

HEMINGS. An integrated tool for hydrological, geomorphological, and environmental modeling

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Abstract The HEMINGS (Hydro-Environmental ModellING tool for GeoSystems) model is a physically-based, distributed hydrological model designed for the integrated simulation of hydrological, geomorphological, and environmental processes in headwater watersheds. Development of the model began with the goal of providing a flexible, modular, and scientifically robust tool capable of explicitly representing the main processes that control runoff generation, infiltration, evapotranspiration, soil erosion, sediment transport, and hydrological connectivity with forestry applications. Since then, HEMINGS has evolved to incorporate new processes, with simulation capabilities progressively presented and validated in scientific publications. This document compiles and systematizes the theoretical formulation, conceptual foundations, and complete computational framework underpinning the model. Its objective is to provide a rigorous description of the physical and numerical foundations of HEMINGS, as well as its modular structure, which allows for its application to a wide variety of problems related to hydrology, erosion, sediment transport, hydrological-forestry restoration, and the assessment of risks associated with extreme events in headwater and mountain watersheds.

Keywords: Hydrological modelling; Headwaters; Soil erosion; Fluvial processes; Forestry; HEMINGS.

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1 Introduction

Watershed management faces major challenges in water resources planning related to the impacts of climate change and global change, particularly in semiarid regions and headwater catchments (Giorgi and Lionello, 2008; Simonneaux et al., 2015; Cantalejo et al., 2024). Topographic forcing drives runoff generation, soil loss, and rapidly responding fluvial transport processes that can produce significant downstream impacts, posing a major challenge for water resources management (Nadal-Romero et al., 2013; Millares and Moñino, 2020). In addition, snow dynamics play an important role in both hydrological and erosion and sedimentation processes (Langlois et al., 2005; Millares and Moñino, 2020), particularly in semiarid environments. From a fluvial perspective, the torrential nature of rivers in headwater catchments controls sediment transport processes, with sediment being stored downstream in the form of alluvial fans (Santi et al., 2011). This storage, in turn, constitutes an important source of sediment inputs to the main river, with significant implications for morphodynamic processes and interactions with infrastructure (Turowski et al., 2010; Millares and Moñino, 2018).

Interest in assessing these processes in mountainous regions has increased significantly in recent years because of the sensitivity of these areas to climatic variability. Changes in hydroclimatic forcing in mountainous environments can produce seasonal variations in runoff, ranging from spring snowmelt to direct winter runoff, with substantial variations

in sediment yields (López-Moreno, 2005; López-Moreno et al., 2008). Estimating these impacts is essential for developing effective adaptation and mitigation strategies (MITECO, 2020). The increasing frequency of other extreme events, such as droughts and wildfires, is expected to further accelerate desertification processes, resulting in major changes in land-use patterns and substantial increases in runoff and erosion rates. Assessing these scenarios requires forecasting methodologies and tools capable of reproducing the physics of the underlying hydrological and hydraulic processes, as well as the uncertainty associated with the driving forces (Raje and Mujumdar, 2010). The question of how watersheds should be managed and which corrective, mitigation, and adaptation measures are most appropriate in the context of climate and global change must be addressed from a probabilistic, risk-based, and integrated perspective.

Over the past several decades, advances in computational capacity, the availability of geospatial information, and Earth observation techniques have helped overcome many of the limitations associated with traditional approaches. In contrast to hydrological calculation methods based on lumped representations of processes, current approaches seek to explicitly incorporate the spatial and temporal variability of precipitation, topographic characteristics, soil properties, vegetation cover and its temporal evolution, and the dynamics of surface flows. Among these tools, physically based, distributed hydrological models are particularly relevant because they combine runoff generation with erosional and fluvial processes throughout the watershed system. These

models provide a physically based representation of the processes governing runoff generation and propagation, soil erosion, and sediment transport, facilitating the identification of critical areas and the assessment of management, mitigation, and adaptation measures. This capability enables a more realistic description of interactions among the atmosphere, land surface, and subsurface, thereby improving our understanding of watershed hydrological response (Beven, 2012). In addition, these models provide a more comprehensive representation of the watershed and the interactions among its parameters and state variables.

Despite their limitations, including the large number of parameters required (Immerzeel and Droogers, 2008), difficulties in representing the spatial variability of soil properties (Beven, 2006), challenges associated with calibrating and validating distributed model outputs Takken et al. (1999), and uncertainty related to geomorphological thresholds within the watershed (Millares et al., 2012, 2019), these models have considerable potential as decision-support tools for watershed management (Van Rompaey et al., 2005; Bussi et al., 2014; Fatichi et al., 2016; Millares et al., 2019, 2022,?). Explicit spatial representation makes it possible to identify runoff source areas, erosion hotspots, and regions that are particularly sensitive to environmental changes—features that are difficult to capture using lumped or semidistributed approaches (Abbott et al., 1986; Refsgaard and Storm, 1996). Because the parameters have a direct physical interpretation, variables related to vegetation, soil properties, and meteorological conditions can be modified to assess watershed responses under different environmental scenarios (Fatichi et al., 2016). These models also offer important advantages for studying extreme hydrological events. Their ability to explicitly represent the spatial dynamics of runoff generation and propagation improves the simulation of flash floods, torrential flows, and high-intensity precipitation events, which are particularly relevant in headwater and mountainous catchments (Moradkhani and Sorooshian, 2009). Similarly, they facilitate the analysis of the mechanisms controlling hydrological and sediment connectivity, providing valuable information for natural hazard management and the conservation of fluvial ecosystems (Millares et al., 2022).

Several models of this type have been applied to watersheds with different characteristics and climatic conditions worldwide (e.g., MIKE SHE (Abbott et al., 1986), SWAT (Arnold et al., 1998), WEPP (Flanagan et al., 2001), LISEM (De Roo et al., 1996), WiMed (Herrero et al., 2014), and HEMINGS (Millares et al., 2019)). It is important to emphasize that no model is universally superior; each tool involves a trade-off among physical realism, structural complexity, spatial and temporal scales, parametric uncertainty, and computational cost (Fatichi et al., 2016).

In addition, particularly in headwater and mountainous catchments, assessing geomorphological impacts and their implications for decision-making requires an understanding of the continuum of erosion and sediment transport processes from a broad, watershed-scale perspective (Owens, 2005). From soil erosion caused by raindrop impact, through concentrated flow in rills and gullies and the downstream transport of eroded material, to hydrodynamic processes within river channels, sediment transport must be understood as a sequence of interconnected processes. The continuity and (dis)connectivity of these processes among subsystems, together with sediment availability throughout the drainage

network, control erosional responses to anthropogenic impacts and global change (Hoffmann, 2015). However, studies are often conducted at the local scale, focusing on specific systems (hillslopes, channels, lakes, structures, etc.) using widely differing methodologies and scientific disciplines, with limited integration among them.

Empirical or parametric models, such as the Universal Soil Loss Equation (USLE), including its revised (RUSLE) and modified (MUSLE) versions (Wischmeier and Smith, 1978; Renard, 1997), even when combined with an additional calculation of the downstream sediment delivery ratio (SDR) (Ferro and Porto, 2000; Van Rompaey et al., 2001), do not account for some dominant processes, such as gully erosion or small-scale landslides, despite their importance (Casali et al., 1999; Poesen et al., 2003; Millares et al., 2019). Furthermore, they do not account for the spatial and temporal variability of all driving forces at the event scale or under continuous simulation conditions (Orlandini and Rosso, 1996; Olivera and Maidment, 1999; Millares and Moñino, 2018; Millares et al., 2018), nor do they provide information on flow, transport, and sedimentation processes along the drainage network (Smith et al., 1995; Johnson et al., 2000).

At the river scale, hydrodynamic models enable the simulation of the physics of water movement across the land surface while accounting for the spatial variability of flow depth and velocity. They are particularly useful for analyzing floods and sediment transport along river reaches. In this field, advances in numerical methods and improvements in computational capacity over recent decades have led to significant progress (Wu, 2007), enabling researchers to constrain the magnitude and spatial distribution of sedimentation and bed scour processes, assess infrastructure-related impacts, and investigate the influence of vegetation on morphodynamic processes, among many other applications.

Complementarily, the development of remote sensing technologies, satellite imagery, and drone-based airborne platforms has transformed modern hydrology and hydraulics by providing increasingly frequent and spatially resolved observations of the Earth's surface over extensive areas. Remote sensing techniques enable the acquisition of spatially distributed information on precipitation, soil moisture, evapotranspiration, vegetation cover, flood extent, water levels, river morphology, and sediment transport. This capability has driven the development and application of distributed hydrological and hydraulic models, improving the representation of the spatial and temporal heterogeneity of the processes controlling runoff generation, erosion, and fluvial dynamics (Patel et al., 2025). The availability of time series from missions such as Landsat, Sentinel, MODIS, SMAP, GPM, GRACE, and SWOT (Kossieris et al., 2024) makes it possible to assess how watershed hydrological conditions evolve in response to climate change, extreme events, land-use changes, and geomorphological alterations.

The combination of these tools currently constitutes one of the fundamental pillars of modern hydrology. Integrating numerical models with data derived from remote sensing not only improves the representation of natural processes but also enhances the predictive capabilities of modeling systems and reduces the uncertainty associated with their application. Implementing these tools in real-world settings as decision-support systems enables a more scientifically rigorous assessment of extreme events, the evaluation of future climate and land-use change scenarios, and the design of

sustainable management strategies for water resources and fluvial ecosystems.

The HEMINGS model, Hydro-Environmental ModelING tool for GeoSystems (Millares et al., 2019, 2022) (Fig. 1), is a distributed, physically based hydrological model developed to simulate hydrological, erosional, fluvial, and environmental processes, with a particular focus on headwater catchments at the event scale.

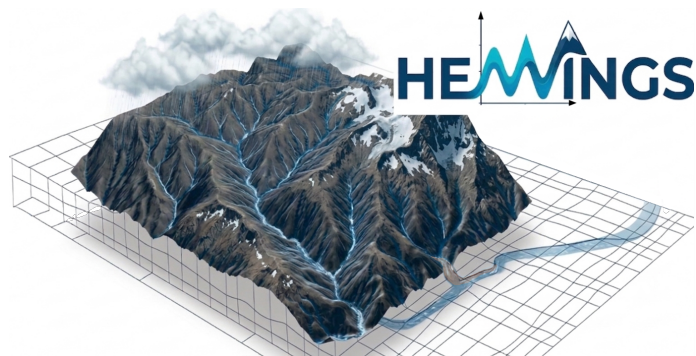


Figure 1. Conceptual diagram and logo of the HEMINGS model.

The finite-volume numerical scheme developed for overland water and sediment flows incorporates several simulation options for runoff routing, including the kinematic wave, diffusion wave, and diffusion wave with an additional diffusive term. For hydrological calculations, the model accounts for the spatial distribution of precipitation using an inverse distance weighting (IDW) scheme, snowfall, infiltration processes based on an optimized Green-Ampt model (Green and Ampt, 1911; Esteves et al., 2000), rainfall interception and distributed evapotranspiration processes derived from satellite imagery (Gavilan et al., 2007; Van der Knijff et al., 2000), and baseflow contributions at specific locations along the river network (Millares et al., 2009).

For hillslope erosion calculations, the model simulates raindrop splash erosion (Ben-Hur and Wakindiki, 2004) and concentrated-flow erosion processes while accounting for variable rill density (Mancilla et al., 2005). In addition, the HEMINGS model can assess gully erosion and headward gully extension using field observation data (Millares et al., 2019). The model incorporates a fluvial sediment erosion and transport module with several bedload transport formulations (Wong and Parker, 2006; Parker, 1990; Powell et al., 2001) and total-load formulations (Bagnold, 1966; Engelund and Hansen, 1967; Brownlie, 1981), depending on the dominant transport regime, as well as bed evolution governed by erosion and deposition processes.

The one-dimensional numerical scheme involves lower computational costs than two- or three-dimensional approaches, enabling the HEMINGS model to perform Monte Carlo simulations to address the substantial uncertainty associated with hydrological modeling. Figure 2 shows the results of 100 Monte Carlo simulations conducted in the Guadalfeo River catchment (Granada, Spain), in which the spatial variability of the parameters k_s and ψ , as well as the variability of river sediment grain size, was evaluated. The results show empirical probability density functions of peak discharge and sediment transport volume obtained using two bedload transport models.

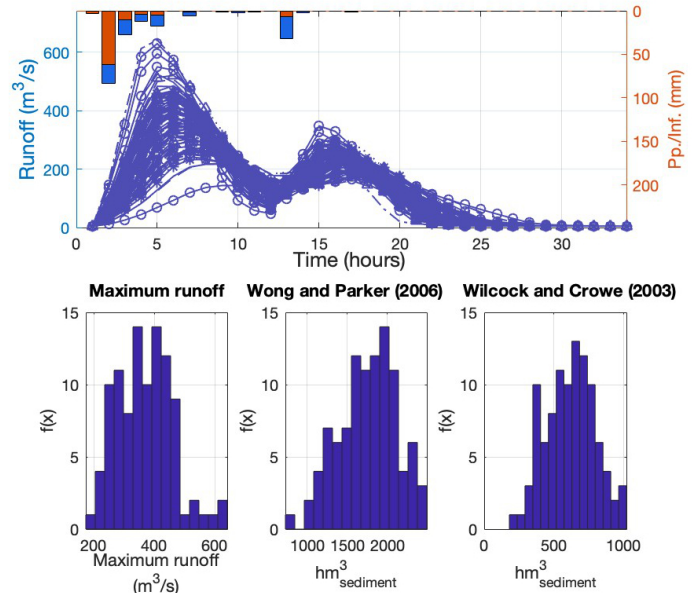


Figure 2. Monte Carlo simulations of runoff and sediment transport processes using the HEMINGS model.

