A CHARACTERIZATION OF LANDSLIDE OCCURRENCE IN THE KIGEZI HIGHLANDS OF SOUTH WESTERN UGANDA

BY

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Abstract

The frequency and magnitude of landslide occurrence in the Kigezi highlands of South Western Uganda has increased, but the key underpinnings of the occurrences are yet to be understood. The overall aim of this study was to characterize the parameters underpinning landslide occurrence in the Kigezi highlands. This information is important for predicting or identifying actual and potential landslide sites. This should inform policy, particularly in terms of developing early warning systems to landslide hazards in these highlands. The present study analysed the area's topography, soil properties as well as land use and cover changes underpinning the spatial-temporal distribution of landslide occurrence in the region. The present study focussed on selected topographic parameters including slope gradient, profile curvature, Topographic Wetness Index (TWI), Stream Power Index (SPI), and Topographic Position Index (TPI). These factors were parameterized in the field and GIS environment using a 10 m Digital Elevation Model. Sixty five landslide features were surveyed and mapped. Soil properties were characterised in relation to slope position. Onsite soil property analysis was conducted within the landslide scars, auger holes and full profile representative sites. Furthermore, soil infiltration and strength tests, as well as clay mineralogy analyses were also conducted. An analysis of the spatial-temporal land use and cover changes was undertaken using satellite imagery spanning the period between 1985 and 2015.

Landslides were noted to concentrate along topographic hollows in the landscape. The occurrence is dominant where slope gradient is between 25˚ and 35˚, profile curvature between 0.1 and 5, TWI between 8 and 18, SPI >10 and TPI between -1 and 1. Landslides are less pronounced on slope zones where slope gradient is <15˚ and >45˚, profile curvature <0, TWI <8 and >18, SPI <10 and TPI >1. Deep soil profiles ranging between 2.5 and 7 meters are a major characteristic of the study area. Soils are characterized by clay pans at a depth ranging between 0.75 and 3 meters within the profiles. The study area is dominated by clay texture, except for the uppermost surface horizons, which are loamy sand. All surface horizons analysed had the percentage of sand, silt and clay ranging from 33 to 55%, 22 to 40% and 10 to 30% respectively. In the deeper horizons, sand was observed to reduce drastically to less than 23%, while clay increased to greater than 50%. The clay content is very high in the deeper horizons exceeding 35%. By implication, such soils with a very high clay content and plasticity index are considered as Vertisols, with a profound influence in the

occurrence of landslides. The top soil predominantly contains more quartz, while subsurface horizons have considerable amounts of illite/muscovite as the dominant clay minerals, ranging from 43% to 47 %. The liquid limit, plasticity index, computed weighted plasticity index (PI_w), expansiveness (ε_{ex}) and dispersion ranging from 50, 22, 17, 10 and 23 to 66, 44,34,54 and 64, respectively also have strong implications for landslide occurrence. Landslides are not normally experienced during or immediately after extreme rainfall events but occur later in the rainfall season. By implication, this time lag in landslide occurrence and rainfall distribution, is due to the initial infiltration through quartz dominated upper soil layers, before illite/muscovite clays in the lower soil horizons get saturated.

Whereas forest cover reduced from 40 % in 1985 to 8% in 2015, cultivated land and settlements increased from 16% and 11% to 52% and 25% respectively during the same period. The distribution of cultivated land decreased in lower slope sections within gradient group < 15˚ by 59%. It however increased in upper sections within gradient cluster 25˚ to 35˚ by over 85% during the study period. There is a shift of cultivated land to the steeper sensitive upper slope elements associated with landslides in the study area. More than 50% of the landslides are occurring on cultivated land, 20% on settlements while less than 15 % and 10% are occurring on grassland and forests with degraded areas respectively.

