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ORIGINAL PAPER

Analysis of topographic parameters underpinning landslide occurrence in Kigezi highlands of southwestern Uganda

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Abstract

An assessment of the influence of topography on landslide occurrence in the Kigezi highlands of southwestern Uganda was conducted. Whereas the frequency and magnitude of landslides in these highlands are on the increase, the topographic attributes underpinning landslide occurrence are not well understood. Sixty-five landslide scars were surveyed and mapped to produce landslide distribution maps. Specific topographic parameters, namely slope gradient, profile curvature, topographic wetness index (TWI), stream power index (SPI), and topographic position index (TPI), were assessed on landslide slope sites. The attributes were parameterized in the field and GIS environment using a 10-m DEM. Landslides were noted to concentrate along narrow topographic hollows, as opposed to broad concave slopes in the landscape. The occurrence is dominant in slope zones where slope gradient, profile curvature, TWI, TPI, and SPI are 25°–35°, 0.1–5, 8–18, −1–1, and > 10, respectively. It was established that profile curvature and slope gradient are the most and least significant topographic parameters in landslide occurrence ($R^2 = 0.802$, p value = 0.088 and $R^2 = 0.5665$, p value = 0.057), respectively. An understanding of these topographic underpinnings would serve to identify and predict potential landslide zones within the landscape and enhance landslide hazard mitigation.

Keywords Topographic parameters · Landslide occurrence · Kigezi highlands

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1 Introduction

Landslides are hydro-geotechnical processes most prevalent in highland and mountainous terrain of tropical and subtropical regions (Broothaerts et al. 2012; Kirschbaum and Zhou 2015; Kirschbaum et al. 2015). They are triggered under multiple geological and morpho-hydrological conditions (Kitutu et al. 2009; Susana et al. 2017). East African highlands are considered to be prone to landslides due to their humid climate, steep topography, and high weathering rates (Knapen et al. 2006; Kitutu et al. 2009). The highlands have experienced a number of landslide occurrences, some of which have been catastrophic (NEMA 2014). Several studies have been conducted on landslide occurrence in these highlands (Bagoora 1988; Knapen et al. 2006; Kitutu et al. 2009, 2011; Mugagga et al. 2012; Mertens et al. 2015). However, identifying the factors controlling landslide distribution and determining their impacts may be difficult (Claessens et al. 2007; Liesbet et al. 2015).

Landslide studies in the East African highlands have focused on the role of anthropogenic factors, rainfall amounts, and slope materials as casual factors (Knapen et al. 2006; Kitutu et al. 2009; Mugagga et al. 2012). There has been limited focus on the topographic parameters underpinning landslide occurrence in these highlands. Topography is often considered as the most important parameter in landslide occurrence (Loos and Elsenbeer 2011). Topographic characteristics of any region have a greater influence on landslide occurrence than other parameters, including soil and land cover (Loos and Elsenbeer 2011; Broothaerts et al. 2012). According to Gao and Maro (2010), topography has been widely reported to bear a close association with landslides. The exact control of topographic settings on landslides is usually analyzed through spatial overlay of landslide distribution maps on topographic layers (López-Davalillo et al. 2014). The major topographic parameters that influence landslide occurrence include slope gradient, aspect, curvature, roughness, distance from drainage network and discontinuities (Broothaerts et al. 2012). Other complex topographic parameters important in landslide occurrence include the topographic wetness index, stream power index, slope length, and topographic position (Infascelli et al. 2013). These parameters vary in space depending on terrain, necessitating site-specific examination to gain an understanding of the underpinnings of landslide occurrence (Kirschbaum et al. 2016).

Although the frequency and magnitude of landslides in the Kigezi highlands of southwestern Uganda have increased (NEMA 2014), topographic parameters underpinning landslide spatial distribution are not well understood. An understanding of the individual and collective role of these parameters should increase community awareness of the landslide risk. Increased community knowledge of the landslide risk could help alleviate the vulnerability and enhance resilience of local communities to landslide hazards. The present study therefore examines the influence of selected topographic parameters on landslide occurrence in this part of the country. The interplay of topographic parameters and their implications for hillslope hydrology and soil characteristics is inferred.