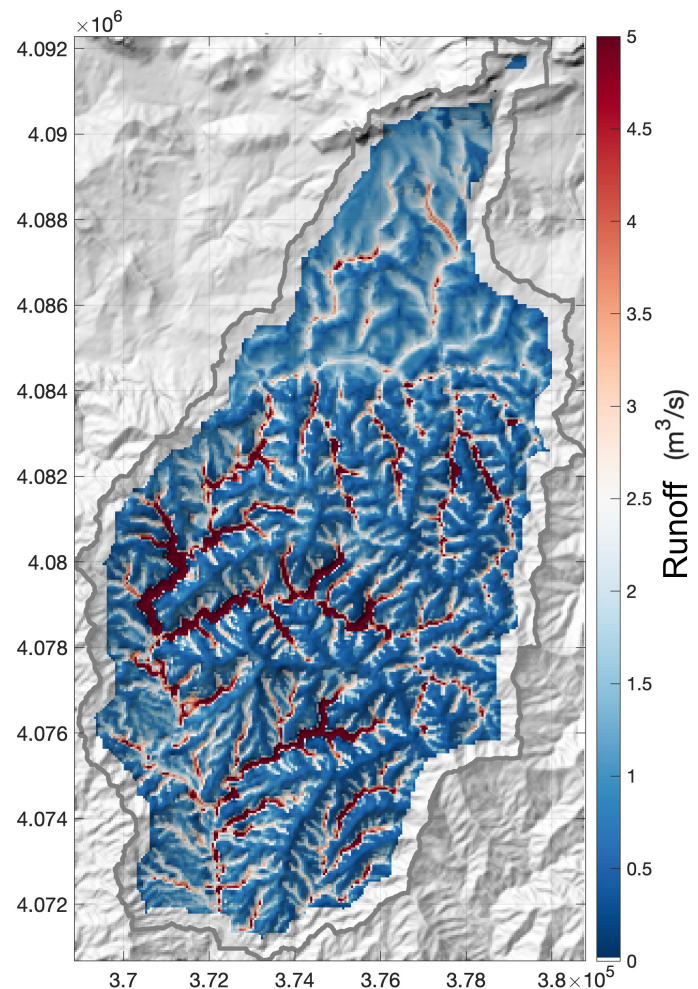


Figure 3. Comparison of two-dimensional (left) and one-dimensional (right) hydrological modeling results obtained with the HEMINGS model.

Optionally, the HEMINGS model also includes the capability to perform two-dimensional simulations. This subroutine, which is currently under development, offers advantages for simulating runoff in watersheds because it enables a more de-

tailed representation of surface water movement, improved flood modeling in river channels, and the representation of depressions, secondary channels, terraces, and other topographic features. Figure 3 shows results obtained with the one- and two-dimensional versions of HEMINGS in the Guadalmedina River catchment (Málaga, Spain). As shown, the two-dimensional model represents floodplains and therefore enables a more comprehensive analysis of interactions between the watershed and the fluvial system.

2 The hydrological model

The hydrological processes module forms the functional core of HEMINGS and serves as the component around which the model's other modules are organized. Its purpose is to simulate—using a distributed, physically-based formulation—the key processes governing the hydrological cycle within a catchment: from precipitation interception and soil infiltration to runoff generation, subsurface flow, evapotranspiration, and water routing through the drainage network. Accurately representing these processes is essential for reproducing the spatiotemporal dynamics of water availability and the hydrological flows that drive the functioning of headwater catchments.

The hydrological data generated by this module provide the foundation for the geomorphological processes simulated by HEMINGS. Specifically, the magnitude, spatial distribution, and temporal evolution of runoff determine the initiation and progression of water erosion, sediment transport, gully formation and evolution, and headward erosion processes. Similarly, hydrological variables influence the catchment's hydrological and sedimentary connectivity, controlling the transfer of water, soil, and sediment from hillslopes to the river network.

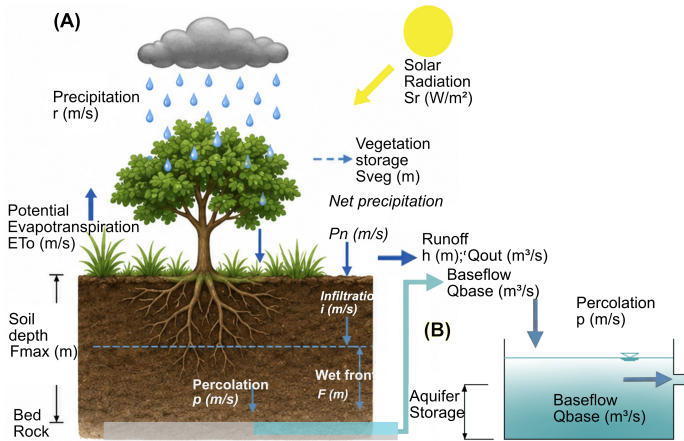


Figure 4. Schematic diagram of the hydrometeorological processes considered in the HEMINGS model.

Figure 4 shows a schematic overview of the hydrometeorological and hydrological processes considered across the digital elevation model. Based on the balance calculated for each grid cell, the model routes surface runoff downstream and contributes subsurface flow and baseflow at specific user-defined discharge points within the river network.

The following sections describe in detail the theoretical concepts of each of the hydrological processes considered.

2.1 Hydrometeorological processes

2.1.1 Precipitation

The HEMINGS model incorporates the Inverse Distance Weighting (IDW) method, utilizing hourly precipitation data and the coordinates of meteorological stations within the study catchment. The IDW method estimates the interpolated value at an unobserved location using a weighted average of surrounding observations, calculated as follows:

$$P(x_0) = \frac{\sum_{i=1}^n w_i P(x_i)}{\sum_{i=1}^n w_i} \quad (1)$$

where $w_i = 1/d_i^p$, $P(x_0)$ is the interpolated precipitation at the target point, $P(x_i)$ is the observed precipitation at station i , d_i is the distance between the target point and station i , p is the power parameter, and n is the number of stations used. The parameter p controls the influence of distance; high values increase the weight given to the nearest stations, resulting in surfaces that are more localized and less smooth.

The IDW method requires few parameters and is simple to implement, making it suitable for event-based hydrological simulations (Kurtzman et al., 2009). Furthermore, unlike geostatistical methods, it does not require assumptions of stationarity or the modeling of spatial covariance structures when there is a high density of rain gauges. Finally, the IDW method can reasonably reproduce the spatial variability of convective events, although it faces limitations in complex, mountainous terrain, where it must be adjusted for elevation gradients.

2.1.2 Snowfall

The HEMINGS model employs a simple and widely used method to estimate precipitation in the form of snow. The model estimates snowfall based on the spatial distribution of the daily mean air temperature—interpolated using an IDW method, similar to the approach used for precipitation—via the following expression:

$$P_s = \begin{cases} P(x_0), & T_{med}(x_0) \leq T_{threshold}, \\ 0, & T_{med} > T_{threshold}, \end{cases} \quad (2)$$

where P_s is snowfall (mm water equivalent), P is total daily precipitation (mm), $T_{med}(x_0)$ is the interpolated daily mean air temperature ($^{\circ}\text{C}$), and $T_{threshold}$ is the threshold temperature for snowfall ($^{\circ}\text{C}$), which can be calibrated to reflect local climatic conditions. The default value for $T_{threshold}$ in the SWAT model is 1.0°C (Neitsch et al., 2011). Liquid precipitation is calculated as

$$r^t = P(x_0) - P_s \quad (3)$$

2.1.3 Evapotranspiration

Evapotranspiration (ET) represents the combined loss of water through soil evaporation and vegetation transpiration. The HEMINGS model incorporates a simplified hourly evapotranspiration model based on satellite imagery data. To do so, it draws upon the work of Itenfisu et al. (2003) and Gavilan et al. (2007), following an approach similar to that used in the WiMMED model (Herrero et al., 2014). The model proposed by Gavilan et al. (2007) establishes simplified relationships between reference evapotranspiration (ET_{o_i}) and

radiometric variables. Instantaneous evapotranspiration can be expressed as

$$ET_{oi} = \frac{R_n - G}{10^6 \lambda} \cdot f_c \quad (4)$$

where ET_{oi} is in mm/s , R_n is the net surface radiation ($W m^{-2}$), G is the soil heat flux ($W m^{-2}$), and λ is the latent heat of vaporization ($MJ kg^{-1}$), expressed as $\lambda = 2.501 - 0.002361 T_a$ (where T_a is the air temperature in $^{\circ}C$). The factor f_c represents the percentage of coverage and can be derived from the vegetation cover factor, C_f , calculated using the expression proposed by Van der Knijff et al. (2000):

$$C_f = \exp\left(-2 \frac{NDVI}{1 - NDVI}\right) \quad (5)$$

where C_f represents the cover factor and $NDVI$ is the Normalized Difference Vegetation Index, estimated from

$$NDVI = \frac{\rho_{NIR} - \rho_R}{\rho_{NIR} + \rho_R} \quad (6)$$

where ρ_{NIR} is the near-infrared reflectance and ρ_R is the red reflectance. The cover factor is given by the expression $f_c = 1 - C_f$, representing the fraction of the land surface covered by vegetation or terrain features.

Net radiation is calculated as

$$R_n = (1 - \alpha)R_s^{\downarrow} + R_L^{\downarrow} - R_L^{\uparrow} \quad (7)$$

where α is the surface albedo, which can be estimated as a weighted combination of satellite image spectral reflectances— $\alpha = \sum_{i=1}^n w_i \rho_i$, where w_i are spectral weighting coefficients, ρ_i are the reflectances of each spectral band, and n is the number of bands used— $R_s^{\downarrow}$ is the incident solar radiation, $R_L^{\downarrow}$ is the atmospheric longwave radiation, and $R_L^{\uparrow}$ is the radiation emitted by the surface. Downward atmospheric longwave radiation is estimated using the Stefan-Boltzmann law, $R_L^{\downarrow} = \varepsilon_a \sigma_{SB} T_a^4$, where ε_a is the atmospheric emissivity coefficient, T_a is the air temperature (K), and σ_{SB} is the Stefan-Boltzmann constant ($5.67037 \times 10^{-8} W m^{-2} K^{-4}$). Similarly, the longwave radiation emitted by the surface can be estimated using $R_L^{\uparrow} = \varepsilon_s \sigma_{SB} T_s^4$, where ε_s is the surface emissivity coefficient and T_s is the surface temperature (K), which is derived from satellite image thermal bands and must be corrected according to the remote sensor employed.

Soil heat flux, G , can be estimated using empirical relationships based on net radiation, surface temperature, and vegetation cover. In areas with moderate to high vegetation cover, soil heat flux typically accounts for between 5% and 20% of the instantaneous net radiation. In simplified approaches for hourly timescales, $G = c_g R_n$ may be used, where c_g is an empirical coefficient ranging from 0.05 to 0.10 for dense vegetation and from 0.15 to 0.30 for bare soil.

Hourly evapotranspiration is ultimately obtained by integrating the latent heat flux:

$$ET_h = \frac{3600 \cdot \lambda ET_{oi}}{\lambda \rho_w} \quad (8)$$

where ET_h is in $mm h^{-1}$ and ρ_w is the density of water.

