

From badland to bushland? Analysis of geomorphic process dynamics and vegetation development in a sub-humid *calanchi* area based on high-resolution UAS data (2014–2024)

Manuel Stark^{a,*}, Annalisa Sannino^{b,1}, Martin Trappe^a, Jakob Rom^a, Jakob Forster^a, Georgia Kahlenberg^a, Florian Haas^a, Francesca Vergari^b

^a Chair of Physical Geography, Catholic University Eichstaett-Ingolstadt, Eichstaett 85072, Germany

^b Sapienza University of Rome, 00185 Rome, Italy

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ABSTRACT

Badlands are among the most rapidly developing landscapes whose evolution is often controlled by threshold-driven geomorphic processes. This study employs quantitative, multi-temporal analysis to examine the spatio-temporal evolution (2014–2024) of a sub-humid *calanchi* badland in the upper *Val d'Orcia* (Italy). We investigate geomorphic dynamics over five consecutive two-year periods, focusing on the interplay between surface processes, vegetation development, and precipitation. Vegetation patterns were mapped semi-automatically from high-resolution orthomosaics (0.05 m), while geomorphic change detection analysis was carried out on DSMs of 0.25 m resolution. Despite stable mean slope gradients, specific sectors exhibited intensified geomorphic activity, designated 'geomorphic hotspots'. Annual erosion rates vary between -0.4 cm and -4 cm. The topographic evolution is primarily driven by irregular high-magnitude events. Water erosion showed a significant spatial reduction of 50% from 2014 to 2024. Furthermore, we document a geomorphic process not previously described in badlands: gravitational bulging, a subsurface sediment deformation induced by water infiltration, swelling of clays and overburden pressure. Vegetation establishes particularly in moist, low gradient depositional areas and evolves fragmented to a continuous structure. Although the erosional area decreased over the monitoring period, erosion rates remained relatively constant, suggesting a transition from high-frequency to high-magnitude, event-driven processes during the observation period. This behaviour suggests that the system operates close to a geomorphic threshold, where intrinsic dynamics, rather than changes in mean climatic conditions, control landscape evolution. Overall, the period 2014 to 2024 represents a phase of progressive badland stabilisation controlled by intrinsic geomorphic dynamics under a normal precipitation regime.

1. Introduction

Soil degradation and the loss of fertile soil are among the most critical environmental challenges worldwide (Poesen, 2018; Quinton and Fiener, 2024). These processes are particularly severe in regions with a long history of human activity, such as areas affected by intensive agriculture or overgrazing, where badlands have developed.

Badlands are among the landscapes with the highest soil erosion rates globally and form striking erosional morphologies (Nadal-Romero and García-Ruiz, 2018). They occur across a wide range of climatic zones (Ballesteros-Cánovas et al., 2017; Boardman et al., 2003; Nadal-

Romero et al., 2011; Kaiser et al., 2014; Zhang et al., 2016), providing valuable natural laboratories for studying landscape evolution (Yang et al., 2021). Depending on annual precipitation, they are commonly classified into three groups: (1) arid badlands (<200 mm yr⁻¹), (2) semi-arid badlands (200–700 mm yr⁻¹), and (3) humid badlands (>700 mm yr⁻¹) (Gallart et al., 2002). In Mediterranean regions, where strong seasonal contrasts in rainfall occur, water availability and denudation processes jointly control landscape stability (Caraballo-Arias et al., 2015). In Italy, the terms *calanchi* and *biancane* denote two morphologically distinct types of badlands (Del Monte, 2017). *Calanchi* (singular: *calanco*) develop on steep clayey slopes, often capped by resistant

* Corresponding author.

E-mail address: mstark@ku.de (M. Stark).

¹ The first two authors contributed equally to this work.

layers, and are characterised by dense networks of gullies and sharp ridges. *Biancane* (singular: *biancana*), by contrast, form rounded clay hills up to about 20 m high, with bare, sun-exposed slopes subject to intense sheet and rill erosion.

In central Italy, erosion processes mainly occur during humid months (spring and autumn), with both rainfall intensity and distribution controlling geomorphic activity (Aucelli et al., 2016; Della Seta et al., 2009; Neugirg et al., 2016). Heavy rainfall events trigger rapid runoff and the formation of gullies, rills, and pipes, whereas low-intensity but persistent rainfall can favour deposition, producing a “cut-and-fill” behaviour driven by rainfall sequence rather than climate trends (Faulkner, 2008). Beyond linear erosion, badlands also experience mass movements such as translational and rotational landslides (Ciccacci et al., 2008). Moreover, during extreme rainfall events enhanced stream power may completely remove depositional signatures, as well as vegetation. In contrast, during periods of low intensity but persistent rainfall, deposition may outweigh erosion (Neugirg et al., 2016). Located in the upper catchments, badlands typically act as major runoff sources and sediment suppliers. They are also referred to as ‘geomorphic hotspots’ (GHS) within geomorphic systems whose connectivity varies with precipitation and vegetation cover (Della Seta et al., 2009; Cantón et al., 2018; Faulkner, 2008).

The rapid evolution of remote sensing techniques has significantly improved our understanding of badland dynamics (Colacicco et al., 2025). Methods such as terrestrial and airborne LiDAR, structure-from-motion (SfM-MVS) photogrammetry from UAS imagery, historical aerial photography, and satellite data have been widely applied to monitor topographic changes (e.g. Stark et al., 2020; Llena and Vericat, 2023; Palmeri et al., 2024; Sannino et al., 2026). Among these, UAS-based SfM-MVS stands out for its cost-effectiveness and flexibility (Eltner and Schneider, 2015), enabling detailed quantification of geomorphic processes even in inaccessible areas such as badlands (Neugirg et al., 2016; Zhang et al., 2025).

Vegetation plays a complex role in controlling erosion processes. It can mitigate splash erosion, enhance soil infiltration capacity, and stabilize surface horizons (Marignani et al., 2008; Vergari et al., 2013), yet under certain conditions it may also facilitate shallow landslides or piping (Khairulmaini, 1996; Romero-Díaz et al., 2007; Della Seta et al., 2009). In Mediterranean badlands, vegetation dynamics are highly variable and tightly linked to rainfall patterns and substrate characteristics (Pulice et al., 2012; Bollati et al., 2016). Despite extensive research, the interaction between geomorphic processes and vegetation development in sub-humid *calanchi* badlands remains poorly understood.

Several indices have been developed to monitor vegetation dynamics (Xue and Su, 2017). Many of these indices are calculated from RGB-band ratios, providing a cost-effective method for evaluating spatial and temporal vegetation distribution using consumer-grade UAS data. One such index is the Triangular Greenness Index (TGI), a remote sensing metric used to estimate leaf chlorophyll content and nitrogen status in various crops, derived from the spectral reflection of specific wavelengths within the visible spectrum (Hunt et al., 2013). Because of its strong correlation with chlorophyll content, it can be used for comparison with the more widely used and established multispectral indices (Fuentes-Peñailillo et al., 2018). To date, the TGI has been used primarily in agricultural, ecological and forestry studies (e.g. Lussem et al., 2018; Starý et al., 2020; Zeng et al., 2022; Kandpal and Kumar, 2023). Given the simplicity of data acquisition and index calculation, we see great potential for applying the TGI in geomorphological research as well. This is particularly relevant in hard-to-access landscapes, such as badlands, where UAS data can be used to simultaneously reconstruct topography and calculate the TGI.

Taking advantage of a medium-term, high-resolution topographic dataset spanning 10 years, a relatively long period for UAV-based photogrammetric monitoring, this study aims to investigate the morphoevolution of a representative sub-humid *calanchi* badland located in

central Italy, in relation to possible variation in vegetation and precipitation dynamics over a ten-year period (2014–2024), based on a two-year interval (and one four-year interval). High-resolution UAS-RGB imagery and ground control points (GCPs) were used to generate digital surface models (DSMs) and digital orthomosaics, enabling both geomorphic change detection and vegetation analysis through the Triangular Greenness Index (TGI).

Specifically, this study aims to:

1. Quantify the magnitude and spatial distribution of geomorphic changes in a sub-humid *calanchi* badland over the last decade.
2. Analyse how variations in rainfall temporal patterns and extreme precipitation events influence erosion and deposition processes.
3. Assess the relationship between vegetation cover dynamics and geomorphic activity using RGB-based vegetation indices (TGI).
4. Evaluate the potential of UAS-based SfM-MVS and TGI for monitoring landscape evolution in inaccessible badland environments.

To achieve these objectives, the study followed five main steps:

- (i) geomorphological change detection through DoDs;
- (ii) mapping of geomorphic process signatures;
- (iii) identification of GHS;
- (iv) analysis of rainfall and extreme events; and
- (v) application of the TGI for semi-automatic vegetation mapping.

2. Study site

The study site is part of the *Scalonca* sub-basin (0.10 km²), located in the upper headwaters of the *Val d'Orcia* (Tuscany, Italy) between 501 and 651 m a.s.l. It represents a valuable study site for the investigation of geomorphic processes in *calanchi* badlands. The Upper Orcia Valley has undergone a strong land reclamation during the last century that heavily transformed the landscape (Brandolini et al., 2018; Vergari et al., 2025). Approximately 22% of the *Scalonca* basin has been levelled and allocated for agricultural use or reforestation. In 1975, the area surrounding the main *calanchi* heads was reforested with conifers (*Cypressus arizonica*, *Cypressus sempervirens*, *Pinus nigra* and *Fraxinus ornus*), covering an area of approximately 0.1 km² (Bazzoffi et al., 1997). Despite erosion control measures such as slope levelling, transverse wooden structures and check dams (Colica, 2000), the *calanchi* headcuts and slopes continues to show a high level of geomorphological activity.

Here, erosion rates were previously derived using a variety of methods at different spatial and temporal scales in numerous studies like Della Seta et al. (2007 & 2009), Ciccacci et al. (2008), Vergari et al. (2013), Aucelli et al. (2016), Neugirg et al. (2016) and Stark et al. (2020). The site offers unique opportunities to examine both the recent geomorphodynamics and the long-term effects of erosion and deposition activity. The topography of the area is defined by steep and sparsely vegetated slopes (Fig. 1) with a mean inclination of 14° in the main valley and 27° in the tributaries. High slope gradients and sparsely vegetated surfaces consisting of unconsolidated marine and fluvial sediments represent the ideal conditions for the rapid geomorphological development (Nadal-Romero et al., 2007) of the *Val d'Orcia* badlands.

From a geologic perspective, the study site was defined by the Late Miocene-Quaternary extensional tectonic phase, which led to the formation of graben structures, such as the *Radicofani* Graben. During the Pliocene and Pleistocene, this tectonic depression was progressively filled with marine and continental deposits, primarily clays, sands and conglomerates (Pascucci et al., 2006). A subsequent phase of regional uplift, associated with large-scale crustal arching, caused these Plio-Pleistocene sediments to be elevated well above present sea level. Currently, these deposits crop out at elevations of up to 900 m a.s.l., particularly near the *Mt. Amiata*, *Radicofani* volcanic neck and *Mt. Cetona*, in the southern sector of the *Radicofani* Graben, where the uplift has been most intense (Vergari et al., 2013). The *Scalonca* catchment

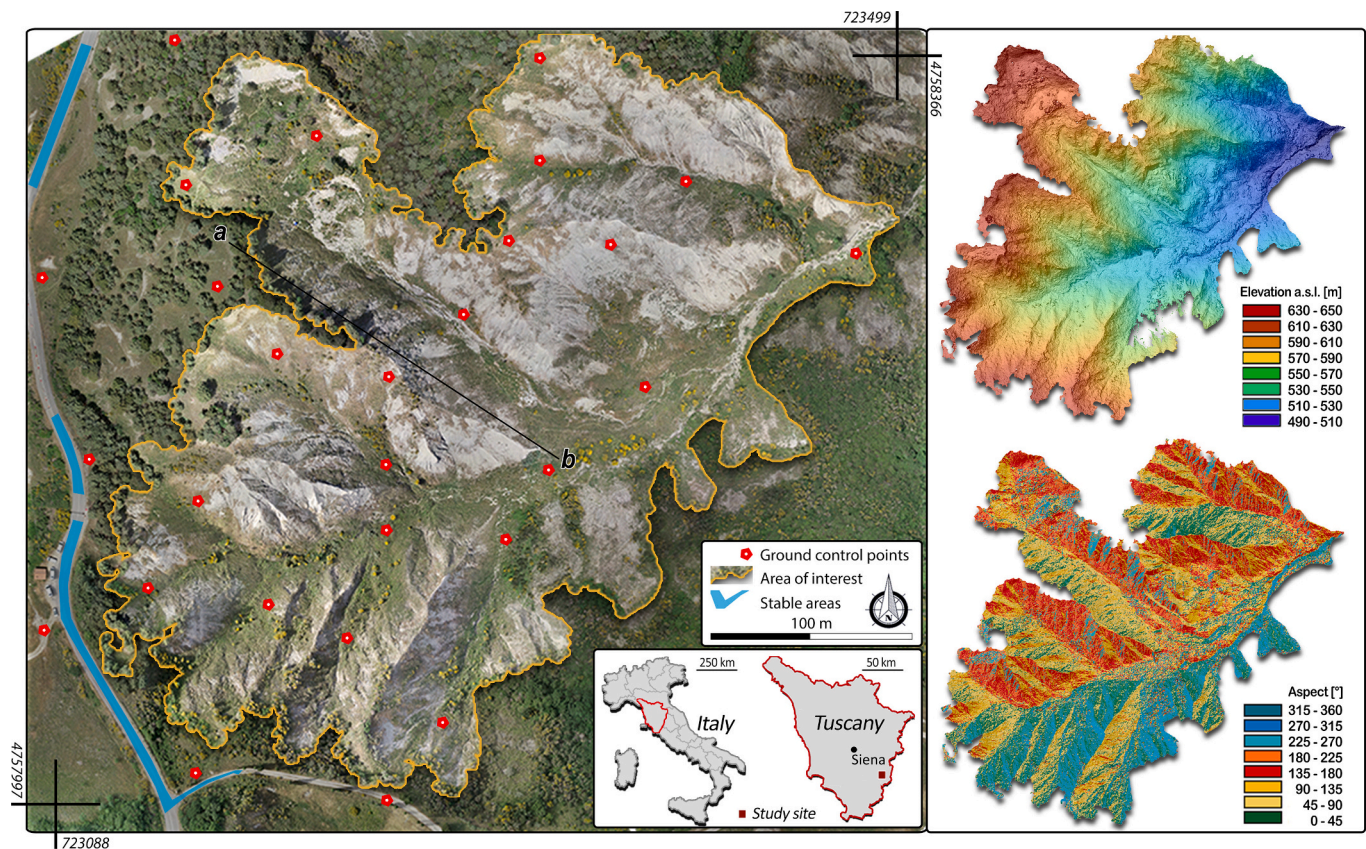


Fig. 1. Left: Study site with location of the geological profile of Fig. 12 (black line a-b). The GCPs of all surveys are located within 5 m of the red dots. Additional GCPs outside the area shown here were used for the 2014 and 2016 surveys. Right: Distribution of slope (top) and aspects (bottom).

features alternating marine claystones, silty claystones, siltstones and sandstones with some intercalated gravelly deposits with a clayey to silty matrix (referred to as pudding lenses by Vergari et al., 2011). These rocks are inclined to the west (15–20° W - SW) at the western flank of the main anticline of the Radicofani basin (Acocella et al., 2002). According to Pascucci et al. (2006) these rocks belong stratigraphically to the *Globorotalia margaritae* to *Gl. punctulata* zone of the Early Pliocene (Zanclean). The gravelly deposits were derived from the southeastern or western flank of the Radicofani basin, they were transported by sub-aqueous gravity flows into the offshore clays in the central basin (Pascucci et al., 2006).