2 Materials and methods

2.1 Study area

The study was conducted in the non-volcanic Kigezi highlands of southwestern Uganda situated between 01°21′25″ and 0°58′08″ south and 29°43′30″ and 30°05′51″ east. The study concentrated on the non-volcanic Kigezi highlands because most studies on landslide occurrence in Uganda have largely focused on Mount Elgon in Eastern Uganda (e.g., Muwanga et al. 2001; Kitutu et al. 2004; Knapen et al. 2006; Claessens et al. 2007; Mugagga et al. 2011; and Mugagga et al. 2012) which is volcanic in nature and therefore with a different geological, ecological, and topographic setting. The topographic and ecological setting of these non-volcanic Kigezi highlands is different from Mt Elgon landscape due to their mode of geological formation and therefore with a different response landslide triggers. This study therefore aimed to explore landslide occurrence in the uplifted non-volcanic environments of Kigezi highlands due to their unique, ecological, soil properties, and topographic setting. The topography of Kigezi highlands is similar throughout the landscape comprising mainly extensive flat-topped ridges and hills, broken by short numerous steep-sided deep subsidiary strike valleys separated by fluted spurs, usually 3–6 km (Ollier 1969). The topography is extremely rugged, consisting of narrow steep convex (20°–45°) and gentle (10°–15°) slopes (Bagoora 1993). The landscape has steep slopes and deep narrow valleys with an amplitude of 600–700 m or greater with gentle pediment slopes (NEMA 2008). Since the topography is almost the same throughout the landscape, a representative area was ideal to describe the entire landscape. Rukiga catchment was therefore selected on the basis of its unique topography which is extremely rugged, consisting of narrow steep convex slopes and topographic hollows between hills, many of which constitute drainage lines. This topography is what is observed everywhere else in the Kigezi highland region. Rukiga catchment was selected on the basis that it is also an area where landslide scars are still visible and therefore easy to map and characterize. In other parts of the highlands, landslide scars are no longer visible on the landscape due to the high rates of vegetation regeneration.

The geology is composed of sedimentary rock system of the Precambrian age (Bagoora 1989). They are collectively grouped into the Karagwe-Ankolean system, underlain by metamorphic gneiss and granite intrusions, as well as shale and phyllite, which have given rise to clay deposits (Bagoora 1993). The Kigezi highlands which are largely non-volcanic in nature are covered by non-volcanic soil properties (Bagoora 1988). The behavior of such non-volcanic soil materials to incoming rain is different from the volcanic soils. The climate of Kigezi highlands is warm to cool humid characterized by a bimodal rainfall pattern with annual rainfall of 1092 mm (UBOS 2017), which can be classified as moderate (NEMA 2010). Rainfall, however, increases to 1250–1540 mm or more in high-altitude areas of greater than 2000 m above sea level (NEMA 2017). The main rainfall seasons are from mid-February to May with a peak in March–April, and September to December with a peak in October/November (NEMA 2014). The vegetation cover of these highlands was until about a century ago characterized by montane forests (Bagoora 1993). Human interference has led to serious degradation and in some cases depletion of vegetation cover (NEMA 2014). Presently, the common vegetation cover in the highlands is characterized by eucalyptus trees, pines, shrubs, and thickets (NEMA 2016).

2.2 Mapping landslide dimensions and spatial distribution

The Rukiga catchment (Fig. 1) was delineated using water delineation tools in ArcGIS. The drainage network and watershed information were generated automatically from a 10-m digital elevation model (DEM) using the Arc Hydro tools in ArcGIS 10.1. A landslide inventory was carried out with the guidance of the local community to identify, map, and characterize landslides. Many of the landslide scars in the area are still visible

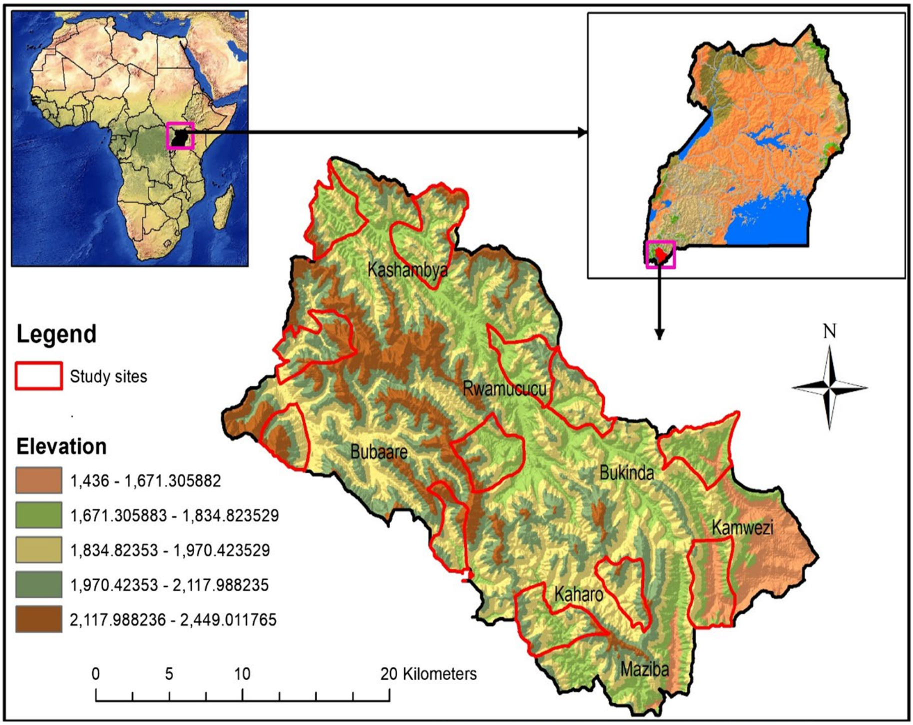

Fig. 1 Location of the study area indicating study sites

and therefore easy to map and characterize. Some of the scars have been concealed by the high rates of vegetation regeneration in the catchment. Analysis of landslide geometry and spatial distribution was done using both field surveys and GIS desktop techniques. Detailed field investigations were carried out to establish the magnitude and spatial distribution of landslides in the Kigezi highlands. Field surveys, capture of coordinates, and measurements of landslide dimensions were done using tape measures and hand-held GPS receivers (Fig. 2). Coordinates for the mapped landslide scars were exported to ArcGIS 10.1 software to produce landslide distribution map for the study area. Spatial relationships between landslide sites and topographic surface maps were derived. Historical landslide data and records were obtained from aerial photographs of scale 1:20,000 for 1985, 1995, and 2005, and Kabale District Local Government Environmental Reports for 2008, 2012, and 2015, as well as interaction with the local community. Landslide scar details from the aerial photographs were scanned and digitized, and their locations were analyzed using ArcGIS software.

