



Tracing Catchment-Wide Erosion Using Pu-Isotopes: Explorative Tests for the Erlenbach (Alptal) Watershed

GEO 511 Master's Thesis

Author: Noelle Talina Schumacher, 21-721-618

Supervised by: Prof. Dr. Markus Egli, Dr. Ilja van Meerveld

Faculty representative: Prof. Dr. Markus Egli

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Abstract

Quantifying soil erosion is crucial, as it is associated with the loss of fertile topsoil, impacts on aquatic ecosystems and damage to infrastructure and reservoir storage. In alpine regions, this can be difficult, as erosion processes are highly variable across space and time. This thesis aims to determine long-term erosion rates in the Erlenbach catchment (Alptal, Switzerland) and to evaluate the suitability of plutonium isotopes ($^{239+240}\text{Pu}$) as tracers for estimating soil erosion at the catchment scale.

For this purpose, long-term monitoring data on sediment transport, recorded at the measuring station at the outlet, were compared with isotope-based estimates. Erosion rates were calculated, using suspended sediment and bedload data, as well as Pu isotopes from soil and sediment samples. Soil samples were analyzed using various modeling approaches, while a steady-state mass balance approach was applied to river sediments, assuming a constant atmospheric input.

The long-term erosion rate derived from suspended sediment is approximately $-1.76 \text{ t ha}^{-1} \text{ yr}^{-1}$ and agrees well with the Pu-based erosion rates from soil (averaging -0.8 to $-2.3 \text{ t ha}^{-1} \text{ yr}^{-1}$) and sediment samples (averaging -1.22 to $-3.25 \text{ t ha}^{-1} \text{ yr}^{-1}$). In contrast, bedload results in a higher erosion rate ($-9.06 \text{ t ha}^{-1} \text{ yr}^{-1}$), indicating that it captures processes such as riverbed and bank erosion and is not directly comparable to soil erosion. No statistically significant differences were found between the various Pu-based models, indicating good agreement among the results despite different model assumptions.

These findings suggest that analyzing river sediments at the catchment outlet is an effective approach to assess average soil erosion across the entire catchment with relatively few samples. However, uncertainties remain, regarding the assumption of a steady-state equilibrium between atmospheric input and erosion-induced output, which can be further influenced by temporary sediment storage and delayed transport processes within the catchment. Overall, the thesis shows that Pu isotopes provide a suitable and reliable method for estimating long-term soil erosion in Alpine catchments.

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1 Introduction

Soil erosion is a natural geomorphic process that is influenced by human activities. Precipitation is a key driver, as raindrop impact detaches soil particles which are then transported downslope by overland flow (Ali et al., 2021; Tamene et al., 2022). Human activities such as deforestation, overgrazing, intensive agriculture and soil compaction can further accelerate erosion by reducing vegetation cover and soil stability (Ali et al., 2021; Andualem et al., 2023; Mushi et al., 2019; Sestras et al., 2023). Climate change intensifies these processes, as many regions are seeing an increase in heavy rainfall and changes in runoff patterns (S. Dowell et al., 2024; Upadhayay et al., 2022).

In alpine regions, erosion is further enhanced by steep topography, shallow soils with low infiltration capacity and seasonal dynamics. Snowmelt and rain-on-snow events can generate high runoff within a short period of time and mobilize large amounts of sediment (Molnar et al., 2008; Rickenmann & McARDell, 2007). The consequences include the loss of fertile topsoil, increased sediment input into water bodies, impacts on aquatic ecosystems, reduced reservoir capacity and damage to infrastructure (Andualem et al., 2023; Barberena et al., 2023).

Reliable quantification of soil erosion is therefore essential for assessing ecosystem stability, planning mitigation measures and supporting sustainable land and water management, including the implementation of the Sustainable Development Goals (SDGs) (Alewell et al., 2013; Ali et al., 2021; Sestras et al., 2023).

Various methods are available to measure soil erosion. Direct methods include, for example, erosion pins for recording small-scale surface lowering (Mushi et al., 2019) or erosion plots in which soil loss is measured under defined conditions (Pampalone et al., 2022). At the catchment scale, retention basins can be used to collect and quantify transported material, while sensor-based systems such as acoustic or seismic devices allow continuous monitoring of sediment transport (Nitsche et al., 2011; Rickenmann et al., 2012; Rickenmann & McARDell, 2007). In addition, physically based models are commonly applied but rely on numerous input parameters and assumptions (Bhattarai & Dutta, 2007). Despite these approaches, significant challenges remain. Field measurements are often time-consuming and costly, limiting their application in large or complex catchments. The strong spatial and temporal variability of erosion processes makes it also difficult to obtain representative results (Andualem et al., 2023; Zhai et al., 2025).

In addition to these conventional methods, isotope-based tracer techniques have become widely used for quantifying soil erosion. These approaches rely on fallout radionuclides, which differ in their origin, transport pathways and temporal applicability. Commonly used isotopes include cesium-137 (^{137}Cs), lead-210 ($^{210}\text{Pb}_{\text{ex}}$), beryllium-7 (^7Be) and plutonium ($^{239+240}\text{Pu}$) (Mabit et al., 2008). Pu isotopes, which were mainly released into the atmosphere during global weapons testing in the 1950s and 60s, were deposited on the Earth's surface through precipitation and bind strongly to fine soil particles. They offer particular advantages in alpine regions due to their comparatively homogeneous distribution and long half-life (Alewell et al., 2013).

The application of plutonium (Pu) as an erosion tracer is based on the comparison of radionuclide inventories. For soil samples, erosion rates are derived by comparing the Pu inventory at a sampling site with that of a stable reference site (S. Dowell et al., 2024). In contrast, an alternative and less established approach is the analysis of river sediment samples collected at the catchment outlet. In this case, the measured Pu activity reflects a mixture of material originating from different parts of the

catchment and thus provides an integrated signal of upstream erosion processes (Chen et al., 2023; Percich et al., 2022). Variations in Pu activity can then be used to estimate catchment-scale soil erosion rates. Therefore, isotope-based methods enable the reconstruction of long-term soil redistribution from a single sampling (Huang et al., 2025; Mabit et al., 2008).

In this context, the following master's thesis examines the long-term erosion rate in the Erlenbach catchment in the Alptal. The area is characterized by steep channels with a pronounced step-pool morphology and a rapid hydrological response to precipitation (Rickenmann & McArdell, 2007). The aim of this thesis is to use $^{239+240}\text{Pu}$ isotopes from soil and sediment samples to determine an erosion rate and to assess the reliability of this relatively new method for an alpine catchment. The analysis of river sediment samples collected at the catchment outlet is of particular interest, as it provides a catchment-integrated signal of soil erosion with relatively low sampling effort.

The Erlenbach catchment provides an ideal setting for this study, as long-term monitoring data of both bedload transport (since the 1980s) and suspended sediment (since 2003) are available. By combining Pu isotope analyses from both soil samples at erosional sites and sediment samples collected at the catchment outlet, this study aims to critically evaluate whether the outlet-based approach provides a robust estimate of catchment-scale soil erosion. The comparison with independent monitoring data further allows for a direct assessment of the method's performance under real field conditions.

This leads to the following key research questions:

1. What are the long-term erosion rates calculated from the existing sediment transport data of the Erlenbach?
2. What are the long-term erosion rates based on Pu isotopes ($^{239+240}\text{Pu}$) from soil and sediment samples?
3. To what extent do the two methods agree and what differences or limitations arise in their comparison?

2 Background

2.1 Soil Erosion and Sediment Transport

2.1.1 Fundamental Processes and Influencing Factors of Soil Erosion

According to Ali et al. (2021), soil erosion refers to a geomorphic process that typically involves three stages. Soil particles are first detached from the soil matrix, then transported by surface runoff, wind or gravity, and eventually deposited again. The detachment of particles is primarily caused by the kinetic energy of raindrops and surface runoff. The impact of individual drops leads to the destruction of soil aggregates and the detachment of fine particles (Ali et al., 2021). Vegetation cover is an important protective factor, as it absorbs the impact energy of raindrops, increases the roughness of the surface and promotes infiltration through root channels. If this protective layer is missing, for example due to clearing, overgrazing or intensive tillage, erosion increases significantly (Ali et al., 2021; Chen et al., 2023; Mushi et al., 2019).

Sandy soils have low cohesion, but often allow high infiltration, which results in less surface runoff. Clay-rich soils on the other hand, have higher cohesion, but can become heavily sealed when saturated, which can promote surface runoff. Larger raindrops increase kinetic energy and thus the ability to detach particles, leading to more intense erosion. The duration of precipitation also influences erosion dynamics, as longer rainfall increases soil saturation and makes overland flow more likely (Ali et al., 2021; Kolli et al., 2021).

Overland flow is a central mechanism that links detachment with transport. It occurs either when precipitation intensity exceeds the infiltration capacity of the soil or when the soil is already completely saturated (Upadhayay et al., 2022; Zimmermann et al., 2012). In particular, the flow depth, velocity and discharge volume determine which particle sizes can be mobilized (Mushi et al., 2019). Topography also plays a key role in erosion processes. With steeper slopes, surface runoff flows faster, leading to higher shear stress and a greater transport capacity of the water. Longer slopes also allow for greater accumulation of runoff and thus higher erosion rates in the lower slope area (Ali et al., 2021).

2.1.2 Coupling of Soil Erosion and Fluvial Sediment Transport

The soil particles mobilized on the slopes do not usually remain where they were detached, but are carried into the water network. In steep, sparsely forested catchments with short transport distances, the connection between the slope and the river is often very direct. Surface runoff can quickly transport the eroded material into the channel. In flatter or more structured landscapes, a larger proportion of the sediment will be deposited on the slope or in topographical depressions, reducing the amount that enters the river (Antoniazza et al., 2023; Chen et al., 2023; Guillory et al., 2025).

As soon as the sediment reaches the channel, it becomes part of the fluvial transport system. A basic distinction is made between suspended load and bedload. Fine particles such as silt and clay can be held in the water column by turbulence and transported over long distances. Coarser particles such as sand, gravel or boulders, move as bedload along the riverbed. Which form of transport dominates depends on the grain size, the flow velocity and the hydraulic capacity of the river (Badoux et al., 2014). A key factor here is stream power. It determines whether sediment is mobilized, transported or deposited. If the flow velocity decreases, for example due to a lower gradient or a widening of the channel, the transport capacity decreases and deposition occurs. Such deposition areas serve as temporary or long-term sediment storage (Andualem et al., 2023).

This is where the concept of sediment delivery ratio (SDR) comes into play. The SDR is defined as „the ratio of sediment yield to total surface erosion“ (Bhattarai & Dutta, 2007, p. 1641). It therefore indicates the proportion of the material eroded on the slopes that reaches the outlet of the catchment. In most catchments, the SDR is significantly less than one, because some of the sediment is redeposited along the transport route. Steep slopes increase transport efficiency and thus the SDR. As the size of the catchment area increases, the SDR generally decreases, as longer transport distances increase the likelihood of intermediate storage (Ali et al., 2021; Zhai et al., 2025).

In alpine catchments, sediment transport is strongly influenced by hydrological events. According to Antoniazza et al. (2022), heavy rainfall events lead to rapid increases in runoff and high transport rates in the short term, with easily accessible material being mobilized in particular. Snowmelt on the other hand, has a slower effect. Rain-on-snow events, in which high precipitation intensities hit already

saturated soils and melting snow cover, are particularly effective. In such situations, both runoff volume and transport capacity are maximized.

Soil erosion and sediment transport are therefore closely linked processes that have spatial and temporal effects beyond individual events. Some of the eroded material is transported immediately, while some is temporarily stored and later remobilized. The sediment loads measured at the outlet of the catchment therefore represent the result of a multitude of redistribution, storage and transport processes (Antoniazza et al., 2022; Rickenmann, 2020, 2024). To estimate long-term erosion rates, approaches are therefore needed that can take this temporal and spatial integration into account. One such possibility is the use of fallout radionuclides such as $^{239+240}\text{Pu}$, which serve as tracers for soil and sediment transport (Loba et al., 2022).

2.2 Pu Isotopes as Erosion Tracers

2.2.1 Properties of Pu Isotopes

Definition and Origin

Pu isotopes ($^{239+240}\text{Pu}$) are anthropogenic radionuclides. In research, it is usually the sum of the activities of the two isotopes ^{239}Pu and ^{240}Pu that is measured, as these are often analyzed together due to their similar radioactive properties (S. M. Dowell et al., 2023; Loba et al., 2022). The primary source of these isotopes is global fallout from above-ground nuclear weapons testing. Between 1945 and 1980 520 atmospheric tests were carried out worldwide, with the most intense phase occurring in the 1950s and 60s (Figure 1). These explosions propelled radioactive material into the stratosphere, where it was distributed worldwide. Ultimately, Pu returned to the Earth's surface mainly through precipitation (Alewell et al., 2013; S. M. Dowell et al., 2023; Loba et al., 2022; Zhang, 2023).

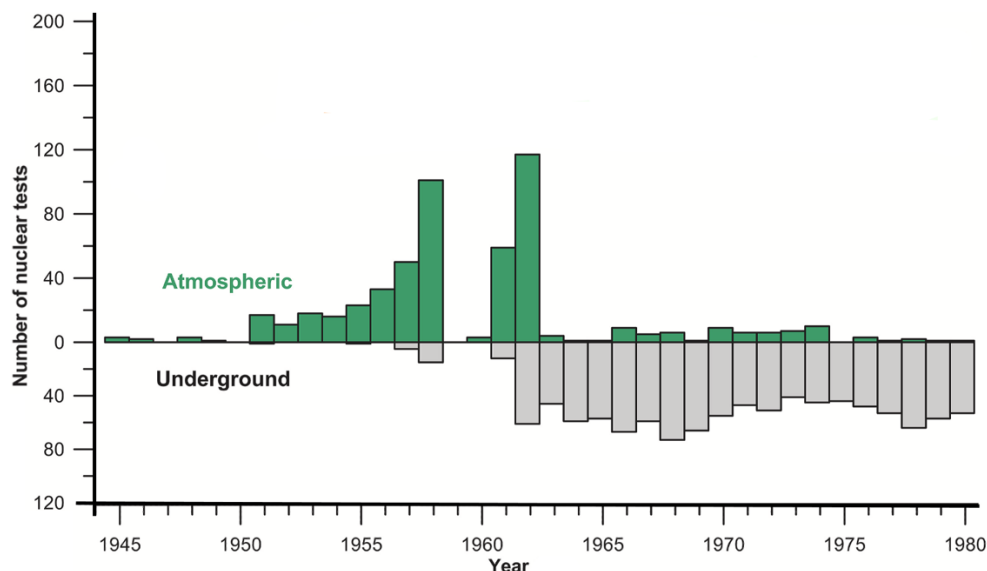


Figure 1. Number of atmospheric and underground nuclear weapon tests (Figure modified from Severi et al. (2023)).

Geographical Distribution and Deposition

Pu was deposited worldwide, although its distribution shows clear regional differences. As around 90% of nuclear tests took place in the northern hemisphere, concentrations are significantly higher there

than in the southern hemisphere (S. M. Dowell et al., 2023; Loba et al., 2022). An important parameter is the isotope ratio ($^{240}\text{Pu}/^{239}\text{Pu}$), which averages around 0.18 for global fallout and serves as a "fingerprint" to distinguish Pu from other sources (Alewell et al., 2013; Loba et al., 2022; Meusburger et al., 2016). Other events, such as reactor accidents (e.g. Chernobyl in 1986 or Fukushima in 2011), nuclear weapons production, or waste reprocessing play a minor role in the global distribution, since Pu, unlike cesium, has a low volatility. This means that in the event of an accident, it usually remains in the solid fuel debris and is deposited more regionally (Alewell et al., 2013; Loba et al., 2022; Zheng et al., 2013). Consequently, Pu in the Swiss Alps originates almost exclusively from global fallout of historical nuclear tests.

Characteristics and Binding Behaviour

A key property of Pu as an erosion tracer is its very high temporal stability in the environment. The two relevant isotopes have extremely long half-lives: ^{239}Pu decays with a half-life of 24'110 years and ^{240}Pu with a half-life of around 6'561 years. Since the deposition from nuclear weapons testing only occurred around 60 to 70 years ago, 99% of the activity originally deposited in the soil is still present today. As a result, the original fallout signal is hardly weakened and is particularly well suited for investigating long-term erosion processes (Alewell et al., 2013; S. Dowell et al., 2024; S. M. Dowell et al., 2023; Loba et al., 2022). Another key property is the strong binding of Pu isotopes to fine soil particles such as clay, loam, silt and organic matter (Alewell et al., 2013; Loba et al., 2022; Mabit et al., 2008). This adsorption is almost irreversible under natural conditions, meaning that Pu is rarely transported in dissolved form (S. Dowell et al., 2024; Loba et al., 2022).

2.2.2 Application of Pu Isotopes in Soil Erosion Assessment

The use of Pu isotopes ($^{239+240}\text{Pu}$) to quantify erosion processes is based on comparing radionuclide inventories between a sampling site and a local reference site. The inventory describes the total activity of a radionuclide per unit area (usually in Bq m^{-2}) and thus represents the amount of fallout stored in a soil profile. The reference site must be a stable location in the terrain where neither soil erosion nor deposition has occurred since the beginning of global atmospheric fallout (S. M. Dowell et al., 2023; Meusburger et al., 2016; Zhai et al., 2025).

The basic principle for determining erosion is simple:

- Net erosion: A lower Pu inventory at the sampling site compared to the reference site indicates that Pu-enriched soil has been eroded since fallout (Meusburger et al., 2016; Zhai et al., 2025).
- Net deposition: A higher inventory indicates deposition of additional Pu-enriched material from other areas (Meusburger et al., 2016; Zhai et al., 2025).

The suitability of Pu as an erosion tracer is based on its strong binding to fine soil particles. Changes in inventories are therefore almost exclusively caused by physical processes such as erosion, transport and deposition of soil material, while chemical rearrangements play a negligible role. Differences between sample and reference inventories can thus be interpreted directly as a measure of soil erosion or deposition (S. M. Dowell et al., 2023; Meusburger et al., 2016; Zhai et al., 2025).

A key advantage of using Pu isotopes is that it allows erosion and deposition rates to be recorded simultaneously in a single sampling. While conventional methods usually only reflect short-term transport processes or local soil losses, radionuclide inventories provide a time-integrated picture of soil redistribution over several decades (S. Dowell et al., 2024; S. M. Dowell et al., 2023; Huang et al., 2025; Mabit et al., 2008). This makes the Pu-based approach a particularly powerful tool for reconstructing long-term erosion dynamics in complex catchments (Huang et al., 2025).

2.2.2.1 Upscaling from Point Measurements to Catchment Scale

Each soil sample, used to determine the $^{239+240}\text{Pu}$ inventory, initially provides only a point measurement (Andualem et al., 2023; Zhai et al., 2025). This means that the measured value describes the erosion or deposition rate at this specific location (Mabit et al., 2008; Zhai et al., 2025). However, since soil movements are variable and can differ greatly even over short distances, a single sample is not sufficient to understand what is happening in an entire catchment (Huang et al., 2025; Zhai et al., 2025).

To address this, upscaling approaches are used to transfer these point-based results to larger spatial scales. In practice, this works with the help of geographic information systems (Huang et al., 2025; Mabit et al., 2008). The catchment area is digitally divided into smaller, preferably homogeneous units (Andualem et al., 2023; Bhattarai & Dutta, 2007). These units are grouped according to characteristics that significantly influence erosion, such as slope gradient, soil type and land use. Representative samples are then collected for each unit, and the measured erosion rates are extrapolated to areas with similar characteristics, providing a spatially integrated picture of soil redistribution (Andualem et al., 2023; Kolli et al., 2021; Mabit et al., 2008).

A key advantage of this approach is its ability to identify spatial patterns and erosion "hotspots". This is a useful basis for planning soil protection measures (Andualem et al., 2023; Huang et al., 2025; Mabit et al., 2008). However, a major disadvantage is the amount of time and effort required (Mabit et al., 2008; Zhai et al., 2025). Since the landscape is often complex, many samples must be taken to reliably represent the natural diversity. Each sample must be prepared and measured in the laboratory, which makes the process costly (Huang et al., 2025; Mabit et al., 2008; Zhai et al., 2025). In addition, there is always the risk that small errors in point measurements will lead to large inaccuracies in the overall result when transferred to a large area (Huang et al., 2025).

2.2.2.2 Catchment Outlet Sampling

An alternative method for studying erosion on a catchment scale is to take sediment samples directly at the catchment outlet (Ali et al., 2021; Barberena et al., 2023). This approach assumes that the transported sediment represents a mixture of material derived from different parts of the catchment. As a result, samples taken at the outlet provide an integrated signal of the erosion processes occurring upstream (Ali et al., 2021; Andualem et al., 2023; Antoniazza et al., 2023; Chen et al., 2023).

For this purpose, permanent measuring stations with special flow channels (flumes) and automatic sampling devices are often installed at the catchment outlet to collect samples automatically at certain water or turbidity levels (Barberena et al., 2023; Commelin et al., 2022; Hancock et al., 2017). By analyzing the Pu isotope signature in these sediments, conclusions can be drawn about their origin: high Pu activities indicate erosion of surface soils where the fallout is concentrated. Low activities, on

the other hand, suggest contributions from deeper soil layers. Therefore, higher Pu activities in river sediments indicate lower erosion rates, whereas lower activities suggest higher erosion rates (Chen et al., 2023; Percich et al., 2022).

The method is efficient, as it provides an integrated picture of the net soil loss of the entire catchment using relatively few samples at the outlet (Barberena et al., 2023; Sestras et al., 2023). However, a major issue is the sediment delivery problem: not all material that is eroded ever reaches the outlet, as some of it is redeposited in depressions, at the foot of slopes or in wide, flat areas within the catchment (Antoniazza et al., 2022, 2023; Bhattarai & Dutta, 2007; Zhai et al., 2025). Furthermore, without additional techniques (such as "sediment fingerprinting"), it is not possible to determine the exact spatial origin of the sediments, for example, which slope they come from (Chen et al., 2023; Lizaga et al., 2024). Technical issues, like traps not being able to catch the finest particles, like clay, can also mess with the measurements (Guillory et al., 2025; Lammer et al., 2025).

2.2.3 Other Radionuclides for Soil Erosion Assessment

There are various fallout radionuclides used to quantify erosion and deposition processes over different time scales. While plutonium ($^{239+240}\text{Pu}$) is becoming increasingly important, cesium-137, lead-210 and beryllium-7 are the most commonly studied alternatives (Loba et al., 2022; Mabit et al., 2008).