Landslides in Kigezi highlands are triggered by a complex interaction of multiple- factors, including dynamic triggers and ground condition variables. Topographic hollows are convergence zones within the landscape where all the parameters interact to cause landslides. Topographic hollows are therefore potential and actual landslide sites in the study area. Characterized by deep soil horizons with high clay content dominated by illite/muscovite minerals in the sub soils and profile concave forms with moderately steep slopes, topographic hollows are the most vulnerable slope elements to landslide occurrence. The spatial temporal patterns of landslide occurrence in the study area has changed due to increased cultivation of steep middle and upper slopes. Characterized by deep soil horizons with high clay content dominated by illite/muscovite minerals in the sub soils and profile concave forms with moderately steep slopes, topographic hollows are the most vulnerable slope elements to landslide occurrence.The spatial-temporal patterns of landslide occurrence in the study area has changed due to increased cultivation of steep middle and upper slopes. A close spatial and temporal correlation between land use/cover

changes and landslide occurrence is discernible. The understanding of these topographical, pedological and land use/cover parameters and their influence on landslide occurrence is important in land management. It is now possible to identify and predict actual and potential landslide zones, and also demarcate safer zones for community activities. The information generated about the area's topographic, pedological and land cover characteristics should help in vulnerability mitigation and enhance community resilience to landslide hazards in this fragile highland ecosystem. This can be done through designating zones for community activities while avoiding potential landslide zones. It is also recommended that, tree cover restoration be done in the highlands and the farmers encouraged to re-establish terrace farming while avoiding cultivation of sensitive steep middle and upper slope sections.

Keywords: landslides, topographic, soil properties, land use/cover changes, Kigezi highlands

DECLARATION

I, DENIS NSEKA (Student number: 215218345) hereby declare that this thesis for the degree of PhD (Environmental Geography) is my own work, composed and written by myself. It has never been previous presented anywhere for assessment of any academic award or published in any peer reviewed journal. All materials used from other sources are duly appreciated and properly acknowledged.

04th/November/ 2018

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DENIS NSEKA

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List of abbreviations and acronyms

Acronyms	**Meaning**
NEMA:	National Environmental Management Authority
FAO:	Food and Agriculture Organisation
DEM:	Digital Elevation Model

RS:	Remote Sensing
GIS:	Geographical Information Systems
SWAT:	Soil Water and Analysis Tools
NMU:	Nelson Mandela University
SAGA:	System for Automated Geoscientific Analyses
UTM:	Universal Transverse Mercator
GPS:	Global Positioning System
API:	Aerial photograph interpretation
SPSS:	Statistical Package for Social Scientists
UBOS:	Uganda Bureau of Statistics
USCS:	Unified Soil Classification System
USGS:	United States Geological Survey
TWI:	Topographic Wetness Index
SPI:	Stream Power Index
TPI:	Topographic Position Index
UNMA:	Uganda National Meteorological Authority
MAK:	Makerere University

CHAPTER ONE

General Introduction

1.1 General Introduction

Landslides are a global environmental hazard that are especially prevalent in mountainous and highland regions of the tropical world (Petley *et al*., 2005; Petley, 2008; Broothaerts *et al*., 2012; Kirschbaum and Zhou, 2015). Landslides are among the most widespread geological threats to lives and cause destruction of property globally (Broothaerts *et al*., 2012). In recent times, the risk from landslide related disasters in highland and montane ecosystems has increased both in frequency and intensity (Kirschbaum and Zhou, 2015). This is due to a combination of several attributes including geological, morphometric, climatic, and anthropogenic that directly or indirectly cause slope instability (Ayalew *et al*., 2004). Projections based on demographic changes in the midst of increasing variability in climatic conditions indicate that the situation may change for the worst (Petley, 2008). The expected increase in landslide disasters in future is due to over exploitation of natural resources, rapid deforestation, change in climate, an increase in hillslope population and uncontrolled excavation (Knapen *et al*., 2006 and Kitutu *et al.*, 2009). An increase in the occurrence of landslides globally has affected human life and damage to socio-economic infrastructures. This heightens the anxiety on how to sustain livelihoods in montane and highland ecosystems (Petley *et al*., 2005).