2.3 Topographic parameterization

A 10-m DEM for the region was used to extract major topographic attributes in the TOPMODEL framework, where the role of topography in catchment hydrology can be explored (Quinn et al. 1995). The DEM was interpolated from 10 m interval contours based on 1:20,000 topographic maps, using the Topo to Raster tool in ArcGIS 10.1. Topographic

Fig. 2 Field investigations to map and measure landslide scars and their distribution

parameterization was performed using ArcGIS 10.1 and Systems for Automated Geoscientific Analyses (SAGA) GIS 2.3.1 software, following Conrad et al. (2015). Topographic analyses involved the computation and establishment of selected topographic parameters, namely slope gradient, slope shape (profile form curvature), topographic wetness index (TWI), stream power index (SPI), and topographic position index (TPI). These parameters were prioritized due to their implications for hydrological and pedologic properties that influence landslide occurrence.

Surface profile curvature, identified during field surveys as the dominant curvature form closely associated with landslide occurrence, as opposed to plan curvature, was focused on in the present study. According to Carson (1985), profile curvature affects the driving and resisting stresses within a landslide in the direction of motion. Curvature values were used to identify peaks, ridges, spur slopes, planar regions, topographic hollows, and pits. The TWI as a surrogate for saturation levels (Grabs et al. 2009) and susceptibility to landslide occurrence was established using the formula:

$$\mathrm{TWI} = \ln\ (\alpha / \tan \beta)$$

where α is the specific catchment area, and $\tan \beta$ is the slope angle (°).

High TWI values occur in concavities with the requisite upslope contributing area. The SPI which is closely related to the TWI is a measure of erosive power of water flow and closely related to the TWI was calculated using the formula $\mathrm{SPI} = \ln(\alpha * \tan \beta)$. The erosive power was used to identify potential sites for landslide occurrence. Both the TWI and SPI were used to quantify flow intensity and accumulation potential within the highlands in the study area.

The TPI which measures the difference between elevation at the central point (ZO) and the average elevation (Z) around it within a predetermined radius was calculated using the formula: TPI = ZO − Z. The Weiss (2001) method of landform classification was used, and six discrete classes were defined. Following Seif (2014), TPI was used to determine topographic slope positions in relation to landslide hazard zones and automate landform classification. Maps of the SPI, TWI, and TPI of the catchment were generated and overlain with landslide scars to derive relationships with landslide distribution.

Statistica software package was used for inferential statistics to describe and compare topographic parameters vis-a-vis landslide distribution. To determine the influence and significance of topographic parameters on landslide occurrence and distribution, a linear regression model was fitted and p values were determined. Linear regressions were used to determine the coefficients of determination (R^2) for topographic parameters in relation to landslide occurrence.

2.4 Analysis of soil profile and topographic characteristics

Given the variations between soil horizons with topographic characteristics and landslide occurrence in the study area, onsite soil profile analysis was performed along different slope positions, gradient clusters, and topographic configurations. The slope positions considered in soil profile analysis included valley bottoms, lower middle, upper middle and uppermost, as well as along spurs and topographic hollows with different slope gradients. Profiles in the upper slope sections were dug up to a depth of 1–1.5 m, while those in the middle slopes ranged between 2 and 4 m and the lower slope soil profiles were greater than 5 m. Soil depth was classified into three broad groups of shallow, medium, and deep. Soil horizons were analyzed in detail with a specific focus on depth.

3 Results

3.1 Landslide characteristics and distribution

The spatial distribution of landslides in the study area is illustrated in Fig. 3. The landslide scars in the catchment varied from small (12.5 m) slides to longer, complex flows that extended to more than 890 m. In most of the observed landslide scars, channel morphometry is shallow and narrow with the depletion/rupture zone either devoid or partially covered by a thin veneer of debris. The average width, depth, and area of the landslide scars ranged from 0.9 to 17.5 m, 0.5 to 5.3 m and 125 to 6000 m^2, respectively ("Appendix").

The estimated volume of hillslope materials displaced by individual landslides varied widely. Small landslides displaced approximately 62.5 m^3, while large occurrences displaced close to 30,000 m^3 of material ("Appendix"). A unique attribute about landslides in these highlands is the larger failure zone than the toe, as opposed to the Mt Elgon slopes, Eastern Uganda, where landslide scars are smaller at failure zones and larger toward the toe (Knapen et al. 2006; Kitutu et al. 2009, 2011). This landslide characteristic is attributed to the topographic characteristics of the study area where landslides follow hollows between spur slopes. The influence of topography on landslide characteristics is unraveled in the discussion section. Most of the studied slides initiated at mid-slopes rather than at the shoulder or top of the slope. The crown and areas close to the main scarp are commonly marked by the presence of acute, open tension cracks that are between 30 and 100 cm wide and were observed at or close to the head scarps of most of the slides. In most of the observed landslide scars, the bedrock was not exposed.