^{137}Cs is the most validated tracer worldwide for medium-term periods of approximately 50 years. It is an artificial radionuclide (half-life: 30.2 years), released by nuclear weapon tests in the 1950s and 60s and by reactor accidents. In soils, it binds strongly to clay minerals and organic matter (Alewell et al., 2013; Mabit et al., 2008). However, its use is limited by radioactive decay, with more than 70% of the original activity already lost (Loba et al., 2022; Mabit et al., 2008). In addition, Chernobyl fallout has led to a heterogeneous deposition in parts of Europe, particularly due to uneven accumulation on snow-covered areas, resulting in a patchy distribution in the soil (Alewell et al., 2013; Loba et al., 2022).

Unlike cesium, lead-210 ($^{210}\text{Pb}_{\text{ex}}$) is a natural radionuclide (half-life: 22.3 years) produced from the decay of radon-222 and continuously deposited from the atmosphere. It enables the reconstruction of erosion rates over longer periods of up to 100 years and is independent of historical individual events such as bomb tests. However, measuring it is technically very challenging, since it is detected at a low energy. The analysis is also often subject to large uncertainties of 30-50% (Mabit et al., 2008).

Beryllium-7 (^7Be) is a natural isotope with a very short half-life (53.3 days). It therefore serves as a tracer for short-term processes. It is produced in the upper atmosphere by cosmic radiation and concentrates in the top few millimeters of the soil surface after fallout. A disadvantage is that its application is almost exclusively limited to bare soil, as vegetation cover intercepts the fallout (Mabit et al., 2008).

Compared to these tracers, Pu isotopes offer several advantages for erosion studies in mountainous environments. In the Alps, Pu originates almost exclusively from global nuclear weapons testing and is therefore more homogeneously distributed than ^{137}Cs . In addition, the very long half-lives ensure high temporal stability, which is important for future use (Alewell et al., 2013; Loba et al., 2022). Furthermore, Pu is subject to lower selectivity during transport than ^{137}Cs , which preferentially binds to the finest particles and is therefore often disproportionately washed out. Plutonium thus provides a more robust picture of actual soil movement (Meusburger et al., 2016).

3 Study Area: The Erlenbach Catchment

3.1 Geographical Location, Topography and Land Use

The Erlenbach catchment is located in the central Swiss pre-alps in the Alptal, with an area of 0.74 km² and has been used over several decades to monitor and analyse hydrological processes. The Erlenbach is a small mountain stream that flows into the river Alp as a tributary. Topographically, the catchment is characterized by a steep relief, ranging from 1110 to 1655 m a.s.l., with the main river channel having an average gradient of approximately 18% (Molnar et al., 2008; Rickenmann, 2020; Rickenmann & McArdell, 2007; Turowski et al., 2009). The catchment area is almost entirely vegetated, with land use divided into two main categories. About 39-40% of the area is forested, while the remaining 60% consists of alpine meadows and moorland. Less than 1% of the area is considered to be vegetation-free (Rickenmann, 2020; Rickenmann & McArdell, 2007; Schneider et al., 2014).

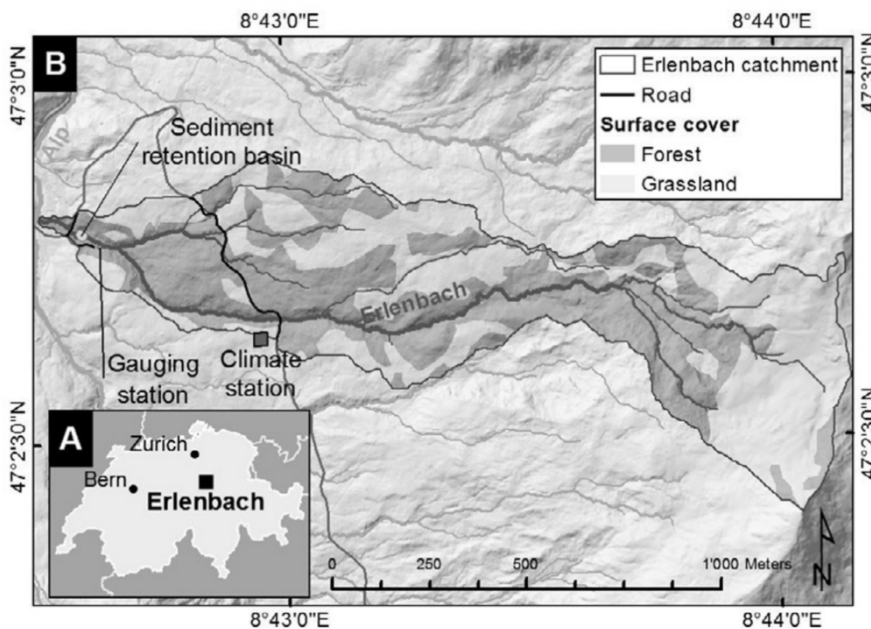


Figure 2. A: Location of the Erlenbach catchment in Switzerland. B: Map of the catchment (Turowski et al., 2016).

3.2 Hydrology and Climate

The Erlenbach catchment has a nivo-pluvial runoff regime, meaning that it is influenced by both winter snow storage and year-round rainfall (Smith et al., 2023). There are around 161 rainy days on average throughout the year. The average annual precipitation in the Alptal is very high at around 2300 mm, while the mean annual runoff is approximately 1770 mm (Nitsche et al., 2011; Rickenmann & McArdell, 2007; Schneider et al., 2014; Turowski et al., 2009).

Between November and April 30-40% of the annual precipitation falls as snow. The snow cover is highly variable and can reach water equivalents between 100 and 800 mm per year. The snowmelt phase usually begins between early March and mid-April, with rain-on-snow events occurring almost every year which can intensify runoff peaks (Rickenmann & McArdell, 2007). During the summer months, convective thunderstorms dominate the hydrological events. These often short but violent events cause sharp increases in runoff. Such events are the main cause of the massive bedload transport in

the Erlenbach (Rickenmann & McArdell, 2007; Schuerch et al., 2006; Stambach, 1988). The highest risk of floods occurs when late snowmelt is combined with heavy rainfall (Stambach, 1988).

The Erlenbach is characterized by its short response time to precipitation events. The time delay between maximum rainfall intensity and peak discharge is usually not long (Schleussinger, 2025; Turowski et al., 2009). For example, a reaction time of only about 10 minutes was documented by Schleussinger (2025) during an intense flood in July 2024.

Table 1. Main characteristics of the Erlenbach catchment according to Rickenmann & McArdell (2007) & Turowski et al. (2009).

Catchment Parameter	Value
Catchment Area (km ²)	0.74
Catchment Elevation Range (m a.s.l.)	1110-1655
Forest Cover (%)	40
Wetland and Grassland Cover (%)	60
Unvegetated Land Cover (%)	<<1
Mean Annual Runoff (mm)	1770
Mean Annual Precipitation (mm)	2300
Mean Annual Temperature (°C)	4.5

3.3 Geology

The upper Alptal valley is geologically divided into two parts. The western flank of the valley consists of Wägitaler flysch from Cretaceous deposits, the layers of which run parallel to the slope. This makes them fairly stable compared to the eastern flank. The eastern flank of the valley, where the Erlenbach is located, consists of Wägitaler flysch from Tertiary deposits and overlying Schlieren flysch. This flank is more unstable due to the high proportion of soft rocks (clay, marl and slate) (Stambach, 1988).

What is special about the flysch in the Erlenbach catchment is the alternating layers of different rock types: On the one hand, there are "hard" layers, which make up about 30% of the rock. These consist of weathering-resistant flysch rocks (sandstones and conglomerates) and tend to be brittle when subjected to pressure. On the other hand, there are "soft" layers, making up the remaining 70%. These consist of clays, marls and shales. These rocks deform under pressure and have very low resistance to weathering. The matrix of loose material in the Erlenbach therefore consists mainly of a mixture of clay, silt and fine sand, in which larger stones and blocks are embedded. As a result, a large proportion of the loose material can be transported in suspension in the water and therefore does not require high flow velocities to be mobilized (Stambach, 1988).

3.4 Morphology and Sediment Processes

The main channel of the Erlenbach is characterised by a step-pool morphology (Figure 3). Steps are usually formed by the deposition of particularly large stones, known as keystone. These act as anchor points behind which smaller material accumulates. Stability is increased by the "jammed state". Here the stones form chain-like structures across the channel, which wedge themselves together and can only be mobilized at extremely high flows (Molnar et al., 2008; Turowski et al., 2009). According to Turowski et al. (2013) these structures tend to break up at discharge rates between 7 and 10 m³/s.

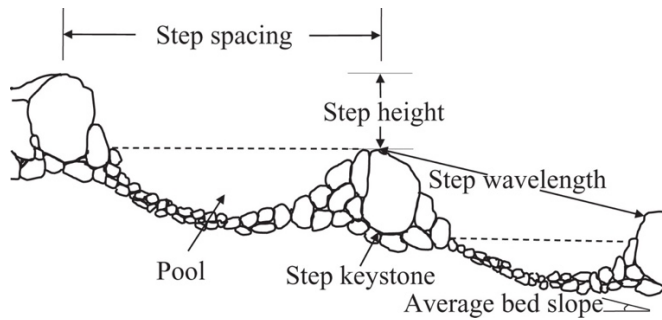


Figure 3. Longitudinal section of a step-pool sequence (Kalathil & Chandra, 2019).

Sediment transport in the Erlenbach is not linearly related to discharge, rather it is controlled, for example, by the availability of loose rock (Rickenmann, 2024; Rickenmann et al., 2012). Schuerch et al. (2006) showed that sediment input to the Erlenbach varies strongly along the channel and is dominated by a few landslides, which causes sediment supply to fluctuate by up to an order of magnitude within a year. The catchment is therefore considered supply-limited, meaning that transport capacity often exceeds the available sediment (Rickenmann et al., 2012).

Another characteristic feature is the time lag between runoff and transport, which usually occurs as clockwise hysteresis. Significantly more sediment is moved during the rising limb of a flood than during the falling limb with the same runoff, as the easily mobilized material on the bed is quickly carried away at the beginning. This effect can only be reduced during very intense events, for example when new material is suddenly made available due to bank erosion (Rickenmann, 2024).

3.5 Monitoring infrastructure

The Erlenbach catchment is an internationally important research site where stream processes have been recorded for decades. Hydrological monitoring is based on two stations: The upper station is located at the transition from the natural stream bed to the artificial channel and is calibrated specifically for small to medium discharges. The lower station at the outlet of the retention basin uses a double triangular profile to provide reliable data even during flood events (Rickenmann, 2020, 2024).

The Swiss Plate Geophones system forms the heart of the bedload monitoring system, with six steel plates installed in the bed, under which sensors register the vibrations of particles rolling over them (Rickenmann, 2024; Rickenmann et al., 2014). All bedload is then collected in a retention basin with a capacity of around 2000 m³ (Figure 4 & 5), the sediment volume of which is measured regularly to determine the long-term bedload transport. In addition, three automatic moving basket samplers enable direct physical sampling during a flood (Rickenmann et al., 2012).

Since very fine particles such as clay and silt are often washed through the retention basin (Molnar et al., 2008; Rickenmann & McArde, 2007), suspended matter monitoring is carried out at the upper measuring station using an automatic sampler that continuously takes water samples for concentration measurement. These samples are also used for chemical water analysis and to determine electrical conductivity (Stammach, 1988).

The system is complemented by meteorological measurements at a station about 550 m above the basin, which records parameters such as precipitation, air temperature, humidity, wind speed and direction, as well as global radiation (Rickenmann et al., 2012; Stähli et al., 2021).



Figure 4. Filled retention basin at the Erlenbach outlet (Source: own photograph, 09.09.25).



Figure 5. Recently emptied retention basin at the Erlenbach outlet (Source: own photograph, 10.11.25).

4 Data & Methods

4.1 Erlenbach Monitoring Station

To quantitatively estimate long-term soil erosion in the Erlenbach catchment, data sets from the Erlenbach monitoring station were used. The basic methodological idea here is based on the assumption that the sediment measured at the outlet of the catchment reflects the total soil erosion of the catchment, as long as it is integrated over a sufficiently long period of several decades.

This assumption is justified for the Erlenbach, as the catchment is characterized by steep topography and frequent flooding, with an average of over 20 sediment-transporting events per year (Rickenmann & McArdell, 2007). Such conditions lead to high transport capacity and frequent rearrangements of the riverbed (Molnar et al., 2008; Turowski et al., 2009). In addition, so-called log jams are regularly broken up during flood events, which leads to the mobilization of previously stored sediments (Jochner et al., 2015). It is therefore assumed that erosion and sediment export in the Erlenbach have roughly balanced each other out over decades and that permanent deposition within the system is negligible.

To estimate sediment export, two complementary data sets were used: bedload and suspended sediment. Bedload data were provided by the "Eidgenössische Forschungsanstalt für Wald, Schnee und Landschaft (WSL)", while suspended sediment data were obtained from the "Nationale Daueruntersuchung der Schweizer Fließgewässer (NADUF)". Bedload transport primarily represents the transport of coarse sediment fractions and includes material from slope and surface erosion as well as from the riverbed and banks. Suspended sediment transport, on the other hand, mainly captures fine-grained material (clay, silt and fine sand) that is directly related to surface soil erosion processes and is supported by turbulent forces in the water (Badoux et al., 2014; Rickenmann, 2020; Stammach, 1988; Turowski et al., 2009).

4.1.1 Bedload Data

4.1.1.1 Data Basis

In 1982, a sediment retention basin with a capacity of 2000 m³ was constructed. This retention basin enables the collection of the entire sediment load leaving the catchment (Rickenmann et al., 2012). The bedload data set is based on repeated measurements of the sediments deposited in the retention basin. Each measurement contains the cumulative sediment volume since the previous survey.

The data set used in this study covers the period from June 1983 to July 2025. Surveys of the retention basin are carried out at least once a year. Additional surveys are also carried out after major flood events. The intervals between the measurements are therefore irregular (Rickenmann et al., 2012). For this study period, 73 individual measurements are available, which corresponds to an average measurement frequency of slightly less than two per year.

Studies of grain size distribution show that about 50-60% of the deposited material contains particles with diameters greater than 10 mm and about 30% consists of particles with diameters greater than 30 mm (Rickenmann et al., 2012). These proportions confirm that a significant amount of the measured sediment load is attributable to bedload transport. The grain size distribution varies between individual flood events, but remains relatively stable over the long-term average (Rickenmann & McArdell, 2007).

4.1.1.2 Data Processing

Due to the relatively small dataset, the time series was visually inspected for inconsistencies, outliers and measurement errors, which revealed some notable features. Between September 25, 1997 and April 23, 1998, the measured sediment volume remained constant at 7904 m³, indicating a remarkably long phase with little or no bedload transport. This is plausible for the Erlenbach, where transport is strongly event-driven and longer inactive periods can occur (Rickenmann, 2024; Rickenmann & McArdell, 2007). Due to the long intervals between measurements, comparable phases could also have occurred in other years but were not explicitly recorded due to the timing of the measurements.

Furthermore, there are no measurements for 2017. It cannot be clearly determined whether no measurements were taken that year or if there was in fact no significant sediment accumulation. In the short term, this can influence the annual balance, as the following measurement may integrate transport over several years. However, this uncertainty is of minor importance when considering long-term trends. Also, the temporal assignment of individual events is only possible to a limited extent anyway due to the large measurement intervals.

Particularly striking is the very large increase in volume in the summer of 1984, when the cumulative volume rose from 272.8 m³ to 2401 m³ and further to 2586 m³ within six days. This jump represents the largest increase in the entire time series and is also documented in the literature (Rickenmann & McArdell, 2007; Turowski et al., 2009). It is therefore not interpreted as a measurement error, but as a consequence of exceptionally intense flood events.

The sediment volumes were converted into masses to ultimately get a uniform unit for soil erosion [t ha⁻¹ yr⁻¹] that can be compared with other soil erosion values. This was done using an average bulk density of 1.75 t/m³, which corresponds to typical densities of alpine bedload sediments and is also used in earlier studies on the Erlenbach (e.g. Rickenmann & McArdell, 2007). Until 2006, the volume of the retention basin was mainly determined by rod measurements taken from a rubber boat in a 1-

meter grid, which has an average uncertainty of approximately 2.5 m³ for the calculated volume difference between two measurement points. Only sediment volumes of more than 50 m³ are considered reliable (relative error ≤ 5%). Since 2006, tachymetric methods and terrestrial laser scanning have been used, which has significantly increased measurement accuracy. Comparative measurements showed deviations of only 0.5 m³ for the whole area of 500 m² (Rickenmann et al., 2012).

Since the time intervals between the individual measurements vary greatly, the mass transported was standardized to an annual rate:

$$\text{Transport [t yr}^{-1}] = \frac{\Delta\text{Mass [t]}}{\Delta\text{Time [yr]}}$$

Here, ΔMass refers to the sediment mass accumulated between two measurements and ΔTime refers to the corresponding time interval in years. For comparability, these values were then related to the catchment area of 74 ha:

$$\text{Transport [t ha}^{-1} \text{ yr}^{-1}] = \frac{\text{Transport [t yr}^{-1}]}{74 \text{ ha}}$$

Annual values were aggregated based on hydrological years defined from October 1 to September 30. For example, the hydrological year of 2003 covers the period from October 1, 2002, to September 30, 2003. Hydrological years are preferred in alpine catchments because hydrological processes follow a seasonal pattern.

4.1.2 Suspended Sediment Data

4.1.2.1 Data Basis

The suspended sediment data set describes the transport of fine sediments. Suspended matter monitoring is carried out at the upper measuring station shortly before the sediments reach the retention basin. The time series used in this study covers the period from January 6, 2003, to December 30, 2024. Since the measurement intervals have a weekly temporal resolution, the data set contains over 1'000 entries. In addition to sediment transport, the complete data set contains a wide range of hydrological information, such as electrical conductivity, oxygen, pH, nutrient and major ions (e.g. calcium, magnesium and sulphate) concentrations. For this study, the following variables were used:

1. End date and time of the measurement interval
2. Duration of the interval
3. Total volume of water discharged during the interval
4. Average suspended matter concentration during the interval
5. Average discharge during the interval

4.1.2.2 Data Processing

The suspended sediment data set was first checked for completeness and consistency. Subsequently, all entries with missing or undefined values were removed. No obvious outliers could be identified in the remaining data set. Although there are a few cases of very high suspended matter concentrations,

such as a value of 4340 mg/L in July 2007, these can be reasonably explained by heavy precipitation and flooding events and are therefore interpreted as realistic extreme events.

The duration of the individual measurement intervals also show a certain degree of variability. While most intervals correspond to one week, there are a few shorter intervals of less than five days and longer intervals of more than nine days, with one interval of 18 days standing out in particular. However, this variability does not pose a methodological problem, as all calculations are based on the respective interval length and the sediment transport is then normalized to an annual rate. This ensures that the measurements remain comparable over time, regardless of the length of the measurement intervals.

The suspended matter mass for each measurement interval was calculated from the mean suspended matter concentration C [mg/L] and the total water volume V [m³] that flowed during the interval using the following formula:

$$\text{Mass [t]} = \frac{C \cdot V \cdot 1000}{10^9} = C \cdot V \cdot 10^{-6}$$

The subsequent standardization to annual transport rates and the normalization to a catchment area of 74 ha were carried out in the same way as for the bedload data set. As with the bedload data set, the hydrological year was also used for aggregation to annual values. This not only enables seasonal processes to be represented in a consistent hydrological cycle but also improves the comparability between bedload and suspended sediment transport.

4.1.3 Hourly Data Temperature, Precipitation and Runoff

To characterize and describe the catchment, as well as to investigate correlations between runoff, precipitation and sediment transport, an hourly data set on temperature, precipitation and discharge, provided by the WSL, was used. The meteorological measurements were taken at the Erlenhöhe station, located 550 m above the retention basin within the catchment. Runoff data were collected at the monitoring station at the catchment outlet. The time series covers the period from January 1, 1983, to December 14, 2025.

The ventilated air temperature was used to determine the average monthly temperature, as the non-ventilated temperature only has data up to March 2020. This means that the latest temperature developments would not be covered. In addition, ventilated measurements generally provide more reliable air temperature values by reducing radiation-induced measurement errors, particularly under daylight and low-wind conditions (Ságová et al., 2025). The ventilated air temperature also contains missing values in the time series, which were removed. These gaps occurred in 1983, 1984, 1988 and 1990. However, due to the long overall data series, such gaps have only a minimal impact on the representation of long-term monthly averages.

To calculate the long-term average of the monthly runoff, hourly discharge values Q [L s⁻¹] were converted to daily runoff [mm d⁻¹] using the catchment area A [m²] according to the following equation:

$$\text{runoff [mm d}^{-1}\text{]} = \frac{Q \cdot 86'400}{A}$$

Daily precipitation totals [mm d^{-1}] were calculated by summing the hourly precipitation values [mm] for each day.

Since the bedload data set does not contain any information about the discharge, the discharge values Q at the catchment outlet were also used for comparison. For this purpose, the mean discharge was calculated for each measurement interval of the bedload data and assigned to the corresponding bedload transport rates.

4.2 Plutonium-Based Approach

In addition to the long-term hydrological measurement data from the Erlenbach Monitoring Station, this study also uses Pu isotopes to estimate soil erosion in the catchment. The basic methodological idea behind this approach is that plutonium deposited from the atmosphere (mainly from nuclear weapons tests in the 1950s and 60s) binds firmly to soil particles (S. Dowell et al., 2024). During erosion processes, this Pu-containing soil material is mobilized and transported out of the catchment through the network of waterways. The Pu isotopes measured in the sediment at the catchment outlet thus reflect the long-term soil erosion of the entire catchment, as the transported sediment represents a mixture of material originating from different parts of the whole catchment (Alewell et al., 2013; Huang et al., 2025; Mabit et al., 2008; Percich et al., 2022).

Three different types of samples are combined for the quantitative determination of soil erosion:

1. Sediment samples from the catchment outlet, which reflect the long-term integrated erosion of the entire catchment.
2. Soil samples from erosional sites, which provide specific information on the spatial variability of erosion and serve as a comparison to the sediment samples.
3. Reference soil samples from areas free of erosion and accumulation, which represent the original atmospheric Pu inventory in the Erlenbach catchment.

Erosion rates are derived using two complementary approaches based on Pu inventories. For soil samples, erosion rates are determined by comparing the Pu inventory at each sampling site with that of the reference sites. A reduced inventory relative to the reference indicates soil loss, whereas higher values reflect deposition. For river sediment samples, erosion rates are estimated using a steady-state mass balance approach. In this case, the measured Pu activity in the sediments is related to a long-term atmospheric input, allowing the calculation of catchment-scale erosion rates under the assumption that Pu input and erosion-related removal are in equilibrium over long time periods. Together, these approaches provide complementary information, with soil samples capturing local erosion patterns and river sediments reflecting an integrated, catchment-wide signal that can be compared with long-term sediment export derived from monitoring data.

