied to sub-basin 21 over the 2009-2015 period to assess the relative importance of SWAT-Amazon input parameters and their interactions, identifying uncertainties from the model's formalisms. First-order indices (S_i) and total-order indices (T_i) were computed to quantify each parameter's individual contribution (with S_i) and its total effect (with T_i), including interactions with other parameters, on the overall variance of the model output. The NSE criterion was applied for



global outputs (Fig. 9a, b), and the interannual daily average for time-series outputs (Fig. 9c, d). The analysis focused on a selected set of SWAT-Amazon parameters, based on calibration experience and physical understanding, rather than exhaustively testing all parameters. This analysis was conducted for sub-basin 21.

6.1.1 Water routing

The analysis focused on the parameters set $(n, B, h_f, C_{nfp}, k_{fp})$, with n accounting for all the resistance term $\sqrt{S_b}/n$. Results show a greater sensitivity of the model to (h_f, C_{nfp}, k_{fp}) , which control flood wave attenuation by the floodplain (Fig. 9a). The interannual time-series analysis (Fig. 9c) reveals greater sensitivity during flood recession than rising waters. The recession is particularly challenging to calibrate due to rapid water discharge drops, where even slight shifts in the timing of floodplain recession cause large discrepancies between simulated and observed flows throughout the entire recession limb. Therefore, the parameter h_f , controlling recession onset, must be carefully assessed. The strong oscillation of Sobol indices for n during low waters (Fig. 9c) reflects the impact of h (through B) on bed roughness and flow resistance when $h < h_{ch}$ (cf. Eq. 32). Small variations of h induce large changes in n during low waters, but with minimal impact on discharge, as shown by the interannual discharge plot in Figure 9b. This effect diminishes quickly as water levels rise.

6.1.2 Sand routing

The sensitivity analysis, using the parameter set $(\beta_s, C_{bk}, d_b, d_s, \eta, k_{ch})$, shows that d_b is the most critical calibration parameter (Fig. 9d, f), as previously highlighted by Fagundes et al. (2023). Calibrated d_b values for sub-basins 19, 5, 14, 23, 22, and 21 are 252 μm , 240 μm , 240 μm , 242 μm , 235 μm , and 220 μm , respectively, matching observed mean diameters at Lagarto (260 μm), Pucallpa (243 μm), and Requena (228 μm). Calibrated d_s values are approximately 80 μm for all sub-basins, except Lagarto (98 μm), are consistent with PSD observations. Since the calibration of the flood recession limb directly influences sand resuspension (Eq. 22), h_f emerges also as a key parameter in sand routing. The influence of k_{ch} is minimal, as the sand load input in sub-basin 21 is sufficient, eliminating the need for riverbed erosion to compensate for sediment supply deficits.