2.2 Precipitation interception

The HEMINGS model includes the calculation of the vegetation canopy's water storage capacity and the amount of precipitation intercepted before reaching the ground. To do this, the model uses $NDVI$, which also allows for accounting for the temporal variability of the vegetation cover. The Leaf Area Index (LAI)—which represents the amount of leaf area per unit of ground surface area and controls the vegetation's water storage capacity—is estimated using a linear relationship based on $NDVI$:

$$LAI = LAI_{\max} \frac{NDVI - NDVI_{\min}}{NDVI_{\max} - NDVI_{\min}} \quad (9)$$

where $LAI_{\max}$ is the maximum Leaf Area Index value, which depends on the vegetation type (with indicative values of 2–4 for grasslands, 3–7 for crops, 4–7 for deciduous forest, 6–10 for coniferous forest, and 8–12 for tropical forest (Arnold et al., 1998)), and $NDVI_{\min}$ and $NDVI_{\max}$ are the normalization limits corresponding to bare soil and dense vegetation, respectively. The model assigns a specific storage capacity per unit of LAI , denoted as S_{LAI} (mm/LAI), determined by the anatomy and architecture of the plant community and inferred from $NDVI$, with approximate values for crops: 0.05 – 0.15 mm/LAI ; shrubland: 0.3 – 0.9 mm/LAI ; conifers: 0.2 – 0.5 mm/LAI ; broad-leaved trees: 0.1 – 0.15 mm/LAI (Keim et al., 2006; Li et al., 2016). The maximum vegetation storage capacity, S_{v_o} , is calculated as:

$$S_{v_o} = S_{LAI} \cdot LAI \quad (10)$$

The precipitation intercepted during each time interval, I^t (m/s), is calculated from

$$I^t = \min(r^t, \frac{S_{v_o}^t}{\Delta t}) \quad (11)$$

where I^t is the intercepted precipitation (m), r^t is the precipitation intensity (m/s), Δt is the time interval, and $S_{v_o}^t$ is the available storage in the plant canopy at time t . Vegetation storage is updated using $S_{v_o}^{t+1} = S_{v_o}^t + r^t \Delta t$ until the maximum storage S_{v_o} is reached. Finally, the effective precipitation reaching the ground is defined as

$$r_n^t = r^t - I^t \quad (12)$$

2.3 Infiltration and runoff

Soil infiltration is estimated using the Green-Ampt model (Green and Ampt, 1911), which is widely used in modeling rainfall-runoff processes. The Green-Ampt expression is given by (Esteves et al., 2000):

$$i^t = k_s \left[\frac{F^t + (h_f + r_n^t \Delta t) \Delta \theta}{F^t} \right] \quad (13)$$

where i^t is the infiltration rate calculated for time step t (m/s), F is the cumulative infiltration at the wetting front (m), k_s is the saturated soil hydraulic conductivity (m/s), h_f is the potential suction head (m), and θ_s and θ_r are the saturated and residual soil moisture contents, respectively. These values are generally estimated based on average soil texture (Rawls and Brakensiek, 1983). The infiltration rate at each time step is calculated using

$$i^t = \min(r_n^t, i^t) \quad (14)$$

and the cumulative infiltration for the next time step is calculated using

$$F^{t+1} = F^t + i^t \Delta t \quad (15)$$

The excess precipitation or runoff in each cell, h_t (m), is calculated using

$$h_t = (r_n^t - i^t) \Delta t \quad (16)$$

2.4 Baseflow

The saturated-zone groundwater flow model incorporated into HEMINGS is based on the scheme proposed for the WiMMed model (Herrero et al., 2014; Millares, 2008). The baseflow contribution to the river network from the unsaturated zone and aquifer storage is formulated using a lumped model consisting of two interconnected linear reservoirs. Figure 5 provides a schematic representation of the baseflow contribution model.

Distributed soil percolation—the excess from the soil infiltration process—recharges a first, rapid-response reservoir associated with lateral flow or interflow. This reservoir, which has a threshold h_{b1} (m), features two outflows: the river contribution flow, q_{b1} (m/s), and the deep recharge flow, q_{perc2} (m/s), which feeds the slow-response aquifer. This latter aquifer also has a threshold h_{b2} (m) and a single outlet linked to the contribution to the river network. The reservoirs are associated with surface areas or homogeneous aquifer zones defined either within or outside the study catchment.

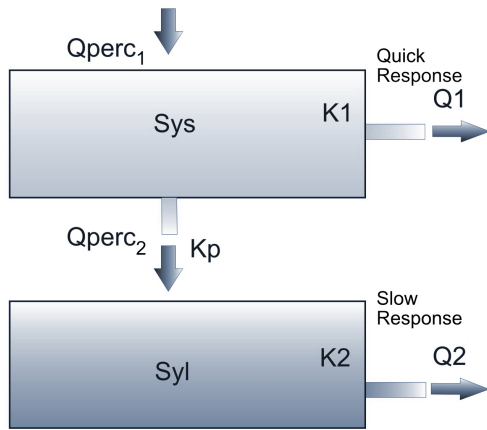


Figure 5. Conceptual linear reservoir model proposed for lateral flow and baseflow contributions.

The response associated with lateral flow—or rapid-response flow—is given by the expressions

$$q_{b1} = \frac{k_p}{k_p + k_1} \left(-\frac{h_{b1}(t+1) - h_{b1}(t)}{dt} Sy_1 + q_{perc1} - \frac{h_{b1}}{k_1} \right) \quad (17)$$

where h_{b1} is the stored water level in reservoir 1 (m), k_p and k_1 (s) are the reservoir recession or storage constants, and Sy_1 is the specific yield of the aquifer zone. The parameters k_p and k_1 can be obtained through recession curve analysis, as described in Millares et al. (2009), while the value of Sy_1

can be estimated based on the geological configuration of the selected aquifer zone. The deep recharge flux, q_{perc2} (m/s), is calculated using the equation

$$q_{perc2} = -\frac{h_{b1}(t+1) - h_{b1}(t)}{dt} Sy_1 + q_{perc1} - q_{b1} \quad (18)$$

Contributions associated with deeper, slower-responding aquifers are estimated using

$$h_{b2}(t+1) = h_{b2}(t) e^{-\frac{dt}{k_2 Sy_2}} + k_p q_{perc2} \left(1 - e^{-\frac{dt}{k_2 Sy_2}} \right) \quad (19)$$

and

$$q_{b2} = q_{perc2} - \left(-\frac{h_{b2}(t+dt) - h_{b2}(t)}{dt} Sy_2 \right) \quad (20)$$

where q_{b2} (m/s) is the slow-response flow contribution to the river network (m/s), k_2 is the storage constant of the slow-response reservoir (s), and Sy_2 is its specific yield. The final baseflow contribution is applied at specific points along the river network—selected during model configuration—using the expression

$$Q_{base} = (q_{b1} + q_{b2}) \cdot A_T \quad (21)$$

where Q_{base} is the contributed baseflow (m^3/s) and A_T is the total area of the associated aquifer zone (m^2).

2.5 Hillslope runoff routing

Runoff in hillslope areas is modeled using an approach based on the diffusive wave equation. This scheme provides a good representation of the Saint-Venant equations for hydrological purposes in areas with both gentle and steep slopes (Orlandini and Rosso, 1996; Jain and Singh, 2005; Gottardi and Venutelli, 2008). The runoff model follows the numerical solution provided by Jain and Singh (2005), which is based on finite volume methods. The conservative form of the one-dimensional equation for surface water within a volume ν bounded by a surface s is given by

$$\frac{\partial \mathbf{U}}{\partial t} + \frac{\partial \mathbf{F}}{\partial x} = \mathbf{G} \quad (22)$$

where

$$\mathbf{U} = \begin{pmatrix} h \\ 0 \end{pmatrix}, \quad \mathbf{F} = \begin{pmatrix} uh \\ h \end{pmatrix}, \quad \mathbf{G} = \begin{pmatrix} r - f \\ s_0 - s_f \end{pmatrix} \quad (23)$$

Here, h is the flow depth (m); u is the depth-averaged flow velocity (m/s); x is the distance in the flow direction (m); t is time (s); r is the net precipitation rate (m/s) ($r = r_n$); i is the infiltration rate (m/s); S_o is the bed slope; and S_f is the friction slope. This expression is integrated using finite volumes across each of the cells covering the entire domain. For a cell, the equation can be expressed in integral form by applying Gauss's divergence theorem, as follows:

$$\int_{\nu} \frac{\partial h}{\partial t} d\nu + \oint_s (\mathbf{u}h \cdot \boldsymbol{\omega}) ds = \int_{\nu} (r - i) d\nu \quad (24)$$

where ν represents the control volume, s is the surface of the control volume, h is the runoff depth, $\mathbf{u}$ is the velocity vector, $\boldsymbol{\omega}$ is the outward-pointing normal vector to the surface, and $r - i$ represents the source term corresponding to net rainfall and infiltration. Resolving the flux vector in the normal

direction allows the two-dimensional problem to be treated as a local one-dimensional problem (Jain and Singh, 2005). Reducing surface runoff to a one-dimensional model has limitations, as it simplifies the role of microtopography, the interaction between vegetation and rill formation, and the direction of subsurface flow (Liu et al., 2004; Nord and Esteves, 2005). However, two-dimensional model-based approaches are limited by the difficulty of validating relationships in natural soil and by the associated computational costs at the catchment scale.

The runoff model operates at the event time scale, utilizing adaptive Δt calculations based on the Courant-Friedrichs-Lewy condition and the Digital Elevation Model (DEM), and employing variable spatial resolution as the basis for calculating runoff flow. The basic equation for 1D water flow in each DEM cell, discretized using the finite volume method (FVM), is given by the expression (Zhao et al., 1994)

$$W^2 \frac{\partial h}{\partial t} + W \sum_{\varepsilon=1}^{sd} F_{\varepsilon} = W^2 G \quad (25)$$

where sd is the total number of contributing upstream cells. The model assumes a D_8 structure, meaning there are 8 possible directions around the peripheral cells. Following the criteria of Jain and Singh (2005), a lateral distribution within the cell is assumed in the case of a diagonal flow direction. Numerical integration via FVM for a DEM is performed using a two-step, second-order *predictor-corrector* scheme described by Zhao et al. (1994) and Jain and Singh (2005). Therefore, for each cell, the value of h in the first predictor step for each Δt is

$$h^* = h^{t-1} + \frac{\Delta t}{W^2} (Q_{in}^{t-1} - Q_{out}^{t-1}) + \Delta t (r^t - i^t) \quad (26)$$

with

$$Q_{in}^{t-1} = \sum Q_{out}^{t-1} \quad (27)$$

using time-superscript notation throughout the section. The outflow rate at the prediction stage, Q_{out}^* ($\text{m}^3 \text{s}^{-1}$), is calculated as $Q_{out}^* = q^* W$. Considering turbulent flow over a rough surface, q^* is expressed using Manning's equation for flow resistance:

$$q^* = \frac{1}{n} h^{*5/3} S_f^{*1/2} \quad (28)$$

This approach, which accounts for the diffusion wave, results in $S_f^* = S - (h_{j+1}^* - h_j^*)/l$. The final water depth value for each cell at time t is obtained from the corrector step using the expression:

$$h^t = 0.5 \left[h^{t-1} + h^* - \frac{\Delta t}{W^2} (Q_{in}^* - Q_{out}^*) + r^t - i^t \right] \quad (29)$$

The Courant-Friedrichs-Lewy condition for this explicit scheme is given by:

$$\max \left[\frac{(u + \sqrt{gh}) \Delta t}{l} \right] \leq 1 \quad (30)$$

where u is calculated as $u = q_{out}/W$.

The numerical solution of the runoff equation was compared with the analytical solution proposed by Eagleson (1970) and described in detail by Kazezyilmaz-Alhan and Medina (2007). This comparison was based on the case study proposed by Gottardi and Venutelli (2008), which involves a river reach with a slope of 0.001 and a length of 200 m, a spatial resolution of $\Delta x = 5$ m, a unit width, and a Manning roughness coefficient of $n = 0.03$. Precipitation is assumed to be uniform, with an intensity of 60 mm/h. Based on these parameters, the numerical model was applied using both kinematic and diffusive wave assumptions, and the results were compared with the analytical solution. The results from the analytical solution (magenta) and the model using the kinematic and diffusive approaches (blue and red, respectively) are shown in Fig. 6. The modeled runoff agrees well with the analytical solution, although there is a time lag in the peak discharge value—a feature previously identified by Gottardi and Venutelli (2008).

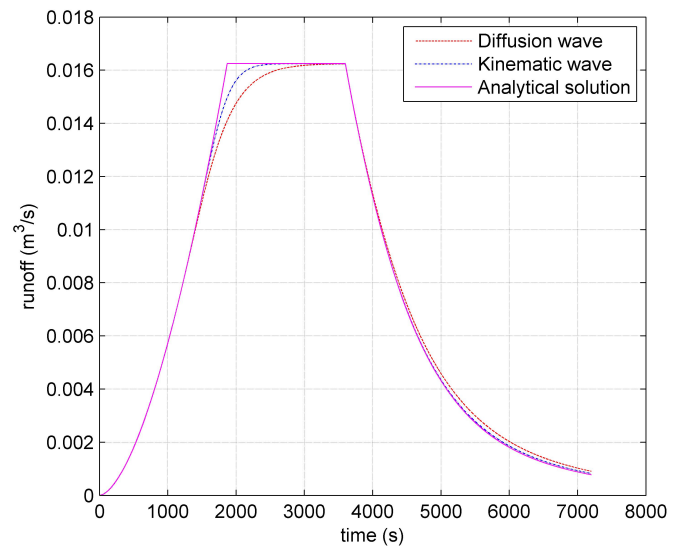


Figure 6. Runoff hydrograph for the analytical solution and the kinematic and diffusive wave approaches for the case study proposed by Gottardi and Venutelli (2008), implemented within the numerical scheme of the model (Millares et al., 2019).

2.6 Fluvial routing

Selecting cells corresponding to hillslopes and fluvial systems within a Digital Elevation Model (DEM) is a fundamental problem in geomorphology and computational hydrology. The hydrological module of the HEMINGS model distinguishes between hillslope cells and river channel cells—either manually, based on an accumulated flow threshold value, or by applying automated criteria to separate the two systems, such as the Topographic Position Index method (Weiss, 2001) or Geomorphons (Jasiewicz and Stepinski, 2013).

Flow routing along the river network follows the same numerical scheme used for hillslope runoff. In this case, the flow moves through a specific geometric cross-section—typically triangular or trapezoidal, with either fixed or variable width along the network—where the hydraulic radius Rh_c^t is calculated. Hydraulic parameters, such as Manning's flow resistance coefficient (n), differ from those used for hillslopes; generally, they are significantly lower, in accordance with recommendations for natural rivers or channels (Chow, 1959; Jung et al., 2010).