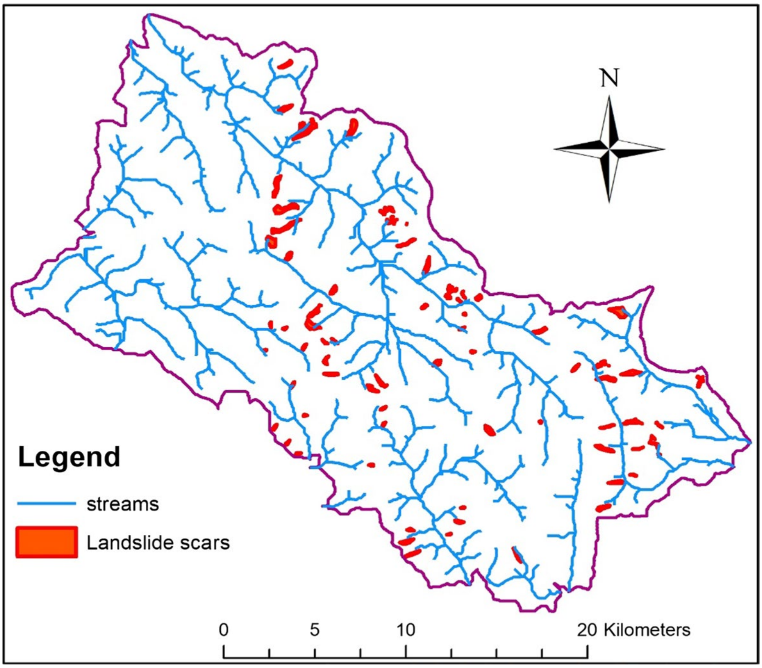

Fig. 3 A landslide distribution map for Rukiga catchment

3.2 Topographic characteristics and landslide occurrence

3.2.1 Slope gradient

More than 60% of the landslides in the Kigezi highlands were noted to occur on slope gradients between 25° and 35° (Fig. 5), yet this gradient category accounts for only 28% of the area's topography. The occurrences were less pronounced within slope gradient categories of <15° and >45° (Fig. 4). Out of 65 landslide scars surveyed, 60% occurred on zones with gradient between 25° and 35° (Fig. 5). Landslide occurrences increased from 4% on moderate slopes to 22% on moderately steep slopes, sharply increased to 60% on steep slopes, and fell to 10% and <4% on very steep and precipitous slopes, respectively. This distribution and the statistically significant relationship between landslide incidence and slope gradient ($R^2=0.675$, p value = 0.057) demonstrate the critical role of gradient as a control on landslide occurrence.

3.3 Slope curvature

Profile curvature values computed for the catchment ranged from −3 to +5 (Fig. 4). Landslide occurrence is dominant in zones where profile curvature values lie between +0.1 and +5, along topographic hollows. This is further confirmed by a strong statistical relationship ($R^2=0.802$, p value = 0.088) between profile curvature and landslide occurrence (Fig. 5). Only 5% of the landslides occurred in zones with profile curvature values between

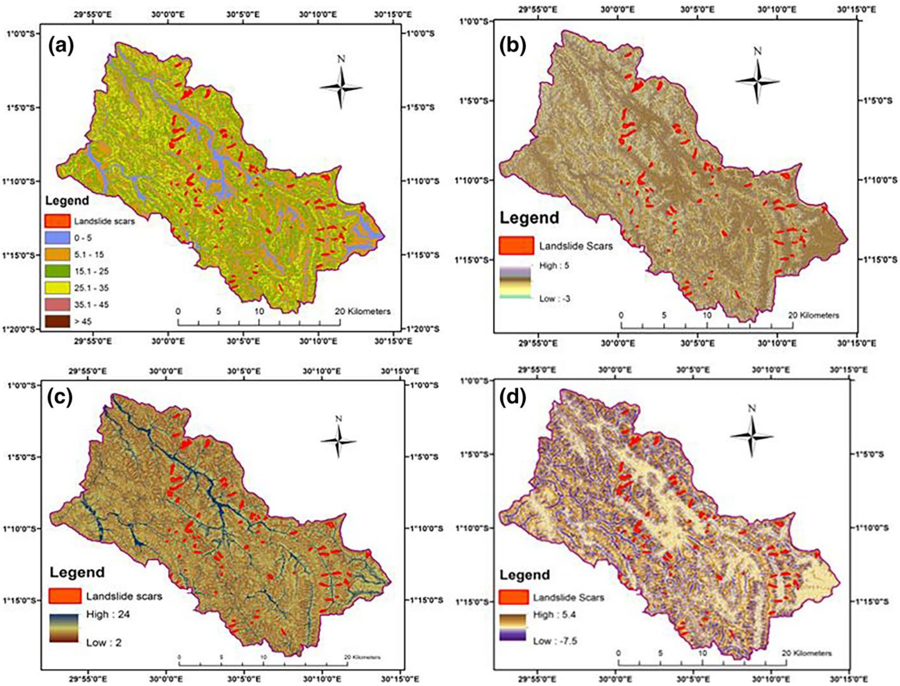

Fig. 4 Maps for selected topographic parameters and landslide distribution; **a** slope gradient, **b** profile curvature, **c** TWI, and **d** TPI

−0.2 and +0.1, which are conspicuous concave zones. Slope zones with profile convex curvature forms with values lower than −0.2 mainly spur slopes and hilltops are devoid of landslide occurrence (Figs. 4 and 5).