The distinctive landscapes and its geomorphological appearance (e. g. leaning electricity pylons, cracks in roads) indicate a rapid evolution of the landform and a high degree of ongoing geomorphological processes. Annual precipitation is highly variable, with a range of 500–1100 mm (Aucelli et al., 2016) and a mean of 700 mm (Brandolini et al., 2018). Precipitation is concentrated in autumn and winter with a maximum in November and a second smaller peak in spring. November represents the month with the largest number of consecutive rainy days (Vergari et al., 2013). Conversely, the precipitation minimum is in July due to the hot and dry summer (Aucelli et al., 2016). These pronounced semi-humid conditions with dryness during the summer months followed by heavy rainfall in autumn provide ideal conditions for severe erosion by water on the steep slopes of the *calanchi* (Della Seta et al., 2009). In spring, on the other hand, further precipitation often triggers landslides on the waterlogged soils (Neugirg et al., 2016).

3. Material & methods

3.1. Data survey and processing

During a ten-year period, five UAS surveys were conducted over the

Scaltona sub-basin, in order to assess morphological changes, identify erosion processes and process areas with enhanced geomorphic activity (GHS).

3.1.1. Geological and laboratory methods

For the upper *Scaltona* catchment individual rock types (claystone-sandstone sequences, siltstone-sandstone sequences and gravelly deposits) were mapped to reconstruct the relations of their occurrences to different mass movements. Additionally, measurements of strike and dip of the layers were conducted as well as for some selected joints.

A parent material sampling ($n = 88$) was included to support the surficial geological mapping. Grain sizes were analysed by sieving (gravel, sand) and using the pipette analysis for the fines. Additionally, to estimate selected petrophysical properties (hydraulic conductivity, water absorption capacity, swelling capacity) further samples ($n = 18$ – see Table S4) were taken by core samplers in the field. The hydraulic conductivity was estimated by using the falling-head-method (DIN, 1989 - 18130-1). From simple swelling tests after Hiltmann et al., 1998 the maximum swelling capacity was calculated after treating the core samples by excess water.

3.1.2. UAS and dGNSS survey

The imagery acquisitions were carried out with four different UAS between May 2014 and May 2024 (Table S1). The utilisation of appropriate flight patterns in conjunction with ground control data (GNSS) (Table S2) facilitates the processing of UAS data into highly accurate (10 m^{-2}) and high-resolution DSM (0.25 m) (Stark et al., 2021). Therefore, we specified high image overlap ratios of $>80\%$ and ensured that the maximum flight altitude was limited to below 120 m above ground. This resulted in variable ground sample distances (GSD), ranging from 1 to 5 cm/pixel, dependent on the survey design (flight height and pattern) and type of UAS. Metal reflector discs with a

diameter of 10 cm, placed on red carpets of 40 × 40 cm, were distributed throughout the study site and used as GCPs. We aimed for a consistent distribution of GCPs across all years (see Fig. 1), paying particular attention to their distribution across a wide range of altitudes and cardinal points. The GCPs were measured with a Leica GS09 differential global navigation satellite system (dGNSS) (2014 & 2016) and a *Stonex S9i* dGNSS (2018, 2022, 2024) in rover-base mode with an accuracy of <2.5 cm horizontally and < 4 cm vertically.

3.1.3. Image analysis (SfM-MVS)

All image data were processed with Metashape Pro (Agisoft LLT, 2023, Version 2.0.1). The GCPs were used to improve the scaling and external orientation of the data (UTM zone 32 N; EPSG 25832). In order to exclude tie-points with low accuracy and to minimise the influence of erroneous tie-points, the point clouds were filtered on the criteria and thresholds given in Fig. 2. Between each filtering step, we applied lens parameter optimization based on Brown's camera model (Duane, 1971), which compensates for most linear and nonlinear forms of lens distortions (Aguilar et al., 2013). Following the extensive refinement of the point clouds, which included the removal of outliers, poorly projected and reconstructed points, and larger vegetation such as trees and tall bushes, the point densities of the individual data sets varied between 302 (2016) and 284 (2024) points/m². Next, we used a 3D point distance filter to achieve a homogeneous point density of 0.05 m across all data sets. To minimise persisting systematic discrepancies (Sanz-Ablanedo et al., 2020) and to correct positional offsets we applied the iterative closest point algorithm method (ICP) to all point clouds (reference: 2024-point cloud) in the software SAGA LIS 9.6.1 (Besl and McKay, 1992). To ensure a robust adjustment, pre-computed normal vectors were employed for each point and only areas with similar orientations were considered for the error minimisation. To exclude geomorphological active areas from the co-registration process, the maximum 3D distance between two corresponding points was set to 0.25 m. Consequently, only those points situated at a distance below 0.25 cm distance, designated as “geomorphologically stable areas”, are

considered for the co-registration. Furthermore, for each data set, the Digital Orthophoto Maps (DOM) were calculated with a pixel size of 0.05 m. The co-registered point clouds were then converted into 2.5D rasterised DSMs using a grid resolution of 0.25 m. This grid size was chosen to ensure a sufficient number of points per cell (>20), given an average point spacing of 0.05 m. Within each grid cell, point elevations were aggregated using the minimum z-value. This approach reduces the influence of residual vegetation points (i.e., scattered points remaining after photogrammetric filtering of RGB imagery) and provides a more reliable approximation of the underlying ground surface. The DSMs are used to calculate the DoDs for the respective periods while the DOMs form the basis for mapping geomorphological process signatures and calculating the TGI.

3.1.4. TGI computation

The DOMs form the basis of the TGI calculation. The TGI was calculated in SAGA GIS (Laserdata; version 9.6.1) in accordance with the following formula:

$$\text{TGI} = G - 0.61 \cdot R - 0.39 \cdot B$$

with R = red colour channel, G = green colour channel and B = blue colour channel. Generally, vegetation pixels are characterised by higher TGI values than bare ground areas. Suitable thresholds for vegetation segmentation are determined based on index values and comparison with true colour information of the DOM. In the two-stage process, the boundary between vegetated and non-vegetated areas is first defined, and the optimal threshold value is then determined within this boundary. In a second step, the value range is narrowed down further by analysing the individual pixel values, and a suitable threshold value is determined for each data set. The thresholds vary only slightly between 14 and 16. The TGI grids are then processed into binary data sets based on these thresholds (0 = no vegetation; 1 = vegetation).

3.1.5. Precipitation

To assess the influence of different types of precipitation on the

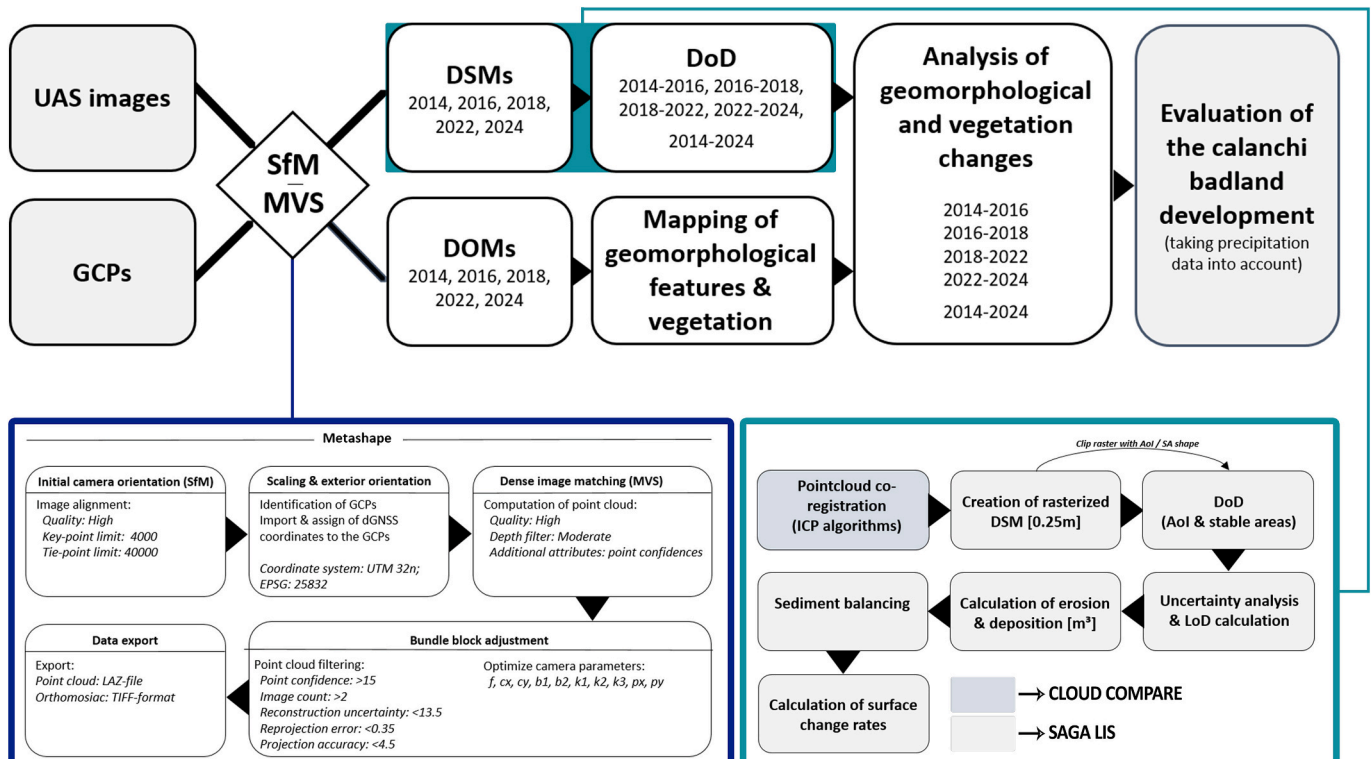


Fig. 2. Graphical presentation of the general workflow (top), SfM-MVS processing (bottom, left) and change detection analysis (bottom, right).

erosion dynamics, we used precipitation data from the *Radicefani* weather station, operated by the Regional Hydrological Service (*Regione Toscana – Servizio Idrologico Regionale (SIR)*, 2025). The station is located approximately 6 km southwest of the study area, and it provides data with an original temporal resolution of 15 min.

3.2. Data analysis

3.2.1. DSM accuracy

In order to distinguish and exclude model errors (due to random noise) from real surface changes and, particularly important in this study, to minimise the influence of unfiltered ground vegetation on the sediment balancing, a threshold value was calculated for each DoD and an uncertainty value, further referred to as *level of detection* (LoD), was defined as follows:

$$\sigma_{\text{DoD}} = \sum \sigma_{\text{StableAreas}}$$

where σ_{DoD} represent the mean of standard deviations (StD) of the height values of all stable area cells. Only areas that showed no observable changes during the entire study period, so called ‘morphologically unchanged/stable surfaces’ (Bennett et al., 2012; Westaway et al., 2000), were considered for the uncertainty assessment of the DSMs. Due to the high geomorphological activity in the entire study site and beyond, those areas are very limited. Only a section of the road in the western part of the area of interest (AoI) and some isolated flat areas in the uppermost parts of the study site could be considered as stable areas (see Fig. 1). We used a significance threshold at a critical t-value (t_{crit}) of 1.96 from Student's t-distribution, corresponding to a confidence level of 95%. The LoD was calculated based on the following equation:

$$\text{LoD} = t_{\text{crit}} \cdot \sigma_{\text{DoD}}$$

3.2.2. Change detection analysis

3.2.2.1. Geomorphological dynamics. After the accuracy analysis the DSMs recorded at two different points in time are utilised to calculate the height (surface) changes (DoD) and subsequently the DoDs are used to calculate and analyse the spatial distribution of positive and negative surface and volume changes (deposition, erosion, and net volume) during the respective observation periods. The volume of the eroded and deposited substrate is calculated according to:

$$V_{\text{change}} = \sum_{i=1}^n \text{DoD}_i \cdot A_{\text{pixel}}$$

with n = total number of raster cells; A_{pixel} = raster cell area. The annual surface vertical change rates ($\text{Change}_{\text{an}}$) are calculated according to:

$$\text{Change}_{\text{an}} = (V_{\text{change}} / nA_{\text{pixel}}) / \text{number of years}$$

with V_{change} = cumulated volume of cell changes.

3.2.2.2. Mapping of geomorphic processes signatures. The georeferencing of DOMs (see Section 3.1.4) resulted in a RMSE below 0.2 m for all registrations. The mapping of geomorphic process signatures was performed in QGIS by manually delineating polygons for the same interval period reported above. Beside the orthomosaics several terrain parameters were used to support the mapping, (e.g. slope, roughness, flow accumulation). The t0 mapping is based on the interpretation of the 2014 orthomosaic and topographical information from the DSM. It thus shows the cumulative effects of various geomorphological processes that were active prior to the first survey and whose traces are still visible in the terrain.

To interpret the process activity between two survey dates (t_1 and t_2), information from the DoD ($t_2 - t_1$) was used in addition to the above-

mentioned parameters. To analyse the geomorphological development of the area in terms of process dynamics, the process signatures and associated landforms were first mapped. In a subsequent step, the processes were categorised according to their respective inferred causes (Table 1). The proportions of the mapped process signatures/landforms were then calculated relatively to the total basin area.

Some exceptions were made for the i) channel-type debris flow, ii) gravitational bulging, iii) sheet wash/ rilling and iv) gullying, due to the nature of the processes. The considerable spatial overlap between process types often hinders a clear differentiation/ identification of the geomorphological processes responsible. In the case of debris flow-like processes, the deposits are typically subjected to repeated mobilisation and relocation (e.g. by sheet wash/rilling), resulting in secondary deposition and erosion areas. Bulging represents a deformation of the slope and the landforms associated with it represent in-place bulging rather than mobilized materials. The colluvial areas for sheet-wash and rill-erosion were combined due to the small spatial scale of the signatures and the mutual overlap of the process areas of both process types.

3.2.2.3. Vegetation dynamics. Following the creation of vegetation maps, the respective absolute and relative share of vegetation-covered areas was calculated relative to the size of the study area and the percentage change in vegetation cover within the individual observation periods was determined.