2.7 Runoff routing options

In addition to a diffusion-wave-based scheme—described previously, where the condition $S_f^* = S - (h_{j+1}^* - h_j^*)$ is assumed for runoff flow in hillslope and channel cells— the HEMINGS model allows for simulation using a kinematic-wave-based scheme, where $S_f^* = S$. Both the kinematic and diffusive approximations are known for their numerical robustness regarding convergence and stability, yielding stable, convergent simulations across a wide range of terrain slopes throughout the watershed. However, the kinematic wave is particularly suitable for steep slopes, whereas the diffusion wave is better suited for slopes ranging from gentle to steep (Yeh et al., 1998).

Furthermore, the kinematic-wave scheme significantly reduces computational cost and is recommended for larger watersheds or long-duration precipitation events. Furthermore, the model offers the possibility of including a diffusive term in the diffusion-wave scheme takes the following form

$$h_2 = h_1 + \frac{\Delta t}{W^2} (Q_{in} - Q_{out}) + r_n \Delta t - i \Delta t + K \frac{h_3 - 2h_2 + h_1}{W^2} \quad (31)$$

where h_1 corresponds to the hydraulic head or water level in the cell, h_2 represents the hydraulic head in the cell immediately downstream, h_3 corresponds to the second cell downstream, Δt is the simulation time step, W is the DEM cell size, Q_{in} represents the inflow to the cell, Q_{out} represents the outflow from the cell, r_n is the source term associated with net precipitation, i represents the infiltration rate, and K is the hydraulic diffusivity coefficient (m^2/s).

The diffusive term corresponds to a finite-difference approximation of the second spatial derivative. This formulation allows for the representation of diffusive flow propagation in the downstream direction. The forward-looking diffusive formulation enables a more stable representation of the spatial propagation of hydraulic gradients, particularly in floodplains, distributed surface flow, and urban runoff. Furthermore, it avoids artificially modifying the original DEM through topographic filling procedures, thereby preserving the original microtopography and maintaining actual natural depressions (Huang and Lee, 2016).

3 Hillslope erosion

The intensity of erosional and transport processes in headwater catchments drives very high rates of soil loss, resulting not only from climatic factors but also from human-induced alterations and abrupt changes in vegetation cover and land use.

On hillslopes, where the diffusive nature of flow constrains water and sediment movement (Simpson and Schlunegger, 2003; Hopp and McDonnell, 2009), erosional processes are highly varied and exhibit complex interrelationships when formulating catchment-scale models. Generally, raindrop impact and concentrated flow are considered the dominant processes in erosion modeling (Nearing et al., 1989; Merritt et al., 2003). Other processes—such as small-scale landslides, rockfalls, subsurface soil washing, and gully erosion—can significantly increase erosion rates (de Vente et al., 2008), yet our understanding of them remains limited. Gully erosion,

in particular, has been responsible for substantial soil loss in semi-arid environments (Wasson et al., 2002; Martínez-Casasnovas, 2003; Casali et al., 2009; Vigiak et al., 2011; Polo et al., 2022), accounting for over 95% of total sediment production in some cases (Poesen et al., 2003).

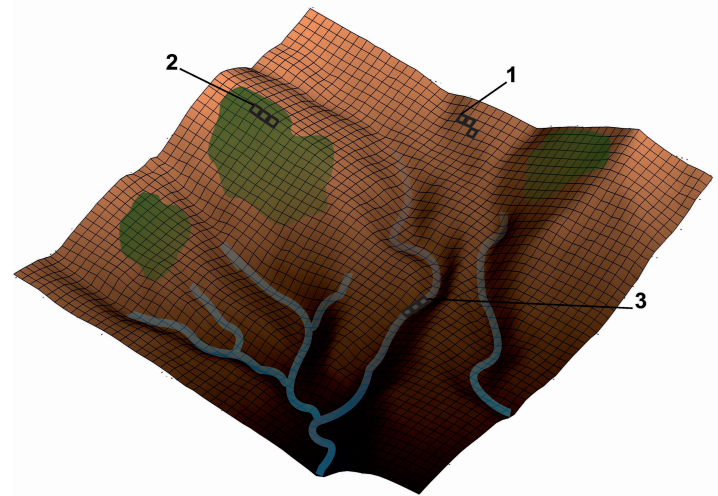


Figure 7. Conceptual diagram of the runoff and erosion model and its various elements across the digital elevation model (DEM): hillslope cells with rill and inter-rill erosion (brown, label 1), hillslope cells with rills, inter-rill areas, and gullies (green, 2), and channel cells (blue, 3) (Millares et al., 2019)

Predictive models addressing gully formation processes have focused on local changes in gully morphology (Sidorchuk and Sidorchuk, 1998; Sidorchuk, 1999), upstream headcut migration (Alonso et al., 2002; Flores-Cervantes et al., 2006; Campo-Bescós et al., 2013; Rengers et al., 2016; Vanmaercke et al., 2016), and the topographic conditions conducive to gully formation (Montgomery and Dietrich, 1992; Prosser and Abernethy, 1996; Millares et al., 2012; Torri et al., 2012; Torri and Poesen, 2014; Rossi et al., 2015). However, from a physical modeling perspective, the influence of gully formation on sediment yield has received less attention (Merritt et al., 2003; Borja et al., 2018). Questions remain regarding how various parameters—such as soil properties, vegetation cover, topography, land-use changes, and their spatial distribution—interact with the stages of rill and gully formation, including their initiation, development, and relationship with other surface and subsurface runoff processes (Bocco, 1991; Torri and Poesen, 2014). In most cases, sediment supply and the responses associated with the aforementioned erosive processes have not been analyzed at the catchment scale (de Vente and Poesen, 2005).

The conceptual framework of the erosion module in HEMINGS is presented in Figure 7. As shown, the hillslope is divided into areas based on different erosive processes. The distribution of hillslope and channel cells across the digital elevation model is determined by the drainage network, taking into account the study area's hydrological regime. In this regard, a distinction is made between fluvial processes and hillslope diffusive processes—which exhibit significant physical differences regarding erosion and transport—following previous proposals (Willgoose, 2005a,b).

From a modeling perspective, the differences can be summarized as follows:

1. Hillslope cells (Fig.7, labels 1 and 2), which include erosive processes such as raindrop impact, erosion/sed-

imentation driven by concentrated flow in rills and gullies, and gully headcut erosion.

2. Channel cells, which include only transport and sedimentation (Fig.7, label 3)

This configuration allows for representing the spatial heterogeneity of the soil and land use while accounting for the spatial distribution of forcing agents. The DEM-based numerical approach enhances the model's ability to interact with raster data, such as remote sensing imagery or interpolated maps. The erosion module includes a dynamic calculation of rill density throughout the simulation, taking into account flow and terrain conditions. Furthermore, gully erosion modeling—which could be compared to the approach implemented in the EUROSEM model for surfaces with high rill density or deep rills (Morgan et al., 1998)—is addressed here in greater detail.

Based on this spatial conceptualization, the model generates and distributes erosion and sedimentation rates for each cell along the drainage network, accounting for the spatial distribution of various hydrological forcing agents (precipitation, snowfall, rainfall interception, evapotranspiration, etc.). In terms of process physics, the conceptual framework can be broken down into splash erosion and slope-flow erosion—manifesting as rills and gullies—as well as headward erosion.

3.1 Raindrop erosion

The erosion rate due to processes related to raindrop impact on hillslope cells is calculated using

$$D_i^t = f_c [r_n^t]^2 S_{di} K_i \quad (32)$$

where D_i is the splash erosion rate ($kg/m^2/s$), $f_c = 1 - C_f$, and S_{di} (Liebenow et al., 1990; Ben-Hur and Wakindiki, 2004) is estimated using

$$S_{di} = 1.05 - 0.85e^{-4 \sin(\alpha)} \quad (33)$$

where α represents the slope angle of the cell.

3.2 Rill erosion

Figure 8 shows the geometric parameters considered for modeling gullies and rills. The various parameters are detailed, such as the gully head width w_g , its depth H_g , its length L_g , the rill width $w_r = 2h_r$, and the cell size W .

The runoff depth within a cell, h^t —calculated for all cells of the DEM at a given time step—is distributed among the rill and gully erosion processes. This simplification enables catchment-scale modeling at a reasonable computational cost. A coefficient λ distributes the flow among the gullies, accounting for the gully density per cell (N_g) and the gully head width (w_g), using the expression $\lambda = 1 - N_g w_g / W$. The relative flow depth in rills, gullies, or channels is obtained from the following expressions:

$$h_r^t = \left[\frac{\lambda Q_{out}^t}{N_r^t u_r^t} \right]^{1/2} \quad (34)$$

$$h_g^t = \left[\frac{2(1-\lambda) Q_{out}^t n}{S_g^{0.5} N_g} \right]^{3/8} \quad (35)$$

$$h_c^t = \left[\frac{Q_{out}^t n}{S^{0.5} w_c^t} \right]^{3/5} \quad (36)$$

where u_r is the flow velocity in the rill, estimated using the empirical relationship $u_r = 3.52(\lambda Q_{out}^t / N_r^t)^{0.294}$ (Govers, 1992).

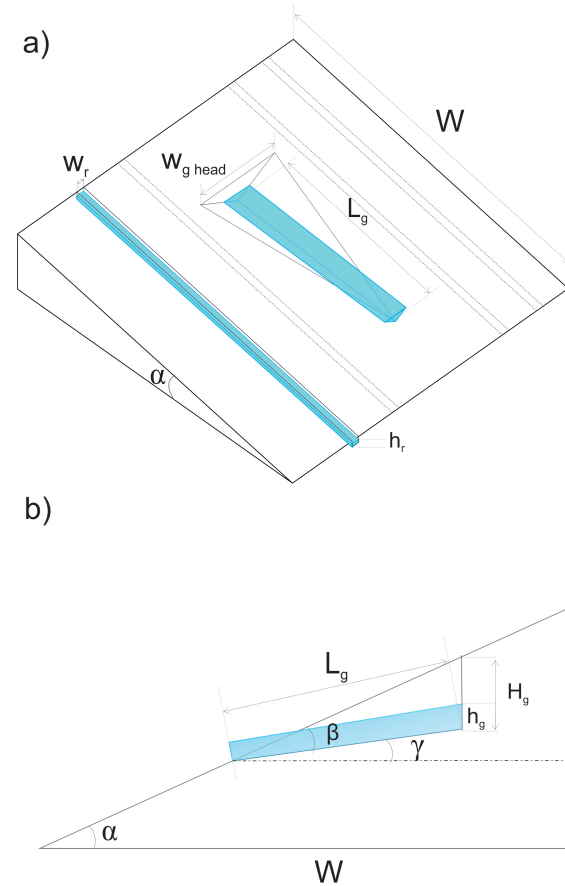


Figure 8. Conceptualization of the hillslope erosion processes considered in the HEMINGS model (Millares et al., 2019).

The model assumes hydraulic cross-sections based on isosceles triangles for rills and gullies (Fig. 8); consequently, the hydraulic radius is obtained from $R_r^t = h_r^t / (2\sqrt{2})$ and $R_g^t = h_g^t / (2\sqrt{2})$, respectively. For channel cells, the hydraulic radius simplifies to the flow depth, $R_c^t \approx h_c^t$, while the channel width is calculated as

$$w_c = c_1 (Q_{out}^t)^{c_2} \quad (37)$$

In this case, the coefficient c_1 and the exponent c_2 are 4.0216 and 0.44, respectively (Leopold and Maddock, 1953). The HEMINGS model calculates rill erosion based on a variable number of rills within the cell during the event. Rill density, N_r^t , is considered time-variant and depends on water discharge, soil moisture, soil roughness resulting from tillage practices, slope, and cell vegetation cover (Mancilla et al., 2005), based on the expression

$$N_r^t = W \left[0.66 + 4.14e4 \ln(q^t) + 0.91 \ln(M_o^{t-1}) + 2.04 \ln(S) - 0.37 \ln\left(\frac{Vr}{Rr}\right) \right] \quad (38)$$

where M_o^{t-1} is the antecedent soil moisture content in the top 15 cm ($M_o = I_{cum}/0.15$), Vr represents the crop residue fraction, and Rr is the random soil roughness (Mancilla et al., 2005), which is determined from available land-use and vegetation cover maps (Junta de Andalucía, 2005).

The erosion/deposition rate for groove processes, D_r ($kg/m^2/s$), is calculated from the balance between transport capacity and sediment load (Yalin, 1963; Nearing et al., 1989), according to the expression

$$T_{cr}^t = k_T (\tau_r^t)^{1.5}, \quad (39)$$

and

$$Q_{sr}^t = \frac{\lambda Q_{out}^t}{h_r^t N_r^t}, \quad (40)$$

where $k_T = 0.4433 + 0.2021 \ln(\tau_r^*)$ (Ferro, 1998) y $\tau_r^* = \tau_r / [(G-1)\rho g d_m]$, con

$$\tau_r^t = \rho g R_r^t S \frac{f_r^t}{f_T} \quad (41)$$

where $G = \rho_s/\rho$, $\rho = 1000 \text{ kg m}^{-3}$, $\rho_s = 2650 \text{ kg m}^{-3}$. The factor f_r/f_T represents the relationship between the friction acting on the ground ($f_r = 8gR_r S/u_r^2$) and the total friction $f_T = 1.11 + 42.76(1 - e^{77.3R_r} + 1.85St113.73Vr^3)$ (Gilley and Weltz, 1995). Finally, the value of D_r is calculated from

$$D_r^t = \begin{cases} K_r(\tau_r^t - \tau_{rc}) \left(1 - \frac{Q_{sr}^t}{T_{cr}^t}\right) & \text{for } T_{cr}^t \geq Q_{sr}^t \\ 0.5 \frac{V_s}{q} (T_{cr}^t - Q_{sr}^t) & \text{for } T_{cr}^t < Q_{sr}^t \end{cases} \quad (42)$$

where V_s ($m \text{ s}^{-1}$) is estimated from

$$V_s = \frac{12\nu_m}{d_m C_D} \left[\left(1 + \frac{C_D d_*^3}{108}\right)^{0.5} - 1 \right] \quad (43)$$

where $C_D = 0.168$, $\nu_m = 1.01 \cdot 10^{-6} \text{ m}^2 \text{ s}^{-1}$ and d_* is the dimensionless diameter, estimated as $d_* = d_m [(G-1)g/\nu_m^2]^{1/3}$.