3.4 TWI, TPI and SPI

An overlay of TWI, TPI, SPI, and landslide distribution maps revealed that landslide occurrence was dominant in zones with high TWI values, ranging between 8 and 18 ($R^2=0.763$, p value = 0.077), TPI values between −1 and 1 ($R^2=0.7009$, p value = 0.065) and SPI values > 10 ($R^2=0.741$, p value = 0.0671), in all cases along profile concave zones and topographic hollows (Figs. 4 and 5). Landslide concentration increases from 4% to 58 in the lower, lower-middle, upper-middle slope elements, respectively. It declines rapidly to 4% in the uppermost slope elements (see Fig. 6).

It is noteworthy, however, that there is a convergence among the topographic parameters within the landscape to induce landslides, as high SPI (> 10) and TWI (> 8) values corresponded with profile concave forms mainly along topographic hollows, associated with the highest incidence of landslides in these highlands. This interaction is explored further in the subsequent section.

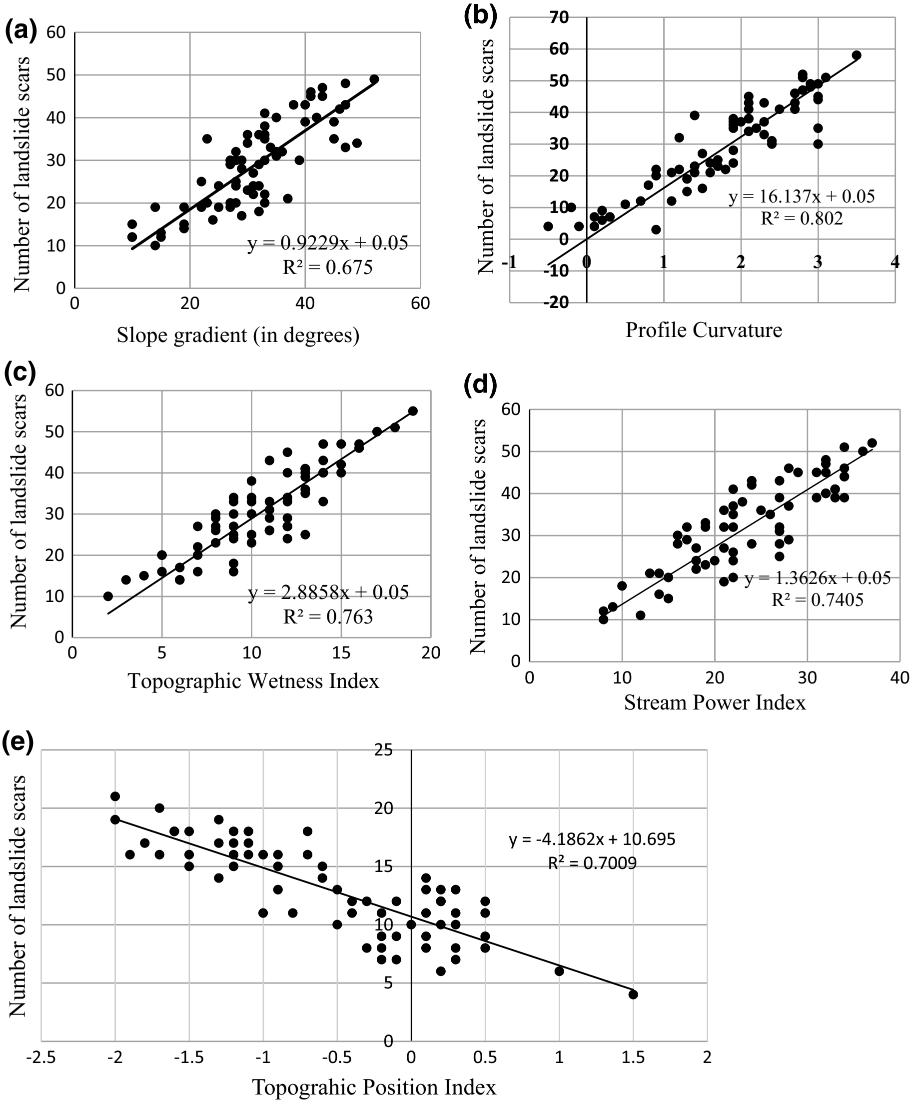

Fig. 5 Relationship between selected topographic parameters and landslide distribution; **a** slope gradient, **b** profile curvature, **c** TWI, **d** SPI, and **e** TPI