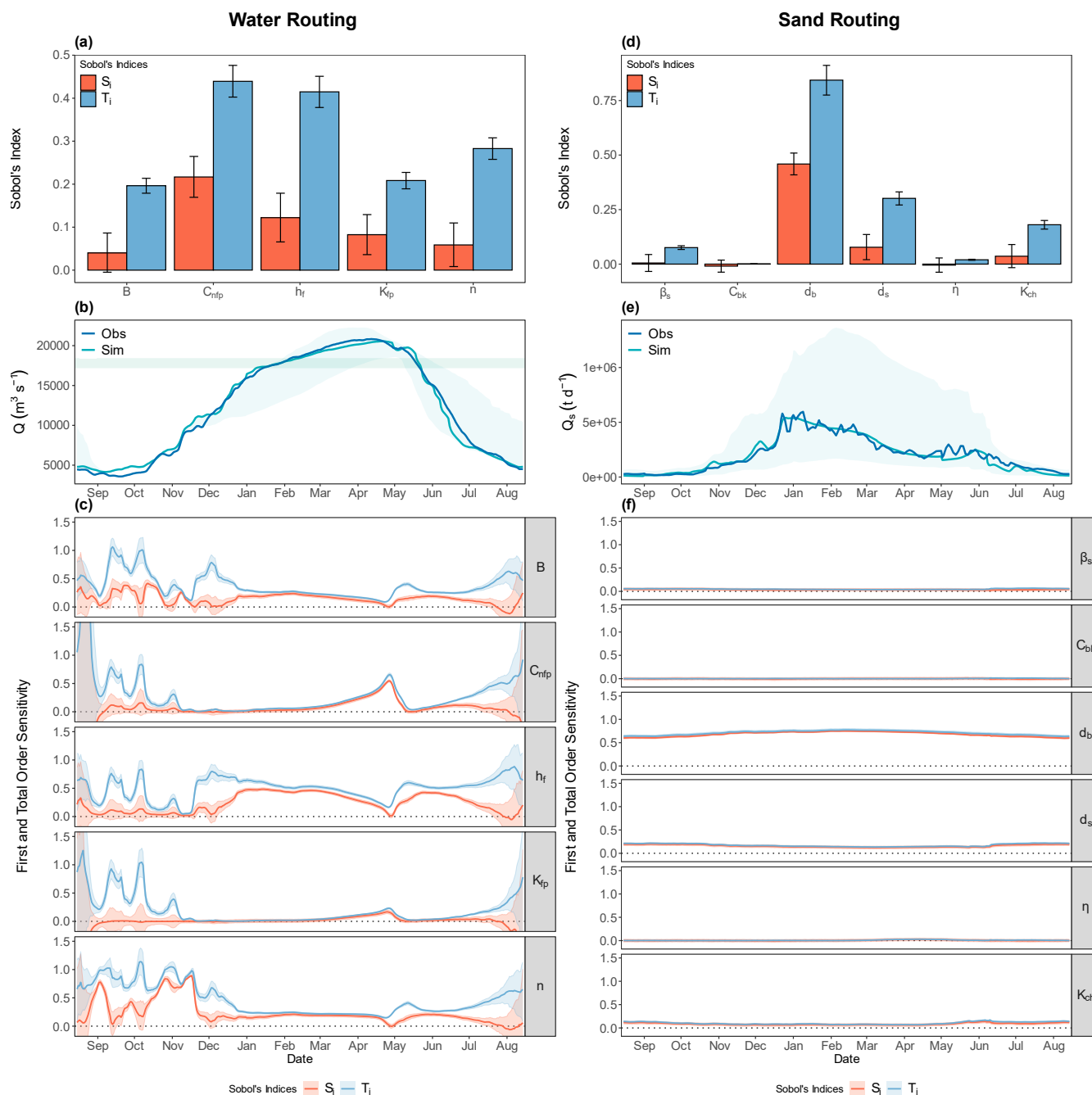


Figure 9: Sobol sensitivity analysis results for water (left) and sand routing (right), applied to sub-basin 21 over the 2009-2015 period. (a, d) present the analysis based on the NSE criterion, while (c, f) summarize the temporal analysis as interannual daily averages. (b, e) show the interannual daily average of the observed flux (blue line), the SWAT-Amazon simulation (cyan line), and the envelope encompassing all simulations performed for the Sobol analysis (light-cyan ribbon). The green horizontal line on (b) represents the approximate discharge triggering floodplain activation.



6.2 Calibration insights from the SWAT-Amazon method

590 In the hydraulic routing method (section 2.4), discharge calibration is insensitive to parameters n and B . Reducing n increases u and Q (Eq. 5), decreasing the water volume in the reach and thus h (Eq. 1). Consequently, u and Q also decrease in proportion. Therefore, when $h < h_f$, discharge calibration relies exclusively of the default SWAT rainfall-runoff model. However, u and h are directly related to the parametrization of n and B and are the primary variables in the transport capacity model utilized here (Section 2.5). Q is also a key variable (Eq. 9). As B is poorly-known, it was excluded from Equations (7) and (9) of the transport capacity model by replacing $B h$ with Q/u and $u B$ with Q/h . With regard to n , another significant source of uncertainty, it should be noted that this variable is not included in the transport capacity equations. This is because u is used and calibrated prior to the Q_s computation.

6.2.1 Calibration strategy for a super station

According to the previous analyses, the calibration strategy for stations with robust, long-term hydro-sediment monitoring is:

- 600 a) Start by calibrating Q in each reach, for $h < h_f$ only, using the SWAT's default hydrologic parameters.
- b) Calibrate Q , considering floodplain effects, using h_f , C_{nfp} and k_{fp} (or θ_{fp}).
- c) Calibrate u and h by adjusting n and B only; Q is unaffected by this calibration.
- d) Check the relationships $Q(h)$ and $u(h)$, revisiting step c if needed.
- 605 e) Compute the $Q_s(h, u, Q)$, independently of n and B . If necessary, adjust Q_s using parameters in Table 1, particularly d_b , the most sensible parameter.

It is important to emphasize that the optimal calibration for water discharge may not align with the best calibration for water level, velocity, and sand load time series. A compromise must be made.

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6.2.2 Calibrating virtual and low-data stations with satellite altimetry

The sensitivity of the model to n and B emphasizes the importance of h (Eqs. 1, 2, 4) in the routing module, particularly for floodplain inundation and dewatering triggers (Section 2.4), as already discussed above. For ungauged reaches (5, 23, 22) in the plain, n and B can be calibrated using relative water levels (hydraulic radius) derived from spatial altimetry and shifted by a constant offset to match simulations (Fig. 10). The Manning coefficient is supposed already known to a reasonable degree, as coherent values of $1/45 \pm 0.0005 \text{ s m}^{-1/3}$ has been found across all the other plain's sub-basins. Given the large sub-basins and significant channel lengths, bed slope errors are assumed to be smoothed, and therefore S_b also well assessed. Thus, velocities can be reliably determined if B is calibrated through h calibration. The calibration for reaches (5, 23, 22) was also guided by looking for the best simulation at Pucallpa and Requena.