The change in vegetation was then derived from the discrepancy between the vegetation cover present at the initial grid (DOM t_1) and the final grid (DOM t_2). Further, areas permanently covered with vegetation are determined by with the formula:

$$V_s = a \times b \times c \times d \times e$$

with $a = 2014$, $b = 2016$, $c = 2028$, $d = 2022$ and $e = 2024$. Cells with $V_s = 1$ correspond to a constant vegetation cover throughout the entire observation period. If a cell in one of the datasets shows the value 0 (=no vegetation), this results in $V_s = 0$, i.e. non-constant vegetation cover.

Next, we examined the changes in land cover, with particular emphasis on the interaction between geomorphological processes and vegetation. Therefore, the two types of maps (vegetation and geomorphological process signatures) were intersected in order to represent the spatial development of vegetation cover and geomorphological process areas, as well as their interaction over the respective time steps.

3.2.3. Precipitation data

Two main types of rainfall parameters, known for their importance for erosion in badlands, were analysed: i) total rainfall sums as a first indicator for rainfall dynamics over a specific period and ii) threshold exceedances as trigger for mass movements (e.g. Peruccacci et al., 2017; Mathys et al., 2003). The following thresholds were used to analyse different types of events (A, B and C): A = 5 mm / h, according to Mathys et al. (2003); B = 5 mm / 15 min for high-intensity short-duration events; C = 60 mm / 5 days representing long-term rainfall events as a potential trigger for denudation processes (Della Seta et al., 2007). Rainfall totals must exceed 5 mm to be counted as an event and between

Table 1

Inventory of mapped geomorphological process types and landforms.

Origin criteria	Process type	Process landforms	
Gravity-driven	Channel type debris flow		
	Gravitational bulging		
	Complex landslide	Scarp	Body
	Translational landslide		
	Rotational landslide		
Water erosion	Sheet wash/ rilling	Sheet wash/ rill erosion areas	Colluvium
	Gullying	Gully erosion areas	

two events, there must be at least a hiatus of 6 h without rainfall. As a combination of intensity and duration, we also established an intensity-duration relationship (ID) (type D) like in e.g. Giannecchini et al. (2012). For the ID relationship (Fig. 3), potentially erosive rainfall events are defined in accordance to Savi et al. (2021) (ID >75% percentile). All potentially erosive events (type A, B, C, D) are analysed for each observation period.

4. Results

4.1. Change detection and geomorphological development

4.1.1. Volumetric assessment of basin-scale net topographic changes

The results of DSMs accuracy analysis are reported in the Table 2. The maximum StD in the stable areas is 0.12 m, while all other datasets show a lower StD, yielding LoDs between 0.08 m and 0.23 m. The thresholded data consistently indicate negative balances, whereas the positive raw values can be partly attributed to imperfect filtration in areas affected by substantial vegetation growth.

The annual volumetric balance varies considerably, ranging from -378 m^3 to -4074 m^3 , with an average annual loss of approximately 736 m^3 over the entire study period (Table 3). Negative changes and absolute volume changes have shown an increase in the second observation period, resulting in higher negative annual balances. The annual mean surface change rates range from -0.4 cm (2018–2022) to -3.9 cm (2022–2024) with a mean value of -0.8 cm for the total observation period of 2014 to 2024. It is important to note that these values represent an hypothetical homogenous average rate of erosion. No consistent trend, neither downward nor upward, can be discerned over time.

4.1.2. Multitemporal geomorphological mapping and spatially distributed topographic changes

The catchment-wide multitemporal analysis allowed the characterization of geomorphological evolution (Fig. 4), both in terms of changes in the magnitude of geomorphic processes and the identification of four GHS (Fig. 5). Over the observation periods, the basin evolved from an initial phase of widespread erosion and hillslope instability (2014–2018) to an intermediate phase of stabilisation (2018–2022), followed by a recent phase of localized instability and complex processes (2022–2024). Throughout the period of monitoring, a marked decrease

Table 2

Stable area DoD values.

DoD	Min. [m]	Max. [m]	Mean [m]	StD. [m]	LoD [m]
2014–2016	−0.294	0.139	−0.118	0.091	0.18
2016–2018	−0.193	0.247	0.161	0.116	0.23
2018–2022	−0.260	0.183	−0.086	0.084	0.16
2022–2024	−0.141	0.238	0.019	0.079	0.15
2014–2024	−0.1	0.122	−0.046	0.043	0.08

in geomorphic active area of approximately 42% was observed (Table 4). Water erosion processes showed the greatest spatial and temporal consistency, whereas processes caused by gravity occurred much more scattered and less frequently. The 2014 (T_0) geomorphological map of Fig. 6a shows the morphological conditions of the study area at the beginning of the monitoring period. At the beginning of the observation period, the catchment was characterised by extensive bare soil, reflecting the cumulative impact of prolonged geomorphological activity. Gravity-driven processes, mainly channel-type debris flows, were dominant and widespread across badland slopes and under translational landslide scarps. Evidence of an old rotational landslide, active since the 1990s (Ciccacci et al., 2003; Stark et al., 2021), was also identified. Surface runoff processes, such as rill erosion, were common, while colluvial deposits were limited.

Henceforth, the focus will be on the effects of dynamic processes that occur during the two-year intervals between data surveys.

Between T_0 and 2016, geomorphological activity slightly decreased, with a clear reduction in debris-flow areas (Fig. 6b). The changes are attributable to a medium-depth translational landslide originating in the uppermost side valley and deposits extending into the main valley. The geomorphological development of the main valley is strongly influenced by sediment input from the tributaries to the northwest (GHS 1 and GHS 2 in Fig. 7), mainly in the form of debris flow-like processes. The main valley shows signatures of strong channel incision directly below the area of deposition, with maximum negative surface changes of -4.47 m over a distance of 220 m (GHS 4). It is striking that the linear incision signatures are interrupted by the transversal patterns related to soil piping processes. This assumption was confirmed by several field inspections. A substantial rotational landslide, originated in the 1990s (Stark et al., 2020), is present in northernmost part of the *calanchi* badlands. Even though there are currently no active rotational movements, secondary geomorphological processes continue to shape and change this area to this day. In addition to linear incisions of varying sizes, ranging from rills to gullies, the northern and north-eastern slopes show signs of minor gravitational mass movements, mostly in the form of translational slides. Most of the negative surface changes are caused by these landslides with magnitudes of 0.1 to 0.4 m in depth.

Between 2016 and 2018, the proportion of bare substrate remained relatively stable (Fig. 6c). Areas affected by water erosion continued to account for the largest share of the total area. The DoD shows minor but widespread negative changes, especially in the southern region of the badland. In the main valley, gully processes were particularly active, causing negative changes of up to -1.7 m alternating with positive changes of up to 1.85 m. The latter were primarily caused by sediment deposits from debris flow-like processes emanating from the side valley of the rotational landslide. In the upper segment of the rotational landslide, signs of ongoing erosion processes are now evident in the form of small-scale translational landslides. The main valley is subsequently filled with easily erodible sediments, which in turn are further shaped by linear erosion processes. As in the previous observation period, the upper parts of the vegetated north-facing slopes show signatures of smaller landslides with deposits at the foot of the side valleys. A detailed examination of the 2018 DSM reveals extensive piping processes along the main valley floor, with maximum surface changes of -1.66 m at the pipe outlets.

The most significant landscape changes occurred in the four-year

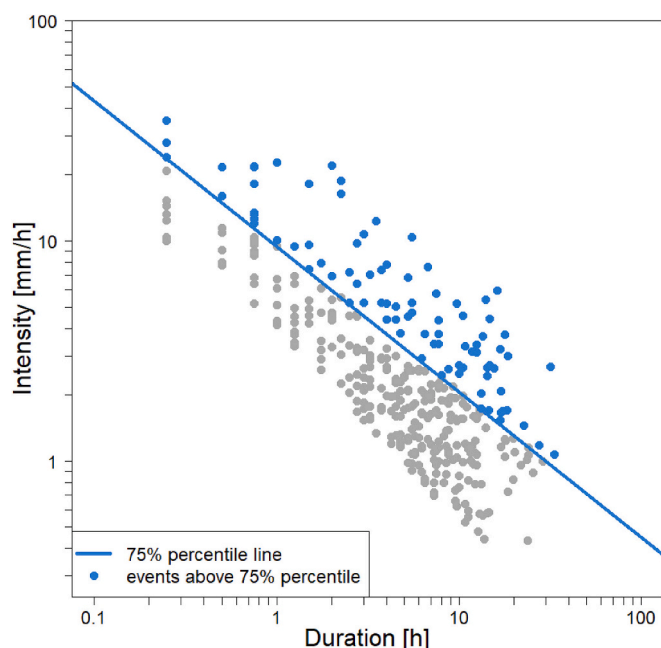


Fig. 3. ID relationship for rainfall data between 2014 and 2024.

Table 3
Results of geomorphic change detection for all periods. It is important to note that the DoD 2018–2022 encompasses a four-year period, whereas all other DoDs comprise approximately two years, with the exception of the DoD 2014–2024, which presents the total alterations over the entire observation period.

DoD		Negative change	Positive change	Overall balance	Balance (annual)	Annual mean surface change rate
		[m ³]				[m]
2014–2016	raw	7689	6242	−1447	−723.5	−0.007
	thresholded	4189	2161	−2161	−1081	−0.010
2016–2018	raw	17,990	3864	−14,126	−7063	−0.067
	thresholded	5754	2078	−3676	−1838	−0.018
2018–2022	raw	7190	17,620	10,430	2607.5	0.025
	thresholded	5170	3660	−1510	−378	−0.004
2022–2024	raw	16,527	8445	−8082	−4041	−0.039
	thresholded	16,557	8409	−8148	−4074	−0.039
2014–2024	raw	28,478	16,517	−11,961	−1196.1	−0.011
	thresholded	1300	5040	−7960	−796	−0.008

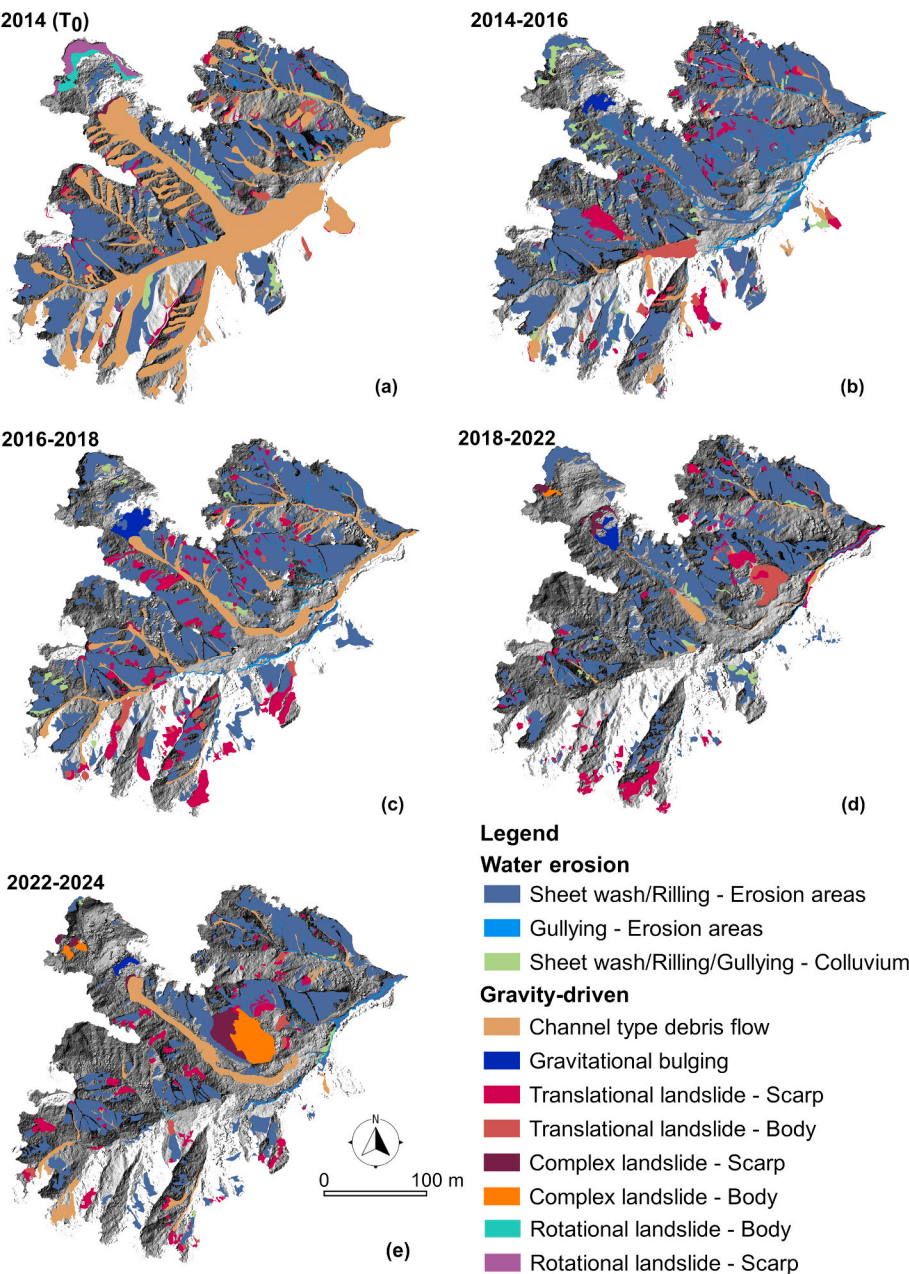


Fig. 4. Multitemporal mapping of geomorphic process signatures and associated landforms from T0 (2014) to 2024. The 2014 map depicts the geomorphological conditions prior to monitoring, whereas the subsequent maps illustrate the effects of active processes during each observation period, derived from pre- and post-event orthophotomosaics and DSMs of Difference (DoDs) analysis.