3.5 Soil depth and topographic characteristics

Shallow soil groups were noted as less than 0.85 m in depth and occured on very steep (35°–45°), and precipitous (> 45°) slopes. Medium and deep soil profiles of 1.5–4 m and greater than 6 m were observed to occupy mid-slopes along topographic hollows and lower slopes, respectively. A relationship exists between soil depth and slope gradient and positions as well as topographic configurations. Soil profile description along the slope profile revealed that soil depth decreased with an increase in slope gradient. Whereas very deep

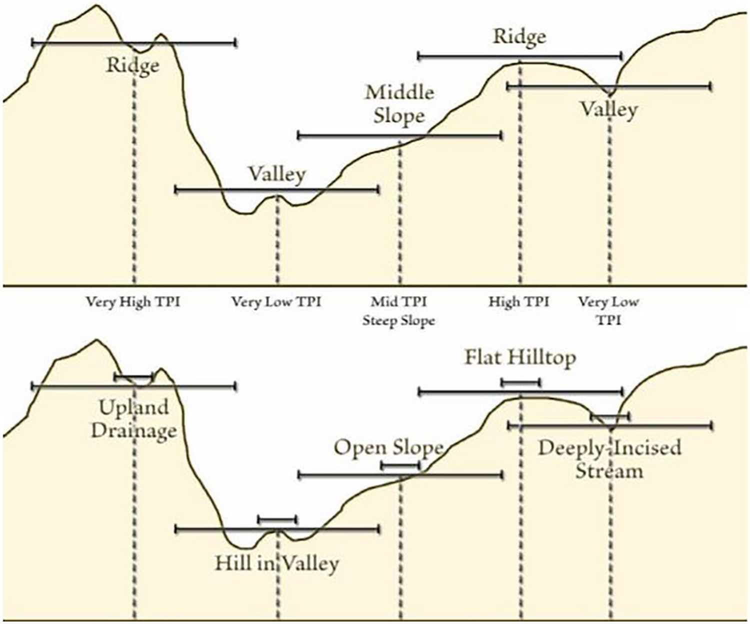

Fig. 6 Slope positions and landforms based on TPI values

soils (>5 m) are found on slope sections with gradients lower than 10°, moderately deep soils (2–5 m) lie on slope angles between 15° and 35°. Very steep slopes with gradients greater than 35° are associated with shallow skeletal soils of less than 1 m. A relationship was also established between soil depth and slope curvature. Very deep, deep, and moderately deep soil profiles (>2.5 m) are found along concave profile forms within topographic hollows. Convex profile forms along hilltops and spur slopes tended to have shallow skeletal soils of less than 1 m.

4 Discussion

4.1 Landslide characteristics and distribution

The morphometric characteristics of the landslide scars observed are important in establishing the magnitude and type of the landslide occurrence. According to the landslide classification scheme by Cruden and Varnes (1996), the most common landslide processes in the study area are rotational slides, where the surface of rupture is curved concavely downward. The geometry of the slide depletion zones varies widely throughout the study area. Whereas the volume of generated debris on the foot slopes of Mt Elgon in Eastern Uganda is dependent on landslide depth (Knapen et al. 2006), it is related to the length and width of the depletion zone in the Kigezi highlands. More than 80% of the landslide scars observed were narrow and shallow because they occurred in small topographic hollows. Such small hollows gather limited materials from the surrounding hilltops and spur slopes and therefore have shallow soil profiles (Raju and Nandagiri 2015). About 10% of the landslide scars were identified as deep seated and wide; they occurred in deeper and wider topographic hollows. Such hollows have the capacity to gather more materials from the surrounding spur slopes and hilltops and are associated with deep soil profiles (Gao and Maro 2010).

4.2 Topographic parameters and landslide occurrence

4.2.1 Slope gradient

A close relationship exists between soil depth and slope gradient (Liang and Uchida 2014). Soil depth reduces with an increase in slope gradient and elevation (Selby 1993). In the present study, it was established through soil profile analyses that slope sections with lower and moderately steep slope gradients had deeper soil profiles than those with very steep gradients. This study found out that landslide occurrence increases with increase in slope gradient but then reduces as the slope gradient increases further. This distribution is due to the variation in soil depth with slope gradient. Landslide occurrence is concentrated on relatively steeper slopes (between 15° and 45°) due to the high downslope component which pulls materials. The concentration of landslides on such slope sections is explained by the accumulation of displaced materials from uppermost very steep and precipitous slopes (Capitani et al. 2013). Sections with gradients of $<15°$ have deeper soil profiles as well as very high soil moisture content, but lack a gradient steep enough to initiate movement of materials. Landslide occurrence is therefore low on such slope elements due to the less downslope force required to move materials (López-Davalillo et al. 2014). Such slope sections are only depositional areas for the landslide debris. Landsliding is a process that occurs within minutes and thus requires a steep gradient to facilitate the movement (NEMA 2012). Considering other factors constant, the minimum gradient required for landsliding in Kigezi highlands was found to be $>15°$.