Finally, implementing an integrated approach combining modelling and remote sensing for monitoring hydro-sediment fluxes would require revisiting hydrological network operations. Specifically, a few gauging campaigns (around four, strategically timed along the annual hydrograph) could be sufficient for model calibration. Additionally, synchronizing gauging with satellite altimetry would enable the construction and simulation of robust rating curves.

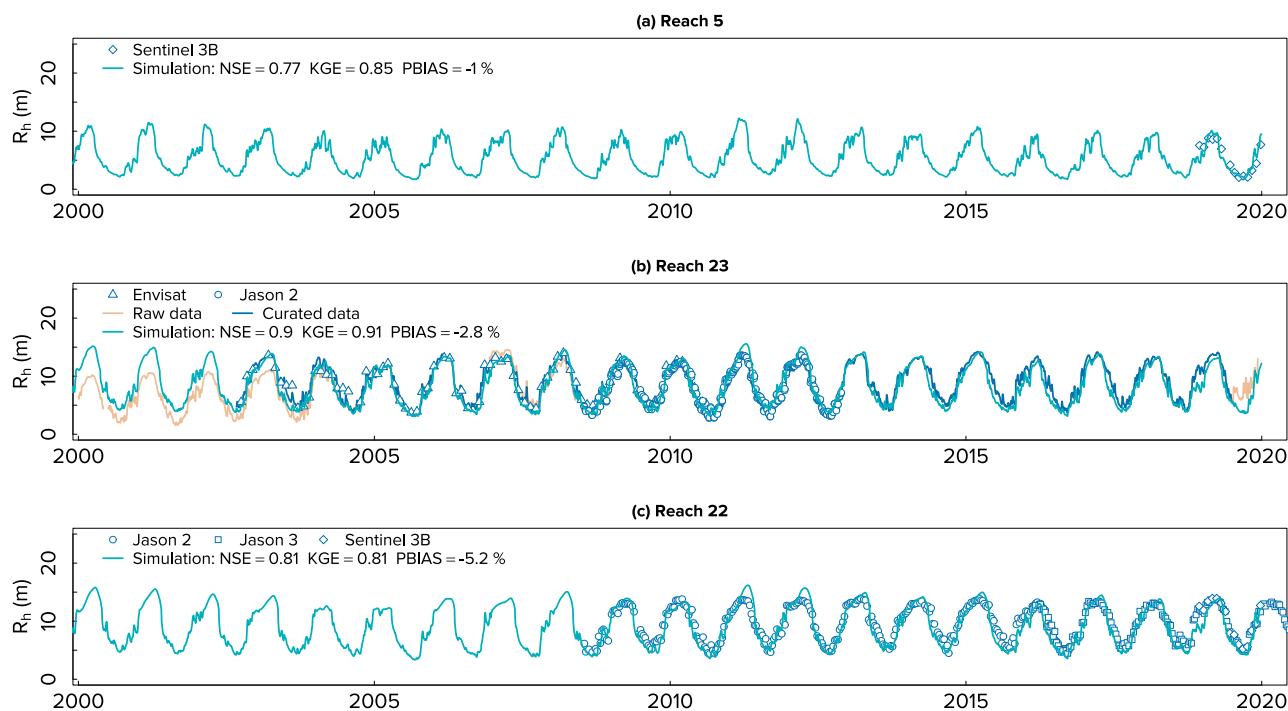


Figure 10: Calibration of ungauged reaches using satellite altimetry: Satellite-derived water levels (blue points) shifted by a constant offset to match simulations: triangles (Envisat), circles (Jason-2), squares (Jason-3), diamonds (Sentinel-3B). Conventional *in situ* observations appear as an orange line, while the curated series incorporating satellite altimetry is in blue. SWAT-Amazon simulations are shown in cyan. (a) Reach 5 (virtual station S3B331-R5), (b) Reach 23 (Contamana), (c) Reach 22 (virtual station JA204-S3B310-R22).



640 6.3 Modelling limitations

This study adopts a hydraulic representation of the river, modelling the floodplain as a reservoir. Rapid exchanges through secondary channels are neglected, and their influence are approximated on discharge propagation through a little adjustment of the Manning's coefficient. Guilhen et al. (2022) used a previous version of SWAT-Amazon with a slightly modified floodplain configuration, allowing 1d floodplain flow through its whole cross-section with a Manning-Strickler equation.
 645 However, in addition to being almost impossible to calibrate, the floodplain flow being extremely sensitive to the floodplain Manning coefficient because of its huge cross-section, the changes in term of results were extremely negligible: less than 1% of the water discharge in the main channel, well below expected model precision.

Another key assumption in SWAT-Amazon is the absence of backflow from the floodplain to the main channel during the rising limb. Observations in the Amazon basin support this, as low-gradient floodplain channels often experience tributary
 650 flow blockage, with the Tapajós River in Brazil being an extreme example. However, during deflooding, floodplain drainage dynamics differ from flooding processes. In SWAT-Amazon, these dynamics are governed by floodplain cross-section geometry. Introducing a Maillet-type law could improve recession representation, but discrepancies in timing appear mainly driven by rainfall variability, masking this effect. Thus, no modifications were made.