4.2.1 Sampling Strategy

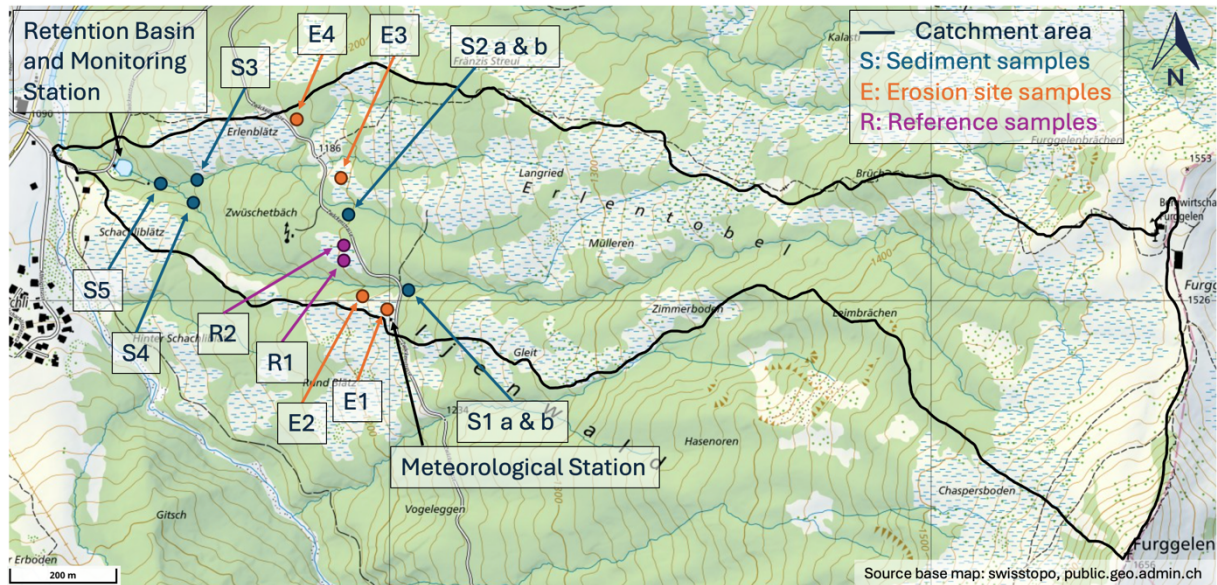


Figure 6. Map of the Erlenbach catchment showing all sampling locations.

Sampling took place over two field days. Soil samples from the erosion sites (E1-E4), the reference soil samples (R1) and the first sediment samples (S1a and S2a) were taken in September 2025. The other sediment samples (S1b, S2b, S3, S4 and S5) were collected on a second field day around two months later, in November 2025. The second reference site (R2) was sampled earlier in the year, in May 2025.

Sediment Samples:

Altitudes of the sediment sampling sites:

- S1: 1213 m a.s.l.
- S2: 1198 m a.s.l.
- S3: 1128 m a.s.l.
- S4: 1126 m a.s.l.
- S5: 1114 m a.s.l.

Sediment samples were taken at a total of five different locations within the catchment area (Figure 6). The samples were taken directly from the riverbed, preferably in shallow water near the banks or directly on the banks. The material taken came from approximately the top 0-5 cm of the sediment layer. Two to five individual samples were taken per location at suitable points a few meters apart. This procedure was used to capture the spatial variability within a location and to increase the representativeness of the samples. During sampling, care was taken to select material with a fine grain size and to avoid larger stones and coarse material.

The sediment samples were taken as close as possible to the outlet of the catchment to ensure that the measured Pu inventories represent the largest possible proportion of the catchment. This is crucial to be able to estimate soil erosion across the entire catchment and not just make statements about individual sub-catchment areas. At the same time, the sampling sites had to be easily accessible, which was mainly the case near the road and influenced the selection of the exact locations.

Most of the sediment samples were taken from the main channel (left stream, viewed downstream; locations S1 and S4; Figure 7b), as this one is hydrologically much more active and carries the majority of the runoff and sediment transport. In addition, samples were taken from a very small side stream with highly loamy sediment and very low runoff (location S2; Figure 7d) to cover this area of the catchment as well. Further samples were taken from the right stream (viewed downstream; location S3; Figure 7a). In addition, samples were taken at the point where the two main streams converge, as close as possible to the catchment outlet (location S5; Figure 7c). This location is intended to capture the most complete integrated signal possible for the entire catchment. It was not possible to take a sample directly at the outlet because various measuring equipment is located there and parts of the stream bed are cemented for the measurements, which prevented sediment sampling.

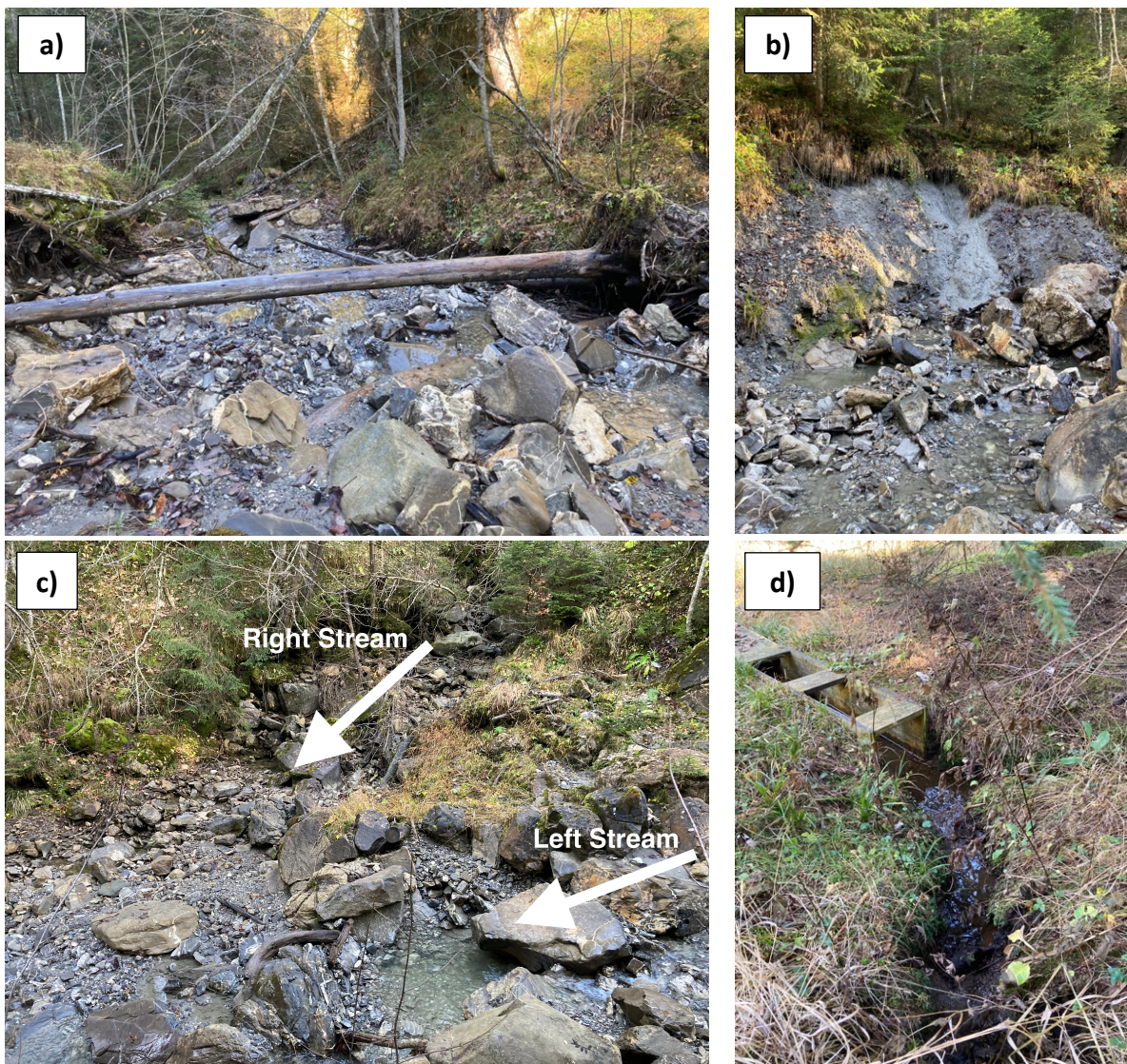


Figure 7. Locations of the sampling sites: (a) right stream; (b) left stream; (c) Confluence point of the right and left stream; (d) small side stream (Source: own photograph).

Soil Samples from Erosion Sites:

To investigate localized soil erosion, soil samples were taken at four locations (E1-E4) where, due to the steep slope, it was assumed that erosion processes would be more pronounced. The sampling procedure was uniform at all sites and followed the same pattern: At each site, four soil profiles were

excavated, each a few meters apart. Each profile was exposed to a depth of at least 30 cm. Five soil samples were then taken from each profile using a soil corer with a fixed volume of 100 cm³. One sample from each of the following depth intervals was taken:

- 0-5 cm
- 5-10 cm
- 10-15 cm
- 15-20 cm
- 25-30 cm

The depth range 20-25 cm was intentionally omitted because the Pu inventory decreases with increasing depth. However, to ensure that there was no deep soil mixing, an additional sample was taken from a depth of 25-30 cm. This approach allows reliable recording of the vertical Pu profile while limiting the number of samples. This resulted in a total of 80 individual soil samples from the four erosion sites (4 sites × 4 profiles × 5 depth intervals).

The soil was very moist and loamy at all sites. The fixed volume of the soil corer allowed the soil density of each sample to be calculated after drying. In addition, the water content was determined by weighing the samples before and after drying.

Table 2. Characteristics of the different erosion sites.

Location	Elevation (m a.s.l.)	Slope	Aspect	Position in Relief	Water Status	Vegetation	Land Use
E1	1209.7	12°	250°W	Upper slope	Wet slope	Permanent pasture	Pasture / Wet meadow
E2	1206	8°	265°W	Mid-slope	Wet slope, waterlogged	Permanent pasture	Pasture / Wet meadow
E3	1192.7	18°	270°W	Mid-slope	Wet slope, waterlogged	Permanent pasture	Pasture / Wet meadow
E4	1177.2	16°	243°SW	Lower slope	Wet slope	Permanent pasture	Pasture / Wet meadow



Figure 8. Location of erosion site 2 (E2) (Source: own photograph).

Reference Samples:

The reference sites (R1 & R2) are located close to each other at an altitude of 1203 m a.s.l. and have a slope of almost 0°. It can therefore be assumed that no significant erosion or accumulation processes take place at those sites. Furthermore, the sites are not used for agricultural purposes. The soil was also very moist and loamy, comparable to the soils at the erosion sites. The samples obtained at the reference sites were used to determine the original atmospheric Pu inventory and form the basis for comparison with the erosion-affected sites.

A total of five reference profiles were dug up (four in May (R2) and one in September (R1)). The creation of several soil profiles makes it possible to calculate average values and makes the results more robust and reliable, as local differences between individual profiles are balanced out. Five soil samples were taken from each profile at the following depth intervals: 0-5 cm, 5-10 cm, 10-15 cm, 15-20 cm and 20-25 cm.



Figure 9. Location of reference site 1 (R1) (Source: own photograph).

4.2.2 Sample Preparation and Measurement

The preparation of the samples for Pu isotope measurement using mass spectrometry, which is described in this chapter, was carried out according to a protocol from the University of Zurich, which is an adaptation of the method described by Ketterer (2015).

Sample Pretreatment:

All samples were first dried in an oven before being mechanically crushed to a particle size of < 2 mm. This was followed by further grinding of the samples in a horizontal mill for 7 minutes, which produced a fine, homogeneous powder.

Removal of Organic Matter:

Exactly 5 g of each ground sample was placed in a glass vial. Empty vials (blanks), standard soil samples, and selected duplicates were treated identically to ensure quality control throughout the entire

process. To remove organic matter, the samples were ashed at 550 °C for over 16 hours. Immediately after, the samples were weighted again and the mass loss was determined as loss on ignition (LOI).

Chemical Extraction and Addition of Tracer:

Firstly, 8-10 ml of concentrated nitric acid (65% HNO₃) was added. After cooling, a precisely defined amount of the plutonium tracer ²⁴²Pu (≈ 4.4 mBq) was added to each sample. This tracer is used for the subsequent quantitative determination of the chemical yield and enables the absolute calculation of the Pu activities in the samples. The samples were then heated overnight at 80 °C, diluted with ultra-pure water, centrifuged and filtered until a clear solution was achieved.

Conversion of Plutonium and Retention of it on TEVA Resin:

To convert plutonium to the desired oxidation state of Pu(IV), iron(II) sulfate was first added to the solution, followed by sodium nitrite. After the chemical conversion, a suspension of TEVA resin was added to the solution. This resin selectively binds Pu, which allows it to be separated from the other elements in the sample solution. The resin was then filtered and washed with various acid solutions to remove impurities. The bound Pu was then separated from the resin and collected in a small, concentrated solution.

Measurement with QQQ-ICP-MS:

The purified Pu fractions were subsequently measured using triple quadrupole inductively coupled plasma mass spectrometry (QQQ-ICP-MS). This technique enables precise determination of Pu isotope ratios (²⁴⁰Pu/²³⁹Pu) as well as the quantification of isotope activities based on the previously added ²⁴²Pu tracer.

4.2.3 Calculation of Erosion Rates

4.2.3.1 Erosion Rates of Soil Samples

Calculation of Pu Inventories

First, the mass depth, which describes the soil mass per unit area, was calculated for each sampled depth interval. This is derived from the measured soil density and the thickness of the respective depth interval. Coarse material with a grain size greater than 2 mm was excluded, as plutonium is predominantly bound to fine soil particles.

Based on the mass depth and the known surface area of the soil corer, the mass of the fine soil fraction was determined for each depth interval. The measured Pu activity (Bq/kg) was then related to the mass depth and converted into a volume-specific plutonium concentration (mBq/cm³), taking into account the thickness of the depth interval.

The Pu inventory of a single depth interval was determined as the product of the calculated Pu concentration and the corresponding thickness. The cumulative Pu inventories of a soil profile were obtained by adding up the inventories of all depth intervals. The total inventory of a profile corresponds to the cumulative inventory up to the maximum sampled depth of 30 cm.

To consider small-scale variability within the sites, several soil profiles were sampled per site. The resulting Pu inventories of the individual profiles were then averaged and used as representative site

values. These averaged inventories formed the basis for the subsequent calculation of long-term erosion rates using different models.

Reference Inventory

Sites R1 and R2, which were assumed to be free of erosion and accumulation, were used as references. The average Pu inventory was calculated from all sampled profiles at these locations using the same procedure as for the soil samples from the erosion sites. The average reference inventory derived from this represents the original atmospheric Pu inventory of the catchment area. This reference inventory is the central benchmark for quantifying soil erosion, as deviations in the inventories at erosion-affected sites are interpreted relative to this reference value.

Model approaches for calculating erosion rates

Three different models were used to estimate soil erosion rates:

(a) Model according to Lal et al. (2013)

The model developed by Lal et al. (2013) is based on the assumption that the vertical distribution profile of atmospheric Pu in undisturbed soils follows an exponential decrease with depth. The Pu concentration $N(z)$ as a function of depth z can therefore be described as:

$$N(z) = N_0 \cdot e^{-\alpha z}$$

where N_0 is the Pu concentration at the soil surface and α is the decay parameter of the depth profile. The parameter α was determined by exponential fitting to the averaged Pu concentrations of the individual depth intervals of the reference profiles.

The total Pu inventory of a soil profile is calculated as the sum of the Pu in all sampled depth intervals. For a site affected by erosion, the inventory loss I_{loss} is defined as the difference between the mean reference inventory I_{ref} and the measured inventory I_{meas} :

$$I_{\text{loss}} = I_{\text{ref}} - I_{\text{meas}}$$

Assuming an uncultivated site where the exponential depth profile is maintained during erosion, the cumulative soil loss L (in cm) can be calculated from the relative inventory loss. According to Lal et al. (2013), this results in:

$$L = -\frac{1}{\alpha \cdot P} \ln \left(1 - \frac{I_{\text{loss}}}{I_{\text{ref}}} \right)$$

P is a particle size correction factor that takes into account the stronger binding of Pu to fine-grained soil material. In this study, calculations were performed for $P=1.0$, 1.2 and 1.5 , in accordance with the recommendations of Lal et al. (2013) and Walling & He (1999). The calculated soil losses in centimeters were then converted into mass losses using the average soil density and divided by the time span since the main atmospheric plutonium fallout (1963 to the sampling year 2025) to obtain average long-term erosion rates in $\text{t ha}^{-1} \text{yr}^{-1}$.

(b) Model according to Walling & He (1999)

In addition, the profile distribution model according to Walling & He (1999) was applied, which was originally developed for estimating soil erosion at undisturbed sites using ^{137}Cs and was transferred here analogously to $^{239+240}\text{Pu}$. The model represents the vertical tracer distribution by describing the cumulative Pu inventory above a certain mass depth.

The cumulative Pu inventory $A'(x)$ up to mass depth x is described by:

$$A'(x) = A_{ref} \cdot (1 - e^{-x/h_0})$$

where

- $A'(x)$ is the cumulative Pu inventory above the mass depth x (kg/m^2),
- A_{ref} represents the mean reference inventory (Bq/m^2) and
- h_0 is a coefficient for describing the depth trend of Pu (kg/m^2)

The parameter h_0 was determined for the reference sites using a solver function in Excel by minimizing the quadratic deviation between measured and modeled cumulative Pu inventories.

For erosion-affected sites, the percentage reduction in the Pu inventory relative to the reference inventory was first calculated:

$$X = \frac{A_{ref} - A_{meas}}{A_{ref}} \cdot 100$$

Assuming that the main atmospheric input of Pu occurred in 1963 and that the profile shape has remained unchanged since then, the average erosion rate Y ($\text{t ha}^{-1} \text{yr}^{-1}$) can be calculated according to Walling & He (1999) as:

$$Y = \frac{10}{2025 - 1963} \cdot h_0 \cdot \ln \left(1 - \frac{X}{100} \right)$$

(c) Model "MODERN" according to Arata et al. (2016)

The MODERN (Modeling Deposition and Erosion Rates with Radionuclides) numerical model, introduced in 2016 by Arata et al., was also used to estimate erosion and deposition rates. Unlike older models, which often assume a rigid theoretical function for the depth distribution of isotopes, MODERN works iteratively. The model directly compares the measured total inventory of a sampling site with the precise depth profile of the reference site to calculate the depth of soil loss or gain (S. Dowell et al., 2024; Loba et al., 2022; Meusburger et al., 2016). It can be applied to cultivated and uncultivated sites and is suitable for all common fallout radionuclides (Meusburger et al., 2016). It also allows the simulation of various scenarios such as plowing, erosion and deposition, which makes the results more realistic in terms of actual land use (S. Dowell et al., 2024).

First, a reference profile that describes the vertical distribution of atmospheric Pu deposition was created from the measured depth profiles of the reference sites. This reference profile was continued downward in the MODERN model using exponential smoothing.

The individual soil profiles were then modeled separately for each erosion site. The four different profiles were included in the calculations for each site. For each profile, the MODERN model compares the measured total inventory with the inventory of the reference profile and determines the point in

the profile where both inventories match. The location of this intersection point indicates the thickness of the soil mass that has been eroded or additionally deposited since the atmospheric fallout.

The soil losses or gains calculated in this way were then converted into average annual erosion or deposition rates ($\text{t ha}^{-1} \text{yr}^{-1}$), taking into account the mass depth, the sampling depth, and the time span between the fallout maximum (1963) and the sampling year (2025). This procedure was applied consistently to all four erosion sites (E1-E4) so that site-specific differences in the modeled erosion rates could be analyzed.

The R scripts used for the MODERN model calculations at each site (E1-E4) are provided in Appendix A (A1-A4).

4.2.3.2 Erosion Rates of River Sediment Samples

For all sediment samples, the isotope ratio $^{240}\text{Pu}/^{239}\text{Pu}$ was determined. The mean isotope ratio was 0.191, which is very close to the characteristic value of global atmospheric fallout from nuclear weapons tests in the northern hemisphere (~ 0.18) (Alewell et al., 2013). Signatures from later specific sources such as reactor accidents (e.g. Chernobyl or Fukushima) can therefore be ruled out. The Pu detected in the sediments therefore originates predominantly from global fallout.

(d) Model according to Hirose et al. (2008) – Atmospheric input Q_1 (Japan)

To quantify erosion rates, a stationary approach was used, which assumes a long-term equilibrium between atmospheric Pu input and erosion-related removal. A value of $Q_1 = 0.003 \text{ Bq m}^{-2} \text{yr}^{-1}$ was used as the annual atmospheric input, which was specified for Japan by Hirose et al. (2008) and is based on measurements from the 2000s. This value describes a largely constant Pu deposition that has been observed since the late 1980s and is mainly due to the resuspension of already deposited fallout material from arid regions of East Asia.

The erosion rate was calculated for each sediment sample using the following relationship according to Padilla et al. (2018):

$$E = \frac{Q}{\rho \cdot N_{\text{Surf}}}$$

where E is the erosion rate (m yr^{-1}), Q is the atmospheric $^{239+240}\text{Pu}$ flux ($\text{Bq m}^{-2} \text{yr}^{-1}$), ρ is the sediment density (t m^{-3}) and N_{Surf} is the measured $^{239+240}\text{Pu}$ activity of the sediment (Bq t^{-1}). A sediment density of 1.3 t m^{-3} was used for all calculations. This equation is based on the steady-state mass balance approach proposed by Padilla et al. (2018), which was originally developed for meteoric ^{10}Be but can be applied analogously to $^{239+240}\text{Pu}$. It assumes that atmospheric radionuclide input and erosion-induced removal are in equilibrium in the long term.

The calculated erosion rates were first expressed in mm yr^{-1} and then converted into area-related erosion rates ($\text{t ha}^{-1} \text{yr}^{-1}$) using the sediment density.

(e) Model according to Rosner & Winkler (2001) – Atmospheric input Q_2 (Germany)

Analogous to model (d), a second calculation approach was used to determine the long-term erosion rates, which differs only in terms of the atmospheric Pu input used. An annual input of $Q_2 = 0.0021 \text{ Bq m}^{-2} \text{ yr}^{-1}$ was used, which was reported for southern Germany from long-term measurements taken between 1986 and 1998 (Rosner & Winkler 2001). This value is below the Japanese input (0.003) and reflects regional differences in the current atmospheric distribution of fallout plutonium.

5 Results

5.1 Hydrology and Climate

The average monthly temperature shows a pronounced annual cycle, with the lowest monthly average temperatures in January (-1.6°C) and February (-1.2°C), and the highest in July (14.2°C) and August (14.1°C) (Figure 10a). Precipitation is higher in the summer months than the winter and autumn months. The average monthly precipitation ranges from about 4.9 mm d^{-1} (October) to 8.2 mm d^{-1} (July) (Figure 10b). The average monthly runoff is highest in spring and lowest in winter. It varies between 2.4 mm d^{-1} in February to 8.2 mm d^{-1} in May. Runoff variability is highest in late spring, particularly in May and June, and lowest during the winter months, such as January and February (Figure 10c).