Landslide occurrence was also found to be less pronounced on slope zones with gradients greater than 45°. Erosion processes are more intensive on such slopes, and hence, soils are shallower. Generally, the shallow soils on such slopes are easily washed away in the early stage of a rainfall event leading to exposure of weathered rocks (Loos and Elsenbeer 2011). Thin soils in such sections also imply limited materials for downslope movement and hence low landslide incidences (Nath et al. 2013). This explains why landslide occurrence in Kigezi highlands is low on slope elements with very steep and precipitous gradients. As observed by several scholars (Hungr et al. 2005; Hosseini et al. 2011; López-Davalillo et al. 2014), the typical slope gradient for landslide occurrence is between 27° and 38°. In keeping with these observations, slope gradient of the source areas in the present study is between 15° and 45°, with a concentration between 25° and 35°. The interplay between gradient and other topographic parameters should therefore play an important role in landslide inducement.

4.2.2 Slope curvature

The concentration of landslides in topographic hollows with profile concave forms ranging from 0.1 to 5 is explained by the availability of high moisture content and deep soil profiles. Whereas profile concave forms act as areas of high saturation rates, profile convex forms are high water shedding slopes and are dry (Infascelli et al. 2013). In the present study, it was established through soil profile analysis that topographic hollows consist of deep soils. The thick soils are mobilized from immediate convex topography and have the propensity to collect much infiltrated water during and after rainfall events. Saturation also makes soil materials in hollows unstable, resulting in downslope movement of slope materials (Gao and Maro 2010), particularly where profile concave forms interact with other topographic parameters, particularly those examined in the present study.

Landslides often occur in areas of convergent topography in which subsurface soil water flow paths give rise to excess pore water pressures downslope (Infascelli et al. 2013; Corominas 2014). The most significant statistical relationship obtained in the present study (($R^2=0.802$; p value 0.088) confirms the strong relationship between profile concavity and landslide occurrence. In concave forms, water flow is concentrated in topographic hollows (Grabs et al. 2009). This increases the soil moisture content and the duration of soil saturation (Gu and Wylie 2016). Topographic hollows also control the spatial distribution of saturated zones and the development of critical pore pressures capable of triggering landslides (Raju and Nandagiri 2015). They should therefore be considered hazard areas and potential sites for landslide occurrence in these highlands.

4.2.3 TWI, SPI, and TPI

As noted earlier, landslide occurrence is concentrated in sections with high TWI values ranging between 8 and 18, also located along topographic hollows with profile concave forms. High TWI leads to saturation which reduces the shear strength of the materials, inducing slope failure (Ali et al. 2014). Valley bottoms have the highest TWI values (>18) due to greater amounts of water received from upslope areas. Studies by Chi et al. (2009) and Grabs et al. (2009) also confirm that valley bottoms have the highest saturation rates. This is due to low slope gradient which does not allow water to drain away easily. Such high saturation rates, however, do not result in landslide occurrence within the valley bottoms (Nath et al. 2013). Similarly, it was established in the present study that valley bottoms have the highest saturation rates but are stable. They lack the required threshold slope gradient ($>15°$ for Kigezi highlands) to initiate downslope movement of materials. Therefore, landslide occurrence is predominant in topographic hollows with high TWI values and gradient steep enough to initiate downslope movement of saturated materials.

Topographic hollows are also sections with high SPI, as an indicator of sediment transport capacity. High SPI values indicate areas in the landscape that have a high potential for erosion during and after rainfall events (Buda 2013; Gu and Wylie 2016). A combination of high TWI as an indicator of potential for landslide occurrence and SPI in topographic hollows explains the concentration of landslides in such slope zones (Ferreira et al. 2015). Such zones have high saturation rates, with higher erosive power and therefore high incidence of landslides.

TPI is important for slope stability analysis aimed at identifying preferential zones for landslide occurrence (Tagil and Jenness 2008). Landslides were identified as dominant on slope positions with TPI values between -1 and 1. However, they are less pronounced in slope sections with high TPI values which are also associated with low TWI and SPI values, mainly along hilltops and spur slopes. As noted earlier, landslide concentration is high (58%) along the middle slope elements. Characterized by topographic hollows, the middle slope positions are convergence zones for eroded materials from hilltops and spur slopes (Mokarram et al. 2015). This promotes saturation, which increases susceptibility to landslide occurrence within such slope positions (Seif and Mokarram 2014).

A convergence zone exists within the landscape where the parameters analyzed interact to induce landslides in the study area. Profile curvature was noted as the statistically most significant parameter. Landslides in Kigezi highlands occur along clearly defined lines in the form of topographic hollows with moderately steep slopes, high TWI and SPI, low TPI, as well as profile concave forms. Conversely, on the Mt Elgon slopes of Eastern Uganda, landslides do not necessarily occur along hillslope hollows. They have been noted to occur mainly on even broad concave elements of slope (Mugagga et al. 2012). This points to the differences in topographic parameters that underpin landslide occurrence in different mountain environments of the country.

5 Conclusion

The volume of debris generated by landslides on slopes of the Kigezi highlands is related to the length and width of the depletion zone. The concentration of landslides on moderately steep slope gradients is explained by the accumulation of displaced materials from uppermost steep slopes. Landslide occurrence is concentrated in topographic hollows with moderately steep slopes, high TWI and SPI, low TPI and profile concave forms. Such topographic settings encourage convergence of moisture and accumulation of soil. They are also associated with high saturation rates and erosive power, signifying reduced shear strength of slope materials. Notwithstanding the high TWI values (> 18), hence high saturation rates associated with valley bottoms, they are stable, as they lack the required threshold slope gradient (> 15° for Kigezi highlands) to initiate downslope movement of materials.