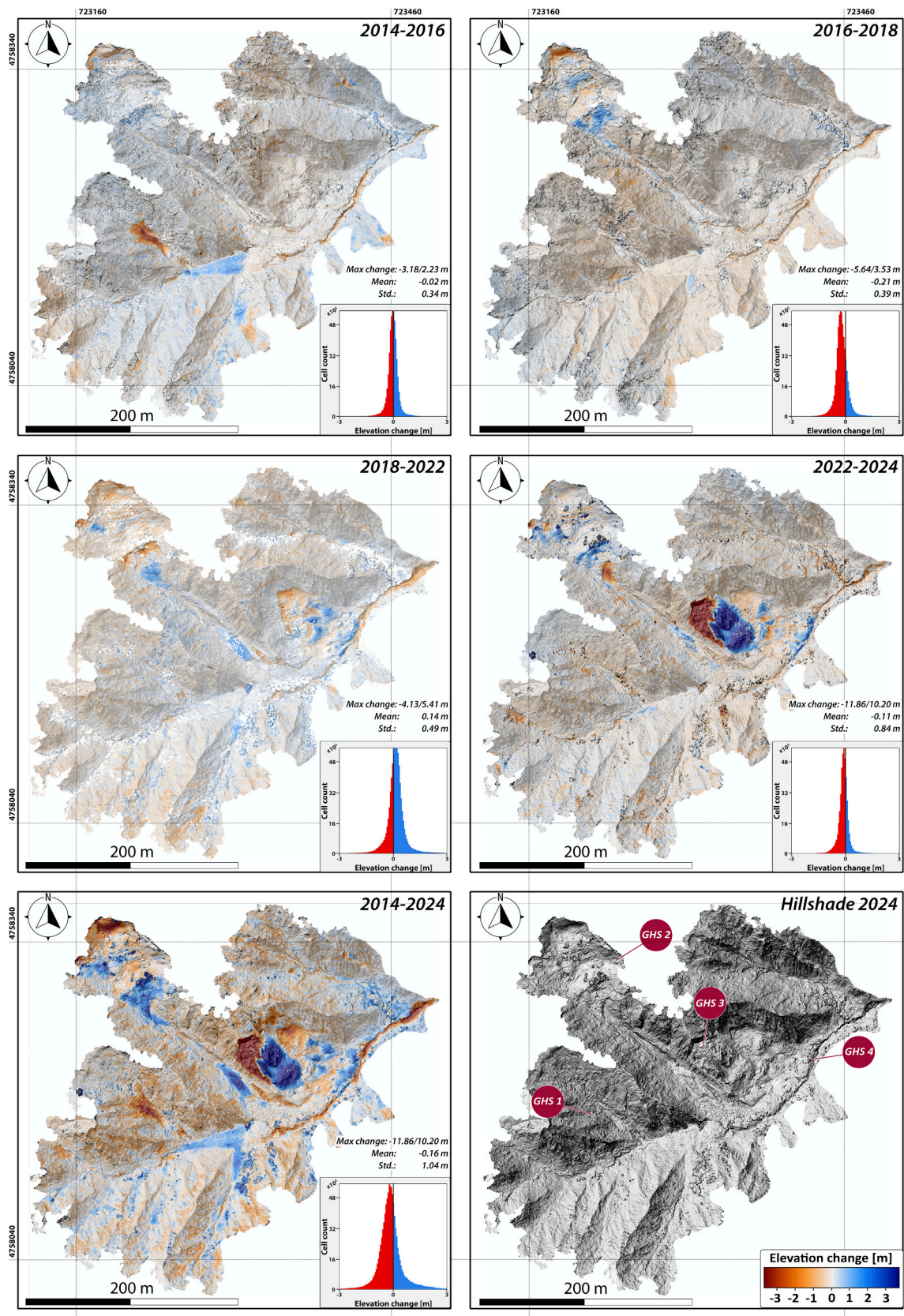
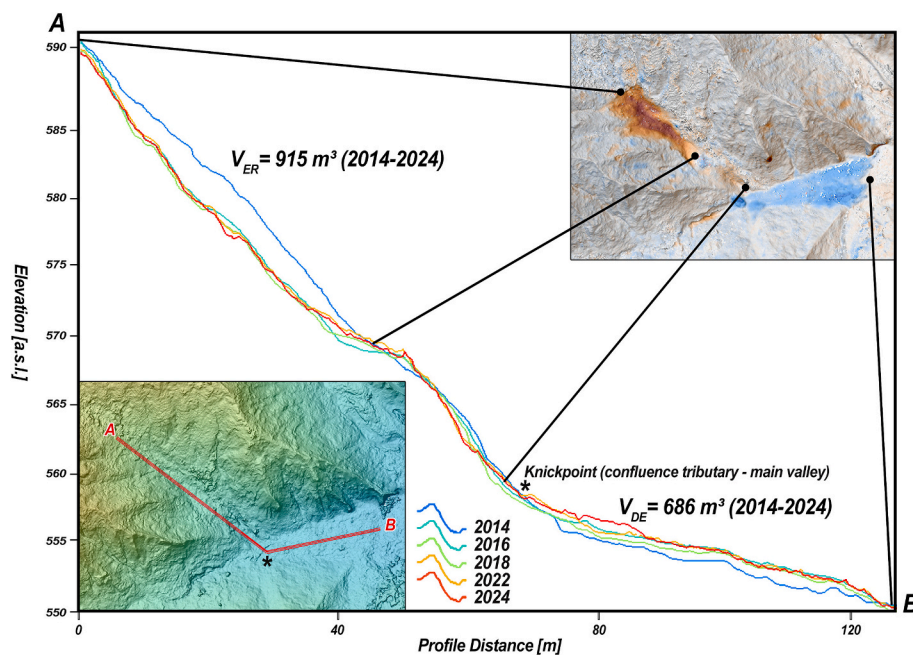


Fig. 5. DoDs (showing the topographic changes above the LoD) with respective histograms and hillshaded DSM 2024 with GHS 1–4 (mean = mean value of DoD; Std = standard deviation of DoD).

Table 4Multitemporal area (%) cover by each geomorphic process signatures and corresponding landforms, from T_0 (2014) through 2024.

Process area (%)	2014 (T_0)	2014–2016	2016–2018	2018–2022	2022–2024
Sheet wash/Rilling	22.74	35.16	31.36	19.18	17.41
<i>Sheet wash/Rill erosion areas</i>	94.31	95.52	98.26	97.01	98.68
<i>Colluvium</i>	5.69	4.48	1.74	2.99	1.32
Gullying	0.18	1.71	0.70	0.24	0.37
Channel type debris flow	24.64	1.95	6.19	1.28	3.99
Gravitational bulging		0.43	0.80	0.47	0.21
Translational landslide	2.22	4.64	6.72	4.47	3.44
<i>Scarp</i>	57.09	64.90	85.94	74.47	84.42
<i>Body</i>	42.91	35.10	14.06	25.53	15.58
Complex landslide	0.11			0.66	3.25
<i>Scarp</i>	100			86.26	38.27
<i>Body</i>				13.74	61.73
Rotational landslide	1.55				
<i>Scarp</i>	58.12				
<i>Body</i>	41.88				
Total geomorphological active area	49.89	43.89	45.77	26.30	28.67

**Fig. 6.** Top: Crossprofiles of GHS 1. Bottom Hillshade2024 (left), DoD 2014–2016 (middle), DoD 2014–2024 (right).

period between 2018 and 2022 (Fig. 6d). The positive balance of the raw DoD is primarily attributable to deposits from the tributaries, which were created by a combination of geomorphological processes, mainly smaller complex landslides and debris-like processes from the upper parts of the tributaries. At the junction between the tributary and the main valley, a drainage channel with pronounced linear erosion has formed on the northern levees, which already existed in 2014. This channel runs in the immediate vicinity of the slope foot in a south-eastern direction and extends into the main valley. A similar development can also be observed in the neighbouring valley to the northeast, where pronounced geomorphological alterations are now evident. These slopes had exhibited only minimal observable changes during earlier observation periods. Most areas of negative changes are situated in regions of extremely steep slopes with inclinations of $>45^\circ$ directly followed by regions of positive surface changes with maximum changes of $+4.71$ m in the gentler slopes ($<35^\circ$ inclination). Some of these depositional areas extend into the main valley, causing surface changes of up to 2.97 m. The steepening caused by the eroded abutment at the lowest parts of the slopes leads to a reduction in the angle of repose and the coefficient of static friction. As in previous periods, the northern slopes

are affected by smaller, shallow translational landslides. The linear incisions in the main valley (GHS 4) continue.

From 2022 to 2024, minor geomorphological changes were observed across the entire study site (see Fig. 6e). Nevertheless, GHS 2 and GHS 3 stand out in particular. The most significant changes can be observed in the central area of the study site (GHS 3) where a relatively large, complex landslide caused surface changes between -11.53 and 8.11 m. Notably, a reactivation of the debris-flow area occurred along the northernmost slope, accompanied by the development of a complex landslide immediately downslope of it. Translational landslides exhibited a marginal decrease and were still detected within the area that was historically affected by rotational landslides, co-occurring with rill-erosion processes. In comparison to the dynamics of earlier periods, the linear incision in the main valley remains at a comparatively moderate level. Only the lowest channel reach at the exit of the badlands (main valley) is affected by linear incision and erosion of the bank areas. A large proportion of the sediment previously available from the upstream areas, mainly from tributary of GHS 2, has already been mobilized in previous years. Consequently, more consolidated and less transportable material is now available. It is particularly striking that in

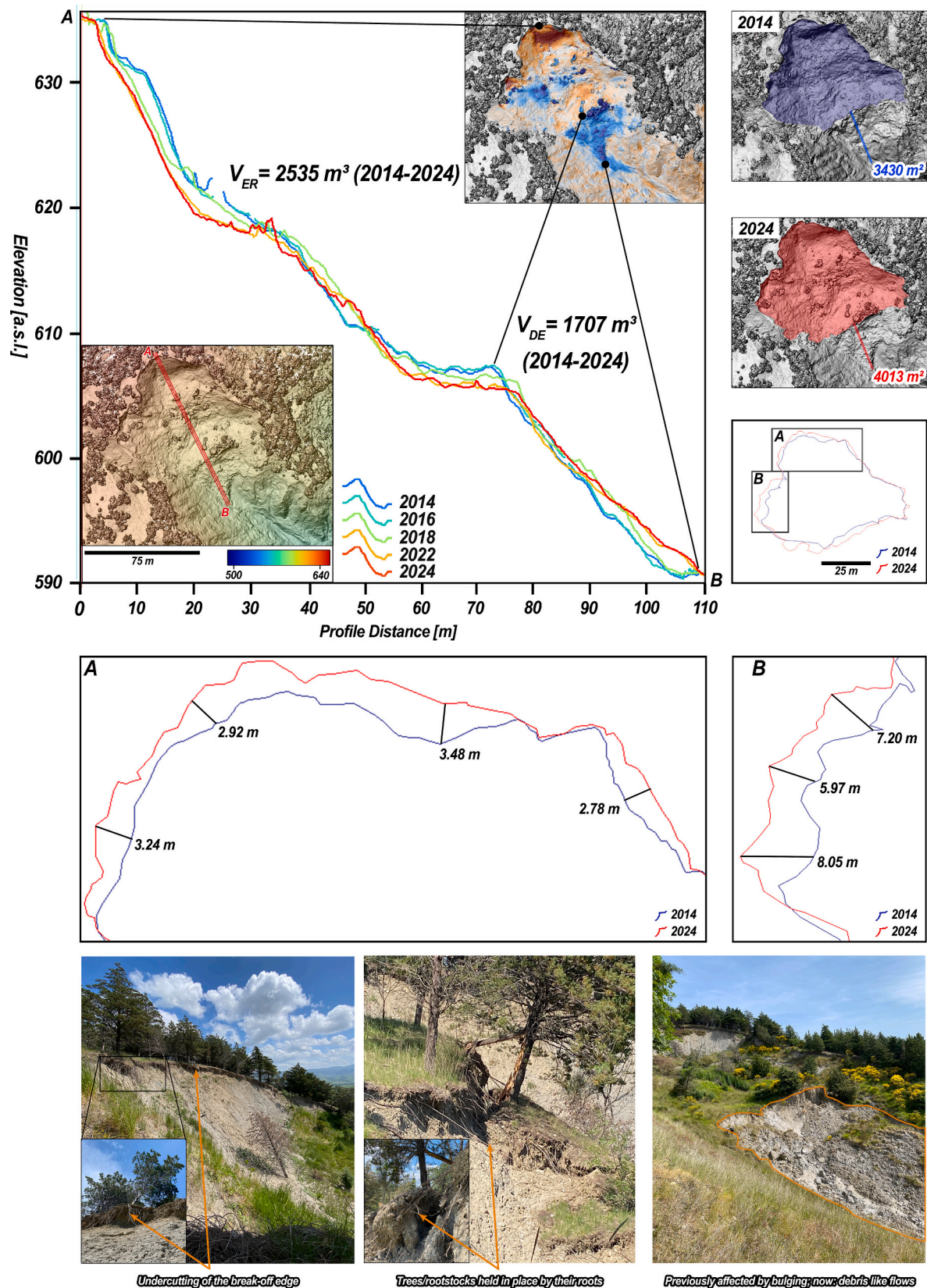


Fig. 7. Longitudinal-profiles and DoD 2014–2024 of GHS 2 (top left) and rotational landslide process area 2014 and 2024 (right column). Center: Mapped *calanco* border 2014 and 2024 showing progressive enlargement of the badland due to regressive erosion processes. Bottom: Images of the site show undercutting at the edge of the *calanco* boundary due to high-frequency water erosion processes (left), fallen trees (centre) and areas affected by bulging in recent years due to high pressure from the rotational landslide area (see the red and blue polygons in the top right top corner of the image).

this side valley, the material input on the north-eastern exposed slopes mainly takes place via small shallow landslides and debris flow-like processes, whereas the south-west exposed slopes mainly show signatures of slope wash and rill erosion.

4.1.3. Identification of geomorphic hotspots (GHS)

The mean slope gradients across the study area show only minor variations over the 10-year study period (see Tables S3 and S4; Fig. S1), thus motivating a more detailed analysis focused on the four GHS: three related to gravitational processes (translational, rotational, and complex landslides) and one to channel incision, with variable activation and persistence through time. Their temporal evolution is described in the following section.

GHS 1: Translational landslide

Between 2014 and 2016, a translational landslide, reaching a maximum depth of 3.18 m, occurred in the middle section of a south-east-facing tributary. The landslide area was predominantly covered with shrubs and grass, and resulted in the transportation of 915 m³ of sediments from the tributary, of which a total of 686 m³ were deposited in the main valley. Repeated site inspections have shown that these deposits consist predominantly of coarse particles (predominantly rounded pebbles) above the sand fraction. An analysis of the surface curvatures in the longitudinal profile (Fig. 4) reveals a transformation from a geometrically relatively simple convex surface with few kink points (in the pre-landslide state) to a concave curved surface with a stepped profile and clearly delineated central drainage channel. In addition to the morphological change, the landslide exposed significant quantities of sediment (vegetation-free areas), resulting in ongoing geomorphological activity in the area in the years following the landslide due to secondary water erosion processes, mainly rill erosion and piping.

GHS 2: Rotational landslide

From 2014 to 2016, the DoD displays only minor geomorphic changes at the upper slope sections of the rotational landslide. The longitudinal profile shows the presence of a plateau in the uppermost and steepest area which was created by the slipping of a soil clod from the reforested area above.

From 2016 to 2018, the area exhibits considerably geomorphological activity with alternating zones of erosion (max. -3.5 m) and deposition. The plateau visible in 2016 was completely eroded by 2018. The ongoing secondary processes have recently re-shaped the vegetation-free surfaces and are controlling the current geomorphic dynamics in this area. Again, like in the previous observation period the most pronounced surface changes occur in the uppermost and steepest areas of the slide surface followed by a relatively flat area with strictly positive surface changes (max. +1.5 m). This is followed by an area of negative surface changes on relatively steep slopes (up to 78° inclination). In the immediate aftermath of the aforementioned downhill trajectory, an area affected by gravitational bulging, characterised by strikingly high positive surface changes, follows.