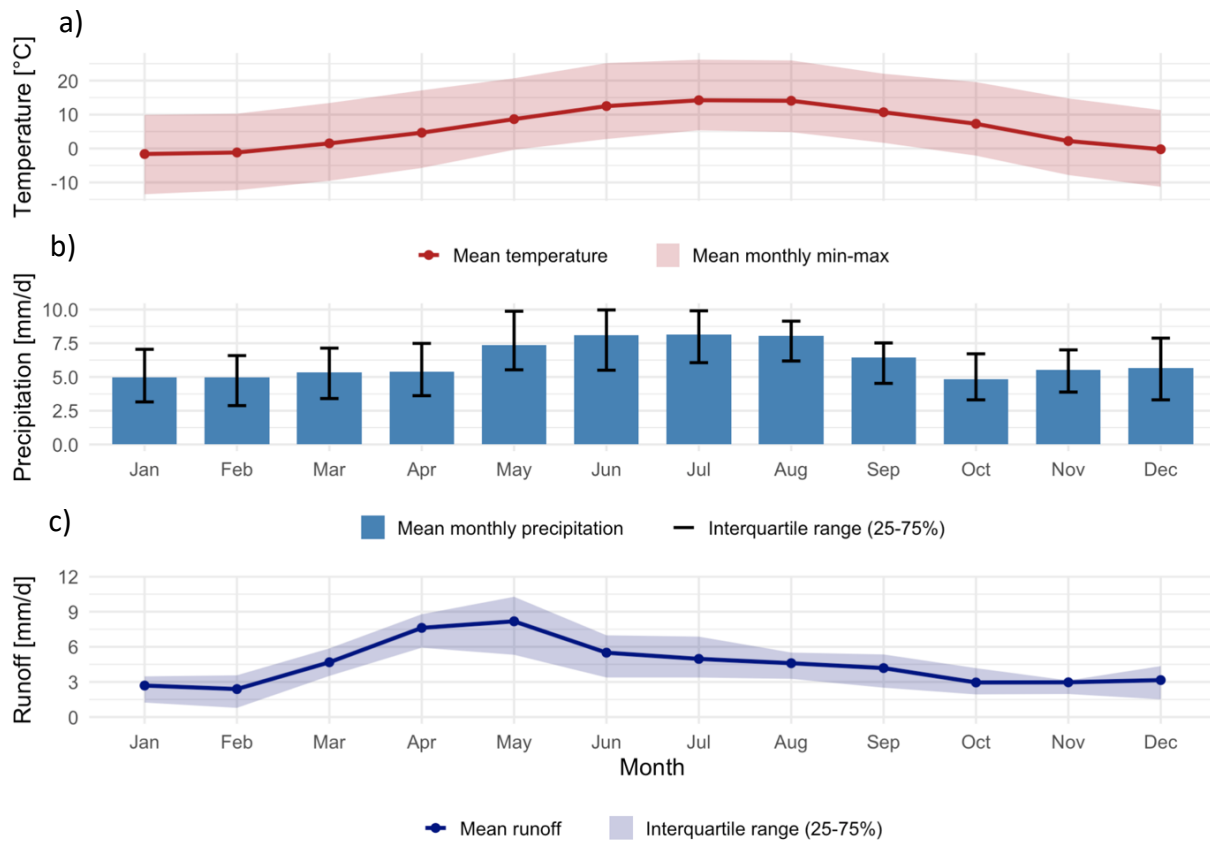


Figure 10. Mean monthly (a) temperature, (b) precipitation and (c) runoff in the Erlenbach catchment (1983-2025).

5.2 Sediment Transport

5.2.1 Bedload Transport

Figure 11 shows the temporal development of bedload transport at the outlet of the catchment. Since bedload transport was not measured continuously, but only at specific points in time, the representation is step-like. Bedload transport is highly variable over time and dominated by a few extreme events. By far the largest event occurred in the summer of 1984. Between 4th of July and the 7th of August, bedload transport reached its maximum value for the entire time series of $540.3 \text{ t ha}^{-1} \text{ yr}^{-1}$, followed by $266.1 \text{ t ha}^{-1} \text{ yr}^{-1}$ in the next six days. Although transport remained high in the following decades, it was less than in 1984. Notable examples include 1993 with $79.7 \text{ t ha}^{-1} \text{ yr}^{-1}$, 2007 with $57.9 \text{ t ha}^{-1} \text{ yr}^{-1}$ and 2010 with $41.7 \text{ t ha}^{-1} \text{ yr}^{-1}$. For most measurement intervals, however, bedload transport was less than $10 \text{ t ha}^{-1} \text{ yr}^{-1}$ and in many cases less than $5 \text{ t ha}^{-1} \text{ yr}^{-1}$.

The cumulative sediment mass increased accordingly. By October 1984, around 4'940 t had already accumulated, meaning that around 18% of the total mass of 28'210 t transported between 1984 and 2025 had already been transported within a few months. Further significant increases in the cumulative curve occurred in 2007 (increase from approx. 16'040 t to 19'392 t) and 2010 (increase from approx. 20'794 t to 23'695 t; Figure 11). After that, the cumulative mass increased much more slowly.

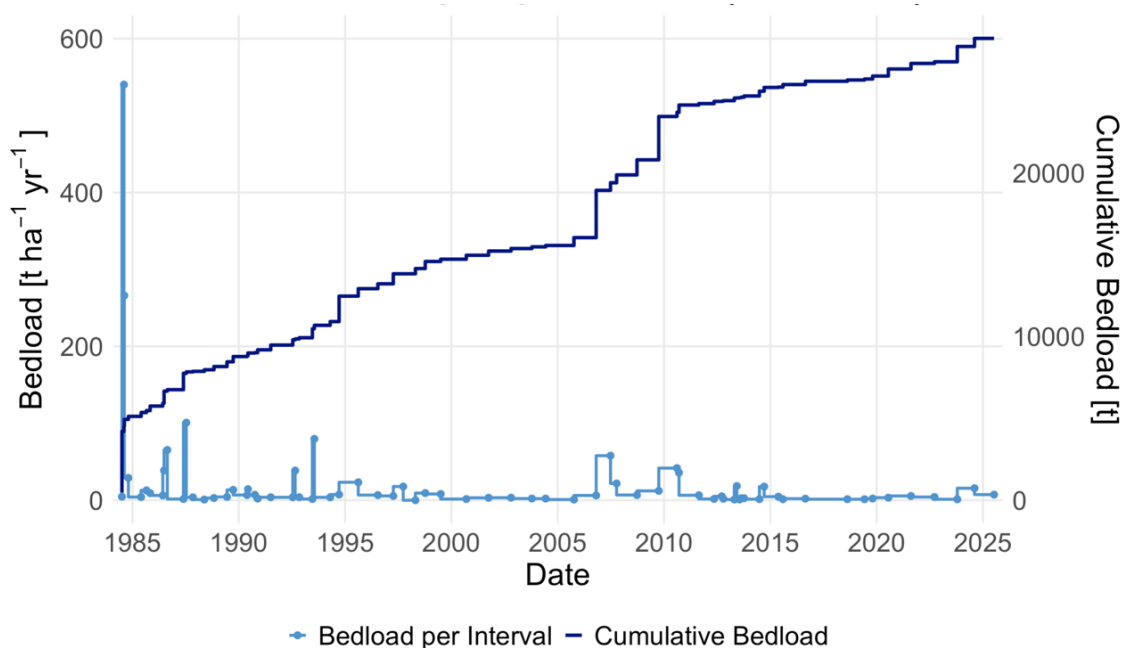


Figure 11. Bedload transport for each measurement interval and the cumulative amount of bedload transport (1984-2025).

The pattern of many low bedload transport periods and few extreme events, is also reflected in the frequency distribution of bedload transport (Figure 12a). The median bedload transport is $4.83 \text{ t ha}^{-1} \text{ yr}^{-1}$, which means that more than half of the measurement periods bedload transport was less than $5 \text{ t ha}^{-1} \text{ yr}^{-1}$. However, due to the high extreme values, particularly from 1984, the mean value is $22.94 \text{ t ha}^{-1} \text{ yr}^{-1}$. The high standard deviation of $71.18 \text{ t ha}^{-1} \text{ yr}^{-1}$ and the positive skew (skew: 5.9) clearly highlight the asymmetry of the distribution as well. For the period since 2003, without the extreme values from 1984, the median is only $3.29 \text{ t ha}^{-1} \text{ yr}^{-1}$, the mean is $8.99 \text{ t ha}^{-1} \text{ yr}^{-1}$ and the standard deviation is at $13.18 \text{ t ha}^{-1} \text{ yr}^{-1}$ (Figure 12b). Again, there is a right-skewed distribution (skew: 2.23).

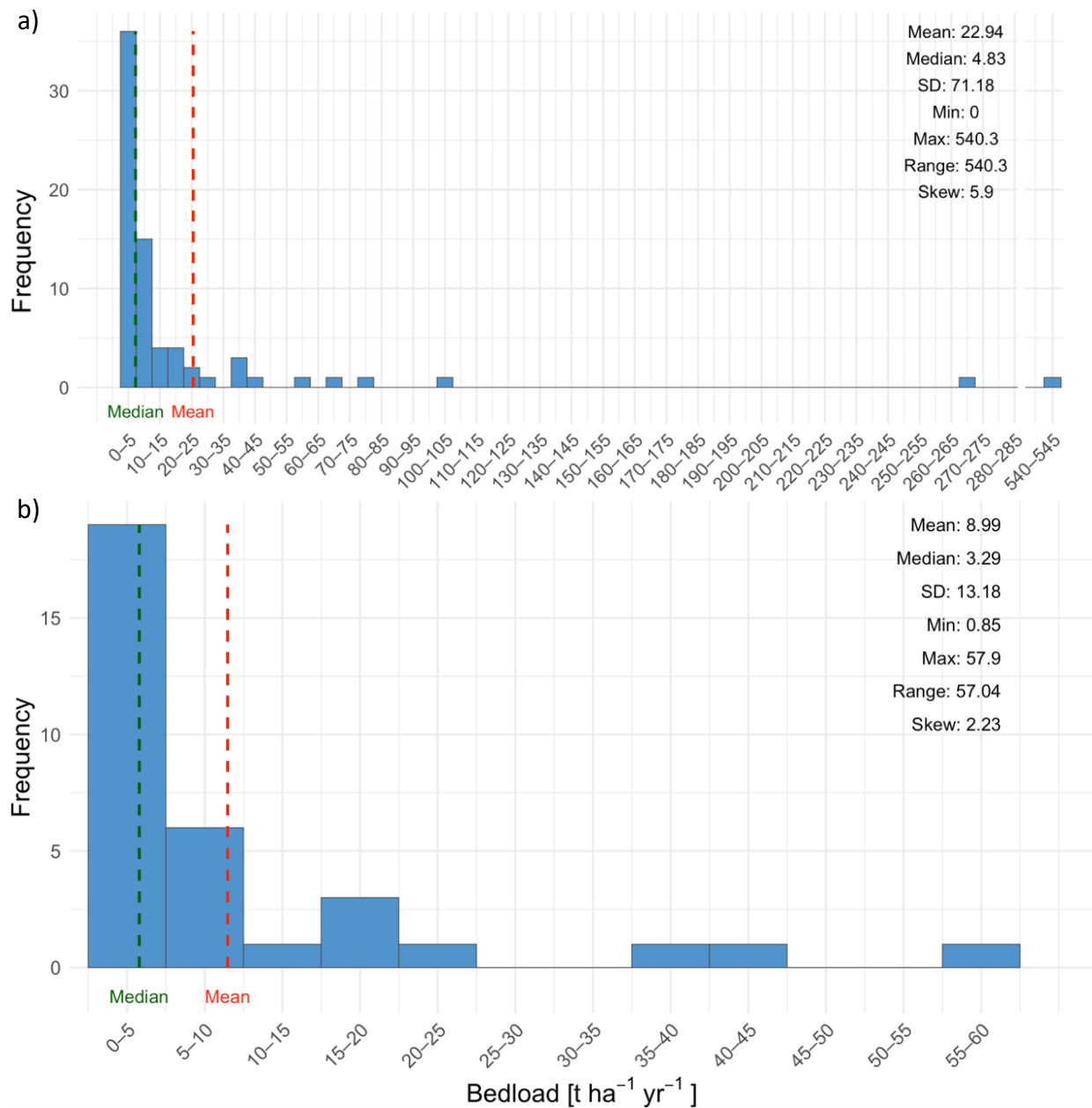


Figure 12. Frequency distribution of bedload transport for two periods (a) 1984-2025; (b) 2003-2025. Note the x axis break in figure a (Xu et al., 2021).

Bedload transport increases with increasing mean discharge (Figure 13) but mean discharge explains only 27% of the variation in bedload transport. The scatter indicates that similar discharges can be associated with different transport intensities. Most measurement points are characterised by a low discharge (approx. $0.03\text{--}0.06 \text{ m}^3 \text{ s}^{-1}$) and low bedload transport rates. However, individual intervals have high transport, e.g., the earlier mentioned August 1984 period with a bedload transport of $540.3 \text{ t ha}^{-1} \text{ yr}^{-1}$ and average discharge of $0.04 \text{ m}^3 \text{ s}^{-1}$.

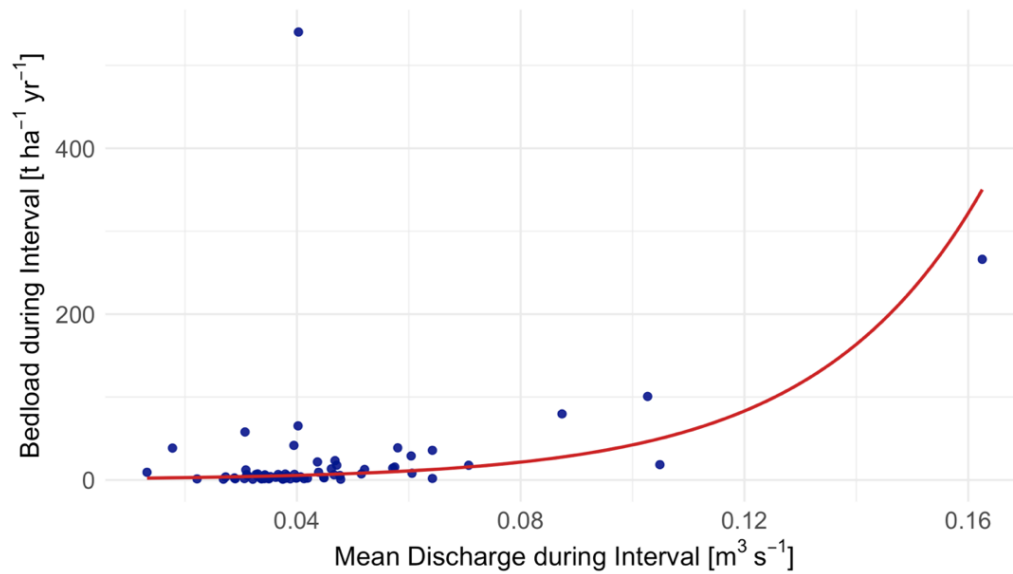


Figure 13. Scatterplot showing bedload per measurement interval versus mean discharge during the interval (blue points), and the exponential fit (red), derived from a linear regression on the log-transformed bedload data (Adjusted R^2 : 0.27; p -value: $2.498e^{-6}$). Each point represents one bedload measurement (1984-2025).

When determining the bedload transport per hydrological year, 1984 ($59.6 \text{ t ha}^{-1} \text{ yr}^{-1}$), 2007 ($45.5 \text{ t ha}^{-1} \text{ yr}^{-1}$) and 2010 ($51.7 \text{ t ha}^{-1} \text{ yr}^{-1}$) stand out as high bedload transport years (Figure 14). In contrast, several years have very low bedload transport, including 1988 ($1.6 \text{ t ha}^{-1} \text{ yr}^{-1}$), 2000 ($1.8 \text{ t ha}^{-1} \text{ yr}^{-1}$), 2006 ($1.1 \text{ t ha}^{-1} \text{ yr}^{-1}$) and 2019 ($1.0 \text{ t ha}^{-1} \text{ yr}^{-1}$).

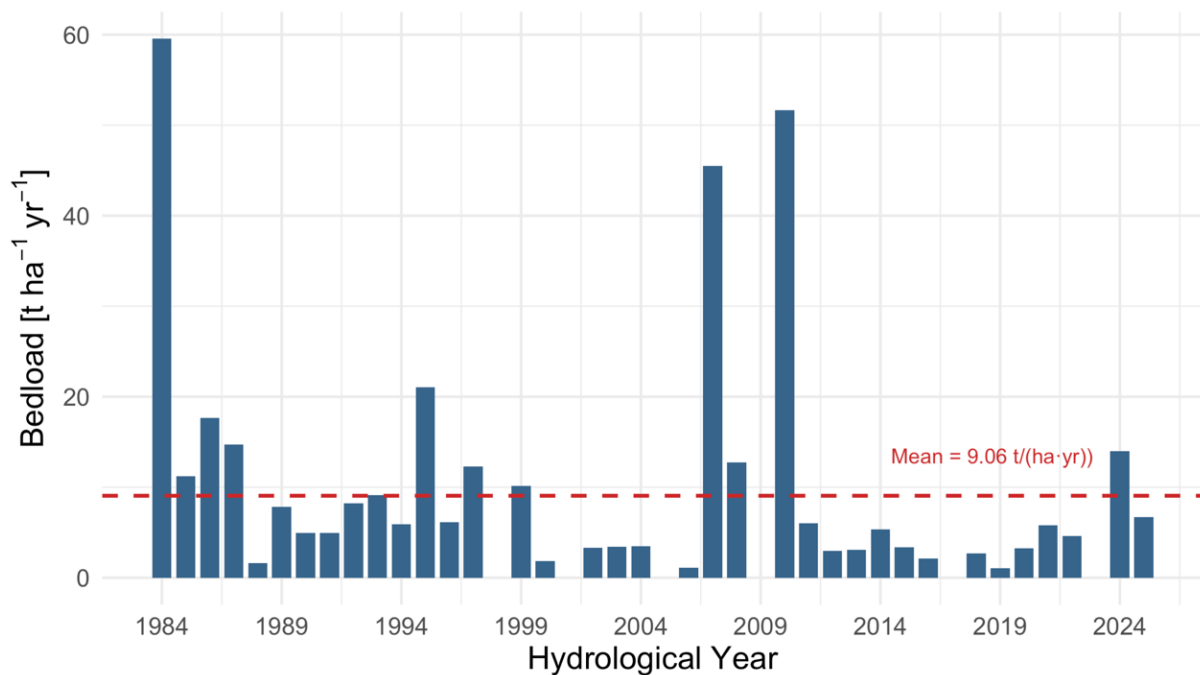


Figure 14. Total bedload transport per hydrological year (October-September) and long-term mean (dashed line: $9.06 \text{ t ha}^{-1} \text{ yr}^{-1}$) for the period 1984-2025. No values are shown for some hydrological years (2001, 2005, 2009, 2017 and 2023) because no regular measurements were taken during these periods and there was more than one year between two surveys. For the hydrological year 1998, however, there was a measurement, but it resulted in a bedload transport of $0 \text{ t ha}^{-1} \text{ yr}^{-1}$. For the hydrological year of 2025 the data are not complete and only cover the measurements until July 2025 (i.e., the last two months are missing).

5.2.2 Suspended Sediment Transport

Most of the weekly suspended sediment transport values are below $5 \text{ t ha}^{-1} \text{ yr}^{-1}$ (Figure 15 & 16). Over 800 measured values (71.4% of all measurements) are in the range between 0 and $1 \text{ t ha}^{-1} \text{ yr}^{-1}$. However, there were some extreme events that reached very high peaks (up to almost $70 \text{ t ha}^{-1} \text{ yr}^{-1}$). Until 2005, only very low values of suspended sediment transport were recorded. Between 2005 and summer 2014, there were several extreme events with a one-week total suspended sediment transport of more than $20 \text{ t ha}^{-1} \text{ yr}^{-1}$. Towards the end of the time series, there is another phase of particularly high suspended sediment transport. In 2024, there were several periods with weekly suspended sediment transport values of over 55 to almost $70 \text{ t ha}^{-1} \text{ yr}^{-1}$, which caused the recent sharp increase in cumulative suspended sediment load. In December 2024, the cumulative curve eventually reached a value of 2862t.

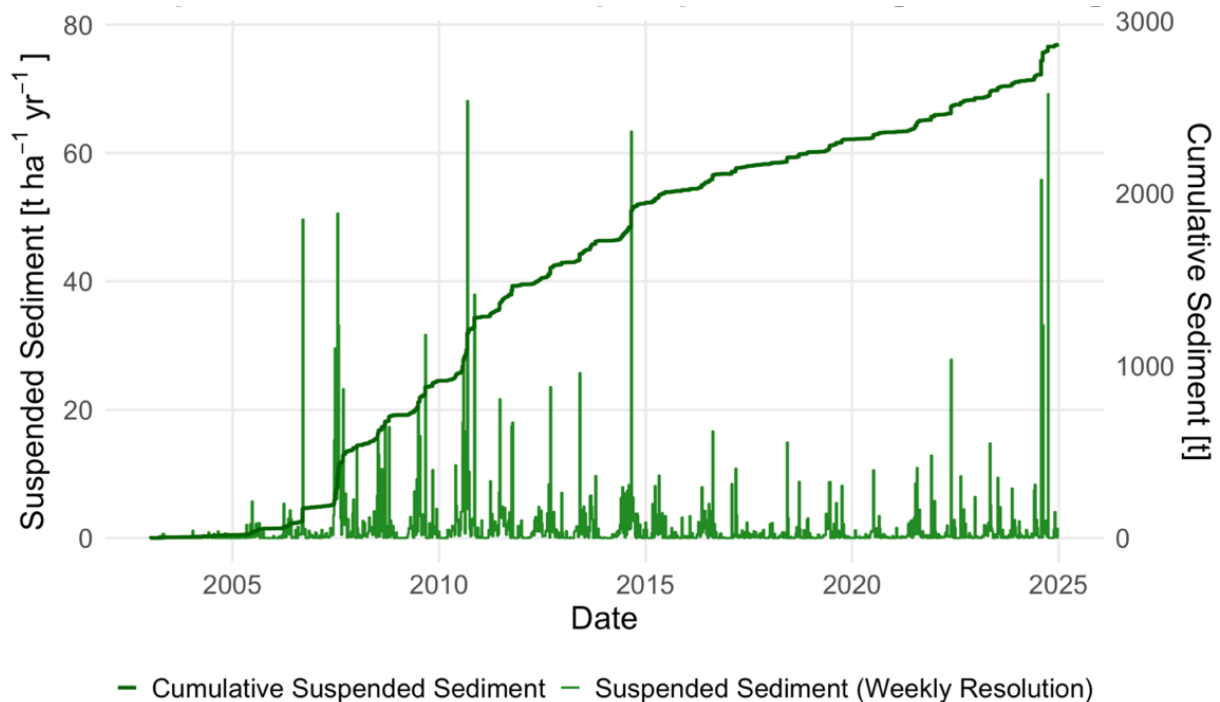


Figure 15. Time series of the weekly suspended sediment transport and cumulative load from 2003 to 2024.