The most affected countries in terms of landslide occurrence and fatalities include India, China, Nepal, Indonesia, and the Philippines (Kirschbaum and Zhou, 2015). The number of damaging landslides have increased worldwide during the second half of 20th century (Broothaerts *et al*., 2012). For example in 2007 a total of 395 fatal landslide events were recorded, inducing a total of 3017 deaths worldwide while the period between 2003 and 2006 experienced 4399 fatal landslides (Petley, 2008) showing an increasing trend in occurrence and fatalities (Fig 1.2). Geographically, whereas 39% of fatal landslide events and 36% of landslide deaths occurred in South Asia, 24% of landslide events and 20% of landslide death occurred in South East Asia (Fig. 1.1). East Asia including China, accounted for 15.2% of landslide events (Petley, 2008; Kirschbaum and Zhou, 2015). In the coming years, landslides are expected to cause more damage to properties due to increasing highland and mountain population with associated degradation (Petley *et al*., 2005; Petley, 2008).

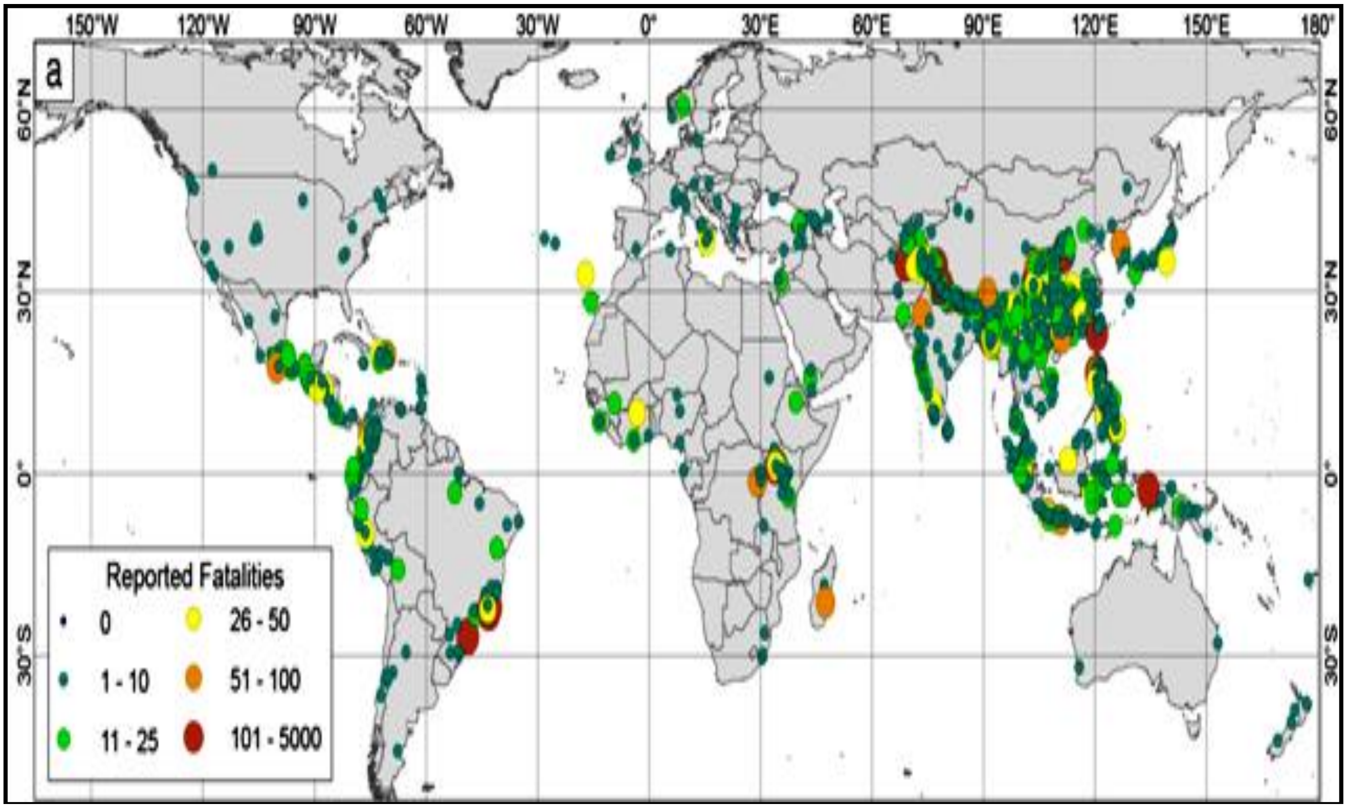

Figure 2.1: Global map of reported landslide events from 2007–2013, showing landslides with fatalities. The size and colour of the data point indicates the number of reported fatalities for each event (after Kirschbaum & Zhou, 2015).

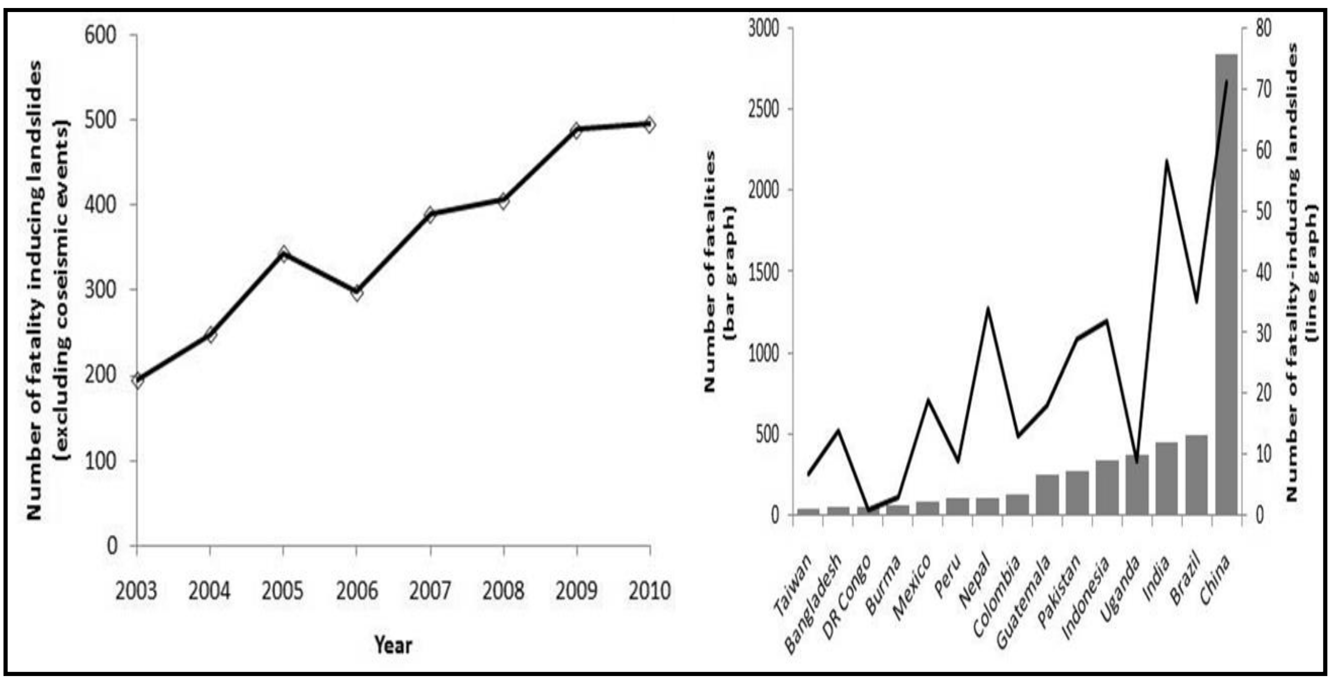

Figure 1.2: Temporal trends in fatalities from landslides and global hotspot countries
Source: American Geophysical Union, 2013.