Floodplain sediment concentration during deflooding was assumed constant. Nevertheless, the unique monitoring of sediment
 655 concentration in a floodplain channel, performed at Lago Grande de Curuai (Brazilian Amazon) shows a decrease from ~800 mg l⁻¹ during low-waters when the floodplain water flows back into the main channel to ~40 mg l⁻¹ when the channel is under the influence of the Amazon River, with a mean concentration of 135 mg l⁻¹ (Moreira-Turcq et al., 2013). This value appears to be relatively close to that calibrated for C_{bk} (for the sand fraction only) at sub-basins 23, 22, and 21 (~180 mg l⁻¹).

A dynamic law linking C_{bk} to water level could refine peak resuspension modelling but requires concentration data for
 660 calibration. Alternatively, Fagundes (2023) treated the floodplain as a homogeneous sediment reservoir, balancing suspended fluxes with settling. However, this method applies only to clay and silt, excluding sand, and the settling law can be challenging to calibrate in the absence of observational data. Given floodplain water storage timescales (months), most fine sediments likely settle before to be remobilized. Furthermore, as underscored in the introduction, sediments resuspended during recession can be centuries to millennia old, indicating long-term floodplain recycling rather than short-term remobilization.

665 Lastly, the decision to distinguish E_{bk} from E_{bed} is supported by calibration results. Allowing streambed erosion ($k_{bed} > 0$) does not properly reproduce the sand resuspension peak, whereas adjusting C_{bk} provides a better fit. This result strongly supports the assumption that sand resuspension is entirely driven by floodplain water recession, *i.e.* by floodplain recycling, banks and bars erosion, rather than riverbed erosion, which remains negligible ($K_{bed} \ll 1$) in the main stem.

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6.4 Hydro-sediment dynamics in Amazonian foreland systems: long-term insights

The robustness of the results enables precise water and sediment balances, distinguishing fine particles from sand, at each virtual and conventional station (Fig. 11). This is achieved at an unprecedented spatiotemporal resolution over the periods 1983–2019 (37 years) for water and sand fluxes and 2000–2019 (20 years) for fine sediments.

6.4.1 Dynamics of flood waves, flooding, and sediment transport

Water and sediment peak fluxes at the Andean outlet occur in February (Fig. 11), declining sharply from March to May as Andean precipitation decreases. The sediment flux at the basin outlet is strongly correlated with Andean discharge (Fig. 12c), confirming the Andes as the primary sediment source. This results from intense erosion along the Eastern Cordillera and Sub-Andean thrust belt, where high precipitation erodes Paleozoic and Mesozoic sedimentary sequences incorporated into Cenozoic deposits eastward as the orogenic prism advances through crustal shortening and thickening (McQuarrie, 2002a, b; Espurt et al., 2008; Gautheron et al., 2013; Pfiffner and Gonzalez, 2013). Additionally, the Central Andes' convex hypsometric profile (Montgomery et al., 2001) promotes deep fluvial incision.

At the lowland outlet, discharge peaks two months after the Andes (April–May) delayed primarily by alluvial plain storage, diffusive flood wave propagation, and runoff from the northern part of the basin, where seasonality, driven by the South American monsoon system (Garreaud et al., 2009), is less pronounced than in the south. This upstream-to-downstream flooding dynamic leads to progressive floodplain inundation (February–April, Fig. 11b, c) and gradual drainage (March–July). Floodplain backflow (Fig. 12a) drives sediment remobilization from April to August (Fig. 12b, d). This secondary sediment source accounts for 22% of the total suspended sediment load at the basin outlet and can have significant impacts on river dynamics, with pronounced effects during major floods (*e.g.*, 2012, Fig. 6c). Floodplain flows, comprising precipitation, river inflows and backflows, are more significant than the water entering the floodplain from the main river (Fig. 12a), reinforcing the assumption that floodplain waters are blocked by the main channel during the rising limb.

In the following analysis, the fine sediment load of the Pachitea Basin was estimated using a rating curve derived from the relationship between water discharge and measured fine sediment load, based on simulated flows. For the other Andean sub-basins (8, 9, 3), fine sediment load was regionalized according to the drainage area of their respective Andean catchments. The Ucayali floodplain consists of three geomorphological compartments (Fig. 2b), each characterized by distinct processes:

Lagarto – Pucallpa (Box 1, Fig. 2b): A major thrust fault system involving the crystalline basement exerts significant tectonic control over the narrow sedimentary basin wedged between the young Shira Mountains and the Fitzcarrald Arch (Espurt et al., 2008; Pfiffner and Gonzalez, 2013; Gautheron et al., 2013). Here, 36% of the $420 \cdot 10^6 \text{ t yr}^{-1}$ of Andean sediment accumulates, with sand deposition ($\sim 100 \cdot 10^6 \text{ t yr}^{-1}$) accounting for 66%. This results in alluvial plain aggradation at 3.7 mm yr^{-1} ($9,750 \text{ t km}^{-2} \text{ yr}^{-1}$), highlighting strong neotectonic control. In this compartment, water storage reaches 1.5 km^3 in March in average, with flood durations (days with $h \geq h_f$) varying from 0 to 56 days in sub-basins 19 and 5, respectively. Consequently, daily-scale Andean flash floods superimposed on the annual monomodal flood wave are rapidly attenuated, while downstream, water



storage increases as the floodplain expands. This process also dampens high-amplitude fine sediment peaks ($-51 \cdot 10^6 \text{ t yr}^{-1}$) through the activation of flood channels. These channels, remnants of the river's past mobility, further enhance the buffering effect. The sudden floodplain expansion at sub-basin 14 as the river course is released from the tectonic constraint imposed by the Shira Mountains reduces fine sediment loads (Fig. 11d), while high meander migration rates (Li et al., 2023) indicate active lateral erosion and deposition.

Pucallpa – Contaya Arch (Box 2, Fig. 2b): This section is a large piggyback basin located between the Andean wedge top and the Moa Divisor thrust fault system, with increased floodplain storage and longer flood durations (78–125 days). In March, it retains 10.4 km^3 of water ($\sim 20\%$ of discharge). Fine sediment load slightly increases ($+7 \cdot 10^6 \text{ t yr}^{-1}$) due to Andean inputs, mainly from the Pachitea sub-basin, where one of the major Andean precipitation hotspots drives intense erosion rates ($\sim 4,020 \text{ t km}^{-2} \text{ yr}^{-1}$, $\sim 1.5 \text{ mm yr}^{-1}$). Upstream sub-basin 22, despite the significant floods experienced, sands are transported with a slight retention ($-4 \cdot 10^6 \text{ t yr}^{-1}$, 4% of load) due to the steeper bed slope than to the following compartment, which sustains transport capacity. Floodplain drainage and sediment resuspension gain importance in this segment.

Contaya Arch – Requena (Box 3, Fig. 2b): the river shifts northeastward into the extensive Marañón Foredeep, a tectonic funnel terminating at the Iquitos Arch. In April, water storage reaches 9.4 km^3 . The maximum flood duration reaches 125 days. The flood impacts both fine sediments and sands (Fig. 11), however with different processes. Fine sediment deposition ($-16 \cdot 10^6 \text{ t yr}^{-1}$, 7% of the load) is driven by prolonged floodplain residence time, while sand sedimentation ($-11 \cdot 10^6 \text{ t yr}^{-1}$) results from energy dissipation when the floodplain is active. This process, never documented before in Amazonian Rivers, captures approximately 14% of the sand flux.

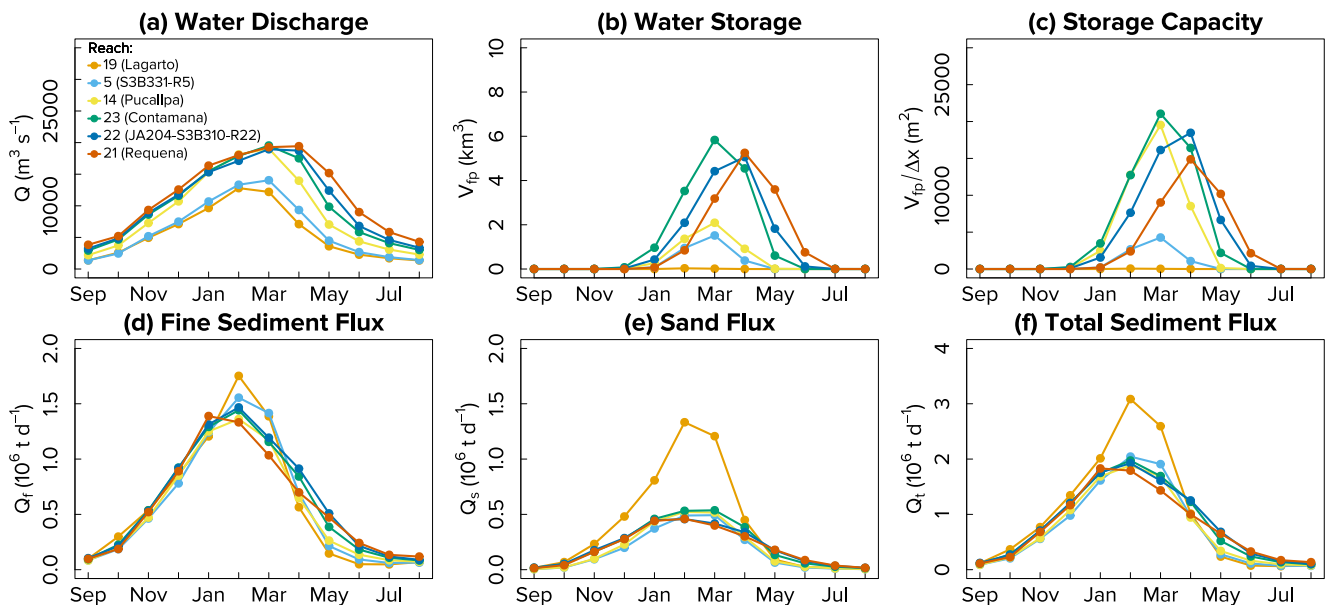


Figure 11: Interannual water and sediment balances along the Ucayali River. (a) Water Discharge (b) Floodplain water storage (V_{fp}) in km^3 . (c) Normalization of V_{fp} by the reach's length (Δx) for cross-sub-basin comparison. (d) Fines suspended sediments. (e) Suspended sand fraction. (f) Total suspended sediment load.