The geomorphological activity continued during the period 2018–2022. The most significant negative alterations occur in the middle and upper segments of the slope adjacent to the tear-off edge. Additionally, a minor translational landslide occurred within the western part of the rotational landslide area. The associated zones of erosion and deposition are clearly visible in the respective DoD (Fig. 7). Within the 4-year period, the maximum changes in this area sum up to -3.36 m with a backward displacement of the tear-off edge of 3.5 m to the north-west. The zone previously affected by bulging shows high erosion values now, caused by water erosion and a channel-type debris flow connecting the area of GHS 2 to the tributary outlet. In contrast to the DoD 2016–2018, there are hardly any positive changes (depositional areas) in middle parts of the depth contour of the respective tributary. Instead, a clearly recognisable, narrowly incised depth contour has now evolved in the deposits from the previous two years.

From 2022 to 2024, geomorphological activity continued with the

formation of two smaller complex landslides in the uppermost slope parts, gradually expanding in north-western direction of the hillslope due to regressive erosion. The area between the upper break-off edge and the plateau below the lower rotational surface increased by 583 m² (approximately 60 m²/y) in the ten-year monitoring period. With a relief energy of over 45 m across a horizontal distance of 110 m, the rotational landslide is situated in one of the most precipitous areas of the badlands. Below the break-off edge, the slope angles increase to over 90°, resulting in undercutting of up to 1 m below the slope's edge. The plateau situated behind the break-off edge, which has been reforested with black pine and Arizona cypress trees, dips towards the south-east, in the direction of the side valley. Large cracks, ranging in width from 1 to 30 cm and reaching depths of up to 40 cm, are forming across the entire plateau area, extending from the break-off edge to the road. This results in the fragmentation of the area into individual clods, with the majority of these clods being prevented from sliding downwards solely by the roots of the reforested trees. The bulging zone is still affected by widespread erosion processes, mainly complex landslides and debris flows, proving the continued instability and sediment transfer downslope. In general, GHS 2 experienced the most continuous and active geomorphic development of all monitored GHS.

Complex Landslide (GHS 3)

GHS 3 represents the area with the most significant topographical changes in the entire observation period. Field surveys conducted in May 2023 and November 2023 enable us to date the respective landslide to this period. Unfortunately, a precise dating is not possible based on the available data. The landslide covers 2275 m², causing a volumetric change of -4020 m³ in the erosion area and 4016 m³ in the deposition area and a surface change between -11.53 and 8.11 m. The slope inclinations in the deposition area show a predominance of values between 20° and 45° after the landslide, with a mean value of 38° (Std. = 16°). Extreme slope inclinations are comparatively rare, occurring only at the scarp in the uppermost part of the process area, with values reaching up to 87°, related to two nearly vertical joints striking SSW-NNE and E-W (Fig. 8, red area). The landslide led to a temporary separation of the steepest slopes from the main valley. This is due to the formation of a notch that extends perpendicular to the direction of the landslide, creating a temporary, natural dividing line between the erosion and deposition areas. This dividing line is visible as a thin grey line between the red and blue areas in the DoD and the longitudinal profiles in Fig. 8. In addition to the significant changes in the erosion area, the complex landslide has also led to notable topographical changes in the deposition area. A significant proportion of the deposited, unconsolidated substrate is located in a topographically steep environment (mean: 37°; max: 83°) in the immediate vicinity of the main valley.

Channel incision (GHS 4)

The GHS 4 study site demonstrates substantial geomorphological alterations throughout the entire study period. The stepped longitudinal profile of the main valley is striking (Fig. 9). The steepest slopes occur in the central third of GHS 4. This area is characterised by the convergence of the two tributary valleys, which exhibit the most substantial sediment input. From 2014 to 2024, the DoD shows clear patterns of erosion across all areas of the depth contour. Particularly notable is the pronounced linear incision in gently sloping terrain areas, which reaches a depth of up to 3.5 m. Alongside the deepening of the primary drainage channel, new, smaller side channels have developed within the sediments of the main valley. In addition to linear incision, existing tributaries also show signs of regressive erosion in the upper parts of the smaller channels. With the exception of the DoD 2018–2022, positive surface changes can only be seen on the northern bank of the main channel in the central area of GHS 4 and are attributable to movements and sediment input of the complex landslide (GHS 3). This is especially visible in the DoD of 2018–2022 and 2022–2024. Although the DoD 2018–2022 shows strong positive changes it is not possible to attribute these to geomorphological processes. The majority of positive elevation changes are caused by the growth of low vegetation, primarily bushes

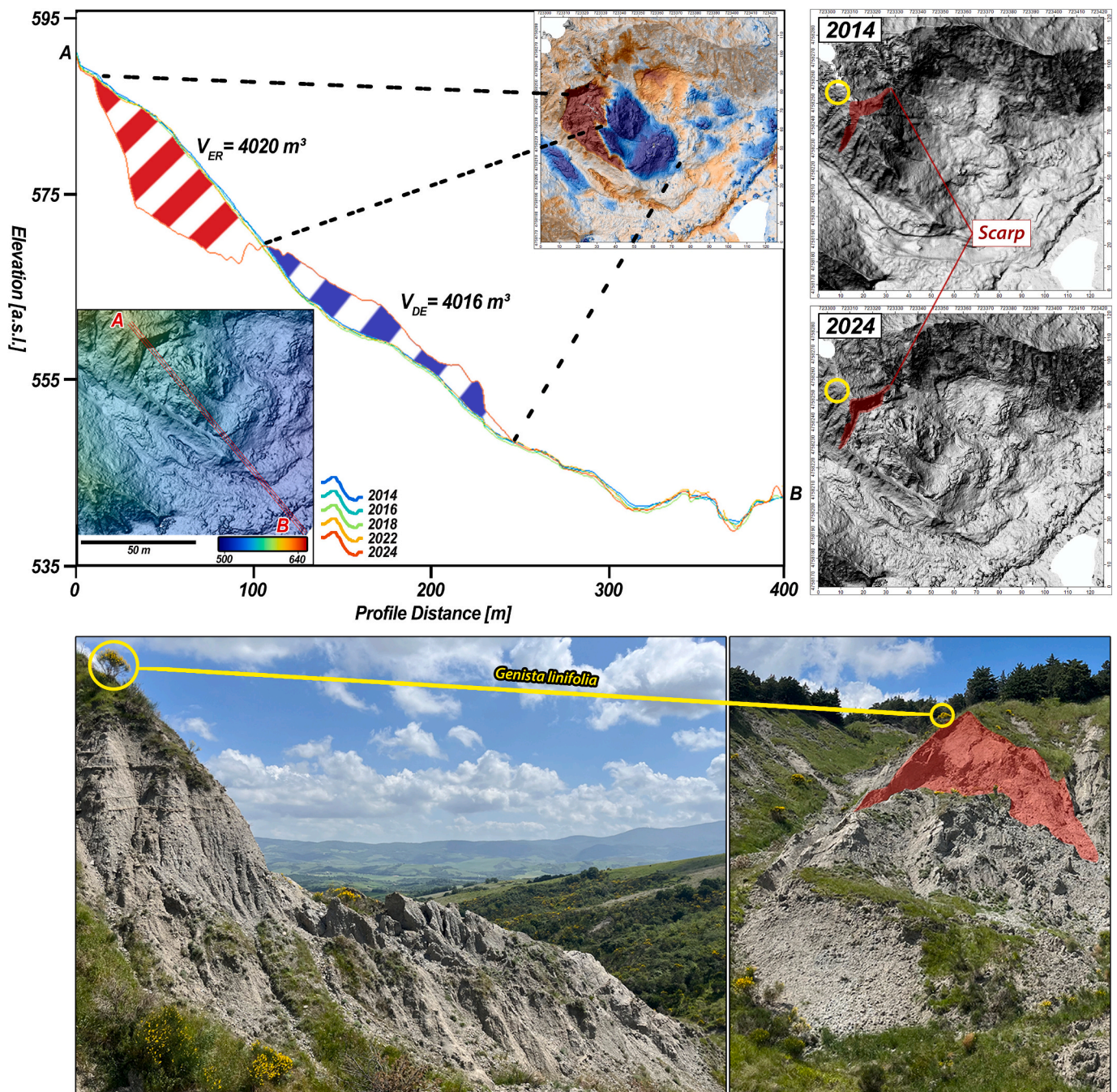


Fig. 8. Longitudinal-profiles and DoD 2014–2024 of GHS 3 (top left), hillshaded DSMs of 2014 (top right) and 2024 (centre right) with the scarp marked as red polygons. Bottom: Images from field in May 2024 in south-east (left) and north viewing direction. The yellow circle marks a *genista linifolia* bush and serves for a better orientation.

and grasses. Since the changes in vegetation in the study area are apparently very significant and highly dynamic in both spatial and temporal terms, this topic will be examined in more detail in the following chapter.

4.2. Geology of the study area and selected rock properties

A distinctive westward dipping sediment profile was recorded along a ridge between two calanchi south of GHS 2 (Fig. 10) in order to determine more precisely the role of material properties in the observed process dynamics. Seven gravel layers with thicknesses ranging from 0,5 m to several m were identified. In the lower portion of the section a

sequence of greyish siltstones, brown sandstones and some fine gravelly sandstones up 40 cm in thickness occur. The coarser rocks are distributed discontinuously. Sometimes they show graded bedding features indicating transportation and deposition by submarine turbidity currents. In the upper part claystones and silty claystones with some indistinct fine sandstones prevail dissected by some gravelly deposits varying laterally in thickness and grain size. Occasionally, small slumping structures (15–20 cm long and 2–5 cm thick fine layered clayey sediments around small pebbles forming a horizontal fold) can be observed. Referring to Blair and McPherson (1999) classification, the size class of the clasts range from pebble to medium boulders.

At the upper scarp of the rotational landslide (GHS 2) a change of

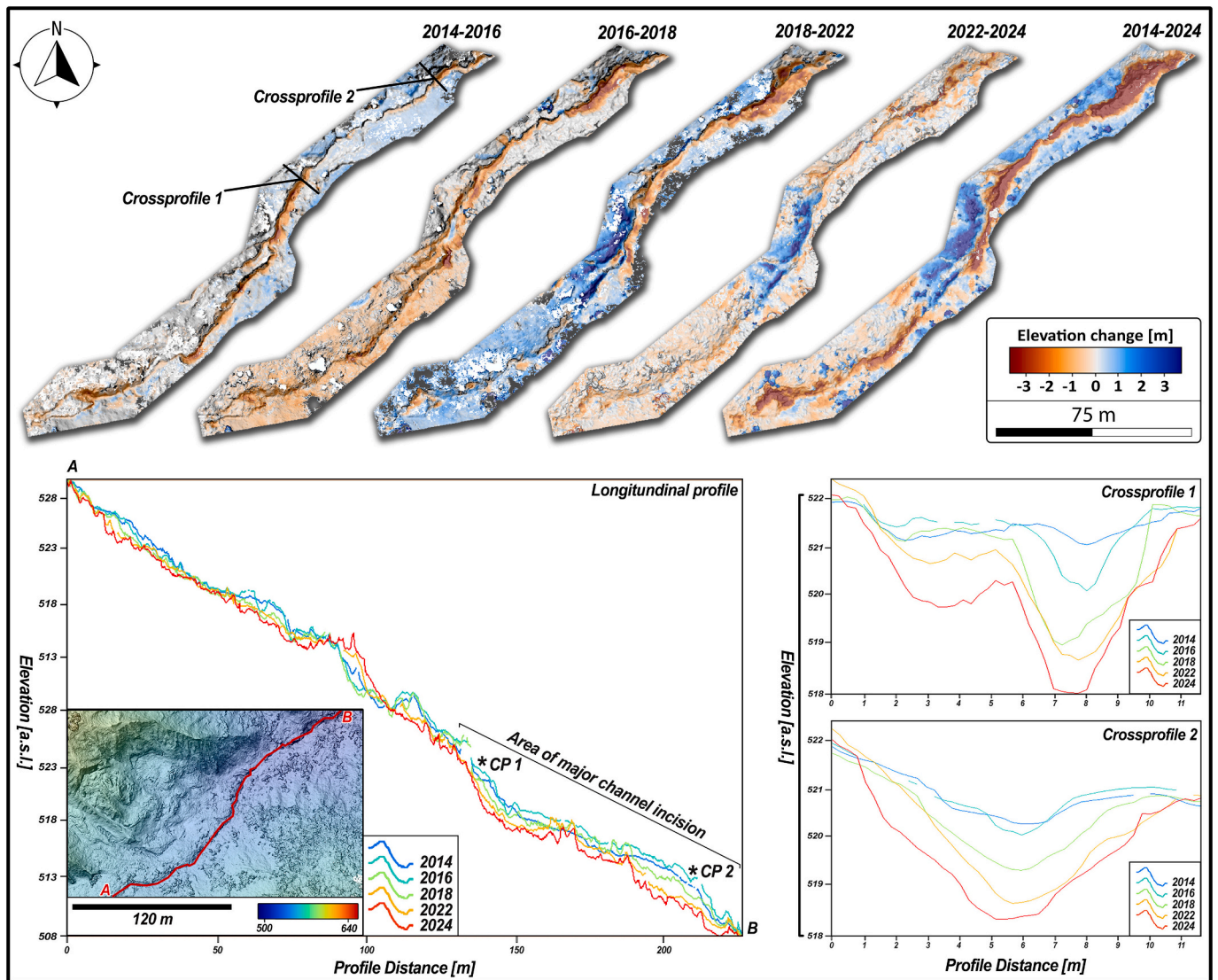


Fig. 9. Multitemporal DoD of GHS 4 (top). Bottom: Longitudinal-profile with locations of longitudinal (red line A-B in the DSM) and crossprofiles (*CP1 and *CP2 in the longitudinal profile). Bottom, right: Multitemporal crossprofiles.

strike and dip of the uppermost strata was observed. The rocks are lying more or less horizontal or they show low dip values (128/10 SW in the western part to 80/10 N in the eastern part). Probably, this dip change was caused by reducing the width of the ridge between the *Scalonca* and the southern *Trucchiole* catchment by erosional processes. Due to the removal of the rock support on both sides of the ridge, the primary inclination of the rocks changed gradually by slow plastic movement of the rocks. Furthermore, the slope beneath the ridge shows a slight swell of the bed rock. As a whole, this context has to be interpreted as a predispositional function of the geology. In connection with additional features (surficial cracks or subsurface joints and faults) several mass movements (e.g. GHS 2 or GHS 3) can be explained.

The clay-rich lithologies of the study area contain more than 50 wt-% clay-sized particles. Although they appear as hard materials in dry conditions, they soften and behave as semi-solid when wetted and exhibit swelling values of 2.5–7.5 Vol-%. These features suggest a relatively low to moderate degree of lithification.

The estimation of hydraulic conductivities coincides with the respective grain size data (Table S4). Claystones of the study area exhibit very low to nearly none water permeabilities (k_f in the range 10–9–10–10 m/s), the intercalated sandstones can be classified as permeable rocks (k_f in the range 10–6–10–8 m/s depending on the clay content).