The East African highlands, due to their wet climate, high population densities, steep topography, and high weathering rates, are prone to landslides (Knapen, 2003; Knapen *et al*., 2006). The highlands have experienced a number of landslide occurrences, some of which have been catastrophic (NEMA, 2014). Landslide studies in these highlands have been done by Bagoora (1988; 1989), Muwanga *et al.* (2001), Breugelmans (2003), Knapen (2003), Kitutu *et al.* (2004), Knapen *et al*. (2006), Claessens *et al*. (2007), Kitutu *et al*. (2009) Babirye (2010), Mugagga *et al.* (2010), Atuyambe *et al*.(2011), Kato and Mutonyi (2011), Kitutu *et al*. (2011), Juventine (2012), Mugagga *et al*. (2012), and Ainomughisa (2015) in Uganda; Christiansen and Westerberg (1999) in Tanzania; Ngecu and Ichangi (1989), Davies (1996), Westerberg and Christiansen (1998), Ngecu and Mathu (1999), Westerberg (1999), Inganga *et al*. (2001) and Ngecu *et al*. (2004) in Kenya; Rapp *et al*. (1972) and Moeyersons (1989, 2003) in Rwanda. These studies have, however, mainly focused on single parameters without analysing the convergence of several attributes in landslide occurrence. Whereas most of these studies have focused on primary parameters especially slope gradient, curvature and soil texture, the influence of complex topographic parameters, clay mineralogy and soil behaviour to water changes have not been well analysed. They have also concentrated on the socio-economic implications of landslides for the communities, with limited attention on the real conditions for landslide occurrence in these highlands. Furthermore, most of these studies have tended to focus on volcanic landscapes, with less attention on the non-volcanic highlands e.g. the Kigezi highlands which are also vulnerable to landslides.

Landslides have become common phenomena in Uganda (NEMA, 2010; 2014). They are associated with extensive damage to property and the environment (Office of the Prime Minister for Uganda, 2005; Kitutu *et al*., 2009; Mugagga *et al*., 2012). They are common in the mountainous and highland areas of Eastern and Western parts of the country especially the Elgon ranges, Rwenzori ranges and Kigezi highlands (NEMA, 2014). Although landslide occurrence has increased in Uganda, limited systematic and scientific research has so far been conducted on the phenomena (Muwanga *et al.,* 2001; Knapen *et al.,* 2006). The increase in landslide occurrence in Uganda's highlands and mountains is attributed to poor environmental management practices and global warming phenomenon which has resulted in high precipitation (NEMA 2010; 2012). Generally, the mountains and highland environments of Uganda are synonymous with landslide hazards (NEMA, 2014). Uganda is therefore among the most affected countries by landslide occurrence and fatalities (Fig. 1.2). The Kigezi highland region of south western Uganda is one of the most affected by landslides due to its steep topography, high rainfall amounts and population densities. The Kigezi highlands are comparable to other mountainous and highland regions in tropical Africa where landslides are common disasters (Bagoora, 1998 and NEMA, 2012).