3.3 Gully erosion

The erosion by carcass-concentrated flow, D_g , is calculated following the same calculation scheme as the one described in equation (42). In this case, instead of K_r , a value of K_g is used to represent the erodibility of the subsoil at the bottom of the reservoir. Erosion by carving is assumed in specific areas in which the number of carcasses, the width and depth of the headland, and the initial length (Fig.??) are previously obtained from field measurements and the analysis of orthophotographs. The model corrects the assigned initial depth to avoid “carcass divers” and so that there is consistency between the internal carcass slope and the cell slope. This is done from the expression

$$\beta = \sin^{-1} \left[H_g \frac{\cos(\alpha)}{L_g} \right], \quad (44)$$

$$\gamma = \alpha - \beta,$$

3.4 Gully headcut assessment

The HEMINGS model incorporates the modeling of headcut advance based on a simplified version of the model proposed by Alonso et al. (2002). Flow is assumed to be uniformly distributed across the width of the gully head. The flow depth at the gully head, h_{gin} , and the inflow velocity, v_{gin} , are calculated using

$$h_{gin}^t = \left[\frac{(1-\lambda)Q_{out}^t n}{N_g S^{0.5} w_g} \right]^{\frac{3}{5}} \quad (45)$$

$$v_{gin}^t = \frac{(1-\lambda)Q_{out}^t}{N_g h_{gin}^t w_g} \quad (46)$$

The inflow angle from the top of the gully, φ_g , is given by the expression

$$\varphi_g^t = \tan^{-1} \left[\left(1 + \epsilon \phi \frac{h_{gin}^t}{H_g} \right) \left[T_e^{t^2} \Omega^t + \ln(T_e^t(1 + \Omega^t)) \right] \right]^{0.5} \quad (47)$$

where the parameters ϵ and ϕ are taken as 2 and 0.3, respectively (Gordon et al., 2007), and $\Omega^t = \sqrt{1 + T_e^{t-2}}$, with T_e calculated as $T_e^t = \sqrt{2gh_{gin}^t}/v_{gin}^t$. The head cut migration, L_{HM}^t , is calculated from

$$L_{HM}^t = v_g^t \sqrt{\frac{K_g \sin^2\left(\frac{\varphi_g^t}{2}\right) Q_g^t}{2(H_g - h_g^t)w_g}} \quad (48)$$

The total contribution of sediment due to the headwater advance in each time interval, D_{HM}^t , is calculated from the expression

$$D_{HM}^t = \frac{H_g^2 L_{HM}^t N_g \rho_s}{W^2 dt} \quad (49)$$

3.5 Transport and deposits on hillslope channels

As described in the previous section, we assume transport and sedimentation occur only in channel cells. These are assumed to be first-order channels with bedrock beds; therefore, no erosion takes place. Consequently, the calculation scheme follows the one described by Equation 42, but only for the condition $T_{ce} < Q^{sc}$.

4 Total sediment yield

Figure 9 shows the soil loss results for the Algeciras catchment (Murcia) for various precipitation events (Millares et al., 2022).

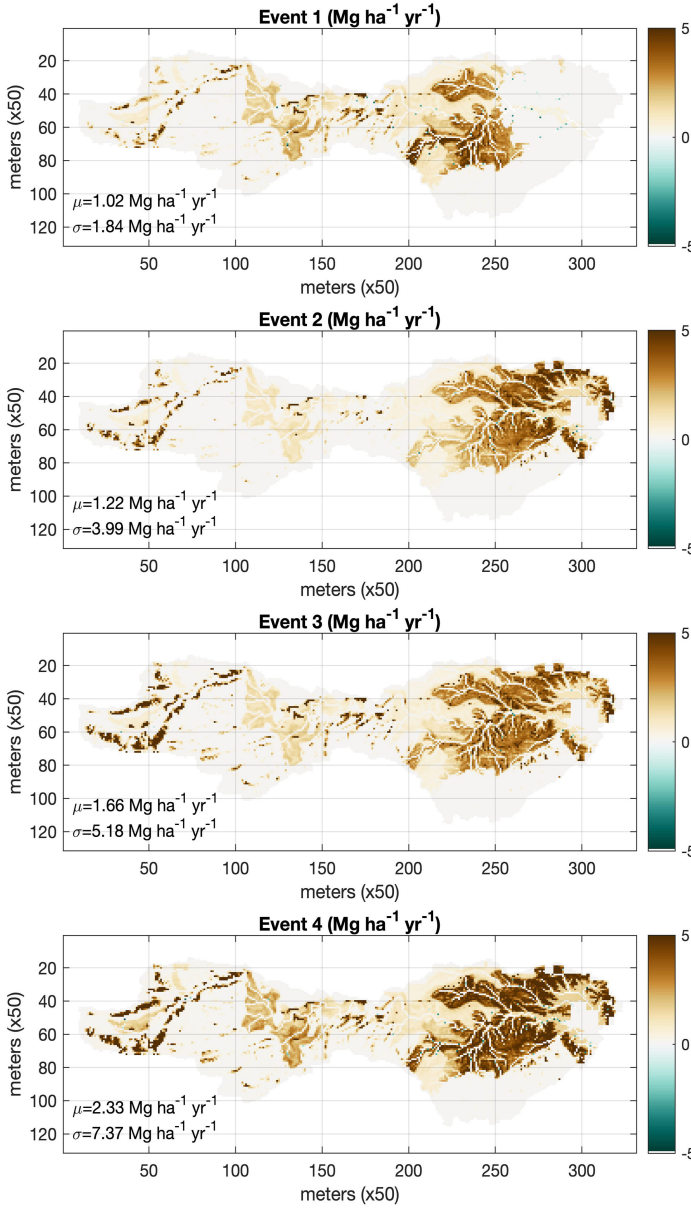


Figure 9. Soil loss estimation using the HEMINGS model in the Algeciras catchment (Murcia, Spain). (Millares et al., 2022)

The total erosion/sedimentation rate in each cell, E^t , corresponds to the sum of the five processes accounted for in the following equation

$$E^t = A_i^t D_i^t + A_r^t D_r^t + A_g^t D_g^t + A_c^t D_c^t + D_{HM}^t \quad (50)$$

which depends on the relative area of each process, given respectively by:

$$A_g^t = \frac{2N_g L_g}{W^2} \quad (51)$$

$$A_r^t = \frac{2N_r^t h_r^t}{W} [1 - A_g] \quad (52)$$

$$A_i^t = 1 - A_g^t - A_r^t. \quad (53)$$

The relative area of the channel cells is defined as:

$$A_c^t = \frac{w_c^t}{W} \quad (54)$$

4.1 Hillslope sediment routing

Regarding sediment transport, the problem can be addressed using a two-step "predictor-corrector" approach—applied in a decoupled manner for both water and sediment (Tayfur, 2002; Wu, 2008). Although flow density can vary depending on sediment concentration, salinity, and temperature, it has been assumed that the sediment load does not affect the flow's hydrodynamic properties, as previously noted (Wu, 2008). The sediment continuity equation for a hillslope within a control volume ν bounded by a surface s can be expressed as

$$\int_{\nu} \frac{\partial(hc_s)}{\partial t} d\nu + \oint_s \frac{\partial(qc_s)}{\partial x} ds = \int_{\nu} E d\nu \quad (55)$$

Taking these considerations into account, the sediment load for each DEM cell can be expressed as

$$c_s^t = \frac{1}{h^t} \left[h^{t-1} c_s^{t-1} + \frac{\Delta t}{W^2} (Q_{s_{in}}^{t-1} - Q_{s_{out}}^{t-1}) + \Delta t E^t \right] \quad (56)$$

where $Q_{s_{out}}$ is the sediment discharge calculated as $Q_{s_{out}} = c_s Q_{out}$.

5 Transport and fluvial dynamics

Sediment erosion and transport in river channels are fundamental processes controlling river morphological dynamics and their spatiotemporal evolution. Mobilization, transport, and deposition determine bed configuration, influence bank stability, and regulate the catchment's sediment connectivity. These processes are essential aspects of studies concerning fluvial hydraulics, river restoration, reservoir management, the design and management of hydraulic structures, and the assessment of global change impacts on river systems (Millares and Moñino, 2018).

Sediment transport assessment requires a sequential methodology that begins by identifying the physical mechanisms responsible for particle movement. From a hydraulic perspective, transport can be classified into: bed load, consisting of particles moving in contact with the bed via rolling, sliding, or saltation; suspended load, comprising particles maintained within the water column by turbulence; and wash load, generally composed of fine sediments whose concentration depends primarily on the supply from the catchment rather than the flow's transport capacity (Van Rijn, 1984).

The dominance of each mode depends on the interaction between flow properties, channel hydraulic conditions, and sediment grain-size characteristics. Correctly identifying transport modes is a prerequisite for selecting appropriate transport models. Consequently, models designed to estimate bed load are generally based on the concept of critical shear stress for the initiation of motion—introduced by Shields (1936)—whereas total load formulations also incorporate mechanisms related to maintaining sediment in suspension and the vertical distribution of concentrations, such as flow velocity.

Once the transport regime has been characterized, the quantitative estimation of solid discharge can be performed using various empirical, semi-empirical, or physically-based mechanistic formulations. Among the most widely used equations for estimating bedload transport are those proposed by Meyer-Peter and Müller (Meyer-Peter and Müller,

1948) (including the update by Wong and Parker (2006)) and Einstein-Brown (Einstein, 1950), as well as those accounting for the sequential contribution of different grain-size classes—specifically designed for gravel-bed rivers, such as the models by Wilcock and Crowe (2003), Parker (1990), or Powell et al. (1996)—and models configured for steep mountain channels, such as those by Schoklitsch (1962) or Takahashi et al. (1987). Formulations by Bagnold (1966), Engelund and Hansen (1967), and Brownlie (1981) are commonly used to estimate total sediment load. Each of these models was developed based on different experimental data and presents limitations regarding the sediment size range, channel slope, flow intensity, and calibration conditions for which it was originally designed.

The existence of a large number of formulations reflects the inherent complexity of sediment transport and the lack of a universal equation capable of adequately representing all fluvial conditions. Numerous comparative studies have highlighted that prediction uncertainty can exceed an order of magnitude, depending on the model used and the characteristics of the channel being analyzed (Wu, 2007; Millares et al., 2014; Conesa-García et al., 2020). Therefore, in engineering and research applications, it is advisable to assess sediment transport by applying multiple models and comparing their results; this approach allows for the analysis of estimate sensitivity, the identification of uncertainty sources, and the selection of the formulation that best reproduces the hydraulic and sedimentological conditions of the system under study.

5.1 Estimation of sediment transport modes

The mechanism by which a particle is transported in a channel can be determined by comparing the flow's shear velocity, u_* , with the particle's settling velocity, w_s (Julien, 2010). Shear velocity represents the intensity of turbulent forces generated by bed shear stress, whereas settling velocity characterizes the tendency of particles to settle due to gravity. Shear velocity is defined as

$$u_* = \sqrt{\frac{\tau_0}{\rho}}, \quad (57)$$

where τ_0 is the shear stress exerted by the flow on the bed and ρ is the water density. Furthermore, settling velocity depends primarily on particle size, shape, and density, as well as fluid properties. This parameter determines the speed at which a grain falls through a fluid at rest and serves as a measure of the particle's resistance to remaining in suspension. The particle settling velocity is calculated using the formula by Rubey (1933)

$$F_1 = \left(\frac{2}{3} + \frac{36\nu^2}{g(2.65 - 1)d_s^3} \right)^{1/2} - \left(\frac{36\nu^2}{g(2.65 - 1)d_s^3} \right)^{1/2} \quad (58)$$

$$w_{s,i} = F_1 \sqrt{g(s_g - 1)d_{s,\text{sup}}} \quad (59)$$

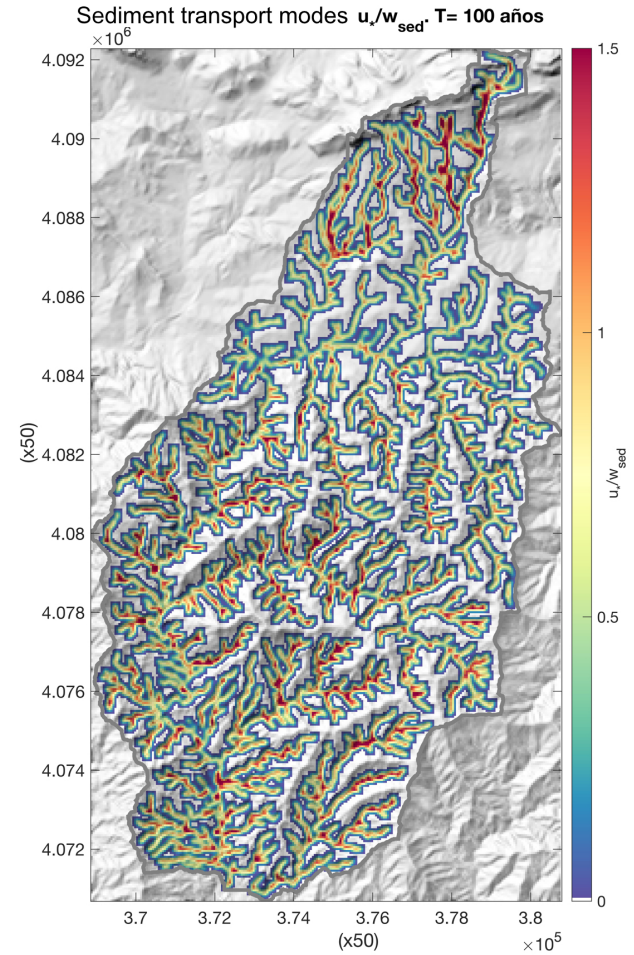


Figure 10. Estimation of fluvial transport modes for the 100-year return period in the Guadalmedina River catchment (Málaga, Spain) (Millares and Nanía, 2026).

where ν is the kinematic viscosity of water, g is the acceleration due to gravity, d_s is the sediment particle diameter, s_g is the specific gravity of the sediment (2.65), and $d_{s,\text{sup}}$ is the representative sediment diameter.