Consequently, diverging flow velocities are indicated for the sedimentary column with the main water flow within the sandstones. Referring to the main dip of the rocks water can percolate where sandstones crop out, the subsurface flow is directed to the west. Thus, swelling of the claystones occur both directly at the rock surface as at near-surface sandstone/claystone contact enabling crack formation and increasing water infiltration. Additionally, at greater depths the cohesive shear strength could be reduced due to the claystone swelling at the contact to interlayered sandstone.

From the geological and petrophysical results two conclusions indicating the geological predisposition for further geomorphological processes can be drawn:

1. The alternating sequence of clay- and siltstones with sandstone layers and gravelly beds leads to seepage at sandstone outcrops creating some groundwater flow along the sandstone layers following the dip. Otherwise, the differential swelling of the bed rock form cracks near the surface improving furthermore the infiltration of water. Thus, these shallow cracks support the formation of translational landslides.
2. The modification of the dip near the upper edges of the *calanchi* or at ridges between two *calanchi* infers a slow plastic movement of the

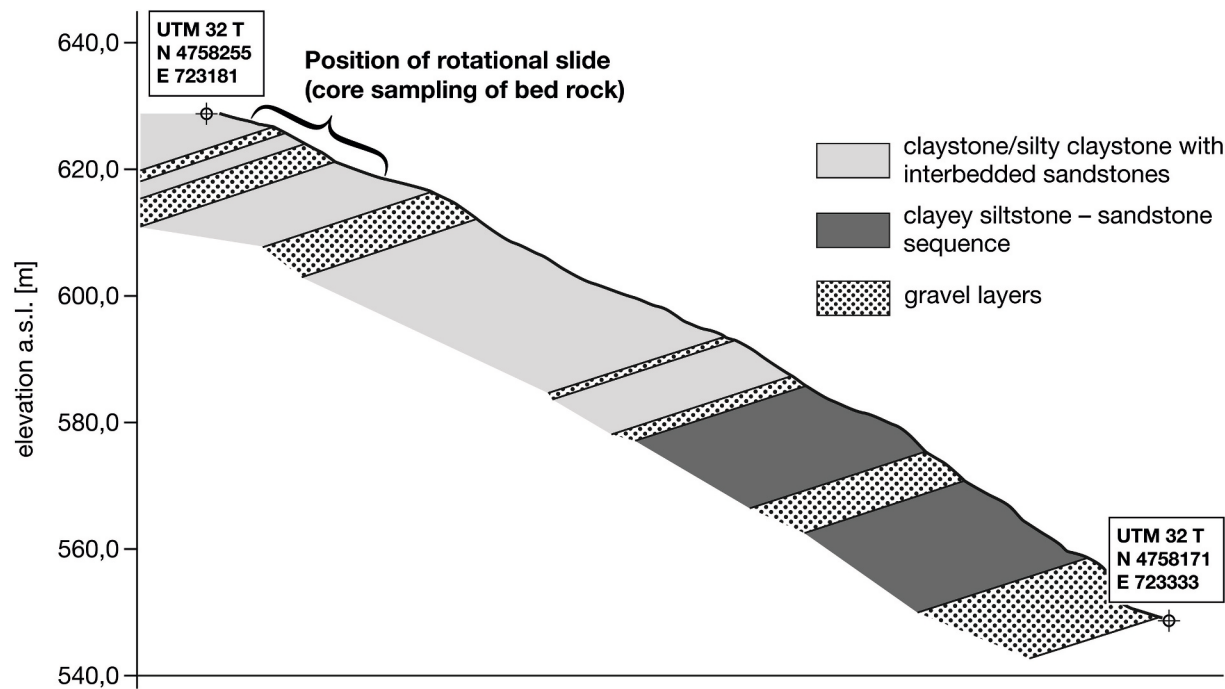


Fig. 10. Geological section along the western crest south of the GHS2 (for localisation see Fig. 1 “profile a-b”).

bed rock (swelling) at the slopes. In connection with pre-existing joints the detachment of rock masses (e.g. rotational landslides or massive rock breakdown) will be favoured.

4.3. Precipitation distributions and patterns

Of all precipitation events characterised as erosive, 165 were classified as type A, 76 as type B, 13 as type C and 88 as type D (Table 5 and Fig. 11). While the variation within type B is relatively low, types A, C and D show greater variations in terms of the number of events in the individual monitoring periods. The most pronounced variation is observed in events A and C, with the enhanced annual precipitation totals following 2018 (>700 mm) also reflected in type A and C events. Noteworthy is the increase in type A events since 2018, as well as the increased frequency of type C events in the period 2018–2022 and type D events in the period 2022–2024. Due to the high daily precipitation sums recorded between the second half of October 2023 and the first two weeks of November 2023 (see Fig. 11), it is assumed that the soil was significantly saturated during this period. These hydrological conditions indicate an increased probability of the landslide event in the late autumn to early winter phase. Despite the variability between the single monitoring periods, considering the whole 10-year study period, no clear intensification of extreme events or changes in the overall precipitation regime were detected.

4.4. Vegetation dynamics through TGI index

The areas classified as vegetation by the TGI calculation show a high degree of consistency with the actual vegetation distribution visible on the DOM. At the beginning of the study period, most of the area was devoid of vegetation.

Over the following decade, vegetation gradually increased (Table 6), evolving from a fragmented to a continuous structure (Fig. 12). Vegetation growth is more pronounced in former geomorphologically active areas, where past geomorphological processes, such as material deposition, have facilitated subsequent vegetation development.

Conversely, areas affected by recurrent processes, such as the upper part of the rotational landslide (GHS2), remain largely vegetation-free. Moderate slopes (<36°) favour vegetation growth, while steeper slopes (>37°) inhibit vegetation, and it's mainly concentrated on north-, south-east-, and east-facing slopes with gradients between 20° and 50° (74% of total vegetation cover). In contrast, north-west-, south-west-, and west-facing slopes generally show sparse vegetation cover. Here, both, flat areas (<20°) and steep slopes (>50°) are largely bare. Between 2014 and 2022, small, vegetation patches appeared within the rotational landslide area but disappeared completely in 2024 after new landslide events. In 2018, vegetation distribution was comparable to 2016 but denser in already vegetated areas, especially in the main valley. The flow paths of the side valleys, however, remained vegetation-

Table 5
Absolute and annual (in brackets) precipitation totals and extreme events. Bold values show the annual maximum values per category.

Period	Total precipitation [mm]	Type A (days with > 5 mm / h)	Type B (days with > 5 mm / 15 min)	Type C (events > 60 mm in 5 consecutive days)	Type D (ID relationship > 75% percentile)
2014/04–2016/04	1256.4 [627.34]	28 [14.00]	12 [6.00]	3 [1.50]	16 [8.00]
2016/04–2018/05	1376.4 [669.98]	27 [13.14]	18 [8.76]	1 [0.49]	17 [8.27]
2018/05–2022/06	2958.0 [730.99]	73 [18.04]	29 [7.17]	8 [1.98]	34 [8.40]
2022/06–2024/05	1507.6 [775.03]	37 [19.02]	17 [8.74]	1 [0.51]	21 [10.80]
2014/04–2024/05	7098.4 [706.36]	165 [16.42]	76 [7.56]	13 [1.29]	88 [8.76]

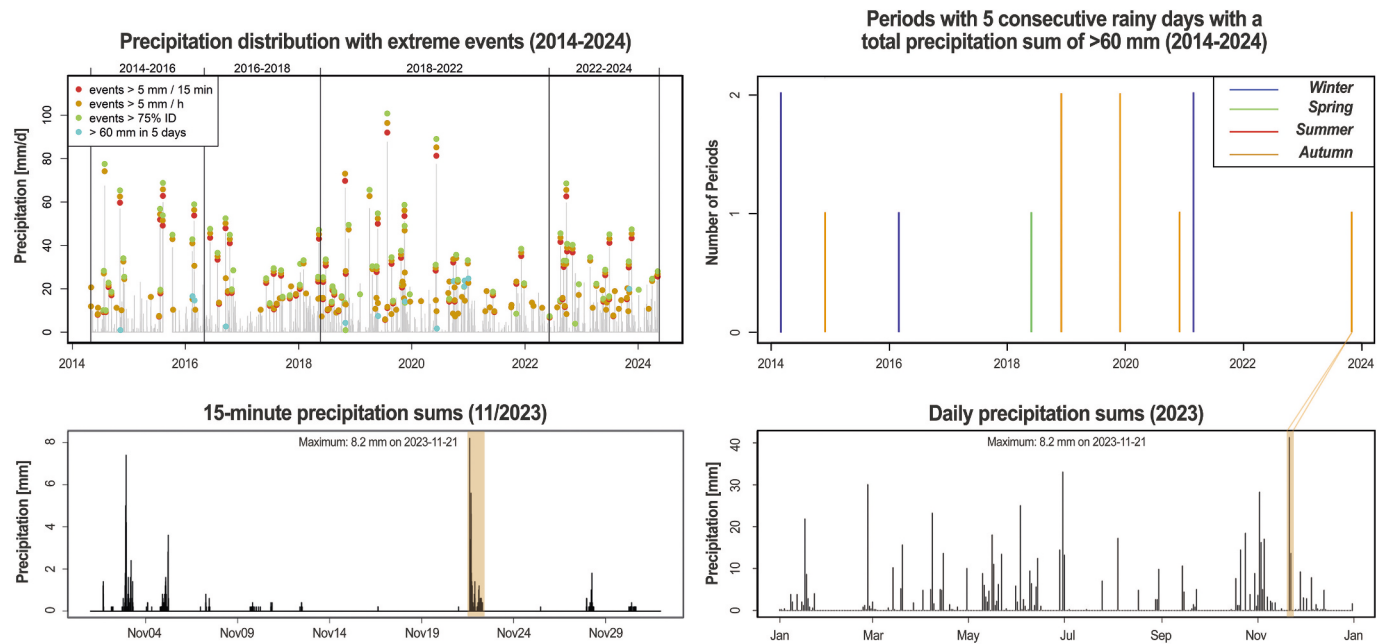


Fig. 11. Rainfall distribution for the entire observation period with extreme events (top, left) and periods of consecutive rainfall (top, right). The black vertical lines show the date of the respective data surveys. Bottom: 15-min precipitation sums for November 2023 (left), and daily precipitation sums for 2023 (right).

Table 6
Vegetation spread and development for all observation periods.

Year	Vegetation		Period	Vegetation change [m ²]
	[m ²]	share of AoI [%]		
2014	35,133	33.5	2014–2016	+1168
2016	36,301	34.6	2016–2018	−634
2018	35,667	34	2018–2022	+4831
2022	40,498	38.6	2022–2024	+5546
2024	46,044	43.9	2014–2024	+10,911

free. By 2022, vegetation cover had further increased: the previously active translational landslides were completely overgrown, and for the first time, vegetation appeared within the drainage channels of the tributaries. In 2024, a slight decline in vegetation cover was observed on north-facing slopes, while a clear contrast persisted between the more vegetated east-facing slopes and the sparsely vegetated west-facing ones. Overall, the period from 2016 to 2024 is characterised by an expansion of vegetation cover, mainly due to the widespread colonization of the main valley and the landslide pathways along the north- and north-east-facing slopes in the southern part of the study area. The most significant growth rates were observed during the period 2018–2022, when the bare substrate decreased by approximately 50% due to the expansion of vegetation (see Table 4).

4.5. Vegetation-geomorphology interaction

The intersection of the geomorphological signatures with the vegetation maps led to the identification of four land cover classes: i) surface characterised by water erosion processes, ii) surface characterised by gravity-driven processes, iii) sparsely vegetated surface and iv) vegetation surface. Transitions between classes in each time step are shown in the Sankey diagram (Fig. 13), where flows are represented by arrows whose width is proportional to the mapped area. While i) and ii) represents unstable surfaces, iii) represents an intermediate stable situation, characterised by sparse vegetation cover and signs of active geomorphological processes. It therefore functions as a transitional category, in which surfaces may either stabilize (e.g., due to growth of vegetation) or destabilise (e.g., due to erosion). Category iv represents

geomorphological stable surfaces.

In general, water erosion increased from 2014 to 2018 but has been undergoing a steady decline since. The increase can be attributed to areas previously dominated by gravity-driven processes (2014 to 2016) and sparsely vegetated areas (2016 to 2018), which have also increased over the course of the observation period. While denudation process signatures show a steady decline after 2018, vegetated areas, by contrast, continuously increase from 2018 and attain their maximum extent in 2024.

By 2016, surfaces initially characterised by gravity-driven activity were increasingly replaced by water erosion processes or transitioned towards partial and full stabilisation. The expansion of sparsely vegetated surfaces is due to the partial stabilisation of areas previously affected by water erosion, to the partial destabilisation of vegetated surfaces and the recultivation of areas affected by gravity-driven processes in the past. In 2018, however, most transitional areas are stabilised (fully vegetated), while a smaller fraction remained unstable or showed renewed denudation activity.

The initial rise in the extent of sparsely vegetated areas, followed by a decline (2016–2018), indicating a progressive stabilisation of the landscape. From 2018 onwards, the expansion of sparsely vegetated terrain occurs mainly at the expense of areas previously affected by water erosion processes, fully vegetated surfaces, and gravity-driven areas. After 2022, all denudation-related categories decrease, while vegetation continues to expand, primarily by occupying formerly sparsely vegetated surfaces. The slight increase in gravity-driven areas observed in 2024 originates from surfaces previously characterised by (sparsely) vegetated terrain and water erosion processes. Sparsely vegetated zones have, for the most part, been stabilised and converted into fully vegetated surfaces, although some areas have persisted.