The proximate and underlying causes of landslides are widely covered in literature (e.g. Ngecu and Mathu 1999; Muwanga *et al.,* 2001; Breugelmans 2003; Knapen 2003; Glade and Crozier 2004; Kitutu *et al.,* 2004; Knapen *et al.,* 2006; Kitutu *et al*., 2009 and Mugagga *et al.,* 2012). Literature is however inconclusive on the geographical significance of the factors leading to landslides in space and time. Landslide occurrence depends on complex interactions among a large number of partially interrelated factors. The main landslide controlling parameters include i) geological makeup, especially the rock units (e.g., mudstone, shale's, phyllite, sandstone, limestone, greentuffes etc.), tectonics and bedrock structure, (ii) topography including elevation, gradient, shape, aspect, curvature, (iii) soil parameters such as soil types, soil texture, soil depth; (iv) land cover and land use distribution; (v) hydrology especially rainfall, soil moisture, drainage density or flow accumulation, flow direction, and infiltration (Muwanga *et al.,* 2001; Kitutu *et al.,* 2004; Knapen *et al*., 2006; Hong *et al*, 2007; Mugagga *et al.,* 2012).

According to Hong *et al*. (2007), slope gradient, soil types and texture are primary-level parameters, while elevation, land cover types, and drainage density are secondary in importance. These parameters however, have a spatial-temporal variation and therefore, site specific investigations are very important to understand the conditions for landslide occurrence in any given region. Most landslide studies on the East African highlands show that anthropogenic factors are the major cause of landslides (Ngecu and Mathu 1999; Muwanga *et al.,* 2001; Breugelmans 2003; Knapen 2003; Glade and Crozier 2004; Kitutu *et al.,* 2004; Knapen *et al.,* 2006). The human impact on the land cover e.g., urban build-up areas, road construction, burning, irrigation, mining, cultivation, and deforestation change hillslope hydrological characteristics and can result into slope failures (Glade, 2003; Beguería, 2006; Mugagga *et al*., 2012; Promper and Glade, 2012).

Although landslide occurrence has increased (Fig. 1.3), the conditions underpinning landslide occurrence in Kigezi highlands are not well understood due to limited research on the subject. Lack of information on landslide occurrence has increased vulnerability of communities to landslide hazards in this highland region. Information on parameters underpinning landslide occurrence in this region is important for predicting and/or identifying potential landslide zones. In order to understand the spatial-temporal distribution of landslide occurrence, all the control parameters need to be evaluated in a holistic sense and interplay determined.

Fig. 1.3: Landslide occurrence in the Kigezi highlands

The fundamental hypothesis of this study was that ‘the spatial-temporal probability of landslide occurrence in the Kigezi highlands is dependent on the topographic, soil and land cover parameters which in turn influence the hydrological response of the hillslopes. Against this background therefore, the present study investigated the spatial-temporal landslide distribution in the Kigezi highlands. The major topographical and soil parameters were investigated to analyse their influence on landslide occurrence. The study also assessed the superimposition of the human factor over the other biophysical parameters in landslide occurrence. The present and past land use/ cover patterns and their influence on spatial-temporal landslide distribution were assessed. Given the fact that landslide occurrence in every region is unique, it is therefore important to characterize the phenomena in a particular region. This would serve to identify variations in the controls on landslide occurrence among the regions.

1.2 Statement of the problem

The disaster risk of Kigezi highland communities to landslide hazards is poorly understood due to paucity of information on landslide underpinnings, particularly the topographical, pedological characteristics, as well as spatial and temporal variations in land use/cover. Most studies on landslide occurrence in Uganda have also 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 ecological and topographic setting. The Kigezi highlands which are largely non-volcanic in nature are covered by soil properties different from those of Mount Elgon (Bagoora, 1997). The present study therefore examines landslide occurrence in the uplifted non-volcanic environments of Kigezi highlands.

The frequency and magnitude of landslide occurrence in the Kigezi highlands have increased (NEMA, 2014). Previous studies on the role of topography in landslide occurrence in this region have mainly focused on the steepness parameter without considering other compound topographic parameters. Attempts made by Bagoora (1988; 1989) and Ainomugisha (2015), focused on the role of primary topographic parameters such as gradient, aspect and altitude. These studies lacked depth in terms of the complexity of topography, especially its role in influencing hillslope hydrology and soil development. The role of topographic configuration has not been widely investigated in these highlands, particularly its effect on surface and subsurface hillslope hydrology. Moreover, general conclusions (e.g., Bagoora 1988) seem to emerge from studies which are much more localized and where study sites were purposively selected (see Carswell, 2002). Both studies (Bagoora, 1988 and Ainomugisha, 2015) concentrated on the steepness parameter and landslide spatial distribution but did not consider compound topographic, soil and land cover parameters underpinning landslide occurrence.