The predominantly low suspended sediment transport values and rare high values result in a strongly right-skewed distribution (skew = 7.22; Figure 16). The median suspended sediment transport is $0.27 \text{ t ha}^{-1} \text{ yr}^{-1}$ and is below the mean value of $1.83 \text{ t ha}^{-1} \text{ yr}^{-1}$. Although the distribution is similar to that for bedload transport (see Figure 12), bedload transport is on average larger than suspended sediment transport.

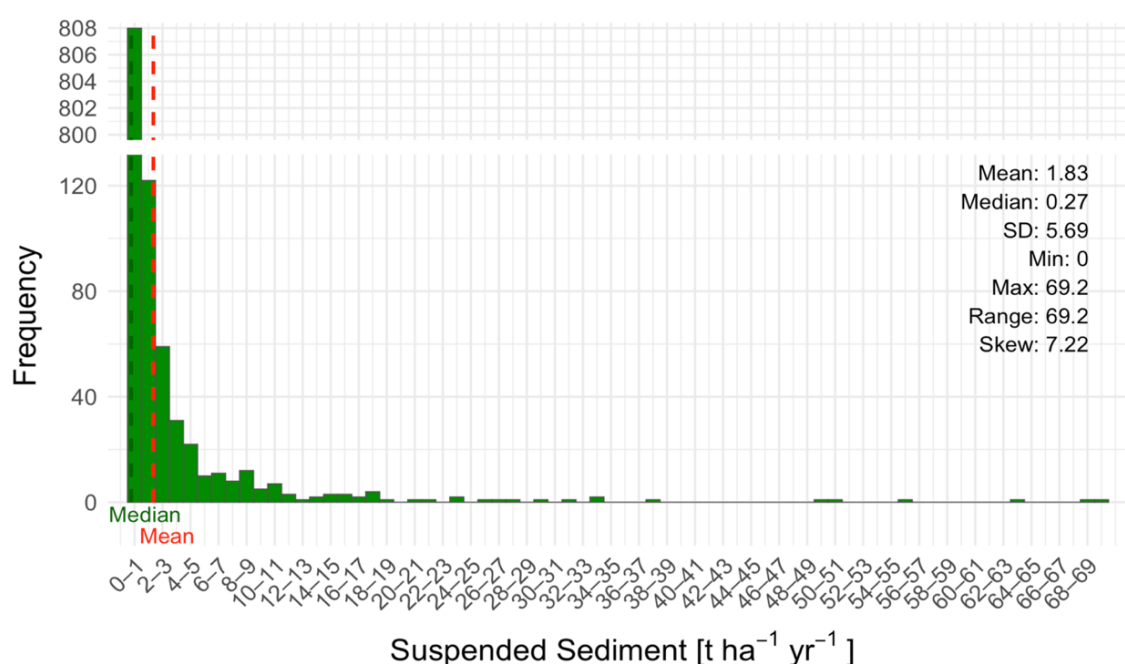


Figure 16. Frequency distribution of suspended sediment transport (2003-2024). Note the y-axis break.

The annually aggregated suspended sediment transport varies greatly between individual years, ranging from very low transport rates in the early 2000s (e.g. $0.07 \text{ t ha}^{-1} \text{ yr}^{-1}$ in 2003) to peaks of $4.27 \text{ t ha}^{-1} \text{ yr}^{-1}$ in 2007 and $4.49 \text{ t ha}^{-1} \text{ yr}^{-1}$ in 2010 (Figure 17). Suspended sediment transport was also high in 2014 at $2.97 \text{ t ha}^{-1} \text{ yr}^{-1}$. The period 2017-2021 had predominantly lower annual suspended sediment transport (below $1.0 \text{ t ha}^{-1} \text{ yr}^{-1}$) but it increased again afterwards, particularly in the hydrological year 2024 with $2.75 \text{ t ha}^{-1} \text{ yr}^{-1}$.

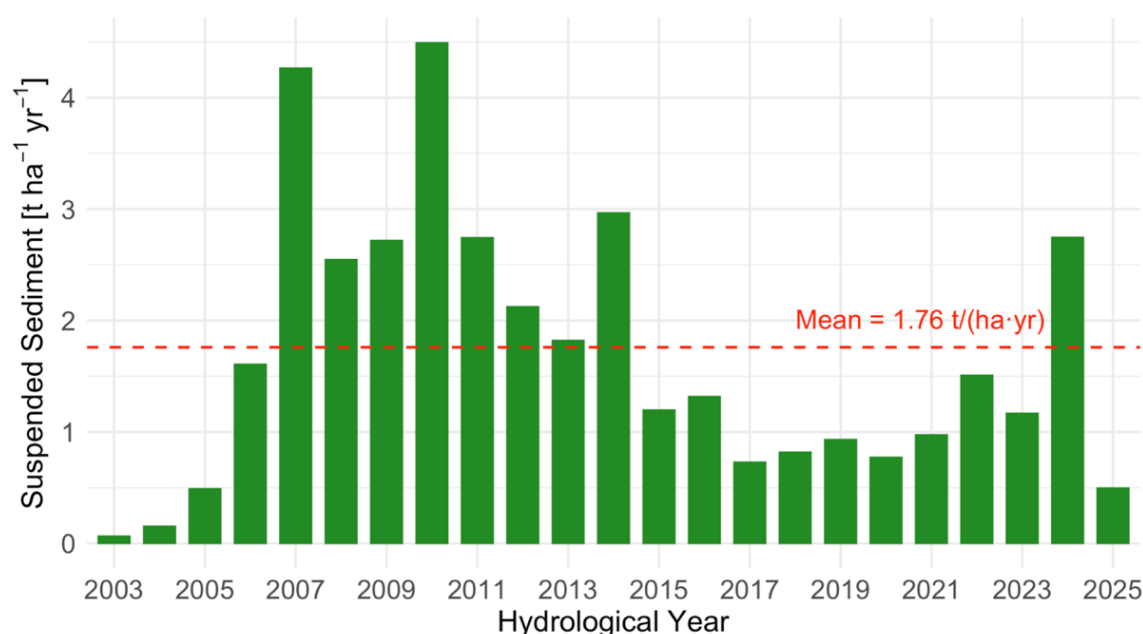


Figure 17. Total suspended sediment transport per hydrological year (October-September) and long-term mean for the period 2003-2024. The horizontal dashed line indicates the long-term mean suspended sediment transport rate of $1.76 \text{ t ha}^{-1} \text{ yr}^{-1}$, calculated from the total exported suspended sediment over the entire observation period (2862 t) divided by the corresponding time span and catchment area. Note that the hydrological year 2025 only contains values from October to December 2024.

When looking at the actual distribution of suspended sediment transport within a year, it is noticeable that there is a clear seasonal distribution, with low monthly averages in the winter months, particularly in February ($0.02 \text{ t ha}^{-1} \text{ yr}^{-1}$), and a maximum in summer (Figure 18). Transport is highest in July, August and especially in September with $0.37 \text{ t ha}^{-1} \text{ yr}^{-1}$. In addition, variation in the monthly average suspended sediment transport is greater in summer than in winter.

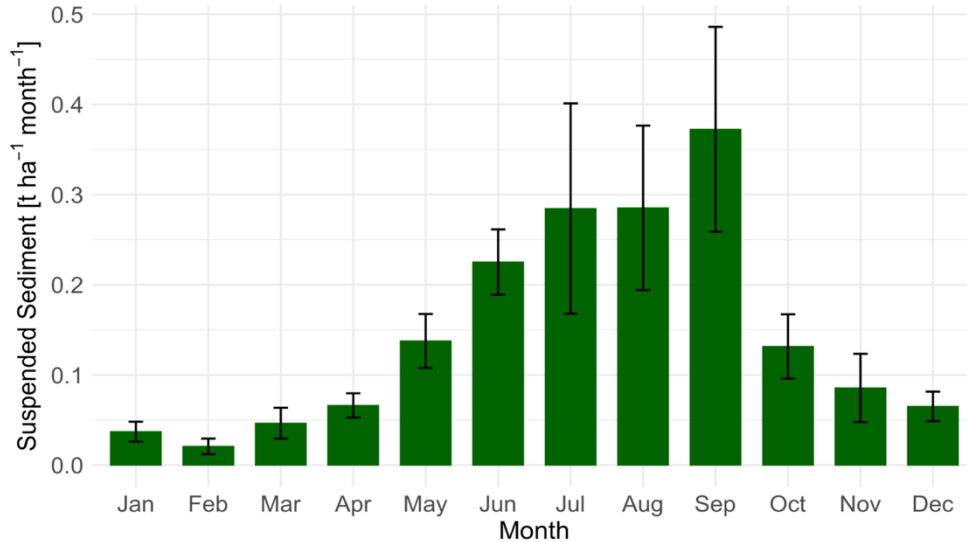


Figure 18. Average monthly suspended sediment with standard error bars in black (2003-2024).

There is a positive relationship between discharge and suspended sediment transport (Figure 19). Most measurement points are concentrated in the range of low to medium discharge and low sediment loads. Nevertheless, there are also some high suspended sediment transport values in this discharge range, including several weeks with a suspended sediment transport between 30 and $50 \text{ t ha}^{-1} \text{ yr}^{-1}$ and extreme values of over $60\text{--}70 \text{ t ha}^{-1} \text{ yr}^{-1}$. There is also some scatter at higher discharge values: while for many weeks suspended discharge remained below $10 \text{ t ha}^{-1} \text{ yr}^{-1}$, for individual weeks it was almost $30 \text{ t ha}^{-1} \text{ yr}^{-1}$.

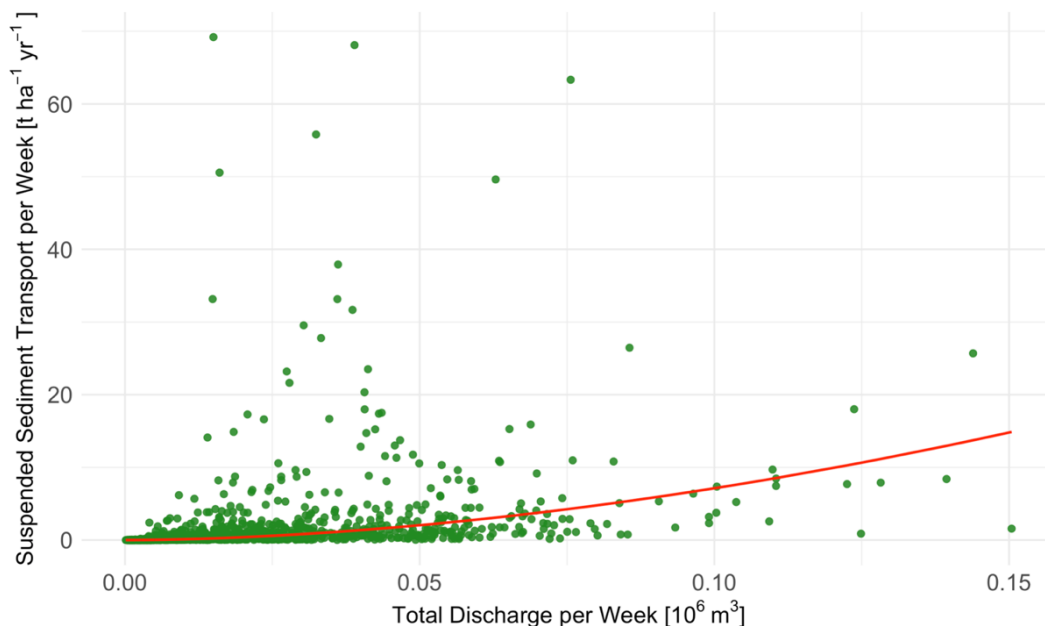


Figure 19. Scatterplot showing the total suspended sediment transport per week versus total discharge during the respective week for the period 2003-2024. The red line shows a power-law fit estimated via linear regression on log-transformed data (Adjusted R^2 : 0.66; $p = 2.2e^{-16}$).

5.2.3 Comparison of Bedload and Suspended Sediment

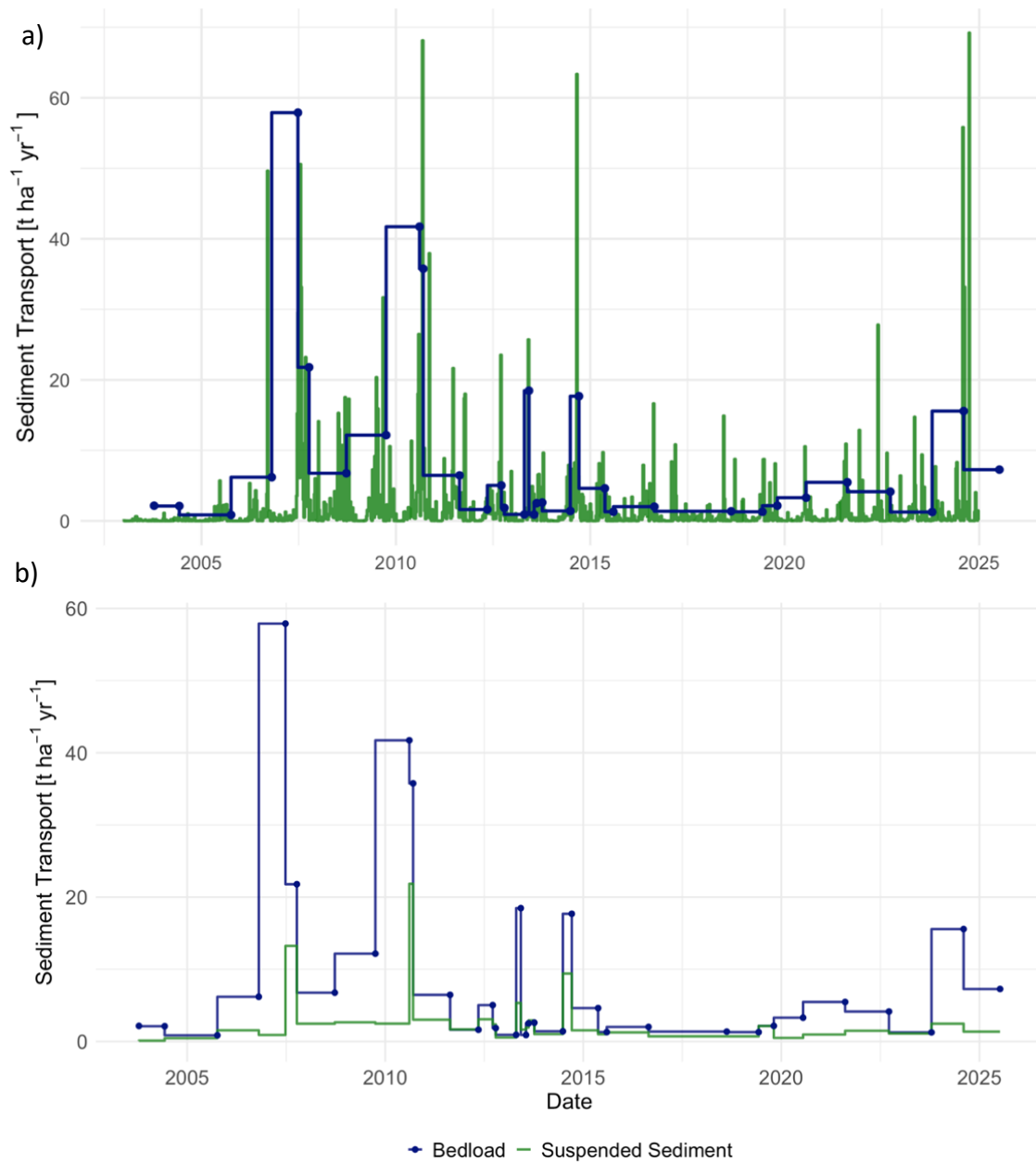


Figure 20. Comparison of the time series of bedload and suspended sediment transport (a) at the measurement frequency, and (b) summed for the same time periods for the period 2003-2025.

The variation in suspended sediment transport is similar to that of the variation in bedload transport (Figure 20) and the two are correlated (adjusted $R^2 = 0.53$; Figure 21). Although bedload transport is larger than suspended sediment transport (practically all points in Figure 21 lie above the 1:1 line), the peak suspended sediment transport values are sometimes as high as the bedload transport during the same time period (Figure 20).

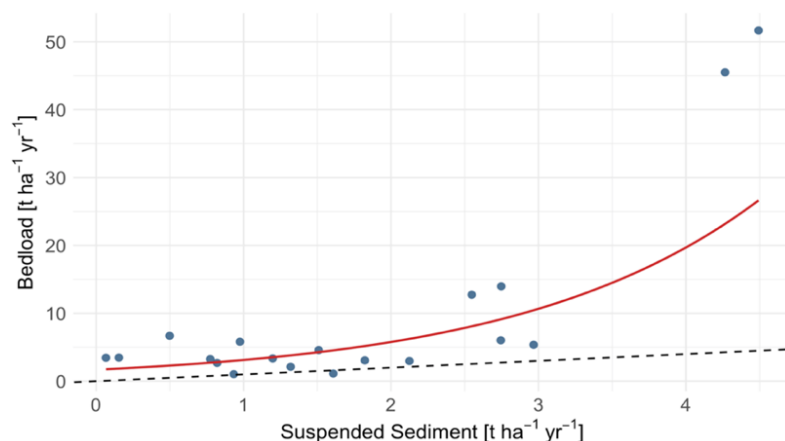


Figure 21. Scatterplot of annual bedload vs suspended sediment transport with an exponential fit in red and the 1:1 line in black (2003-2025) (Adjusted R^2 : 0.53; p -value: 0.0002).

5.3 Pu Isotopes Results

5.3.1 Soil Samples

Pu inventories at the reference site generally decrease with increasing depth, although slightly higher values occur in the second depth interval (Figure 22). A similar pattern is observed at the erosion sites, where inventories also decline with depth but often peak in the second interval, and in some cases (sites 1 and 3) also in the third. From a depth of 20 cm, Pu inventories decrease significantly at all sites. Sites 1, 3 and 4 show clear signs of soil erosion, as Pu inventories are generally lower than in the reference profile. In contrast, site 2 has higher Pu inventories in the uppermost depth intervals, indicating deposition.

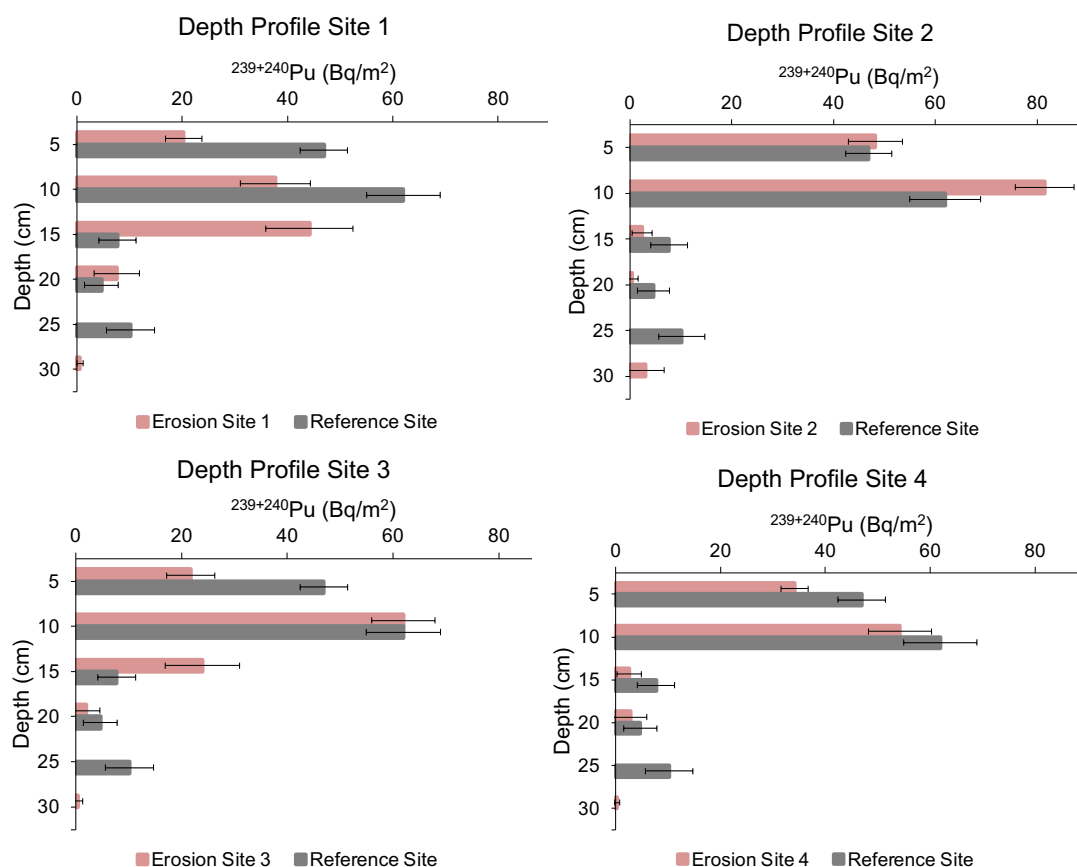


Figure 22. Depth profiles of the soil erosion sites in comparison with the reference site. The error bars show the standard error.

The calculated erosion and deposition rates are given in Table 3. All sites except site 2 show on average erosion across the models, whereas site 2 predominantly indicates slight deposition. Only the MODERN model suggests weak erosion there. This pattern is consistent with the depth profiles shown in Figure 22. The standard deviations indicate variability within the sites, particularly at site 4, which shows a wide range of calculated erosion rates.

Table 3. Soil erosion and deposition rates ($t\ ha^{-1}\ yr^{-1}$) for the sampling sites E1-E4 calculated using different models. Negative values indicate erosion and positive values deposition. Results from the Lal et al. (2013) model are shown for particle size correction factors $P = 1.0, 1.2$ and 1.5 , while h_0 is the profile shape parameter used in the Walling & He (1999) model. SD = standard deviation, SE = standard error.