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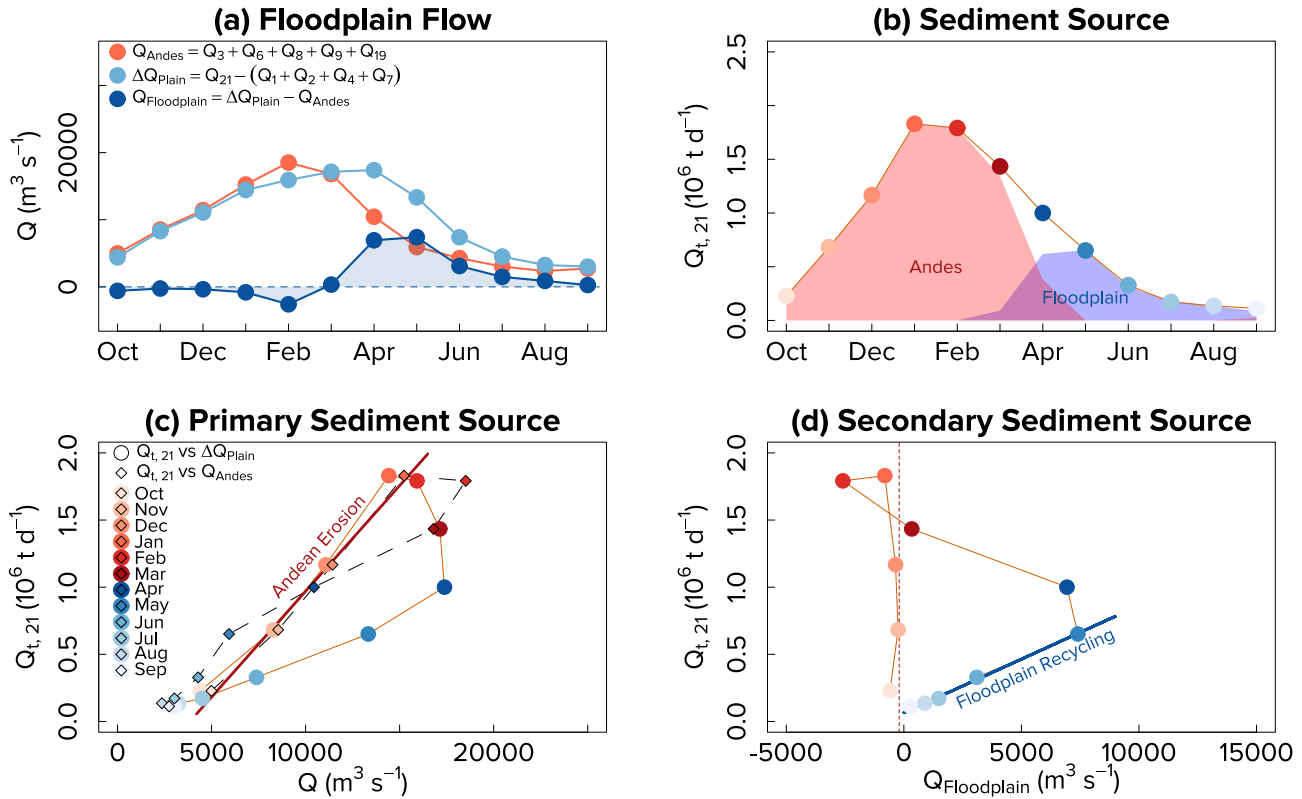


Figure 12: Identification of sediment sources at the interannual scale. (a) Water discharge components: Andean Basin outflow (Q_{Andes}), basin outlet discharge excluding lateral contributions (ΔQ_{plain}) and floodplain flows ($Q_{Floodplain}$), all in $m^3 s^{-1}$. (b) Decomposition of total sediment flux at the basin outlet $Q_{t,21}$ ($10^6 t d^{-1}$) by sediment sources: red for Andean erosion, blue for floodplain recycling. (c) Relationship between $Q_{t,21}$ and ΔQ_{plain} (circles) and Q_{Andes} (diamonds). The red regression line represents the relationship between Q_{Andes} and $Q_{t,21}$ during rising limb months (d) Relationship between $Q_{t,21}$ and $Q_{Floodplain}$. The blue regression line illustrates the relationship between Q_{Andes} and $Q_{t,21}$ during recession months.