The ratio between the two velocities allows for the identification of the predominant transport mode. When the shear velocity is significantly lower than the settling velocity, turbulent fluctuations are unable to keep particles in suspension; consequently, transport occurs primarily along the bed via rolling, sliding, and saltation. According to Julien (2010)

$$\frac{u_*}{w_s} = \begin{cases} < 0.2, & \text{No movement,} \\ 0.2 - 0.5, & \text{Bed load,} \\ 0.5 - 2, & \text{Mixed transport,} \\ > 2, & \text{Suspended load.} \end{cases} \quad (60)$$

Figure 10 shows the dominant transport modes in the Guadalmedina River catchment (Málaga, Spain) for return periods of $T=10$ and 100 years, following the calibration and validation of the HEMINGS model using outflow data from the study area (Millares and Nanía, 2026). A transition is observed between transport dominated by bedload for $T = 10$ years and total load transport for $T = 100$ years.

5.2 Bedload

5.2.1 Wong and Parker (2006) transport model

The [Wong and Parker \(2006\)](#) model is an updated revision of the [Meyer-Peter and Müller \(1948\)](#) model. The condition for the onset of motion is defined by the critical Shields parameter $\tau_c = 0.0495$, based on the correction proposed by [Wong and Parker \(2006\)](#). Unit bedload transport is calculated as

$$q_b = 3.97 (\tau - \tau_c)^{1.5} d_{50} \sqrt{g(s_g - 1)d_{50}} \quad (61)$$

where q_b is the unit sediment transport rate ($m^3/s/m$), valid only if $\tau \geq \tau_c$ (otherwise, $q_b = 0$); Q_b is the total sediment discharge (m^3/s); θ is the Shields parameter; τ_c is the dimensionless critical Shields stress; d_{50} is the median sediment diameter; and B is the width of the river section. The total solid discharge is $Q_{MPM} = q_b B$.

5.2.2 Parker (1990) transport model

The [Parker \(1990\)](#) model is based on the grain-size distribution of the surface sediment or armor layer. The Shields reference stress is $\tau_{rm}^* = 0.0386$, and it is defined as

$$\Phi_{sg} = \frac{\tau_{50}^*}{\tau_{rm}^*} \quad (62)$$

For each grain-size fraction i :

$$\Phi_i = w_i \Phi_{sg} \left(\frac{d_i}{d_{50}} \right)^{-0.0951} \quad (63)$$

with $w=0.7$. The transport parameter W_i is calculated in a piecewise manner.

$$W_i = \begin{cases} 11.9 \left(1 - \frac{0.853}{\Phi_i} \right)^{4.5}, & \Phi_i > 1.59 \\ 0.00218 \exp [14.2(F_i - 1) - 9.28(\Phi_i - 1)^2], & 1 \leq \Phi_i \leq 1.59 \\ 0.00218 \Phi^{14.2}, & \Phi < 1 \end{cases} \quad (64)$$

where f_i is the grain-size fraction of class i , d_i is the diameter of class i , R_h is the hydraulic radius, S is the hydraulic slope, ρ_s is the sediment density, W_i is the dimensionless transport function, and Φ_i is the mobility intensity of fraction i . The total unit transport is:

$$Q_b = \sum_i \left[W_i f_i B \frac{(gR_h S)^{3/2}}{(s_g - 1)g} \right] \quad (65)$$

where B is the section width and Q_b is the sediment discharge in m^3/s .

5.2.3 Powell et al. (1996) transport model

The [Powell et al. \(1996\)](#) model is specifically configured for ephemeral or quasi-ephemeral rivers. In this case, the reference dimensionless shear stress is $\tau_{rm}^* = 0.03$. The relative mobility of each grain-size fraction is calculated as

$$\Phi_i = \frac{\tau_i^*}{\tau_{rm}^* \left(\frac{d_i}{d_{50}} \right)^{-0.74}} \quad (66)$$

If $\Phi \geq 1$:

$$W_i = 11.2 \left(1 - \frac{1}{\Phi_i} \right)^{4.5} \quad (67)$$

The sediment transport rate Q_b (m^3/s) will be

$$Q_b = W_i B f_i \frac{(gR_h S)^{3/2}}{(s_g - 1)g} \quad (68)$$

5.2.4 Wilcock and Crowe (2003) transport model

The [Wilcock and Crowe \(2003\)](#) model assumes that the percentage of sand in the surface layer plays a rolling role that facilitates the onset of motion, thereby leading to a decrease in the reference dimensionless shear stress. The critical reference dimensionless shear stress is expressed here as $\tau_{rm}^* = 0.021 + 0.015e^{-20F_s}$, where F_s is the sand content. The corresponding dimensional stress is calculated from

$$\tau_{rm} = \tau_{rm}^* g D_m \rho_w (s_g - 1) \quad (69)$$

For each fraction:

$$b = \frac{0.67}{1 + \exp \left(1.5 - \frac{d_i}{D_m} \right)} \quad (70)$$

which allows for the calculation of the reference shear stress for each diameter class i and the value of Φ over time, based on

$$\tau_{ri} = \tau_{rm} \left(\frac{d_i}{D_m} \right)^b \quad (71)$$

$$\Phi = \frac{\tau}{\tau_{ri}} \quad (72)$$

In this case, the dimensionless transport rate for each fraction is calculated from

$$W_i = \begin{cases} 0.002\Phi^{7.5}, & \Phi < 1.35 \\ 14 \left(1 - \frac{0.894}{\sqrt{F_i}} \right)^{4.5}, & \Phi \geq 1.35 \end{cases} \quad (73)$$

The transport rate Q_b (m^3/s) is given by the expression

$$Q_b = W_i f_i \frac{B(gR_h S)^{3/2}}{(s_g - 1)g} \quad (74)$$

5.2.5 Takahashi et al. (1987) transport model

The [Takahashi et al. \(1987\)](#) model addresses debris flows with relatively low sediment concentrations, where particle transport is still dominated by the interaction between the water flow and the sediment, without reaching the fully granular behavior of a mature flow. It requires a minimum slope for transport to occur. The model proposes a dimensionless sediment transport rate defined by

$$q_s^* = \frac{2}{3} \frac{(4.2 - 0.3 \times 0.65)}{\cos^2 \theta_s (0.8 \tan \theta_s)^2} \tau_*^{2.5} \quad (75)$$

where θ_s is the slope angle, u^* is the friction velocity, and τ_* is the dimensionless shear stress.

The threshold condition is

$$\tau_* > \cos \theta_s \left[\left(0.65 + \frac{q_s^*}{8.5} \right) (0.8 - \tan \theta_s) - \frac{\rho_s - \rho_w}{\rho_w} \tan \theta_s \right] \quad (76)$$

The transport rate Q_b (m^3/s) is given by the expression

$$Q_b = q_s^* u_* d_{50} B \quad (77)$$

5.2.6 Schoklitsch (1962) transport model

The [Schoklitsch \(1962\)](#) model is a classic bed-load sediment transport formulation developed for mountain channels. Although not originally conceived for debris flows, it has been used in some simplified models of immature debris flows, where solid transport is dominated by flow capacity and a threshold for the initiation of particle motion exists. The model estimates the bed-load contribution (m_s^3/s) based on the expression

$$Q_b = K_{sc} S^\beta (Q - Q_c) B \quad (78)$$

where Q_b is the bed-load discharge (m_s^3/s), Q is the flow discharge (m^3/s), Q_c is the critical discharge for the onset of transport, K_{sc} is an empirical coefficient depending on sediment size and density—estimated as $2.5/s$ with values in the range $0.5 < K_{sc} < 5$ — β is an exponent varying between 1.5 and 2, and S is the channel slope.

The critical discharge can be expressed as ([Lenzi et al., 1999](#); [Millares et al., 2014](#); [Lammer et al., 2025](#))

$$Q_c = 0.26(s-1)^{5/3} (d_{40}/S^{0.78})^{3/2} \quad (79)$$

5.3 Total load

Total-load sediment transport models calculate the movement of the entire sediment mass within the river cross-section, including both bed load and suspended load. In this case, the state variables include—in addition to the bed shear stress, τ_b —the mean flow velocity and the concentration of transported sediment.

5.3.1 Bagnold (1966) transport model

The [Bagnold \(1966\)](#) model is based on the concepts of flow power and transport efficiency. The proposed equation is

$$Q_{ct,Bag} = \frac{\rho_w R_h S U B}{\rho_s} \left(\frac{e_b}{0.63} + 0.01 \frac{U}{w_s} \right) \quad (80)$$

where R_h is the hydraulic radius, S is the energy line slope, U is the mean flow velocity, B is the channel width, e_b is the bed transport efficiency ($0.2 < e_b < 0.3$), and w_s is the particle settling velocity.

5.3.2 Engelund-Hansen (1967) transport model

The [Engelund and Hansen \(1967\)](#) model relates sediment transport to flow power via the Shields parameter. The sediment concentration is calculated as

$$C_w = 0.05 \left(\frac{s}{s-1} \right) \frac{US}{\sqrt{(s-1)gd_s}} \frac{R_h S}{(s-1)d_s} \quad (81)$$

where d_s is the characteristic diameter of the grain-size distribution. The sediment discharge is obtained from

$$Q_{ct,EA} = C_w \times 10^6 \times 0.001 \times \frac{Q}{\rho_s} \quad (82)$$

where Q is the liquid discharge.

5.3.3 Brownlie (1981) transport model

The [Brownlie \(1981\)](#) model uses dimensionless parameters to estimate sediment transport. First, the dimensionless diameter of the particles is defined:

$$d_* = d_s \left(\frac{(s-1)g}{\nu^2} \right)^{1/3} \quad (83)$$

The critical Shields parameter is calculated as:

$$Y_{bw} = \left(\frac{g^{1/2}(s-1)^{1/2}d_s^{3/2}}{\nu} \right)^{-0.6} \quad (84)$$

$$\tau_{*,c} = 0.22Y_{bw} + 0.06 \times 10^{-7.7Y_{bw}} \quad (85)$$

The critical speed of movement initiation is determined by:

$$U_c = \sqrt{(s_g - 1)gd_s} \times 4.596\tau_{*,c}^{0.529} S^{-0.1405} \sigma_g^{-0.1606} \quad (86)$$

where σ_g is the geometric deviation of the granulometric distribution, which can be estimated from the relationship

$$\sigma_g = 0.5 \left(\frac{d_{84}}{d_{50}} + \frac{d_{50}}{d_{16}} \right) \quad (87)$$

The sediment concentration in ppm (mg/L) is calculated when the flow velocity exceeds the critical velocity:

$$C_{ct,Brw} = 7115c_b \left(\frac{U - U_c}{\sqrt{(s-1)gd_{s,\sup,4}}} \right)^{1.978} S^{0.6601} \left(\frac{R_h}{d_s} \right)^{-0.3301} \quad (88)$$

where $c_b = 1.268$ is a calibration coefficient.

Finally, the transported solid discharge is determined from

$$Q_{ct,Brw} = C_{ct,Brw} \times 0.001 \times \frac{Q}{\rho_s} \quad (89)$$

5.4 Fluvial dynamics

The HEMINGS model simulates sediment transport along cells defined as river cells. To do this, it employs the Exner equation, which determines the temporal evolution of the riverbed elevation based on the erosion/deposition balance. In its continuous, one-dimensional form, the Exner equation is written as

$$(1-p) \frac{\partial z_b}{\partial t} = - \frac{\partial q_s}{\partial x} \quad (90)$$

where p is the bed material porosity (dimensionless), z_b is the bed elevation [L], and q_s is the volumetric sediment discharge per unit width [L^2/T], calculated using a suitable bed-load or total-load model.