Soil profile	Model by Lal et al. (2013)			Model by Walling & He (1999) Erosion rate ($t\ ha^{-1}\ yr^{-1}$); h_0	Model MODERN by Arata et al. (2016) Erosion rate ($t\ ha^{-1}\ yr^{-1}$)
	Erosion rate ($t\ ha^{-1}\ yr^{-1}$); $P=1$	Erosion rate ($t\ ha^{-1}\ yr^{-1}$); $P=1.2$	Erosion rate ($t\ ha^{-1}\ yr^{-1}$); $P=1.5$		
E1					
1	-2.15	-1.80	-1.44	-1.44	-2.84
2	0.91	0.76	0.61	-0.03	0.35
3	-1.44	-1.20	-0.96	-1.05	-2.02
4	-4.72	-3.93	-3.15	-3.43	-4.77
Average	-1.85	-1.54	-1.23	-1.49	-2.32
SD	2.32	1.93	1.54	1.42	2.12
SE	2.27	2.07	1.85	1.78	2.17
E2					
1	0.74	0.61	0.49	0.59	0.27
2	-2.56	-2.14	-1.71	0.07	-3.19
3	0.95	0.79	0.63	0.77	0.36
4	1.10	0.92	0.73	0.90	0.43
Average	0.05	0.05	0.04	0.58	-0.53
SD	1.75	1.46	1.17	0.36	1.77
SE	1.97	1.80	1.61	0.90	1.98
E3					
1	1.29	1.07	0.86	1.43	0.51
2	-1.89	-1.57	-1.26	0.08	-2.27
3	-3.20	-2.67	-2.14	-2.96	-3.45
4	-1.96	-1.63	-1.31	-1.49	-2.56
Average	-1.44	-1.20	-0.96	-0.74	-1.94
SD	1.92	1.60	1.28	1.90	1.71
SE	2.06	1.88	1.68	2.05	1.95
E4					
1	-1.45	-1.21	-0.97	-1.05	-2.03
2	-3.61	-3.01	-2.41	0.08	-4.36
3	-2.05	-1.71	-1.36	-1.26	-2.76
4	-8.11	-6.76	-5.41	-4.03	-7.63
Average	-3.80	-3.17	-2.54	-1.57	-4.20
SD	3.01	2.51	2.01	1.74	2.49
SE	2.58	2.36	2.11	1.97	2.35

All models result in erosion rates within a similar range (Figure 23), although some differences are evident. The MODERN model tends to calculate higher erosion values, while the model of Walling & He (1999) generally gives lower estimates. For the model of Lal et al. (2013), erosion rates decrease with increasing particle size correction. Despite these variations, the distributions overlap, and the Kruskal-Wallis test indicates no significant differences between the models ($p = 0.24$; see Appendix B for the R code).

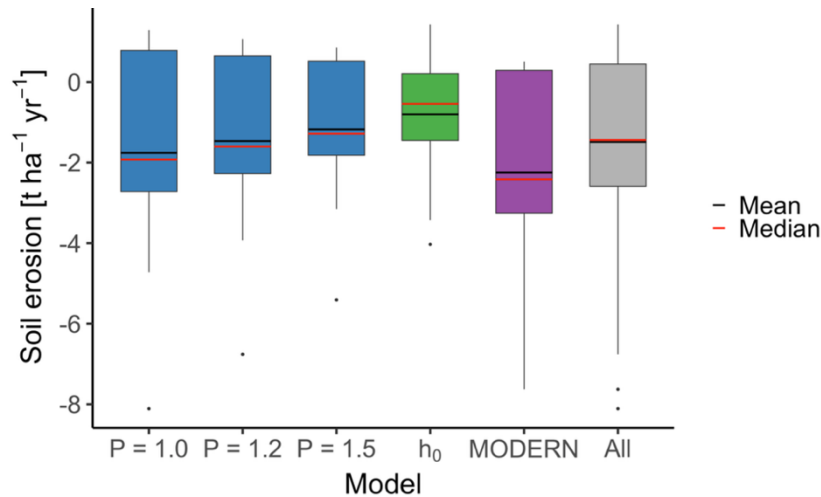
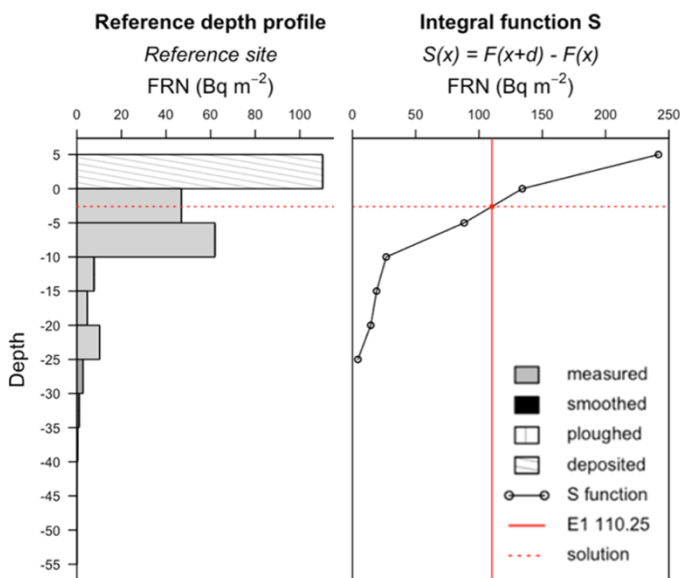


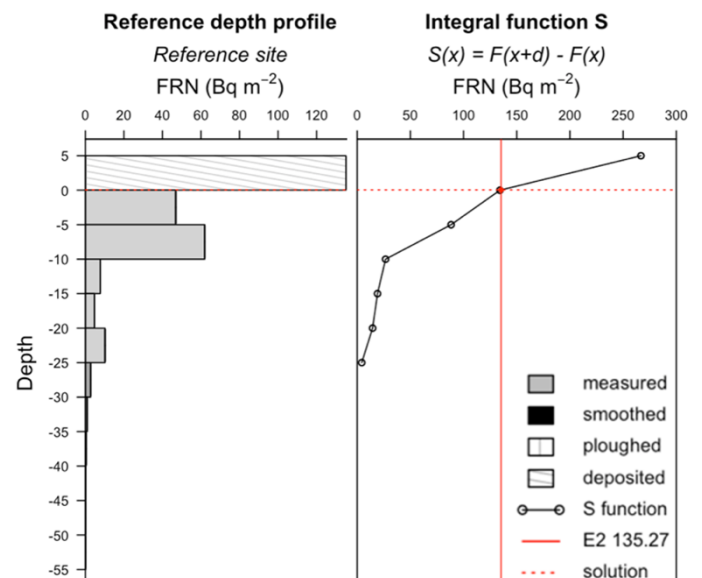
Figure 23. Boxplots of soil erosion rates from all erosion sites combined, estimated using the different models, described in chapter 4.2.3.1. Boxes show the interquartile range, with mean and median indicated, whiskers extending to $1.5 \times \text{IQR}$. The "All" category combines results from all models.

Figure 24 shows one of the model outputs from the MODERN model based on averaged radionuclide inventories (see Appendix A5 for the R code). Soil loss or gain since 1963 is determined from the intersection of the S-function with the measured total inventory (red line), with the corresponding depth (red dashed line) indicating erosion or deposition relative to the reference surface. For the study sites, soil losses of -2.61 cm (E1), -2.63 cm (E3) and -4.39 cm (E4) were calculated, while site E2 shows almost no change (+0.03 cm).

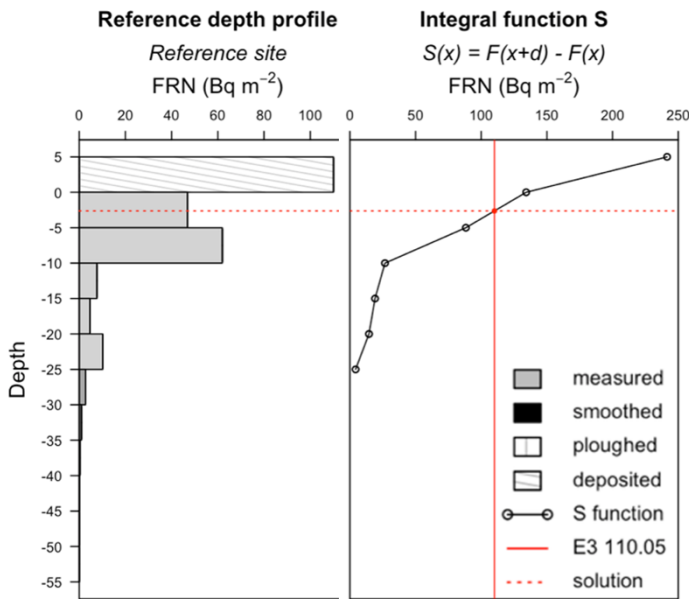
Erosion Site 1



Erosion Site 2



Erosion Site 3



Erosion Site 4

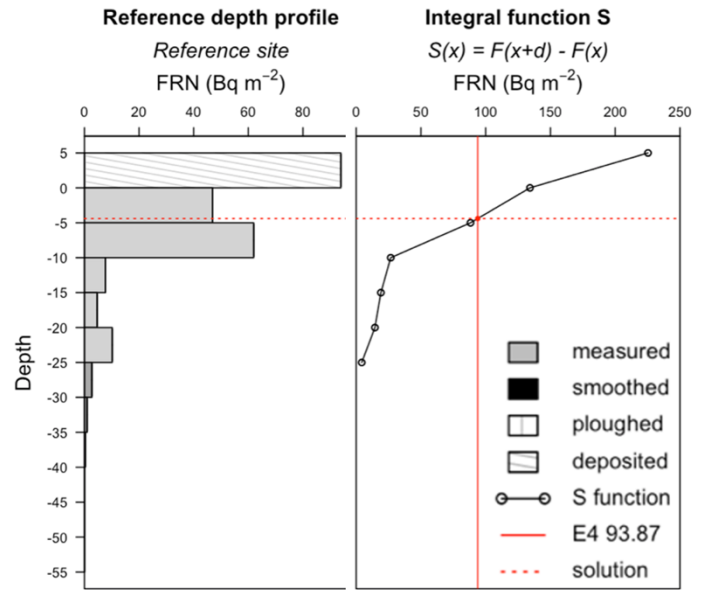


Figure 24. MODERN model outputs for erosion sites E1-E4 showing (left) the measured $^{239+240}\text{Pu}$ depth profile with simulated smoothed reference layers and an added depositional layer, and (right) erosion sites showing the cumulative $^{239+240}\text{Pu}$ inventory as a function of mass depth. The vertical red lines mark the measured total $^{239+240}\text{Pu}$ inventory of the soil profile. The horizontal red lines represent the solution of the S-function (the mass depth in the reference profile at which the cumulative $^{239+240}\text{Pu}$ inventory equals the measured inventory). The offset between this solution depth and the reference surface defines the net soil loss or gain since 1963.

5.3.2 River Sediment Samples

The calculated erosion rates for the river sediment samples are shown in Table 4. Both $^{239+240}\text{Pu}$ inputs result in negative values, which indicates erosion in the catchment. Erosion rates based on Q_1 are slightly higher than those derived from Q_2 but remain of the same order of magnitude.

Considerable variability can be observed between the sites, particularly at site 2a, which shows extremely high values that strongly influence the mean. Site 2b also shows elevated erosion rates. As site 2 represents a small tributary with low water and sediment transport (see Figure 7d), S2a and b were excluded from further analysis to avoid bias from these outliers. Excluding site 2 results in a noticeable decrease in both the mean erosion rate and standard deviation.

Spatially, erosion rates increase towards the outlet of the catchment. The location S5, located closest to the catchment outlet, shows the highest average erosion rate ($-2.04 \text{ t ha}^{-1} \text{ yr}^{-1}$). Slightly further upstream is S4, which is situated on the main sediment-transporting tributary, that also shows high erosion rates ($-1.73 \text{ t ha}^{-1} \text{ yr}^{-1}$). In contrast, S3, a smaller tributary with lower water and sediment discharge, shows considerably lower erosion ($-0.33 \text{ t ha}^{-1} \text{ yr}^{-1}$). Although S1 is located further upstream, it lies on the same sediment-rich tributary as S4 and therefore shows comparatively high erosion rates ($-1.64 \text{ t ha}^{-1} \text{ yr}^{-1}$) (see Figure 6 for the locations of all sampling sites).

A temporal comparison between samples S1a and b, as well as S2a and b, which were taken approximately two months apart, shows no consistent pattern. While the values at site 1 are relatively similar and show only minor differences, there are significant deviations between the two measurements at site 2. However, due to the small number of samples and the relatively short time difference, no clear conclusions can be drawn about a possible temporal trend.

Table 4. Erosion rates ($t\ ha^{-1}\ yr^{-1}$) of river sediment samples calculated using two atmospheric $^{239+240}Pu$ inputs: $Q_1 = 0.003\ Bq\ m^{-2}\ yr^{-1}$ (Hirose et al., 2008) & $Q_2 = 0.0021\ Bq\ m^{-2}\ yr^{-1}$ (Rosner & Winkler, 2001).

River Sediment Samples	Erosion rates ($t\ ha^{-1}\ yr^{-1}$) with Input $Q_1 = 0.003\ Bq\ m^{-2}\ yr^{-1}$ (Hirose et al., 2008)	Erosion rates ($t\ ha^{-1}\ yr^{-1}$) with Input $Q_2 = 0.0021\ Bq\ m^{-2}\ yr^{-1}$ (Rosner & Winkler, 2001)
Sampling Site 1a		
S1a.1	-0.70	-0.49
S1a.2	-2.64	-1.85
S1a.3	-2.39	-1.67
S1a.4	-0.81	-0.57
Average	-1.64	-1.14
Sampling Site 1b		
S1b.1	-1.45	-1.01
S1b.2	-2.73	-1.91
S1b.3	-2.80	-1.96
Average	-2.32	-1.63
Sampling Site 2a		
S2a.1	-0.04	-0.03
S2a.2	-27.99	-19.59
Average	-14.02	-9.81
Sampling Site 2b		
S2b.1	-3.90	-2.73
S2b.2	-8.35	-5.85
S2b.3	-0.04	-0.03
Average	-4.10	-2.87
Sampling Site 3		
S3.1	-0.43	-0.30
S3.2	-0.47	-0.33
S3.3	-0.24	-0.16
Average	-0.38	-0.26
Sampling Site 4		
S4.1	-3.54	-2.48
S4.2	0.00	0.00
S4.3	-2.56	-1.79
Average	-2.03	-1.42
Sampling Site 5		
S5.1	-2.58	-1.80
S5.2	-1.34	-0.94
S5.3	-3.26	-2.29
Average	-2.39	-1.68
Average all	-3.25	-2.28
SD	-5.99	-4.19
SE	-3.64	-3.05
Average without Site 2	-1.75	-1.22
SD	-1.19	-0.83
SE	-1.62	-1.36

Figure 25 confirms the strong influence of outliers, particularly from site 2. When all sites are included, the separation between mean and median indicates a skewed distribution. After excluding site 2, the mean and median become more similar and the variability decreases. This pattern is consistent for both inputs (Q_1 and Q_2). Despite minor differences between the two inputs, a Kruskal-Wallis test shows no statistically significant differences ($p = 0.64$) (see Appendix C for the R code).

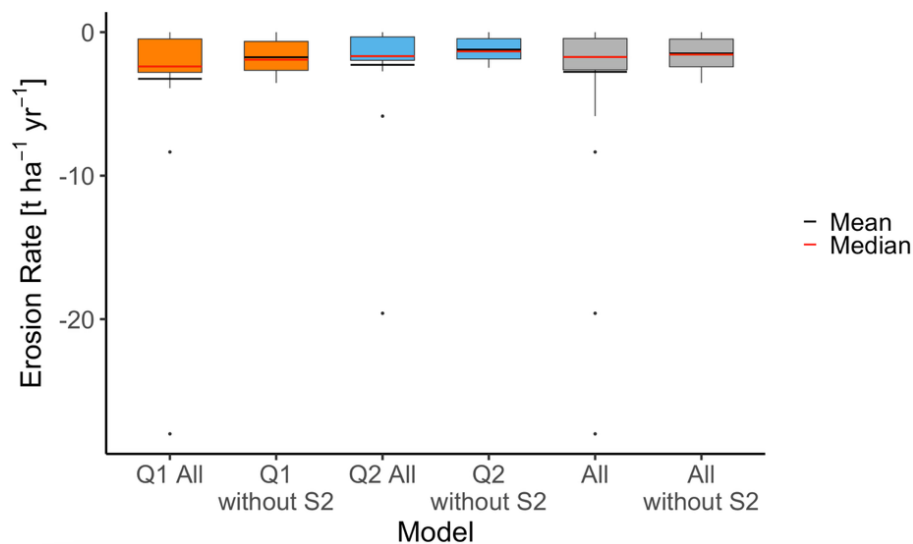


Figure 25. Boxplots of river sediment erosion rates for different atmospheric inputs and site selections. Q_1 and Q_2 , correspond to the two different inputs, reported by Hirose et al. (2008) and Rosner & Winkler (2001). "All" includes all sampling sites, while "without S2" indicates that site 2 was excluded from the calculations.

5.4 Comparison of Monitoring Data and Pu-Isotope Estimates

Figure 26 compares erosion rates calculated from Pu isotope data (soil and sediment samples) with long-term erosion rates derived from monitoring data (see Appendix D for the R code). The modeled values from soil and sediment samples are all similar. Differences between the models were tested using a Kruskal-Wallis test, which showed no statistically significant differences ($p = 0.34$), indicating that the modeling approaches result in comparable erosion estimates. Measured sediment transport (suspended sediment and bedload) was not included in the statistical test due to the lack of distributional information and is therefore used for qualitative comparison only. The long-term erosion rate from suspended sediment measurements ($-1.76 \text{ t ha}^{-1} \text{ yr}^{-1}$) corresponds closely with the modeled values and lies within the interquartile range of the Pu-based results. In contrast, the bedload erosion rate ($-9.06 \text{ t ha}^{-1} \text{ yr}^{-1}$) is considerably higher and clearly outside the range of the modeled and suspended sediment-based values.

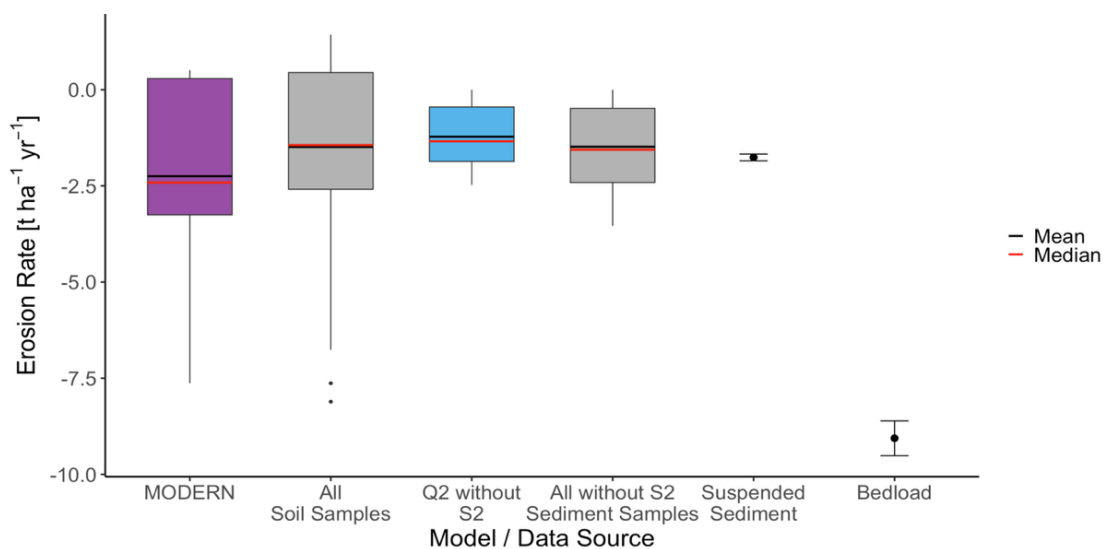


Figure 26. Boxplots and points comparing erosion rates from measurements and models. Points with error bars ($\pm 5\%$) show calculated erosion rates from measured Bedload and Suspended Sediment Data.

6 Discussion

6.1 Controls on Sediment Transport in the Erlenbach Catchment

The sediment transport data in the Erlenbach catchment indicate that both bedload and suspended sediment transport are dominated by a small number of high-intensity events that contribute a disproportionately large share to the total sediment load. This is reflected for example in the strongly right-skewed distributions, where a few high values strongly influence the overall statistics.

Previous studies in the Erlenbach confirm the central role of such extreme events. For instance, long-term analyses by Turowski et al. (2009) show that 41% of the total sediment yield over a period of more than 20 years was transported during just three major flood events. Measurements of bedload transport using geophone and plate impact sensors have also shown that a single flood event in June 2007 generated 96% of the recorded geophone pulses during the corresponding observation period of the retention basin, which illustrates the high transport efficiency of such extreme events (Rickenmann et al., 2012). A similar pattern is also evident for suspended sediment. For example, modeling in the Ardèche catchment showed that 75% of the annual ^{137}Cs flux was exported during the highest 10% of discharge days (Guillory et al., 2025).

This pattern is related to the rapid hydrological response of alpine catchments. Mountain streams such as the Erlenbach respond rapidly to precipitation events, which leads to strong discharge fluctuations and short but high-energy flood peaks (Stähli et al., 2021). The short response time of the runoff to precipitation events can be explained by the characteristics of the soil. Since a large portion of soil in the Erlenbach catchment consists of shallow, loamy layers, it has a low infiltration capacity. This means that these soils become saturated even with small amounts of water and the subsoil can hardly absorb any more water. During heavy rainfall, this leads to rapid surface runoff (Molnar et al., 2008; Rickenmann et al., 2012; Stambach, 1988).

These flood peaks are crucial for the mobilization of large amounts of sediment. Major floods influence not only immediate sediment transport but also sediment availability within the river. In steep mountain streams with a step-pool morphology, the riverbed is often stabilized by keystone structures. Field observations in the Erlenbach by Turowski et al. (2013) show that extreme events can cause significant morphological changes, for example by destroying step structures and causing local erosion over a meter deep. At the same time, extreme floods can also reactivate hillside processes and riverbank erosion, adding additional sediment from the catchment to the river system (Rickenmann, 2020).

The results also show a clear seasonal pattern for suspended sediment transport. The highest loads occur predominantly in late summer, while transport in spring remains comparatively low despite the high mean runoff due to snowmelt. This is further evidence that it is not the seasonal mean runoff, but rather short-term, high-intensity hydrological events that are decisive for sediment mobilization. The hydrology of the Erlenbach catchment is characterized by frequent, high-intensity summer precipitation events. These generate rapid, short-lived discharge peaks that create efficient sediment transport conditions. (Rickenmann, 2024; Rickenmann & McArde, 2007). Consistent with this, the largest sediment transport events in the Erlenbach, analyzed by Turowski et al. (2009), were all triggered by intense summer thunderstorms.

Another factor that explains why sediment transport in spring is relatively low despite high discharge could be the temporal decoupling between sediment supply and sediment transport. The study by Schuerch et al. (2006) in the Erlenbach shows that slope processes and minor landslides frequently deliver material into the channel during winter and spring, for example as a result of increased soil saturation or the additional weight of the snow pack. This sediment can initially be stored in the channel and is only later mobilized by more energetic summer events. According to Antoniazza et al. (2022), major sediment transport can also occur in the spring under certain conditions, such as during rain-on-snow events, when additional precipitation hits an already saturated snowpack, causing a sudden increase in runoff.