5. Discussion

5.1. 10-years badland development

The study area exhibited a decline in the proportion of surfaces impacted by geomorphological processes during the observed period, suggesting a shift towards a phase of stabilisation. Nevertheless, the annual erosion rates remained relatively stable. This suggests that areas

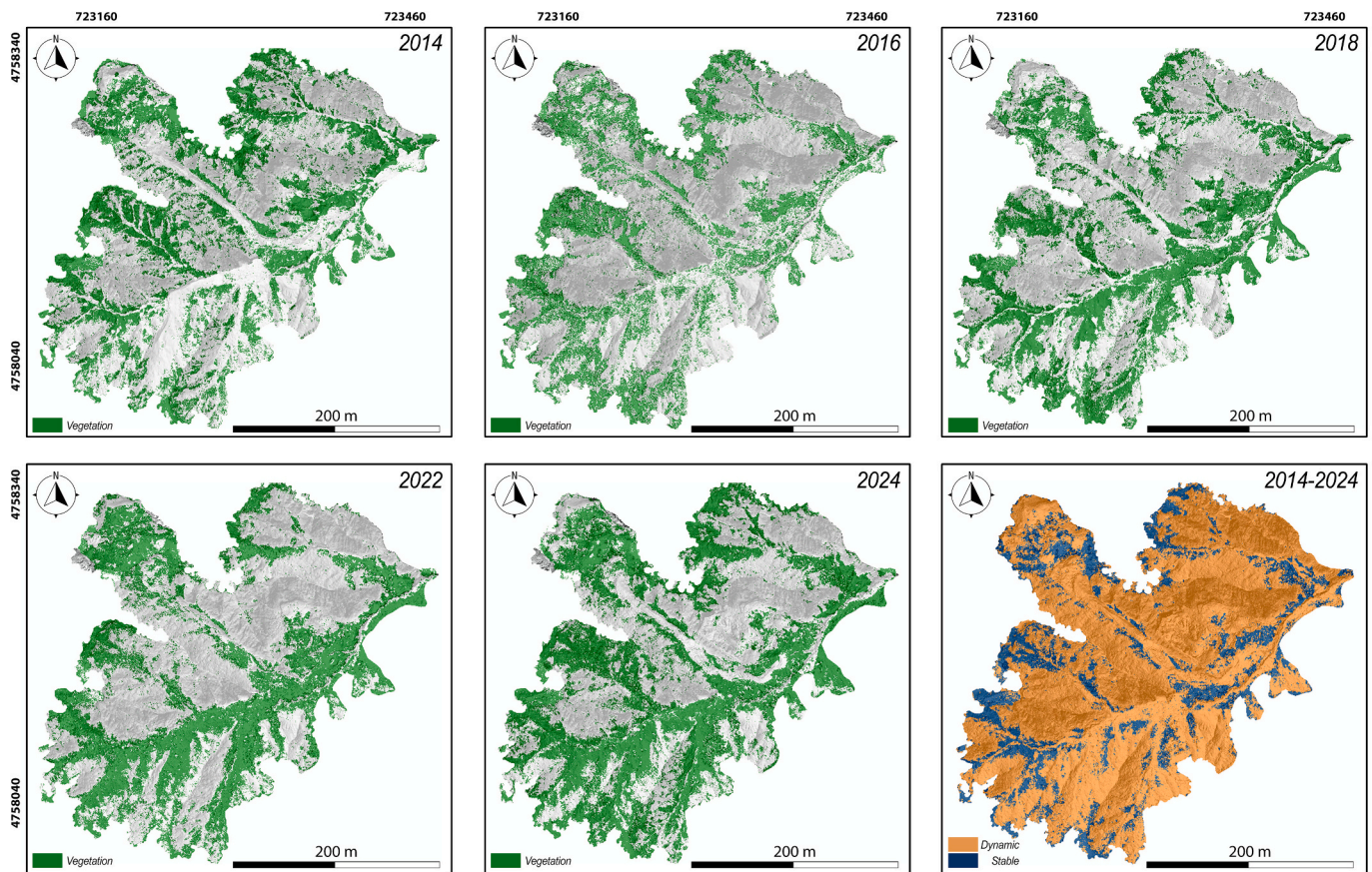


Fig. 12. TGI-based vegetation maps and map of vegetation dynamics in the period 2014–2024 (bottom right).

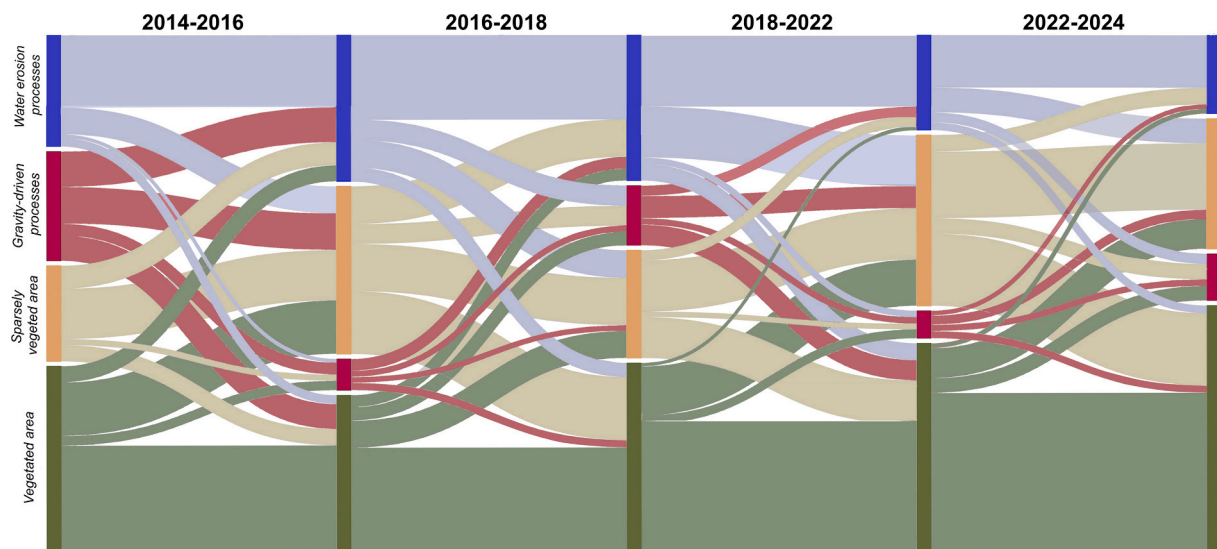


Fig. 13. Sankey diagram illustrating the multitemporal transition (area) of land-cover classes across different time steps.

affected by geomorphological processes now involve a smaller portion of the landscape, indicating that geomorphic activity is increasingly concentrated in fewer but higher magnitude events, as in the case of GHS 3. Moreover, the results of this study emphasise the pivotal role of gravity-driven processes in the long-term geomorphological evolution of sub-humid badlands. This is in agreement with findings by Vergari et al. (2019) and Stark et al. (2020). The vegetation development was predominantly controlled by a reduction in geomorphic activity under

relatively stable precipitation patterns over the study period. Indeed, in a humid badlands context, where annual precipitation is sufficient to sustain full vegetation cover, the main constraints on plant establishment are geomorphological and climatological factors (Martínez-Murillo and Nadal-Romero, 2018). The study presented here also confirmed Della Seta et al.'s (2009) assumption that in southern Tuscany's badlands also piping contribute to denudation, especially in depositional areas, and develop rapidly through collapses.

5.1.1. Geomorphic dynamic across the study site

The tributaries and the main valley exhibit clear evidence of active gully erosion and piping within depositional areas, confirming previous observations in Mediterranean badlands (Faulkner et al., 2004; Faulkner, 2013). The widespread piping processes must be taken into account when interpreting the results, particularly when considering the calculated erosion rates. As these processes do not cause any direct changes to the surface but nevertheless result in sediment removal, the actual rates of material loss are certainly even higher than the values measured on the basis of the DSMs. Among the mapped signatures, one process type still undocumented in badlands emerged: bulging. This gravity-induced deformation process manifests as positive surface changes in the DODs, despite the absence of sediment accumulation. This process appears to be linked to water infiltration into clay-rich substrates beneath the historical rotational landslide (GHS 2), where layered conglomerates and sandstones overlie low-permeability silty clays. Temporary perched water tables likely promote hydro-mechanical weakening, leading to progressive slope deformation, as already observed by Colica (2000). The south-facing slopes show persistent water erosion signatures and limited vegetation development, due to drought stress and reduced soil moisture. These observations are consistent with the results of comparable studies in badlands (e.g. Alexander et al., 2008; Torri et al., 2000 and Pulice et al., 2012). Although several studies attribute particular significance to these processes in the further development of badlands, based on the results presented here, this finding can only be partially agreed with.

Despite the continuous action of water erosion on exposed slopes, the temporal pattern of surface change rates indicates that the recent geomorphic evolution of the study area is primarily controlled by high-magnitude events, particularly channel-like debris flows and translational landslides, rather than by high frequency, low magnitude runoff processes. Evidently, low- and high-magnitude processes have the capacity to interact through time, for instance, when persistent, low-magnitude runoff progressively reduces the safety factor of larger landslides. In the present study this could not be observed as the multi-temporal mapping demonstrates that vegetation growth is clearly decelerating low-magnitude runoff processes. The findings of this study suggest that high-magnitude processes continue to dominate geomorphic development of the badland, while low-magnitude processes are declining. Pittau et al. (2024) draw analogous conclusions and define the period from 2000 to 2014 as a phase of soil cover stability characterised by lower precipitation but with frequent events of high intensity.

Overall, the presented surface change rates are in good accordance with those obtained in comparable studies, e.g. Ciccacci et al. (2003) and (Ciccacci et al., 2008), Della Seta et al. (2009), Aucelli et al. (2016), Neugirg et al. (2016), Stark et al. (2020). According to Clarke and Rendell (2006), erosion rates can be understood as a function of the angle of inclination. In their study, they observed the highest erosion rates on inclines of approximately 35° . This is consistent with the

present study, which has shown that most erosion areas are located on relatively steep slopes with an incline between 24° and 56° . This applies to both gravitational (mean: 38°) and water erosion processes (mean: 45°). The preponderance of vegetation-free areas susceptible to water erosion processes is observed to occur predominantly on slopes that exceed 34° of inclination (Fig. 14).

5.1.2. Lithology and substrate

The specific geological and lithological structures represent a crucial predisposing factor controlling geomorphodynamics in the study area. This conclusion can be illustrated on the basis of the GHS. The characteristic layering of marine and transitional transported sediments in the geological composition of the slope allows water to penetrate through cracks, leading to swelling processes in deeper sediment layers. This creates unstable soil conditions, such as reduced shear forces, which increases the frequency of shallow translational landslides, as in GHS 1. The general slope instability in the upper parts of the side valley of GHS 2, due to e.g. steep, sparsely vegetated slopes, existing E-W faults, gradually leads to the deepening of existing cracks at the upper edge of the *calanchi* when water infiltrates. Consequently, this led to a deep-seated, curved detachment surface that manifests itself in the form of rotational landslides (GHS 2).

Over the course of the year, high water reactivity of clays can lead to significant fluctuations in soil moisture content and to considerable shrinkage and swelling processes, as described in the relevant literature (e.g. Kaiser et al., 2018; Zhang et al., 2023). In the present study, this process (gravitational bulging) was documented for the first time in badlands based on the analysed surface changes (see Fig. 9) for the area immediately below the rotational landslide. The exposed soil surfaces are highly prone to be affected by secondary erosion processes and weathering factors. However, weathering processes (chemical, physical and biological) are particularly crucial for the conversion of solid rock into transportable sediments, resulting in unconsolidated sediments and soft bedrock (García-Ruiz et al., 2013; Kašanin-Grubin et al., 2018; Avcioglu et al., 2022). Especially wetting-drying cycles are crucial in physical weathering and regolith behaviour, which resulted in cyclic deep crust formations in the spring and summer seasons due to the high swelling capacity of samples (Avcioglu et al., 2024).

A little further down the valley (along the geological profile a-b), the sensitivity of substrates with a high clay content is once again impressively demonstrated. The strong reaction with water reduces aggregate cohesion and the internal friction of substrate, thus decreasing slope stability and favouring translational landslides and shallow mass movements. This observation was repeatedly confirmed in the present study, thus supporting the findings of Rengasamy and Olsson (1991), who reported that water accumulation at the boundaries between soil and source material is a factor that promotes clay dispersion and shallow landslides, especially during periods of heavy rainfall. Conversely, and in accordance with Bazzoffi et al. (1997) and Gallart et al. (2013), on

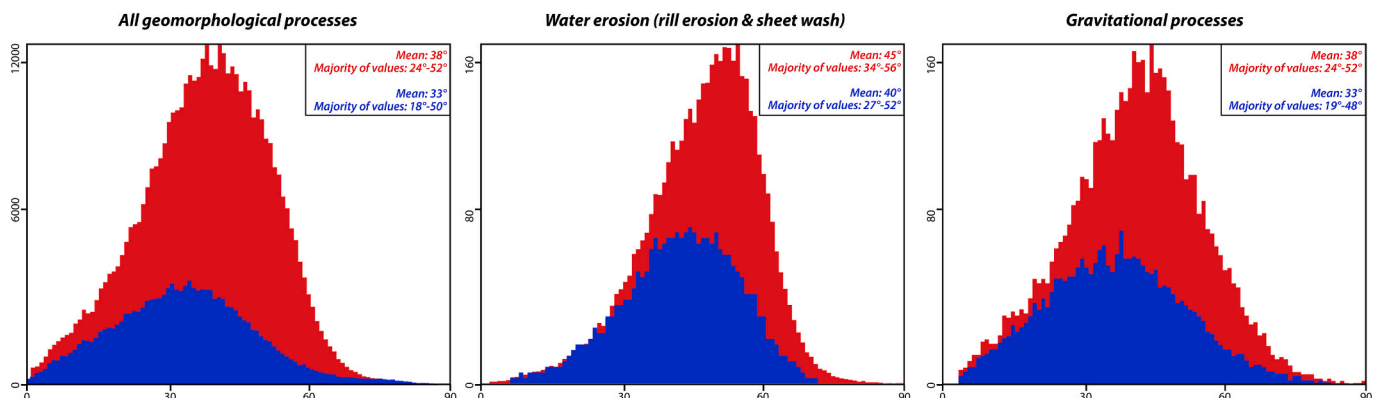


Fig. 14. Distribution of slope inclination values for negative (red) and positive (blue) surface changes in the DoD 2014–2024.

steeper south-facing slopes where vegetation growth is limited, water erosion dominates.

Overall, the strong coupling between lithology and hydro-mechanical behaviour governs slope sensitivity and provides the structural framework within which high-magnitude geomorphic processes are triggered.

5.1.3. Vegetation and precipitation

Vegetation plays an important but spatially and temporally variable role in slope stability, by reducing splash effects (Bryan, 1974) limiting surface runoff, and the potential for erosion. In addition to the mechanical effect of soil stabilisation, extensive vegetation has a positive effect on the microclimate near the ground (Geiger et al., 1995) and hinders the drying out of the soil matrix.

During the study period, vegetation cover increased progressively, mainly colonising depositional areas and north-facing slopes, where higher soil moisture and lower solar radiation favour plant growth (Bazzoffi et al., 1997; Gallart et al., 2013), exerting a geostabilizing effect (Faulkner, 2013). In this sense, many authors showed that the encroachment of vegetation cover in the local sodic-rich dispersive marls and clay deposits leads to a decrease of exchangeable cations concentration and, thus, of clay dispersivity and erosion potential if gradients are sufficiently suitable (e.g. Phillips, 1998; Vergari et al., 2013). Within the scope of the present study, vegetation-free areas were observed to undergo rapid re-vegetation. Due to significant vegetation growth, most scarps and landslide surfaces were no longer evident within a period of two years. Re-colonization is therefore occurring significantly faster than in Pittau et al. (2024), where complete re-colonization of the landslide slopes took up to 17 years. Of course, the rate of recolonisation depends to a larger extent on prevailing moisture (substrate) and precipitation conditions. However, the stabilising effect of vegetation appears to be temporary and strongly controlled by local topographic and substrate conditions.