6.4.2 Long-term trends

No significant trends were observed in flood duration, floodplain water volume, discharge extremes, or sediment fluxes in sub-basins 21, 22, and 23 for the simulation period. This suggests that the effects of global and regional changes at the Ucayali Basin scale are still negligible, contrasting with studies of the broader Amazon Basin (e.g. Marengo, 2004; Gloor et al., 2013; Li et al., 2020; Fleischmann et al., 2023). However, most of these studies focus on a very limited number of sites with long-term data. This highlights the relevance of the integrated approach proposed in this study for generating new long-term analyses of regional trends and providing a more detailed perspective on the Amazon's response to global changes. Espinoza et al. (2009) is the only study on Ucayali discharge evolution, based on inconsistent 1996–2005 data, now curated here. Figueiredo et al. (2024) and Polasky et al., (2025) identified mixed precipitation trends in the Ucayali Basin, respectively for the 1982–2023 and 1980–2022 period, which may explain the lack of clear trends here. Nevertheless, Guimberteau et al. (2013; 2017)



project increased precipitation in the western Amazon by the century's end, potentially leading to greater inundations, sediment deposition, and floodplain recycling, which may alter river mobility, particularly during extreme events. This could impact
745 bedload transport, leading to bed aggradation, which would facilitate overflow and then intensify exchanges.

7 Conclusion and perspectives

This study presents the first integrated approach for monitoring hydro-sediment fluxes in the Amazon basin, combining remote sensing data with hydrological and hydraulic modelling. It provides the first long-term (37–20 years) mass balance estimates for the foreland region and discriminates fine sediments from sand loads. The SWAT-Amazon model, specifically designed
750 for Amazonian rivers, achieves unprecedented accuracy in simulating daily water levels, velocity, discharge, and suspended sand load. This framework enhances the understanding of hydro-sediment dynamics in the Ucayali Basin, with the implementation of virtual stations to complete the conventional network. The Andean Ucayali River exports $455 \cdot 10^6 \text{ t yr}^{-1}$ of sediment (40% sand), corresponding to an erosion rate of $2,350 \text{ t km}^{-2} \text{ yr}^{-1}$ (0.9 mm yr^{-1}). In March, the sedimentary basin retains 19.1 km^3 of water ($\sim 38\%$ of total discharge for the same month), trapping 36% of Andean sediment input (65% sand).
755 Fine sediments are transferred downstream with a net loss of -21% , attributed to the dampening effect of Andean flash floods at the floodplain entrance. The Ucayali delivers $290 \cdot 10^6 \text{ t yr}^{-1}$ of sediment to the Amazon River, 26% of which is sand ($74 \cdot 10^6 \text{ t yr}^{-1}$), making it the second-largest sediment contributor (25%) after the Madeira River, despite a drainage basin four times smaller and a contribution of 6% of the Amazon's discharge to the Ocean.

Floodplains play a critical role in water-sediment routing and as a secondary sediment source (22% of the total suspended
760 load). This study is the first to identify a sand sedimentation process in such large river, where floodplain activation reduces transport capacity, capturing $\sim 14\%$ of the sand flux at peak discharge and thus decoupling water discharge from sand flux. These findings highlight the need to rethink hydrological network operations through the concepts of super and virtual station, combining long-term, high-quality, data with satellite measurements to calibrate and validate this integrated approach. Synchronizing targeted hydro-sediment gauging campaigns with satellite missions would significantly enhance model
765 performance. Moreover, the study underscores the importance of regionally focused and long-term analyses over broad, continental-scale assessments, which often introduce high uncertainties and may misguide sustainable development and mitigation strategies.

Looking ahead, this approach could be extended to other Amazonian basins facing similar hydro-sediment monitoring challenges, with model refinements and targeted calibration-validation campaigns. The assimilation of satellite-derived fine
770 sediment concentrations would significantly improve floodplain modelling by defining Floodplain Hydrological Response Units based on morphology (lakes, channels, etc.). By reconciling fine sediment mass balances (*i.e.*, adjusting discrepancies between modelled and observed satellite data), this method would provide a proxy for water residence time in floodplains. This key variable could facilitate the development of bio-hydro-geochemical models, improving our understanding of the Amazon's role in global biogeochemical cycles and its response to environmental changes.



775 In conclusion, the integration of remote sensing and hydro-sediment modelling offers a powerful and scalable framework for
monitoring hydro-sediment fluxes in large river basins, with significant implications for research, environmental management
and policy-making.