Discretization using finite differences in time (explicit scheme) and finite volumes in space for each river cell of the DEM leads to

$$z_b^{n+1}(i) = z_b^n(i) - \frac{\Delta t}{(1-p)\Delta x} [q_{s,out}(i) - q_{s,in}(i)] \quad (91)$$

where $q_{s,in}(i)$ is the sediment discharge entering the cell from the upstream cell, and $q_{s,out}(i)$ is the sediment discharge leaving the cell toward the downstream cell.

5.5 Analysis of check dams for forest hydrological restoration

The study of sediment transport in mountain rivers is of great importance due to high flow energy, the torrential nature of the flow, and the significant temporal variability of discharge rates. In these channels, flood events can trigger significant morphological changes over short timescales, involving bed erosion and deposition that affect both flow dynamics and infrastructure located within the channel. The HEMINGS model has been configured to analyze fluvial dynamics in mountain river reaches subject to the installation of forest-hydrological control dams and to assess the associated impacts. By considering estimated transport modes, appropriate fluvial transport models for each section, and fluvial dynamics derived from the Exner equation, the forest-hydrological control dam analysis module enables the estimation of sediment volume entering the dam for various return periods, the resulting equilibrium slope (see Fig. 11), the critical shear stress, and the tractive force. Furthermore, it allows for the evaluation of the dam's flow-attenuation effect by comparing inflow and outflow rates.

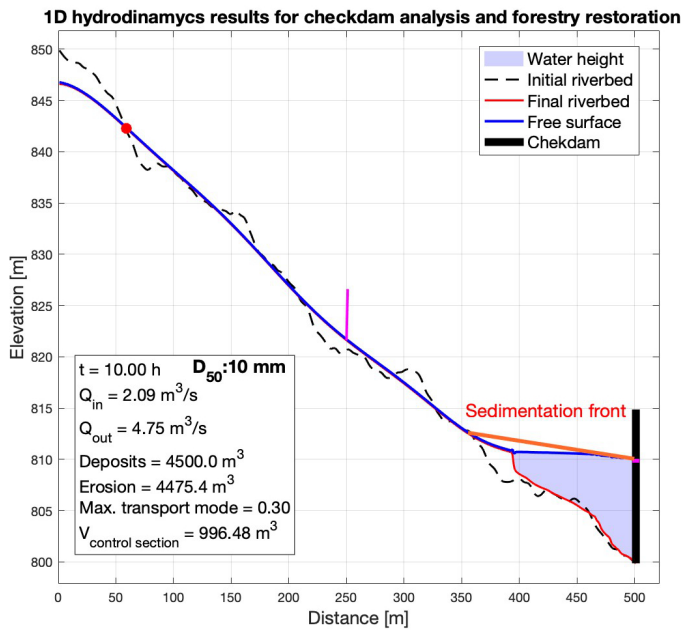


Figure 11. 1D hydrodynamic and geomorphological analysis of a river reach and a hydro-forestry check dam using the HEMINGS model on the Guadalmedina River (Málaga, Spain).

This application is particularly valuable for analyzing, configuring, and assessing the impact of installing a forest-hydrological control dam in a real mountain torrent. The module employs a one-dimensional numerical model that solves a coupled system comprising: 1) the shallow-water equations—based on the conservation of mass and momentum—to describe the flow; 2) the Exner equation to model bed evolution; and 3) a transport model selected according

to the specific transport mode to estimate the sediment discharge rate (Díaz et al., 2008). The input data required for the analysis include: a) the actual longitudinal profile of the channel under study; b) the inflow hydrograph simulated using HEMINGS for a specific return period; c) sediment grain-size data; d) the characteristic cross-section(s) of the reach; and e) the location and specifications of the dam being analyzed. The numerical scheme is based on a Godunov-type method with an HLL approximate solver (Harten et al., 1983; Toro, 2013), suitable for capturing shock waves and supercritical flows typical of torrential rivers. The flow is described by the mass and momentum conservation equations, written in conservative form as:

$$\frac{\partial \mathbf{U}}{\partial t} + \frac{\partial \mathbf{F}(\mathbf{U})}{\partial x} = \mathbf{S}(\mathbf{U}), \quad (92)$$

where the vector of conservative variables and physical fluxes is defined by the expressions

$$\mathbf{U} = \begin{pmatrix} A \\ Q \end{pmatrix} \quad \mathbf{F}(\mathbf{U}) = \begin{pmatrix} Q \\ \frac{Q^2}{A} + \frac{1}{2}gAh \end{pmatrix}. \quad (93)$$

Here, A is the wetted cross-sectional area, Q is the discharge, h is the flow depth, g is the acceleration due to gravity, and x and t are the spatial and temporal coordinates. The source term $\mathbf{S}$ incorporates the bed slope and friction:

$$\mathbf{S} = \begin{pmatrix} 0 \\ gA(S_{bed} - S_f) \end{pmatrix}, \quad (94)$$

where $S_{bed} = -\partial z_b / \partial x$ is the bed slope and S_f is the friction slope, evaluated using Manning's formula

$$S_f = \frac{n^2 Q |Q|}{A^2 R_h^{4/3}}, \quad (95)$$

with n being Manning's coefficient and R_h the hydraulic radius. For mountain cross-sections, a non-linear geometric relationship is used:

$$A = Bh + \theta h^2, \quad (96)$$

where B is the channel bottom width and θ is a coefficient representing the slope of the channel banks. This expression captures the widening of the channel as the flow depth increases, a characteristic feature of torrential rivers. The mean flow velocity and the celerity of gravity waves are calculated as $u = Q/A$ and $c = \sqrt{gh}$. These quantities define the characteristic speeds $\lambda^\pm = u \pm c$, which are used in the HLL solver. Bed evolution is described by the Exner equation—expressing sediment mass conservation—as outlined in the previous section. The domain $[0, L]$ is discretized into N_x cells of size Δx . Time-stepping is performed using a step size Δt that satisfies the CFL condition:

$$\Delta t \leq \text{CFL} \cdot \frac{\Delta x}{\max(|u| + c)}. \quad (97)$$

For each interface $i + 1/2$ between cells i and $i + 1$, the left (L) and right (R) states are calculated. The extreme wave speeds are estimated as $V_L = \min(u_L - c_L, u_R - c_R)$ and $V_R = \max(u_L + c_L, u_R + c_R)$. The numerical flow of HLL is defined according to the sign of V_L and V_R :

$$\mathbf{F}_{i+1/2}^{HLL} = \begin{cases} \mathbf{F}_L, & V_L \geq 0, \\ \mathbf{F}_R, & V_R \leq 0, \\ \frac{V_R \mathbf{F}_L - V_L \mathbf{F}_R + V_L V_R (\mathbf{U}_R - \mathbf{U}_L)}{V_R - V_L}, & V_L < 0 < V_R. \end{cases} \quad (98)$$

This solver is robust for supercritical and transcritical flows, common conditions in mountain rivers. Once the flows in all interfaces have been calculated, the update of the conserved variables follows the explicit Godunov scheme:

$$A_i^{n+1} = A_i^n - \frac{\Delta t}{\Delta x} (F_{1,i+1/2} - F_{1,i-1/2}), \quad (99)$$

$$Q_i^{n+1} = Q_i^n - \frac{\Delta t}{\Delta x} (F_{2,i+1/2} - F_{2,i-1/2}) + \Delta t g A_i^n (S_{bed,i} - S_{f,i}). \quad (100)$$

The following boundary conditions are applied: upstream, a flow hydrograph $Q_{in}(t)$ and the corresponding draft are imposed, and downstream, an outflow flow Q_{out} or a free flow condition is imposed. Boundary fluxes are calculated directly from the imposed variables:

$$F_{1,left} = Q_{in}, \quad F_{2,left} = \frac{Q_{in}^2}{A_L} + \frac{1}{2} g A_L h_L. \quad (101)$$

Figure 11 shows the final simulation results from the hydro-forestry check dam analysis module, applied to an actual mountain river reach in the Guadalmedina River catchment using an inflow hydrograph estimated by the HEMINGS model for a 20-year return period. The analysis employed the Meyer-Peter and Müller sediment transport model (with the correction proposed by Wong and Parker (2006)), a characteristic sediment diameter of $d_{50} = 10$ mm, and a constant cross-section with a 10-meter base and side slopes of $\theta = 0.5$. The figure illustrates the evolution of the riverbed and the slope of the sediment accumulated at the check dam.

6 Sensitivity analysis and calibration

Figure 12 presents sensitivity index box plots showing median values (red lines), 25th and 75th percentiles (box edges), and extreme values (red crosses). The results were derived from different flow and erosion parameters.

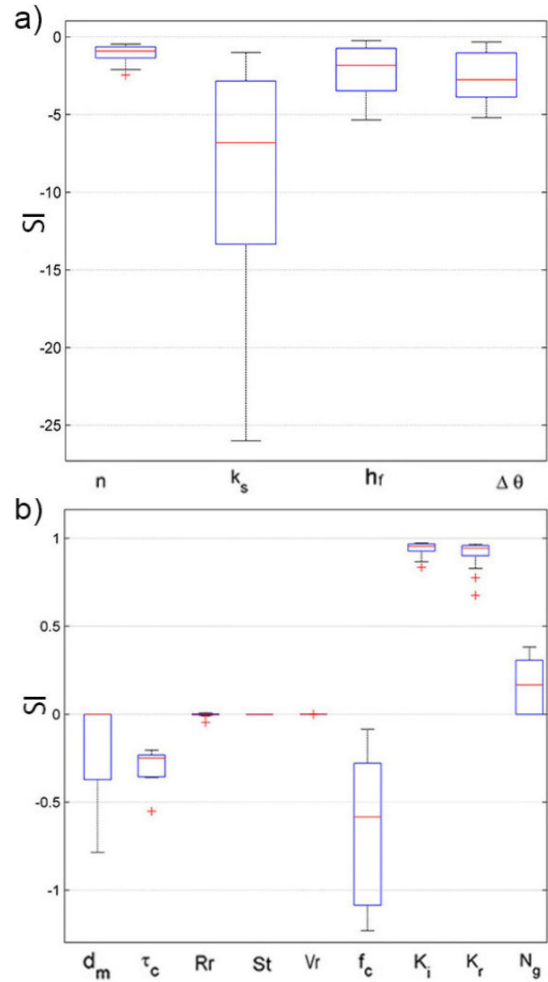


Figure 12. Sensitivity index box plots showing median values (red lines), 25th and 75th percentiles (box edges), and extreme values (red crosses) (Millares et al., 2019).

These results were evaluated in Millares et al. (2019) through a local analysis of the various parameters. The sensitivity index SI (dimensionless), calculated for an increase in a specific parameter, was derived from the expression (Van Griensven et al., 2006)

$$SI = \frac{\frac{R_m(p_1, \dots, p_i + \Delta p_i, \dots, p_m) - R_m(p_1, \dots, p_i, \dots, p_m)}{R_m(p_1, \dots, p_i, \dots, p_m)}}{\frac{\Delta p_i}{p_i}} \quad (102)$$

where R_m is the model output (peak discharge, hydrograph volume, etc.) and Δp_i is the variation of a given parameter p_i , while the other variables remain constant. The sign of SI indicates whether the model responds in the same direction as p_i . For runoff processes, the selected parameters were n , k_s , h_f , and $\Delta\theta$, with respect to the peak discharge modeled at the sub-catchment outlet during the event. For erosion/deposition processes, the selected parameters were d_m , τ_c , R_r , St , V_r , f_c , K_i , K_r , and N_g , associated with the total simulated sediment volume at the catchment outlet. Table ?? summarizes the range of values considered in previous studies (USDA-SCS, 1992; Foster et al., 1995; Romero et al., 2007; Knapen et al., 2007) and the data estimated for the study site. As noted by Knapen et al. (2007), the values of τ_c and K_r can vary by several orders of magnitude among soils with different textures or organic matter properties.

The model results show high sensitivity to variations in the

soil hydraulic conductivity parameter, k_s . This has previously been observed in other studies (De Roo et al., 1996). Regarding erosion modeling (Fig. 12), the model exhibits the greatest sensitivity to soil erodibility parameters. Sensitivity to the soil cover fraction was lower, although it displayed a very wide range. The parameter τ_c shows moderate sensitivity. The role of soil particle diameter in sedimentation processes is noteworthy, which is to be expected given its influence on settling velocity.

6.1 Model calibration

The sensitivity analysis results allow us to optimize the number of parameters subject to calibration. The parameter f_c is relatively easy to estimate from orthophotography and satellite imagery. The value of d_m can be estimated from soil samples (Perez, 1987; Delgado et al., 1993). Although its spatial distribution varies, the range of values considered for d_m is sufficiently broad to evaluate its sensitivity. The values for flow resistance and soil hydraulic conductivity (n and k_s), used to calculate hillslope runoff, and those for erosion (τ_c or K_r), are the most difficult to determine at the catchment scale. To reduce parameter estimation uncertainty, 2,500 simulations were conducted in Millares et al. (2019). For each simulation, peak discharge and sediment volume were compared against field measurements, and the Nash-Sutcliffe efficiency (NSE), correlation coefficient, and mean relative error were calculated.