An important difference between bedload and suspended sediment lies in their relationship to discharge, as evident in the results. While suspended sediment transport shows a relatively strong correlation with discharge, it is much weaker for bedload. This suggests that the transport of suspended sediment is controlled more strongly by the river's transport capacity. In bedload transport, this relationship is less obvious. Coarse particles are often embedded in the riverbed and therefore cannot always be mobilized, even during high discharge. Consequently, similar discharge conditions can lead to very different bedload transport volumes depending on how much transportable material is available (Rickenmann, 2020). Such fluctuations between capacity-limited and supply-limited conditions were also observed by Antoniazza et al. (2022) in the Vallon de Nant (Canton of Vaud).

6.2 Long-term Erosion Signals from Pu Tracers

The vertical distributions of $^{239+240}\text{Pu}$ in the soil profiles studied, show a characteristic depth profile with a maximum near the surface of the soil, which is consistent with results reported in the literature. Instead of a purely exponential decrease with depth, a polynomial profile was observed, in which the highest activities typically occur a few centimeters below the soil surface. Such distributions have also been described in other studies in different regions (e.g. S. Dowell et al., 2024; Matisoff et al., 2011) and are generally attributed to the subsequent migration of the fallout signal. In contrast, freshly deposited radionuclides typically show an exponential decrease with depth (Alewell et al., 2013).

Downward migration can be explained by various post-depositional processes. Studies show that $^{239+240}\text{Pu}$ has a higher mobility in soil compared to ^{137}Cs , as it binds preferentially to organic matter and can therefore be transported more easily within the soil (Meusbürger et al., 2016). Changes in pH and the decomposition of organic matter can further promote this process by converting Pu into more mobile forms. In addition, physical processes such as bioturbation and the transport of fine particles by water infiltration contribute to the vertical redistribution (Alewell et al., 2013). In contrast, ^{137}Cs is more strongly bound to the fine mineral fraction and therefore remains more frequently in the uppermost soil horizons, leading to a more exponential decrease with depth (Percich et al., 2022).

There are significant differences in the calculated erosion rates between the individual soil sampling sites, including local deposition at site E2. Such small-scale variations are typical in mountainous catchments and reflect the spatial heterogeneity of erosion and deposition processes. A key influencing factor is slope gradient: Erosion generally increases with slope steepness (Ali et al., 2021), which could explain why E2, with the lowest inclination (8°), shows the lowest erosion rates. In addition, micro-relief structures can act as sediment traps where material from higher-elevation slope areas accumulates (Li et al., 2020). The highest erosion rates and the greatest local variability were measured at site E4. As shown in Figure 27, this area is used as a kind of forest road, which can lead to soil

compaction and therefore a reduced infiltration capacity. This promotes the formation of surface runoff and erosion processes (Thomaz et al., 2022; Zhai et al., 2025). Furthermore, it can be assumed that the driving itself contributes to the direct removal of soil material. The observed variability between the sites shows that point-specific soil samples can only capture the complex spatial distribution of erosion and deposition to a limited extent (Antoniazza et al., 2023; Zhai et al., 2025).



Figure 27. Location of erosion site 4 (E4) (Source: own photograph).

The use of different erosion models leads to varying results, primarily due to the respective model assumptions. The model by Lal et al. (2013) is based on an idealized exponential depth distribution of the tracer, whereas the approach by Walling & He (1999) uses cumulative mass-depth profiles and thus describes the vertical displacement of the tracer differently. In contrast, the numerical MODERN model by Arata et al. (2016) uses the measured reference profile directly and is not bound to a fixed functional form, making it more sensitive to changes in the profile shape (Meusburger et al., 2016; Dowell et al., 2024). Erosion estimates are also influenced by the Particle Size Correction Factor (P), which accounts for the preferred binding of Pu to fine particles. Higher P values result in lower calculated erosion rates, as part of the inventory loss is attributed to the preferential transport of fine-grained particles and is therefore not fully interpreted as soil loss (Lal et al., 2013). Overall, no single model is clearly superior. However, the MODERN model is regarded as the most advanced, as it makes no rigid assumptions about profile shape, utilizes the entire soil profile and thus represents the measured fallout radionuclides inventories particularly realistically (Meusburger et al., 2016; Dowell et al., 2024).

Furthermore, erosion rates derived from the mass balance approach show a clear dependence on the assumed atmospheric Pu deposition. Naturally, the use of the higher input Q_1 (Hirose et al., 2008) leads to higher erosion rates than the lower value Q_2 (Rosner & Winkler, 2001), since the erosion rate is proportional to the atmospheric input. The differences between the two approaches reflect regional variations in atmospheric fallout input, with Q_1 being strongly influenced by resuspension processes in East Asia. For the Erlenbach catchment, the measured input Q_2 for southern Germany, therefore appears to be more realistic, as it better represents the regional conditions.

The erosion rates calculated from the sediment samples vary along the river and tend to be higher towards the catchment outlet. This pattern reflects changes in sediment sources and connectivity within the catchment. In its lower section, the Erlenbach is characterized by steep banks, where the

stream cuts into the soft subsoil (Stammach, 1988). This process mobilizes Pu-depleted material from deeper layers, which can lead to higher calculated erosion rates. At the same time, depth erosion reduces support at the foot of the slopes and can trigger or intensify slope movements (Figure 28) (Stammach, 1988). The left bank in particular is highly active and continuously supplies material to the stream (Rickenmann, 2020, 2024; Rickenmann & McArdell, 2007; Schuerch et al., 2006).

These processes raise the question of whether Pu in river sediments reflects purely soil erosion. Additional sediment sources such as bank erosion and deeper material could also contribute to the signal. However, the good agreement between erosion rates derived from river sediments and soil samples suggests that hillslope erosion remains a major contributor to the material exported at the outlet. Since plutonium is primarily bound to fine particles, which are mainly derived from soil erosion rather than coarse bedload, the results indicate that hillslope-channel connectivity increases downstream. In combination with higher discharges, which enhance sediment transport efficiency, this leads to a stronger contribution of hillslope-derived material and thus higher erosion rates towards the outlet. This interpretation is further supported by the observation that higher erosion rates occur in sections with greater discharge, indicating that hydrologically active parts of the catchment are more effectively connected to sediment sources (Percich et al., 2022; Turowski et al., 2016).



Figure 28. Active riverbank erosion at location S4 (Source: own photograph).

6.3 Evaluation of the Pu Isotope Approach

The results show that soil erosion in the Erlenbach catchment can generally be estimated accurately using Pu isotopes from river sediments. This is particularly clear from the close match to long-term suspended sediment data, which is reasonable, as Pu preferentially binds to fine-grained particles, that make up most of the suspended sediment (Alewell et al., 2013). However, the results showed that both suspended sediment and bedload transport vary significantly from year to year. To meaningfully compare the monitoring data with the Pu-based results, it is therefore necessary to use long-term averages for both suspended sediment and bedload, provided these are interpreted as measures of soil erosion. In contrast to suspended sediment data, bedload data are only of limited use. Bedload primarily consists of coarser particles that are transported along the riverbed and often originate from the riverbed itself, the banks or slope processes and are therefore not only related to soil erosion (Stammach, 1988; Turowski et al., 2009).

Furthermore, the erosion rates of the soil samples are in the same order of magnitude as those from the river sediments. This provides further evidence that the river sediments accurately represent soil erosion. The soil samples were specifically taken at locations where erosion was expected. If these sites are representative of the eroding areas within the catchment, the integrated signal in the river sediments, which reflects a mixture of material from these areas, should correspond closely to the measured soil erosion rates. In this sense, the river sediments can be interpreted as an average signal of the catchment, and the good agreement indicates that the soil samples capture this average reasonably well.

It would also be possible to extend the erosion rates from the soil samples to the entire catchment through spatial upscaling, which would allow for a detailed representation of local differences in soil erosion. However, this requires a higher number of samples and is therefore more time-consuming (Lal et al., 2013). Nevertheless, the existing soil profiles serve as an important reference to demonstrate that the erosion rates derived from river sediments fall within a plausible range.

In principle, the determination of soil erosion based on Pu measurements in river sediments is therefore also applicable in other catchments under certain conditions. It is crucial that there is active water and sediment transport, as this is the primary mechanism by which plutonium is transported from the slopes into the river system and ultimately to the catchment outlet (Meusburger et al., 2016). Good connectivity between the slope and the channel is therefore essential, whereas in systems with significant temporary storage, the signal may be weakened (Flatley et al., 2025; Padilla et al., 2018). To ensure the most representative assessment of the catchment, samples should preferably be taken at the catchment outlet, as this is where the integrated signal from the entire area is present (Percich et al., 2022). Additionally, a sufficient number of samples is necessary to account for small-scale variability (Loba et al., 2022). In alpine regions, Pu offers clear advantages over Cs, as it originates from global fallout and is more homogeneously distributed (Alewell et al., 2013). Another critical point is the selection of a suitable atmospheric Pu input, as this directly influences the calculation of erosion rates and must be regionally adjusted.

The approach of using Pu isotopes in river sediments offers several key advantages. In particular, a single sampling procedure provides an erosion signal that integrates data over decades (S. M. Dowell et al., 2023), making it especially valuable in catchments without long-term monitoring data. As a result, the fieldwork required is relatively minimal, while at the same time providing a robust estimate of long-term soil erosion across the entire catchment.

The method has also already been successfully applied in various regions. For example, Percich et al. (2022) used sediment traps at the outlets of five catchments in the US to measure $^{239+240}\text{Pu}$ in suspended sediment. In Spain and Portugal, Padilla et al. (2018) used meteoric ^{10}Be to determine the denudation rate of the entire Guadiana Basin. In China, Chen et al. (2023) combined ^{137}Cs measurements on slopes with a fingerprinting technique at the outlet to quantify the contributions of different land uses to sediment export. Bayesian mixture models such as MixSIAR are often used for this purpose, which statistically calculate the contributions of different sources (e.g. farmland vs. bank erosion) (Fang et al., 2022; Upadhayay et al., 2022; Wynants et al., 2021).

One important development is the use of multi-isotope approaches. To understand the dynamics across different time scales, several isotopes can be combined. While ^7Be , with a half-life of about 53 days, reflects short-term erosion events (e.g. after heavy rainfall events), ^{137}Cs and $^{239+240}\text{Pu}$ provide data over decades and $^{210}\text{Pb}_{\text{ex}}$ even over a century. This combination makes it possible to examine

whether erosion rates have changed recently (e.g. due to climate change or changes in land use) compared to long-term averages (Huang et al., 2025; Loba et al., 2022; Mabit et al., 2008). This also shows that isotope-based approaches are not limited to the quantification of long-term erosion rates but can also provide insights into temporal dynamics and sediment sources.

Recent studies have combined radionuclide data with remote sensing, GIS and artificial intelligence. For example, random forest algorithms are used to create maps of sediment origin and distribution for complex terrain from point isotope measurements (Huang et al., 2025). The integration of isotope data for the calibration of physical erosion models (such as SWAT) is also considered a good way to increase the accuracy of predictions in large catchment areas (Guillory et al., 2025; Huang et al., 2025).

6.4 Limitations and Uncertainties

The Pu isotope approach to estimating soil erosion is based on several conceptual assumptions. A key point is that the Pu signal represents an average erosion rate over several decades (S. Dowell et al., 2024). As a result, neither individual years nor specific extreme events can be directly captured. This complicates comparisons with monitoring data, which often cover shorter time periods. A time scale bias arises particularly when rare, high-frequency extreme events (which Pu integrates) are absent during the shorter monitoring period (Wittmann et al., 2024).

Furthermore, a steady state between tracer input and output is often assumed, which is frequently not the case in dynamic mountain catchments (Jautzy et al., 2023). This introduces uncertainty, since the signal measured at the outlet does no longer reflect the total soil erosion of the catchment, but only the portion that is actually transported there. Jautzy et al. (2023) therefore propose carrying out additional analyses (e.g. pair-nuclide approaches like $^{26}\text{Al}/^{10}\text{Be}$) in such areas to determine the degree of imbalance between input and output. Another potential issue is that during major floods, suspended sediment does not necessarily settle but is washed away (Schuerch et al., 2006), which also leads to an underestimation of the exported Pu-relevant mass. In addition, the time lag between hillslope erosion and sediment delivery to the outlet can further contribute to an underestimation of soil erosion.

The validity of the results also depends heavily on the representativeness of the sampling. For river sediments, the sampling location is critical, as samples should ideally be taken at the outlet of the catchment to get the most comprehensive signal possible (Percich et al., 2022). The number and distribution of samples also influence the results. Both river and soil samples must be sufficiently numerous to capture the spatial heterogeneity of erosion and deposition processes. Factors such as slope gradient, micro-relief or vegetation cover lead to small-scale differences that cannot be adequately represented by a small number of samples (Alewell et al., 2013). According to Loba et al. (2022), it is therefore recommended to either collect multiple replicate samples (typically five to ten cores per site) or to collect and homogenize larger sample quantities (approx. 1-2 kg) to reduce the high sample and analysis costs while adequately accounting for spatial variability.

The calculated erosion rates also depend on model assumptions. Different models, such as the approach by Lal et al. (2013), the model by Walling & He (1999) or the MODERN model, are based on different assumptions regarding the vertical distribution of the tracer and therefore lead to different results. Model parameters such as depth coefficients or the particle size correction factor also influence the results.

A particularly important source of uncertainty is the assumed atmospheric Pu input, as this can vary regionally. Likewise, the reference inventory is crucial, since it serves as a baseline against which all soil samples are compared to. Errors or uncertainties at this stage are directly transferred to the erosion or deposition rates. For example, if the reference value is set too high, this can lead to an overestimation of erosion (Loba et al., 2022). Furthermore, determining a representative reference site is often difficult, as small-scale variability in fallout as well as local influences such as vegetation, micro-relief or past disturbances can affect the inventory (Zhang, 2023).

Additional uncertainties arise from the monitoring data used, as measurement gaps and irregular time intervals make it difficult to attribute sediment transport to specific events, which may result in extreme events being over- or underrepresented. Furthermore, the mass calculations are based on the assumption of a constant sediment density of 1.75 t/m^3 , although in reality this varies depending on the sediment's composition. This creates uncertainties in the conversion from volume to mass that directly affect the calculated transport and erosion rates.

Finally, laboratory analyses are also subject to uncertainties. Sample preparation is complex and can influence the measurement results. Furthermore, measurement uncertainties arise particularly at low activity levels. This is especially problematic in regions with low fallout, such as the tropics, where some samples may even fall below the detection limit, thus making a reliable interpretation difficult. Additionally, the required measurement methods are expensive and not universally available (S. M. Dowell et al., 2023).

7 Conclusions and Outlook

The purpose of this study was to determine long-term erosion rates in the Erlenbach catchment and to evaluate the suitability of Pu isotopes ($^{239+240}\text{Pu}$) as tracers for estimating soil erosion at the catchment scale. This was done by comparing sediment transport monitoring data with isotope-based analyses of soil and river sediment samples.

The results show that sediment transport in the Erlenbach is strongly dominated by a few very intense runoff events, which account for a large fraction of the total sediment export. Average erosion rates derived from the monitoring data differ significantly between suspended sediment and bedload transport: Suspended sediment data provide realistic values for soil erosion at approximately $-1.76 \text{ t ha}^{-1} \text{ yr}^{-1}$, while bedload data show significantly higher values, as they additionally capture processes such as riverbed and bank erosion.

The calculation of long-term erosion rates using Pu isotopes shows that the various modeling approaches essentially deliver comparable results. The erosion rates derived from river sediment samples are of a similar order of magnitude as those from soil samples, which underlines the consistency and reliability of the Pu-based method.

The comparison of the methods shows that Pu analyses provide a realistic estimate of long-term soil erosion and correlate particularly well with suspended sediment data. This indicates that the method reliably captures the overall erosion process within the catchment. However, the significantly higher bedload data indicate that these take into account other sediment sources in addition to soil erosion and therefore cannot be directly correlated with soil erosion.

Despite the generally consistent results, this study also reveals several methodological limitations. A key assumption of the Pu-based method is the existence of a long-term equilibrium between atmospheric input and erosion-induced loss. In dynamic alpine catchments, however, this assumption is often only partially valid, as sediment can be temporarily stored and subsequently remobilized with a time lag. Another source of uncertainty arises from the choice of atmospheric Pu input. Different values reported in the literature lead to different erosion rates, which is why it is important to select these as accurately as possible. Similarly, determining a representative reference inventory plays a crucial role, as deviations in this value also influence the calculated erosion rates.

The interpretation of the river sediment samples is also not entirely straightforward. Plutonium is preferentially bound to fine-grained particles and therefore primarily reflects soil erosion. Processes such as bank erosion or landslides contribute to the Pu signal only when they mobilize fine material. Particularly in the lower reaches of the Erlenbach, where severe depth erosion and slope instability occur, increased sediment connectivity can lead to more material being transported from eroded areas, causing the calculated erosion rates to rise slightly. Overall, the good agreement between the erosion rates derived from river sediments and soil samples suggests that soil erosion on the slopes remains the dominant contributor to the measured Pu signal.

Finally, the monitoring data themselves are subject to uncertainties, for example due to irregular measurement intervals or assumptions of a constant sediment density. Nevertheless, they provide a valuable reference for interpreting the isotope-based results.

The results of this study show that the use of Pu isotopes, particularly in combination with sediment samples collected at the outlet of the catchment, is a promising approach for estimating long-term erosion rates. At the same time, however, further research is needed to better validate the method and gain a more precise understanding of its limitations.

One interesting point is the combination of multiple radionuclides with different half-lives. Multi-isotope approaches make it possible to study erosion processes on different timescales and to identify differences between short-term events and long-term trends. Furthermore, sediment fingerprinting methods could be used to more accurately determine the origin of the transported material and to better quantify the contribution of different erosion sources.

Another promising area of research is the integration of isotope data with methods such as remote sensing, GIS and machine learning. Such approaches could help extrapolate point measurements to larger catchment areas and provide a more detailed picture of spatial erosion patterns.

When applying this method to other catchments, it is particularly important to emphasize the significance of connectivity between the slopes and the river. A representative integrated signal at the outlet is only possible in systems with sufficient active water and sediment transport. Future research should therefore focus more closely on determining the geomorphological and hydrological conditions under which the method can be reliably applied.

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9 Appendix

Appendix A: R code for MODERN model calculations

A1: Site E1

```
install.packages("path_to_modern_package", repos = NULL, type="source")
# load the library
library("modeRn")
# creates a reference profile composed of 5 layers of thickness 5 cm
RDP = createReferenceProfile(FRNinv = c(46.92, 61.97, 7.72, 4.66, 10.21), thickness =
    5, name = 'Reference site', falloutTime = 1963, refTime =
2025, massDepth = 177.62)
plot(RDP)
print(RDP)

# creates sampling layers of thickness 30 cm
P1 = createSamplingProfile(FRNinv = 103.40, thickness = 30, name = 'E1 Profile 1',
    falloutTime = 1963, refTime = 2025)
P2 = createSamplingProfile(FRNinv = 153.22, thickness = 30, name = 'E1 Profile 2'
    , falloutTime = 1963, refTime = 2025)
P3 = createSamplingProfile(FRNinv = 110.25, thickness = 30, name = 'E1 Profile 3',
    falloutTime = 1963, refTime = 2025)
P4 = createSamplingProfile(FRNinv = 74.14, thickness = 30, name = 'E1 Profile 4',
    falloutTime = 1963, refTime = 2025)

# creates smoothed layers below the reference profile
RDP_smooth1 = addSmoothedLayers(RDP, P1)
plot(RDP_smooth1, main = 'Simulated depth profile')
print(P1)

RDP_smooth2 = addSmoothedLayers(RDP, P2)
plot(RDP_smooth2, main = 'Simulated depth profile')
print(P2)

RDP_smooth3 = addSmoothedLayers(RDP, P3)
plot(RDP_smooth3, main = 'Simulated depth profile')
print(P3)

RDP_smooth4 = addSmoothedLayers(RDP, P4)
plot(RDP_smooth4, main = 'Simulated depth profile')
print(P4)

# sampling profiles with eroded layers
MODERN_P1 = MODERN(RDP_smooth1, P1)
plot(MODERN_P1)
print(MODERN_P1)
ER1 <-yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 157.15,
sampleDepth = 30)

show(ER1)

MODERN_P2 = MODERN(RDP_smooth2, P2)
plot(MODERN_P2)
print(MODERN_P2)
```

```
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 104.05,
sampleDepth = 30)
```

```
MODERN_P3 = MODERN(RDP_smooth3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 143.70,
sampleDepth = 30)
```

```
MODERN_P4 = MODERN(RDP_smooth4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 144.25,
sampleDepth = 30)
```

```
#mit Deposition
RDP_depP1 = addDepositionLayers(RDP_smooth1, 103.40, 1)
MODERN_P1 = MODERN(RDP_depP1, P1)
plot(MODERN_P1)
print(MODERN_P1)
yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 157.15,
sampleDepth = 30)
```

```
RDP_depP2 = addDepositionLayers(RDP_smooth2, 153.22, 1)
MODERN_P2 = MODERN(RDP_depP2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 104.05,
sampleDepth = 30)
```

```
RDP_depP3 = addDepositionLayers(RDP_smooth3, 110.25, 1)
MODERN_P3 = MODERN(RDP_depP3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 143.70,
sampleDepth = 30)
```

```
RDP_depP4 = addDepositionLayers(RDP_smooth4, 74.14, 1)
MODERN_P4 = MODERN(RDP_depP4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 144.25,
sampleDepth = 30)
```

A2: Site E2

```
install.packages("path_to_modern_package", repos = NULL, type="source")
# load the library
library("modeRn")
# creates a reference profile composed of 5 layers of thickness 5 cm
RDP = createReferenceProfile(FRNinv = c(46.92, 61.97, 7.72, 4.66, 10.21), thickness =
5, name = 'Reference site', falloutTime = 1963, refTime =
2025, massDepth = 177.62)
plot(RDP)
print(RDP)

# creates sampling layers of thickness 30 cm
P1 = createSamplingProfile(FRNinv = 145.08, thickness = 30, name = 'E2 Profile 1',
```

```

        falloutTime = 1963, refTime = 2025)
P2 = createSamplingProfile(FRNinv = 95.66, thickness = 30, name = 'E2 Profile 2'
        , falloutTime = 1963, refTime = 2025)
P3 = createSamplingProfile(FRNinv = 149.44, thickness = 30, name = 'E2 Profile 3',
        falloutTime = 1963, refTime = 2025)
P4 = createSamplingProfile(FRNinv = 152.68, thickness = 30, name = 'E2 Profile 4',
        falloutTime = 1963, refTime = 2025)

# creates smoothed layers below the reference profile
RDP_smooth1 = addSmoothedLayers(RDP, P1)
plot(RDP_smooth1, main = 'Simulated depth profile')
print(P1)

RDP_smooth2 = addSmoothedLayers(RDP, P2)
plot(RDP_smooth2, main = 'Simulated depth profile')
print(P2)

RDP_smooth3 = addSmoothedLayers(RDP, P3)
plot(RDP_smooth3, main = 'Simulated depth profile')
print(P3)

RDP_smooth4 = addSmoothedLayers(RDP, P4)
plot(RDP_smooth4, main = 'Simulated depth profile')
print(P4)

# sampling profiles with eroded layers
MODERN_P1 = MODERN(RDP_smooth1, P1)
plot(MODERN_P1)
print(MODERN_P1)
ER1 <-yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 130.75,
sampleDepth = 30)

show(ER1)

MODERN_P2 = MODERN(RDP_smooth2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 141.25,
sampleDepth = 30)

MODERN_P3 = MODERN(RDP_smooth3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 129.55,
sampleDepth = 30)

MODERN_P4 = MODERN(RDP_smooth4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 128.85,
sampleDepth = 30)

#mit Deposition
RDP_depP1 = addDepositionLayers(RDP_smooth1, 145.08, 1)
MODERN_P1 = MODERN(RDP_depP1, P1)
plot(MODERN_P1)
print(MODERN_P1)
yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 130.75,
sampleDepth = 30)