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Appendices

A1: List of notations

Main subscripts and superscripts

Variable name	Definition
$\cdot_i$	Reach number
$\cdot_t$	Day of the simulation
$\cdot_{t+\Delta t}$	Day after the day of simulation
$\cdot'$	Grain-related bed shear stress relative variable
$\cdot_c$	In-bank flow configuration (<i>i.e.</i> , without floodplain)
$\cdot_{cf}$	Flow with active floodplain

785 Main variables

Variable name	Unit	Definition
h	(m)	Mean water height
z	(m)	Height above the bed
h_f	(m)	Water height that triggers the floodplain activation
h_{ch}	(m)	Water height that ends the additional bed roughness influence
B	(m)	Channel Width
S_b	(m m ⁻¹)	Channel bed slope
S_w	(m m ⁻¹)	Water-surface slope
S_f	(m m ⁻¹)	Energy gradient or friction slope
n	(s m ^{-1/3})	Manning coefficient
ζ_n	(–)	Main channel flow resistance correction factor, used when the floodplain is active
A_h	(m ²)	Wetted area
P_h	(m)	Wetted perimeter
R_h	(m)	Hydraulic radius
V	(m ³)	Volume of water stored in a reach
W_{fp}	(m)	Floodplain width
ΔV_{fp}	(m ³ d ⁻¹)	Water volume that flows during the simulation day from the main channel to the floodplain
V_s	(t)	Suspended sand mass stored in the reach
V_T	(m ³)	Water volume transmitted by the river to the unsaturated water table
V_E	(m ³)	Water volume evaporated from the reach
V_R	(m ³)	Water volume of surface, sub-surface and base runoff in a sub-basin
D_{ovbk}	(t d ⁻¹)	Overbank deposition term
E_{bk}	(t d ⁻¹)	Channel bank erosion term
E_{bed}	(t d ⁻¹)	Riverbed erosion term
Δx	(m)	Reach length
Q	(m ³ s ⁻¹)	Water discharge
u	(m s ⁻¹)	Water velocity
C_s	(t m ⁻³)	Suspended sand concentration
C_{bk}	(t m ⁻³)	Concentration of bank and bar inputs (constant)
C_b^*	(m ³ m ⁻³)	Bedload reference concentration



Q_s	(t d ⁻¹)	Suspended sand load
Q_s^*	(t d ⁻¹)	Transport capacity for suspended sands
ΔQ_s^*	(t d ⁻¹)	difference ΔQ_s^* between the transport capacity Q_s^* and the sand load Q_s
P_s	(–)	Rouse number for the suspended sands fraction
w_s	(m s ⁻¹)	Suspended sand particle settling velocity
d_s	(m)	Arithmetic mean diameter of suspended sands
d_b	(m)	Arithmetic mean diameter of riverbed sands
d_*	(–)	Dimensionless grain size
θ_{cr}	(–)	Critical dimensionless shear stress threshold
θ	(–)	Shield's dimensionless shear stress parameter
f	(–)	Darcy-Weisbach roughness factor
u_*	(m s ⁻¹)	Shear velocity
β_s	(–)	Ratio of suspended sand to eddy diffusivity
α_s	(–)	Ratio between mean concentration and index concentration for sand particles
g	(m s ⁻²)	Gravitational acceleration
ν	(m ² s ⁻¹)	Kinematic viscosity
ν	(m ² s ⁻¹)	Kinematic water viscosity
σ	(–)	Sorting coefficient used to determine k'_s : $k'_s = \sigma d_b$
ρ_w	(kg m ⁻³)	Water density
ρ_s	(kg m ⁻³)	Sediment density
s	(–)	Relative sand density
k_s	(m)	Nikuradse equivalent roughness height
κ	(–)	Constant of Von Kármán
k_{fp}	(–)	Coefficient to determine the floodplain width
θ_{fp}	(rad)	Angle of the floodplain riverward slope (case of floodplain with triangular cross-section)
C_{nfp}	(–)	Coefficient for increasing the flow resistance in the main channel when the floodplain is active
C_{nch}	(–)	Coefficient for increasing the flow resistance in the main channel when low waters
K_{bed}	(–)	Main channel susceptibility to erosion (riverbed only), between 0 and 1
η	(–)	Correction exponent for transport capacity when the floodplain is active

Code availability

The codes developed and used in this study are openly available at: <https://github.com/william-santini/SWAT-Amazon>.

Data availability

The data supporting the findings of this study are available from the following repository (Santini et al., 2018):

790 <https://doi.org/10.6096/DV/CBUWTR>. Additional datasets, such as water level records, suspended sediment concentration time series, and others, are available from the corresponding author upon request and on the CZO HYBAM website: <https://hybam.obs-mip.fr/> (last accessed: 6 June 2025).



Interactive computing environment

795 The R notebooks developed and used in this study are openly available at <https://github.com/william-santini/SWAT-Amazon>.

Author contributions

WS conceived the study, conducted the investigation, developed the methodology, and performed the data analysis. WS and ADY developed the codes (R and Fortran). WS, BC, JLC, JMM, WLC, JR, ADY, and JMC contributed to data interpretation and manuscript preparation. JR, WS, and JMM processed the remote sensing data. WS and JJPA were responsible for
800 hydrological data acquisition. WS supervised the laboratory analyses. JMM, WS, and WLC secured funding and were responsible for project administration and supervision.

Competing interests

The authors declare that they have no conflict of interest.

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