Figure 13 shows the runoff and sediment flux calibration results for the selected parameters. The color plots display the Nash-Sutcliffe model efficiency (first column), correlation coefficients (second column), and mean relative errors (third column) for n and k_s (top row panel) and τ_c and K_r (bottom row). The best-fitting values can be clearly identified via the Nash-Sutcliffe model efficiency coefficient and the mean relative error. Regarding the flow resistance coefficient, n , and the infiltration rate, k_s , a linear relationship is observed with a narrow range of optimal values for the k_s factor—spanning 0.4 to 0.45—which represents nearly half of the original estimated map for k_s (where maximum values ranged from $1.4 \times 10^{-6} \text{ m s}^{-1}$ to $5 \times 10^{-6} \text{ m s}^{-1}$). The values for n exhibit a wider range, from 0.015 to 0.04.

Soil erodibility and critical shear stress showed an increasingly exponential relationship, with a relatively constant value of 4 Pa as soil erodibility values increased. This behavior highlights how changes in soil erodibility below $1.0 \times 10^{-3} \text{ s m}^{-1}$ significantly affect concentrated flow erosion. Above this threshold, changes in critical soil shear stress occur slowly, reaching higher levels with a mean value of approximately 4 Pa. This relationship aligns with that described by Knapen et al. (2007) based on field and laboratory tests.

6.2 Model validation

Based on the preceding analysis, a constant value of $n = 0.04$ was selected for runoff flow, while distributed maps were used for water flow parameters, with ranges of 1.44×10^{-6} – $5 \times 10^{-6} \text{ m s}^{-1}$ for k_s , 3.6–16 mm for h_f , and 0.3–0.45% for $\Delta\theta$. Five simulations were conducted to evaluate the effect of different erosion processes and their relationship with soil erodibility (Table ??). Simulation 1 excluded gully formation processes, accounting only for rill erosion and raindrop impact. Simulation 2 incorporated gully formation processes, using a

constant soil erodibility profile of $K_g = 0.2K_r$ and a significant reduction in rill erodibility. In simulations 3 through 5, the initial surface erodibility (relative to rills) was kept constant, while the soil erodibility profile was modified to $K_g = 0.5K_r$, $K_g = 0.2K_r$, and $K_g = 0.1K_r \text{ s m}^{-1}$.

Table 2. Values of the parameters of the HEMINGS model in different simulation scenarios and $\tau_c = 5 \text{ Pa}$.

Unit	R_c (-)	St (-)	C_f (-)	K_i kg s m^{-4} ($\times 10^4$)	K_r s m^{-1} ($\times 10^{-3}$)	K_g s m^{-1}	d_m mm
1	0.06	0.4	0.7	[4.6; 16]	[0.3; 1.0]	0	0.15
2	0.06	0.4	0.7	[4.6; 16]	[0.4; 0.16]	$0.2K_r$	0.15
3	0.06	0.4	0.7	[4.6; 16]	[0.3; 1.0]	$0.1K_r$	0.15
4	0.06	0.4	0.7	[4.6; 16]	[0.3; 1.0]	$0.2K_r$	0.15
5	0.06	0.4	0.7	[4.6; 16]	[0.3; 1.0]	$0.5K_r$	0.15

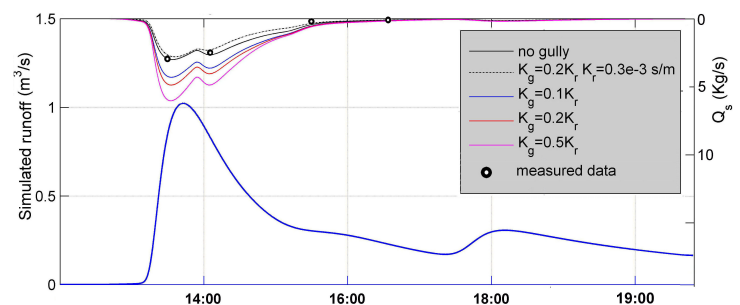


Figure 14. Validation results of the HEMINGS model for discharge (lower curve, left axis) and sediment transport (upper curves, right axis) (Millares et al., 2019).

Figure 14 shows the results corresponding to the five simulations. As can be seen, the modeled erosive response matches the measured data in both timing and shape, exhibiting a peak bimodal sediment flux linked to the geomorphological configuration of the sub-catchment, which comprises two drainage units. Excellent agreement is observed regarding the Nash-Sutcliffe efficiency coefficient for the calibrated parameters of simulations 1 and 2, with $\text{NSE} = 0.89$ and $\text{NSE} = 0.88$, respectively. It is important to note that, for simulation 2, it was necessary to reduce rill erodibility by 70% to compensate for the effect of gully formation. Simulations 3 through 5 yield values that overestimate the measurements when the calibrated rill erodibility values are maintained, showing a proportional increase linked to the internal erodibility of the soil profile.

Figure 15 shows the validation results of the HEMINGS model for the Guadalmedina River catchment (Millares and Nanía, 2026). It can be observed that the model correctly captures runoff flows, with NSE values ranging from 0.76 to 0.92, for events of varying intensity.

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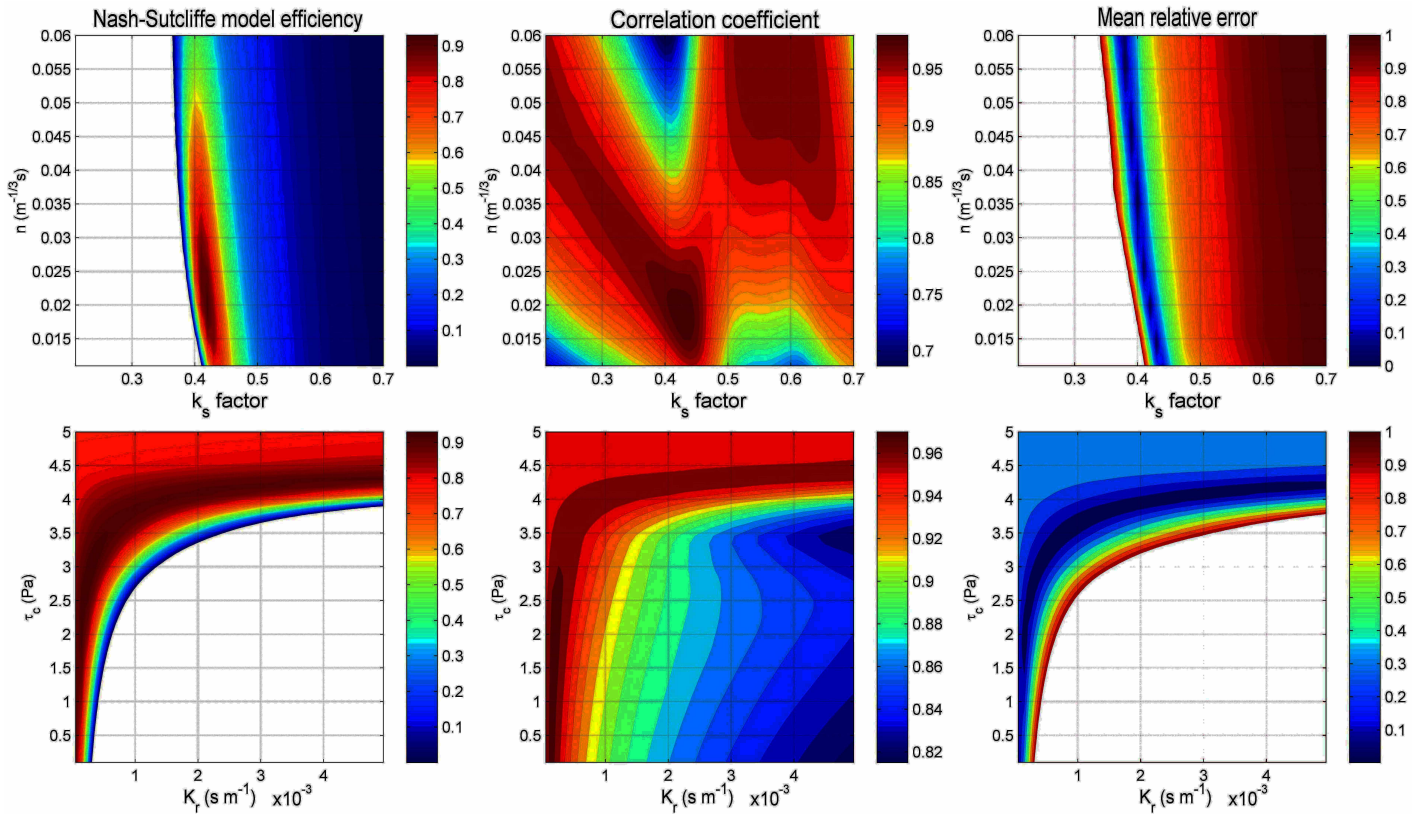


Figure 13. Model calibration results for runoff (top row) and sediment flux (bottom row), based on an objective function derived from measured hydrographs and sedimentographs, specifically selecting Q_{peak} and sediment volume (Millares et al., 2019).

demic and Research Collaboration (SPARC), Ministry of Education, Govt. of India. 3) LIFEWATCH: *Thematic Center on Mountain Ecosystems and Remote Sensing, Deep Learning/Artificial Intelligence, and Electronic Services*. Ministry of Science and Innovation. Spain. 4) CCAMICEM: *Morphological and Systemic Response of Ephemeral Channels to Climate Change: Dynamics, Resilience and Proposals for Action*. Ministry of Industry, Economy and Competitiveness. Spain. 5) H2020-ECOPOTENTIAL: *Improving Future Ecosystem Benefits Through Earth Observations*. H2020-Framework Programme. European Commission.

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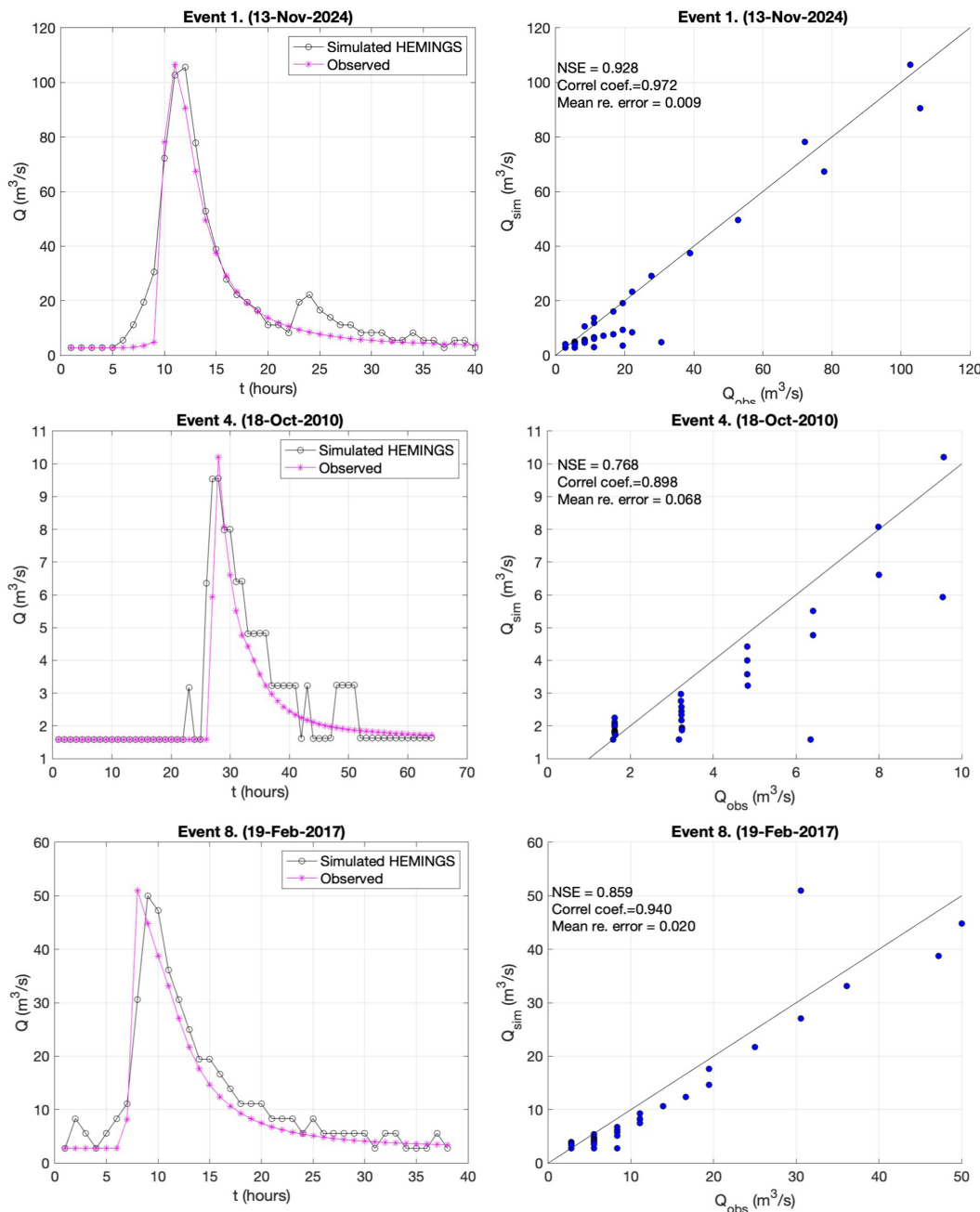


Figure 15. Validation results of the HEMINGS model for different events in the Guadalmedina River catchment (Málaga, Spain) (Millares and Nania, 2026).

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