Considering rainfall, both its amount and timing are key drivers of geomorphological activity and vegetation dynamics (Yang et al., 2021; Marsico et al., 2021). Despite some increasing or decreasing trends, no clear intensification of extreme events or changes in the overall precipitation regime were detected. This suggests that the observed geomorphological changes were not primarily event-driven (i.e. related to exceptional precipitation) but rather reflect intrinsic geomorphological dynamics operating under a normal precipitation regime (Faulkner, 2008). Faulkner suggests that when stream power oscillates around the system critical threshold for geomorphic change, the system becomes highly sensitive, favouring cut-and-fill dynamics through intrinsic threshold exceedance. This mechanism well represents the present conditions of the study area, where changes occur under near-threshold conditions rather than requiring external drivers such as climatic shifts.

Thus, under the present climatic conditions, geomorphological processes induced by episodic high intensity events, occurring within an otherwise normal climatic framework, such as complex landslides or debris flow-like processes, play a key role in landscape development. Indeed, previous studies show that landslides in this area are usually triggered by heavy winter and spring rainfall, when the clay formations become saturated with water and their cohesion and shear strength decrease significantly (Vergari, 2015). Due to their mass and energy, these high-magnitude geomorphological processes have the potential to erode and remove dense, bushy vegetation cover, as well as mobilise fine particles from sediment deposits in valleys and from source material on slopes. The resulting new, unconsolidated (deposits) or unprotected (bare substrate) surfaces can in turn be further developed by secondary processes, including new, mostly smaller mass movements, and lead to significant landscape changes in a relatively short period of time. Additionally, root systems (here: e.g. *genista linifolia* and *pinus nigra*) may locally enhance infiltration and interflow along low-permeability clay layers, reducing soil strength and promoting shallow translational failures.

Even of smaller magnitude, those processes can significantly influence sediment dynamics and contribute to the dominant 'cut-and-fill' cycles (Faulkner et al., 2008). Due to the low transport capacity of the events, the eroded sediments are deposited in the valleys or middle to lower slope areas. Fig. 15 shows a simplified representation of the complex interplay of geomorphological processes responsible for the landscape development of the *Val d'Orcia calanchi* badlands. These processes depend strongly on the vegetation cover and the location (slope/valley) in the catchment area. Furthermore, the strongly pronounced seasonality of the Mediterranean climate, with its marked seasonal fluctuations, has a considerable influence on the observed landscape developments. Low or absence of rainfall combined with high solar radiation in spring, when vegetation begins to sprout, may enhance erosion due to unfavourable microclimatic conditions. In the *calanchi* badlands of the upper *Val d'Orcia*, the aforementioned geostabilising effect due to vegetation is therefore only temporary.

5.2. Methodological limitations

5.2.1. Multitemporal high resolution photogrammetric analysis

Photogrammetric techniques, and in particular structure-from-motion (SfM) approaches, have become widely used in earth-surface sciences due to their ability to generate high-resolution digital elevation models in a cost-effective and flexible way (Westoby et al., 2012; Haas et al., 2016; Torre et al., 2025). However, despite their increasing popularity and applicability, it's not free from limitations that must be carefully considered when interpreting quantitative results. A key limitation lies in the fact that reconstructs the visible surface of the terrain, meaning that ground elevation is not always directly captured, particularly in areas affected by vegetation cover or surface occlusions. As a consequence, vegetation can introduce significant disturbances in point cloud reconstruction, and ground elevation may be poorly represented or partially missing. In addition, although aerial platforms such as UAVs allow efficient coverage of large and remote areas, the quality of the resulting point clouds can still be affected by acquisition conditions, image geometry and local imperfections (Crema et al., 2020).

During the ten-year observation period, three UAS with different characteristics were used, including sensor type, image resolution, and the ability of real-time kinematic (RTK) positioning (see the supplementary materials for details). Since all UAS used are equipped with a mechanical shutter and the data recordings were carried out with the utmost care (James et al., 2019), the expected differences in quality in surface reconstruction (SfM-MVS) due to the use of different UAS are considered negligible (Stark et al., 2021). The results of the accuracy analyses (chapter 4.1.1) with a maximum standard deviation of 0.12 m support this hypothesis. The influences resulting from highly dynamic environmental conditions, such as varying wind and light conditions, can also be classified as negligible after careful quality control of all data sets. However, it should be noted that the uncertainty estimated in stable, flat areas likely represents a lower-bound estimate of the actual uncertainty. In steeper, rougher, and vegetated terrain, uncertainties could increase due to complex morphology (roughness) and slope. Future studies could further investigate the spatial variability of uncertainty through dedicated check-point measurements, GCP residual analyses, or repeated same-day surveys under unchanged terrain conditions.

Due to the separation, classification, and extraction of vegetation from the point cloud using various filters based on object geometry and coloring the influence of vegetation growth on the sediment balancing could also be minimized. However, due to the rapid expansion of vegetation, this must be taken into account. For example, repeated field observations showed that many deposition areas of both, gravitational and water-erosion processes are recolonised with vegetation within one year. This can be particularly relevant when interpreting the results of change detection analyses, such as volumetric estimates of sediment balancing, as vegetation regrowth can locally increase surface

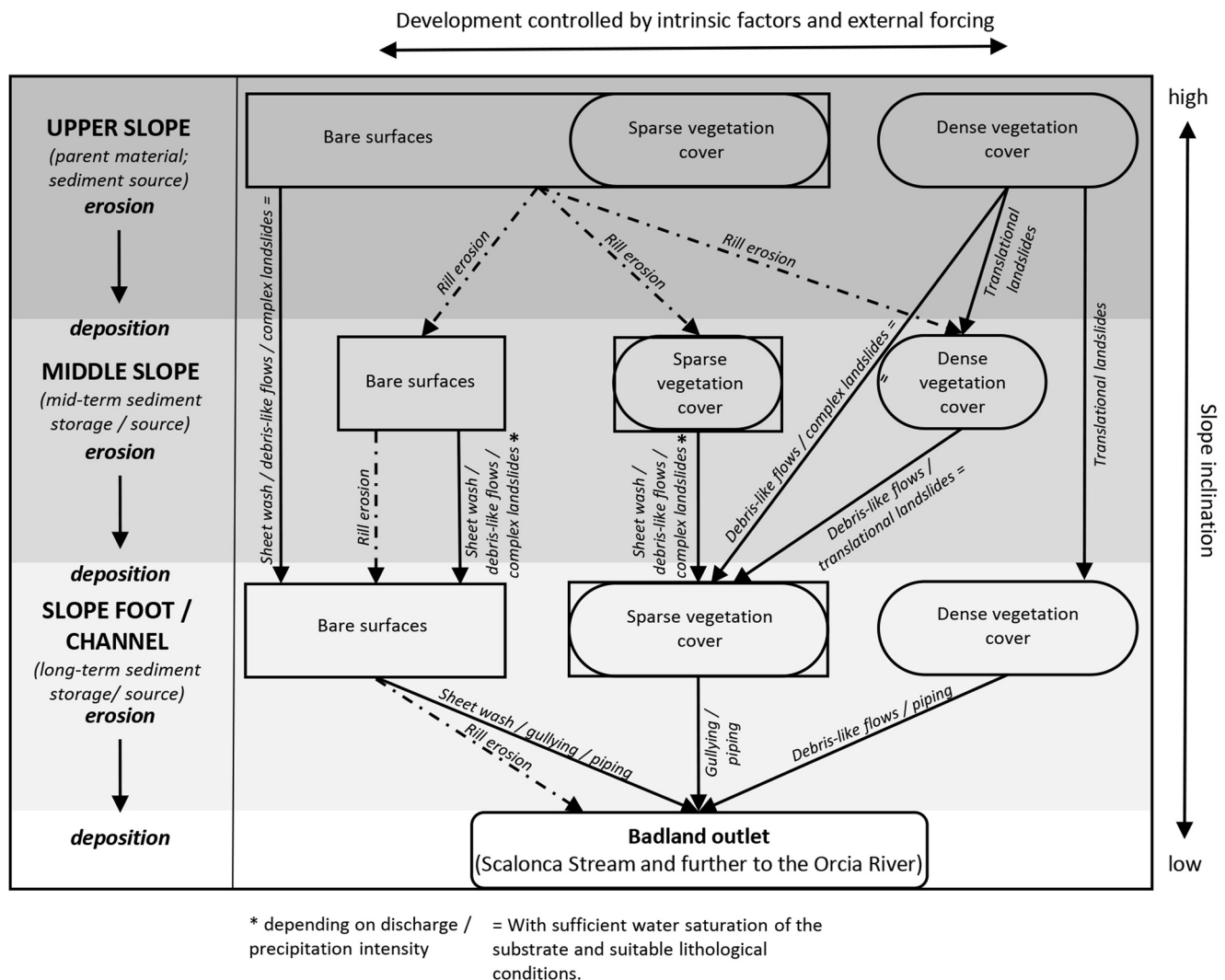


Fig. 15. Simplified representation of the complex interaction of geomorphological processes responsible for the landscape development of the *calanchi* badlands in the upper Val d'Orcia for different precipitation events and soil cover. The dotted line represents moderate precipitation events, and the solid line represents extreme events or prolonged rainfall that exceed intrinsic system thresholds and drive high-magnitude geomorphic responses.

complexity and influence the DEM uncertainty. In the context of the present study, this applies, for example, to some sub-areas of the 2022 dataset where very dense ground vegetation could not be fully removed, unlike sparsely distributed vegetation (bushes and trees), thus contributing to the positive net balance observed in the raw data for that time period (see Table 3). However, most surface changes are due to meadow expansion and fall below the LoD (see Table 2) and can therefore be excluded from the sediment balancing. In this context, the accurate interpretation of identified changes (including metrics, geometries and shapes) is of crucial importance.

Finally, regarding the multitemporal geomorphic mapping of process signatures, it should be noted that some processes may not show detectable signatures, in certain time spans, even when they are not completely inactive. For example, in the case of water-driven erosion processes (e.g., gully and rill systems), cut-and-fill behaviour (Faulkner et al., 2008) may redistribute erosion and deposition along channels, so that alternating incision and infilling can result in weak or near-zero net topographic change between survey dates. As a consequence, across multi-temporal observation window, geomorphic activity may persist without producing a clear or continuous signature in the DoD-derived products, potentially leading to an underestimation of geomorphic activity in the mapping results.

5.2.2. Semi-automatic mapping of vegetation from RGB data

Although the TGI Index represents a promising technique, it is affected by several errors related to both lighting conditions and the phenological and physiological state of the vegetation. For example, shadows caused by clouds or physical obstacles affect the colour value of data and might lead to miscalculations (Hunt et al., 2011; Polat et al., 2024). Furthermore, even vegetation that is not distinctly green, such as *Trifolium hirtum*, which is common in the study area, or plants that are dead from the previous season or undergoing vegetative stress due to the typical Mediterranean summer drought may be misinterpreted because of their brownish tones. These conditions are particularly important to consider when seeking a uniform value across a multitemporal dataset to distinguish between vegetation and bare substrate. Since all datasets were collected in the same season (May to June), such errors are considered low.

6. Conclusion and outlook

This study utilised high-resolution, multi-temporal analyses to investigate the geomorphological dynamics of a subhumid *calanchi* badland in the upper Orcia river basin (southern Tuscany, Italy) over a period of 10 years. The primary focus is on the interaction between geomorphological processes and vegetation, along with precipitation

dynamics. The integration of field-based geomorphological surveys, geomorphological interpretation of multi-temporal DOMs, and comparison with the derived DSMs and DoDs enabled the quantification and characterization of the geomorphological evolution of the catchment.

The period between 2014 and 2024 can be characterised as a phase of badland stabilisation. This tendency is most clearly observable in the decline of geomorphological active regions, particularly with regard to water erosion processes, accompanied by an expansion of vegetated areas. In 2014, the study area was characterised by a significant proportion of bare ground and sparsely vegetated areas. However, 10 years later, the situation has undergone a marked reversal. The most significant increase in the extent of vegetated areas was observed along the primary channel and on the south-south-eastern slopes. Although the erosion process area decreased over time, the erosion rates remained relatively constant. This indicates a shift from high-frequency to high-magnitude processes in the most recent observation period. The precipitation analysis did not reveal any significant trends, indicating that even in the absence of detectable changes in the meteorological regime, substantial landscape-scale transformations and complex mass-movement processes may still occur. In the context of long-term observations, these changes could be interpreted as temporary manifestations of an underlying cyclical behaviour that is driven more by precipitation recurrence periods and interactions in the process regime (process-response reactions) than by a sustained, monotonous precipitation trend.

Due to the small cross-section, which in some cases measures less than 1 m, and the dense vegetation in the lower reaches of the *Scalonca* stream, it is assumed that sediment export from the catchment is limited to very fine grain sizes. The high clay and silt content in the substrate suggests that this seemingly marginal process can have a significant impact on the sediment budget of the catchment.

Quantitative observation of the changes observed in the lower reaches of the *Scalonca* stream proved to be a challenge using the methods used here. This is due to the complex geometries and the dense vegetation cover. The methods presented in this study (UAS SfM-MVS) only allow a limited analysis of these changes. Future work should therefore focus more on event-based studies in order to quantify the sediment input into the *Scalonca* stream and its receiving watercourse, the *Orcia* river, as a result of extreme precipitation events.

Continuous monitoring of these areas, utilizing automated cameras as part of a 4D approach, in conjunction with suspended sediment measurements at the outlet of the badlands, would facilitate the observation and analysis of future geomorphological developments on an event-based basis. In addition, future work should include the use of complementary sensors, such as LiDAR enabling DTM creation (bare ground surface/ elevations). Additionally, the integration of multi-spectral imagery is recommended to facilitate the characterization of vegetation cover and its dynamics. This would contribute to a deeper understanding of the geomorphological evolution of the sub-humid *calanchi* badlands. This is applicable to both the further investigation of the seasonal components of geomorphological systems of Mediterranean badlands and the study of connectivity between slope and channel systems. Given the increasing geomorphological activity in the middle and lower slope areas, we assume that erosion activity will also remain active in the higher slope areas, thereby further strengthening the functional sediment connectivity of the system (Vergari et al., 2019) and to a continuous spatial enlargement of the badland.

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CRedit authorship contribution statement

Manuel Stark: Writing – original draft, Visualization, Project administration, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **Annalisa Sannino:** Writing – original draft, Visualization, Methodology, Investigation, Formal analysis, Data

curation, Conceptualization. **Martin Trappe:** Writing – original draft, Resources, Methodology, Investigation, Data curation. **Jakob Rom:** Software, Methodology, Formal analysis, Data curation, Conceptualization. **Jakob Forster:** Investigation, Data curation. **Georgia Kahlenberg:** Investigation, Data curation. **Florian Haas:** Validation, Supervision, Conceptualization. **Francesca Vergari:** Writing – original draft, Validation, Supervision, Investigation, Conceptualization.

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Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

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Data availability

[Starketal2026_Calanchi_data \(Original data\)](#) (OneDrive folder with access on demand)

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