Distinct differences in topographic underpinnings emerge between Mt Elgon slopes of Eastern Uganda and the Kigezi highlands. Whereas landslides occur mainly on broad concave slope elements and the volume of generated debris on slopes of the former is dependent on landslide depth (e.g., Knapen et al. 2006; Kitutu et al. 2009; Mugagga et al. 2012), they are confined to profile concave hollows and debris volume is related to the length and width of the depletion zone on slopes of the latter. As opposed the landscape of Mt Elgon, where landslides occur on any part of the landscape (Kitutu et al. 2011; Mugagga et al. 2011), in the Kigezi highlands, the landslide path can be easily identified and demarcated within the landscape. In Kigezi highlands, compared to other slope elements in the landscape, topographic hollows are conducive for landslide processes. An understanding of the interaction among the respective topographic parameters and their influence on landslide occurrence is important in landslide hazard mitigation. This would permit the prediction of potential landslide zones and demarcation of safer zones for community land use activities. It would also mitigate community vulnerability to landslide hazards in this fragile highland ecosystem.

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Appendix: Landslide scar geometric characteristics

Landslide scars	Landslide scar dimensions (m)			Area of the landslide scar (m^2)	Volume of the scar (m^3)	Gradient (°) range at slide failure zone
	Average width (m)	Average depth (m)	Length (m)			
1	3.7	1.7	402	1487	2462	25–28
2	9.66	2.3	463.5	4477	6064	30–32
3	17.5	0.74	350	6125	4427	30–34
4	2.1	1.2	602	1264	3773	26–27
5	8.5	2.245	14.1	120	258	30–33
6	10	5	600	6000	30,000	33–35
7	10	5	400	4000	20,000	32–34
8	10	5.3	498	4980	26,394	31–33
9	16.6	4.3	315	5229	22,600	33–35
10	10	0.5	12.5	125	63	23–25
11	5.6	0.85	525.1	2941	1676	42–45
12	5.8	2.3	530	3074	10,508	39–42
13	2.7	1.8	885	2390	3669	33–36
14	3.14	1.52	786	2468	4341	30–33
15	2.95	1.6	784	2313	4518	34–37
16	6.2	2.8	835	5177	14,496	35–37
17	4.33	2.8	752	3256	9413	29–32
18	5	2.5	600	3000	7500	33–35
19	2.5	2.5	653	1633	4081	34–37
20	1.7	1.9	268	456	866	29–31
21	1.2	1.4	198	238	333	24–27
22	0.9	2.1	213	192	403	30–33
23	2.4	2	201	482	965	28–30
24	1.7	1.4	341	580	812	17–19
25	2.3	1.9	189	435	826	20–23
26	2.8	1.7	244	683	1161	26–29
27	1.9	1.2	196	372	447	31–34
28	2.4	1.5	204	490	734	15–18
29	2.8	1.9	302	846	1607	19–22
30	1.9	1.6	194	369	590	29–32
31	2.1	2	219	460	920	26–29
32	1.3	1.8	142	185	332	24–26
33	2.7	2.2	408	1102	2424	29–31
34	2.5	2	386	965	1930	34–37
35	1.6	0.9	125	200	180	26–29
36	1.7	2.1	184	313	657	20–23
37	2.4	2.2	296	710	1563	23–26
38	1.8	1.2	202	364	436	28–31
39	1.4	1.7	182	255	433	30–34
40	2.7	2.2	501	1353	2976	25–28

Landslide scars	Landslide scar dimensions (m)			Area of the landslide scar (m^2)	Volume of the scar (m^3)	Gradient (°) range at slide failure zone
	Average width (m)	Average depth (m)	Length (m)			
41	2.1	1.7	234	491	835	18–21
42	1.5	1.8	267	401	721	28–32
43	2.2	2.2	58	128	281	25–27
44	1.9	2.7	135	257	693	22–25
45	2.8	1.2	196	549	659	30–33
46	2.3	1.9	243	559	1062	28–31
47	4.2	0.9	55	231	208	26–30
48	2.9	1.1	129	374	412	33–35
49	3.2	0.7	231	739	517	29–32
50	3.1	0.8	89	276	221	20–22
51	2.8	2.1	197	552	1158	18–21
52	3.4	1.7	238	809	1376	17–19
53	3.2	1.2	345	1104	1325	32–34
54	2.8	0.8	118	330	264	16–18
55	1.8	1.1	102	184	202	18–20
56	3.6	2.8	189	680	1905	29–32
57	3.9	3.1	213	831	2575	25–28
58	3.2	2.7	96	307	829	27–29
59	1.8	1.2	47	85	102	22–25
60	1.2	0.8	66	79	63	20–22
61	5.9	2.1	138	814	1710	26–29
62	3.6	1.9	123	443	841	32–34
63	6.2	3.2	84	521	1667	29–32
64	7	1.7	73	511	869	23–25
65	4.2	1.3	144	605	786	27–30

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