```

```

RDP_depP2 = addDepositionLayers(RDP_smooth2, 95.66, 1)
MODERN_P2 = MODERN(RDP_depP2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 141.25,
sampleDepth = 30)

RDP_depP3 = addDepositionLayers(RDP_smooth3, 149.44, 1)
MODERN_P3 = MODERN(RDP_depP3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 129.55,
sampleDepth = 30)

RDP_depP4 = addDepositionLayers(RDP_smooth4, 152.68, 1)
MODERN_P4 = MODERN(RDP_depP4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 128.85,
sampleDepth = 30)

```

A3: Site E3

```

install.packages("path_to_modern_package", repos = NULL, type="source")
# load the library
library("modeRn")
# creates a reference profile composed of 5 layers of thickness 5 cm
RDP = createReferenceProfile(FRNinv = c(46.92, 61.97, 7.72, 4.66, 10.21), thickness =
5, name = 'Reference site', falloutTime = 1963, refTime =
2025, massDepth = 177.62)
plot(RDP)
print(RDP)

# creates sampling layers of thickness 30 cm
P1 = createSamplingProfile(FRNinv = 166.95, thickness = 30, name = 'E3 Profile 1',
falloutTime = 1963, refTime = 2025)
P2 = createSamplingProfile(FRNinv = 90.57, thickness = 30, name = 'E3 Profile 2'
, falloutTime = 1963, refTime = 2025)
P3 = createSamplingProfile(FRNinv = 80.14, thickness = 30, name = 'E3 Profile 3',
falloutTime = 1963, refTime = 2025)
P4 = createSamplingProfile(FRNinv = 102.55, thickness = 30, name = 'E3 Profile 4',
falloutTime = 1963, refTime = 2025)

# creates smoothed layers below the reference profile
RDP_smooth1 = addSmoothedLayers(RDP, P1)
plot(RDP_smooth1, main = 'Simulated depth profile')
print(P1)

RDP_smooth2 = addSmoothedLayers(RDP, P2)
plot(RDP_smooth2, main = 'Simulated depth profile')
print(P2)

RDP_smooth3 = addSmoothedLayers(RDP, P3)
plot(RDP_smooth3, main = 'Simulated depth profile')
print(P3)

RDP_smooth4 = addSmoothedLayers(RDP, P4)

```

```

plot(RDP_smooth4, main = 'Simulated depth profile')
print(P4)

# sampling profiles with eroded layers
MODERN_P1 = MODERN(RDP_smooth1, P1)
plot(MODERN_P1)
print(MODERN_P1)
ER1 <-yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 94.50,
sampleDepth = 30)

show(ER1)

MODERN_P2 = MODERN(RDP_smooth2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 88.65,
sampleDepth = 30)

MODERN_P3 = MODERN(RDP_smooth3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 113.40,
sampleDepth = 30)

MODERN_P4 = MODERN(RDP_smooth4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 138.20,
sampleDepth = 30)

#mit Deposition
RDP_depP1 = addDepositionLayers(RDP_smooth1, 166.95, 1)
MODERN_P1 = MODERN(RDP_depP1, P1)
plot(MODERN_P1)
print(MODERN_P1)
yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 94.50,
sampleDepth = 30)

RDP_depP2 = addDepositionLayers(RDP_smooth2, 90.57, 1)
MODERN_P2 = MODERN(RDP_depP2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 88.65,
sampleDepth = 30)

RDP_depP3 = addDepositionLayers(RDP_smooth3, 80.14, 1)
MODERN_P3 = MODERN(RDP_depP3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 113.40,
sampleDepth = 30)

RDP_depP4 = addDepositionLayers(RDP_smooth4, 102.55, 1)
MODERN_P4 = MODERN(RDP_depP4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 138.20,
sampleDepth = 30)

```

A4: Site E4

```
install.packages("path_to_modern_package", repos = NULL, type="source")
# load the library
library("modeRn")
# creates a reference profile composed of 5 layers of thickness 5 cm
RDP = createReferenceProfile(FRNinv = c(46.92, 61.97, 7.72, 4.66, 10.21), thickness =
                             5, name = 'Reference site', falloutTime = 1963, refTime =
2025, massDepth = 177.62)
plot(RDP)
print(RDP)

# creates sampling layers of thickness 30 cm
P1 = createSamplingProfile(FRNinv = 110.35, thickness = 30, name = 'E4 Profile 1',
                           falloutTime = 1963, refTime = 2025)
P2 = createSamplingProfile(FRNinv = 91.57, thickness = 30, name = 'E4 Profile 2'
                           , falloutTime = 1963, refTime = 2025)
P3 = createSamplingProfile(FRNinv = 106.52, thickness = 30, name = 'E4 Profile 3',
                           falloutTime = 1963, refTime = 2025)
P4 = createSamplingProfile(FRNinv = 67.06, thickness = 30, name = 'E4 Profile 4',
                           falloutTime = 1963, refTime = 2025)

# creates smoothed layers below the reference profile
RDP_smooth1 = addSmoothedLayers(RDP, P1)
plot(RDP_smooth1, main = 'Simulated depth profile')
print(P1)

RDP_smooth2 = addSmoothedLayers(RDP, P2)
plot(RDP_smooth2, main = 'Simulated depth profile')
print(P2)

RDP_smooth3 = addSmoothedLayers(RDP, P3)
plot(RDP_smooth3, main = 'Simulated depth profile')
print(P3)

RDP_smooth4 = addSmoothedLayers(RDP, P4)
plot(RDP_smooth4, main = 'Simulated depth profile')
print(P4)

# sampling profiles with eroded layers
MODERN_P1 = MODERN(RDP_smooth1, P1)
plot(MODERN_P1)
print(MODERN_P1)
ER1 <-yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 145.10,
sampleDepth = 30)

show(ER1)

MODERN_P2 = MODERN(RDP_smooth2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 174.75,
sampleDepth = 30)

MODERN_P3 = MODERN(RDP_smooth3, P3)
plot(MODERN_P3)
print(MODERN_P3)
```

```

yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 170.30,
sampleDepth = 30)

MODERN_P4 = MODERN(RDP_smooth4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 211.05,
sampleDepth = 30)

#mit Deposition
RDP_depP1 = addDepositionLayers(RDP_smooth1, 110.35, 1)
MODERN_P1 = MODERN(RDP_depP1, P1)
plot(MODERN_P1)
print(MODERN_P1)
yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 145.10,
sampleDepth = 30)

RDP_depP2 = addDepositionLayers(RDP_smooth2, 91.57, 1)
MODERN_P2 = MODERN(RDP_depP2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 174.75,
sampleDepth = 30)

RDP_depP3 = addDepositionLayers(RDP_smooth3, 106.52, 1)
MODERN_P3 = MODERN(RDP_depP3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 170.30,
sampleDepth = 30)

RDP_depP4 = addDepositionLayers(RDP_smooth4, 67.06, 1)
MODERN_P4 = MODERN(RDP_depP4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 211.05,
sampleDepth = 30)

```

A5: Averaged profiles for sites E1-E4

```

# This script applies the MODERN model to averaged radionuclide inventories for each study
site (E1-E4), rather than individual profiles.
install.packages("path_to_modern_package", repos = NULL, type="source")
# load the library
library("modeRn")
# creates a reference profile composed of 5 layers of thickness 5 cm
RDP = createReferenceProfile(FRNinv = c(46.92, 61.97, 7.72, 4.66, 10.21), thickness =
5, name = 'Reference site', falloutTime = 1963, refTime =
2025, massDepth = 177.62)
plot(RDP)
print(RDP)

# creates sampling layers of thickness 30 cm
P1 = createSamplingProfile(FRNinv = 110.25, thickness = 30, name = 'E1',
falloutTime = 1963, refTime = 2025)
P2 = createSamplingProfile(FRNinv = 135.27, thickness = 30, name = 'E2'
, falloutTime = 1963, refTime = 2025)
P3 = createSamplingProfile(FRNinv = 110.05, thickness = 30, name = 'E3',

```

```

        falloutTime = 1963, refTime = 2025)
P4 = createSamplingProfile(FRNinv = 93.87, thickness = 30, name = 'E4',
        falloutTime = 1963, refTime = 2025)

# creates smoothed layers below the reference profile
RDP_smooth1 = addSmoothedLayers(RDP, P1)
plot(RDP_smooth1, main = 'Simulated depth profile')
print(P1)

RDP_smooth2 = addSmoothedLayers(RDP, P2)
plot(RDP_smooth2, main = 'Simulated depth profile')
print(P2)

RDP_smooth3 = addSmoothedLayers(RDP, P3)
plot(RDP_smooth3, main = 'Simulated depth profile')
print(P3)

RDP_smooth4 = addSmoothedLayers(RDP, P4)
plot(RDP_smooth4, main = 'Simulated depth profile')
print(P4)

# sampling profiles with eroded layers
MODERN_P1 = MODERN(RDP_smooth1, P1)
plot(MODERN_P1)
print(MODERN_P1)
ER1 <-yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 137.29,
sampleDepth = 30)

show(ER1)

MODERN_P2 = MODERN(RDP_smooth2, P2)
plot(MODERN_P2)
print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 132.6,
sampleDepth = 30)

MODERN_P3 = MODERN(RDP_smooth3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 108.69,
sampleDepth = 30)

MODERN_P4 = MODERN(RDP_smooth4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 175.30,
sampleDepth = 30)

#mit Deposition
RDP_depP1 = addDepositionLayers(RDP_smooth1, 110.25, 1)
MODERN_P1 = MODERN(RDP_depP1, P1)
plot(MODERN_P1)
print(MODERN_P1)
yearlyEDRates(MODERN_P1, samplingTime = 2025, falloutTime = 1963, massDepth = 137.29,
sampleDepth = 30)

RDP_depP2 = addDepositionLayers(RDP_smooth2, 135.27, 1)
MODERN_P2 = MODERN(RDP_depP2, P2)
plot(MODERN_P2)

```



```

print(MODERN_P2)
yearlyEDRates(MODERN_P2, samplingTime = 2025, falloutTime = 1963, massDepth = 132.6,
sampleDepth = 30)

RDP_depP3 = addDepositionLayers(RDP_smooth3, 110.05, 1)
MODERN_P3 = MODERN(RDP_depP3, P3)
plot(MODERN_P3)
print(MODERN_P3)
yearlyEDRates(MODERN_P3, samplingTime = 2025, falloutTime = 1963, massDepth = 108.69,
sampleDepth = 30)

RDP_depP4 = addDepositionLayers(RDP_smooth4, 93.87, 1)
MODERN_P4 = MODERN(RDP_depP4, P4)
plot(MODERN_P4)
print(MODERN_P4)
yearlyEDRates(MODERN_P4, samplingTime = 2025, falloutTime = 1963, massDepth = 175.30,
sampleDepth = 30)

```

Appendix B: Soil samples – Boxplot and Kruskal-Wallis test

```

# Load libraries
library(tidyr)
library(dplyr)
library(ggplot2)

# 1. Load soil data

Soil <- read.csv("path_to_data/Boxplot_soil.csv", sep = ";", header = TRUE)
head(Soil)

# 2. Reshape data to long format

Soil_long <- Soil %>%
  pivot_longer(
    cols = everything(),
    names_to = "Model",
    values_to = "Erosion"
  )

# Rename models
Soil_long$Model <- recode(
  Soil_long$Model,
  "modern" = "MODERN",
  "h0"     = "h0"
)

# Add an "All" category
Soil_all <- Soil_long %>% mutate(Model = "All")
Soil_long <- bind_rows(Soil_long, Soil_all)

# Set factor order for plotting
Soil_long$Model <- factor(
  Soil_long$Model,
  levels = c("P1", "P1.2", "P1.5", "h0", "MODERN", "All")
)

# 3. Summary statistics

```

```

stats <- Soil_long %>%
  group_by(Model) %>%
  summarise(
    mean = mean(Erosion, na.rm = TRUE),
    median = median(Erosion, na.rm = TRUE),
    .groups = "drop"
  )

# 4. Define colors

box_colors <- c(
  "P1"      = "#377eb8",
  "P1.2"    = "#377eb8",
  "P1.5"    = "#377eb8",
  "h0"      = "#4daf4a",
  "MODERN"  = "#984ea3",
  "All"     = "grey70"
)

# 5. Create the boxplot

ggplot(Soil_long, aes(x = Model, y = Erosion)) +
  geom_boxplot(aes(fill = Model), width = 0.6, outlier.shape = 16) +

  # Add mean lines
  geom_segment(data = stats,
    aes(x = as.numeric(Model) - 0.3,
        xend = as.numeric(Model) + 0.3,
        y = mean, yend = mean,
        color = "Mean"),
    linewidth = 1) +

  # Add median lines
  geom_segment(data = stats,
    aes(x = as.numeric(Model) - 0.3,
        xend = as.numeric(Model) + 0.3,
        y = median, yend = median,
        color = "Median"),
    linewidth = 1) +

  # Colors
  scale_fill_manual(values = box_colors, guide = "none") +
  scale_color_manual(values = c("Mean" = "black", "Median" = "red"), name = NULL) +

  # Labels
  labs(
    x = "Model",
    y = expression("Soil erosion [t ha"^{-1}*" yr"^{-1}*"]")
  ) +

  scale_x_discrete(labels = c(
    "P1"      = "P = 1.0",
    "P1.2"    = "P = 1.2",
    "P1.5"    = "P = 1.5",
    "h0"      = expression(h[0]),
    "MODERN"  = "MODERN",
    "All"     = "All"
  )) +

```

```

theme_classic(base_size = 28) +
theme(
  axis.title = element_text(size = 32),
  axis.text = element_text(size = 28),
  legend.text = element_text(size = 28)
)

# 6. Kruskal-Wallis test

Soil_test <- Soil_long %>% filter(Model != "All")
kruskal.test(Erosion ~ Model, data = Soil_test)

```

Appendix C: Sediment samples – Boxplot and Kruskal-Wallis test

```

# 1. Load river sediment data

River <- read.csv("path_to_data/Boxplot_river.csv", sep = ";", header = TRUE)

# Rename columns
colnames(River) <- c(
  "Q1_all",
  "Q2_all",
  "Q1/S2_S6",
  "Q2/S2_S6",
  "all",
  "all/S2_S6"
)

head(River)

# 2. Reshape data to long format

library(tidyr)
library(dplyr)
library(ggplot2)

River_long <- River %>%
  pivot_longer(
    cols = everything(),
    names_to = "Model",
    values_to = "Erosion"
  )

# Order for plotting
River_long$Model <- factor(
  River_long$Model,
  levels = c(
    "Q1_all",
    "Q1/S2_S6",
    "Q2_all",
    "Q2/S2_S6",
    "all",
    "all/S2_S6"
  )
)

# 3. Summary statistics

```

```

stats <- River_long %>%
  group_by(Model) %>%
  summarise(
    mean = mean(Erosion, na.rm = TRUE),
    median = median(Erosion, na.rm = TRUE),
    .groups = "drop"
  )

# 4. Boxplot colors

box_colors <- c(
  "Q1_all" = "#FF7F00",
  "Q1/S2_S6" = "#FF7F00",
  "Q2_all" = "#56B4E9",
  "Q2/S2_S6" = "#56B4E9",
  "all" = "grey70",
  "all/S2_S6" = "grey70"
)

# 5. Create the boxplot

ggplot(River_long, aes(x = Model, y = Erosion)) +

  geom_boxplot(aes(fill = Model), width = 0.6, outlier.shape = 16) +

  # Mean lines
  geom_segment(
    data = stats,
    aes(
      x = as.numeric(Model) - 0.3,
      xend = as.numeric(Model) + 0.3,
      y = mean,
      yend = mean,
      color = "Mean"
    ),
    linewidth = 1
  ) +

  # Median lines
  geom_segment(
    data = stats,
    aes(
      x = as.numeric(Model) - 0.3,
      xend = as.numeric(Model) + 0.3,
      y = median,
      yend = median,
      color = "Median"
    ),
    linewidth = 1
  ) +

  # Box colors
  scale_fill_manual(values = box_colors, guide = "none") +
  scale_color_manual(values = c("Mean" = "black", "Median" = "red"), name = NULL) +

  # Axis labels
  scale_x_discrete(labels = c(
    "Q1_all" = "Q1 All",
    "Q1/S2_S6" = "Q1\nwithout S2",

```

```

    "Q2_all"      = "Q2 All",
    "Q2/S2_S6"   = "Q2\nwithout S2",
    "all"        = "All",
    "all/S2_S6"  = "All\nwithout S2"
  )) +
  labs(
    x = "Model",
    y = expression("Erosion Rate [t ha"^{-1}*" yr"^{-1}*"]")
  ) +
  theme_classic(base_size = 28) +
  theme(
    axis.title = element_text(size = 32),
    axis.text  = element_text(size = 28),
    legend.text = element_text(size = 28)
  )
)

# 6. Statistical tests

# Shapiro-Wilk normality test
shapiro.test(River_long$Erosion)

# Kruskal-Wallis test
kruskal.test(Erosion ~ Model, data = River_long)

```

Appendix D: Comparison – Soil samples, sediment samples & monitoring data

```

# 1. Load combined data
Combined <- read.csv("path_to_data/Combined.csv", sep = ";", header = TRUE)
head(Combined)

# 2. Load libraries
library(tidyr)
library(dplyr)
library(ggplot2)

# 3. Prepare model data for boxplots
model_data <- Combined %>%
  select(modern, All, Q2.S1_S6, all_without_S1_S6) %>%
  pivot_longer(
    cols = everything(),
    names_to = "Model",
    values_to = "Erosion"
  )

# Map column names to readable labels
model_data$Model <- recode(model_data$Model,
  modern = "MODERN",
  All = "All Soil Samples",
  Q2.S1_S6 = "Q2 without S2,S6",
  all_without_S1_S6 = "All without S2,S6 Sediment Samples")

# 4. Prepare measured data (Suspended & Bedload)
measured_data <- data.frame(
  Model = c("Suspended Sediment", "Bedload"),
  Erosion = c(
    mean(Combined$Suspended, na.rm = TRUE),
    mean(Combined$Bedload, na.rm = TRUE)
  )
)

```

```

)

# 5% error bars
measured_data <- measured_data %>%
  mutate(
    ymin = Erosion * 0.95,
    ymax = Erosion * 1.05
  )

# 5. Levels for plotting
x_levels <- c(
  "MODERN",
  "All Soil Samples",
  "Q2 without S2,S6",
  "All without S2,S6 Sediment Samples",
  "Suspended Sediment",
  "Bedload"
)

model_data$Model <- factor(model_data$Model, levels = x_levels)
measured_data$Model <- factor(measured_data$Model, levels = x_levels)

# 6. Calculate mean and median for boxplots
stats <- model_data %>%
  group_by(Model) %>%
  summarise(
    mean = mean(Erosion, na.rm = TRUE),
    median = median(Erosion, na.rm = TRUE),
    .groups = "drop"
  )

# 7. Plot
ggplot() +
  # Boxplots for model data
  geom_boxplot(
    data = model_data,
    aes(x = Model, y = Erosion, fill = Model),
    width = 0.6,
    outlier.shape = 16
  ) +
  # Mean lines (black)
  geom_segment(
    data = stats,
    aes(
      x = as.numeric(Model) - 0.3,
      xend = as.numeric(Model) + 0.3,
      y = mean,
      yend = mean,
      color = "Mean"
    ),
    linewidth = 1
  ) +
  # Median lines (red)
  geom_segment(
    data = stats,
    aes(
      x = as.numeric(Model) - 0.3,
      xend = as.numeric(Model) + 0.3,
      y = median,

```

```

    yend = median,
    color = "Median"
  ),
  linewidth = 1
) +
# Points + error bars for measured data
geom_point(
  data = measured_data,
  aes(x = Model, y = Erosion),
  size = 3,
  color = "black"
) +
geom_errorbar(
  data = measured_data,
  aes(x = Model, ymin = ymin, ymax = ymax),
  width = 0.2,
  color = "black"
) +
# Fill colors for boxplots only
scale_fill_manual(
  values = c(
    "MODERN" = "#984ea3",
    "All Soil Samples" = "grey70",
    "Q2 without S2,S6" = "#56B4E9",
    "All without S2,S6 Sediment Samples" = "grey70",
    "Suspended Sediment" = "white",
    "Bedload" = "white"
  ),
  guide = "none"
) +
# Legend for mean/median
scale_color_manual(
  values = c("Mean" = "black", "Median" = "red"),
  name = NULL
) +
# Axis labels
scale_x_discrete(labels = c(
  "MODERN" = "MODERN",
  "All Soil Samples" = "All\nSoil Samples",
  "Q2 without S2,S6" = "Q2 without\nS2",
  "All without S2,S6 Sediment Samples" = "All without S2\nSediment Samples",
  "Suspended Sediment" = "Suspended\nSediment",
  "Bedload" = "Bedload"
)) +
labs(
  x = "Model / Data Source",
  y = expression("Erosion Rate [t ha"^-1*" yr"^-1*]")
) +
theme_classic(base_size = 18) +
theme(
  axis.title = element_text(size = 24),
  axis.text = element_text(size = 20),
  legend.text = element_text(size = 20),
  legend.title = element_text(size = 20)
)

# 8. Kruskal-Wallis test (model data only)
kruskal.test(Erosion ~ Model, data = model_data)

```

Declaration of Independence

I hereby declare that the submitted thesis is the result of my own, independent work. All external sources are explicitly acknowledged in the thesis.

DeepL Translator was used to support translation between German and English. ChatGPT-5 was used to assist with parts of the R coding, including error correction, and to improve the wording of some text written by me. All text parts assisted by AI have been checked by me, the author of this thesis.

M. Schumacher

Signature

Zurich, 22.04.2026

Place, Date