Unlike the previous studies which concentrated on single parameters, the present study characterized the major conditions for landslide occurrence including topographic, soil and land use/cover attributes in a conceptual framework. An evaluation was done to determine the convergence of these parameters in landslide occurrence. This has been a major omission in most landslide studies in the Kigezi highlands as well as other highland and mountainous environments in Uganda. The present study developed a holistic conceptual model to explain landslide occurrence and susceptibility for the study area. The present study therefore provides greater insights into our understanding of landslide occurrence in the region and contributes to lessening the vulnerability and enhances disaster risk reduction of communities to landslide hazards.

1.3 Aim and objectives of the study

1.3.1 Overall aim of the study

The overall aim of this study was to characterize the parameters underpinning landslide occurrence in the Kigezi highlands. This information is important for predicting or identifying actual and potential landslide sites. This should inform policy, particularly in terms of developing early warning systems to landslide hazards in these highlands.

Specific Objectives

1. To analyse the topographical parameters underpinning landslide occurrence in the Kigezi highlands of south western Uganda. This objective was achieved by examining selected topographic parameters that influence landslide occurrence, namely slope gradient, profile curvature, topographic wetness index, stream power index, and topographic position index. These parameters were obtained using a 10m DEM in ArcGIS 10.1 as well as field measurements using clinometers and GPS receivers.

2. To assess the influence of soil properties on landslide occurrence in Kigezi highlands of South Western Uganda. This objective was achieved through field investigations, collection of soil samples at different depths and points along the slope profile for determination of soil physical properties, soil-water infiltration and laboratory analysis. Laboratory analyses comprised shear box, Atterberg limits, sieve and hydrometer analysis as well as XRD analysis.

3. To assess land use/ cover changes and their implications for landslide occurrence in Kigezi highlands of South Western Uganda**.** This objective was achieved through analysis of the spatial-temporal land use and cover changes using satellite images spanning 1985 to 2015. The Landsat imagery data used for land use and cover classification included; Land sat 5TM, 7ETM+ scenes, Landsat 8 OLI/TIR of 30m spatial resolution, all obtained from path 173 and row 061. Other data sources included; aerial photographs, Google Earth images, topographic maps, historical and local government records, as well as field investigation conducted to verify the land cover distribution and landslide occurrence in the highlands.

4. To develop a conceptual model for landslide occurrence in the study area. This objective was achieved by integrating all the parameters identified in a conceptual framework. The inherent topographic, soil and land use/ cover parameters were integrated to explain landslide occurrence in the study area. This conceptual model is important for predicting or identifying potential landslide sites in the study area.

1.4 Research questions

1. How does the topography of the highlands influence landslide occurrence?
2. Why are landslides occurring along topographic hollows and not on other slope elements?
3. How does soil characteristics of the highlands influence landslide occurrence?
4. What is the influence of land use and cover changes on landslide occurrence?
5. Which land use and cover category is most affected by landslide occurrence?

1.5 Description of the study area

1.5.1 Location

The study was conducted in the non-volcanic Kigezi highlands of South-Western Uganda situated between 01˚ 21' 25 " and 0˚ 58' 08" south; and 29˚ 43' 30" and 30˚ 05' 51" east. The highlands are bordered by the Republic of Congo to west, Rwanda to the South and Ankole highlands to the East. Rukiga catchment (Fig.1.4) was delineated for detailed study of the topographical, soil properties, land use/cover characteristics and their influence on landslide occurrence.

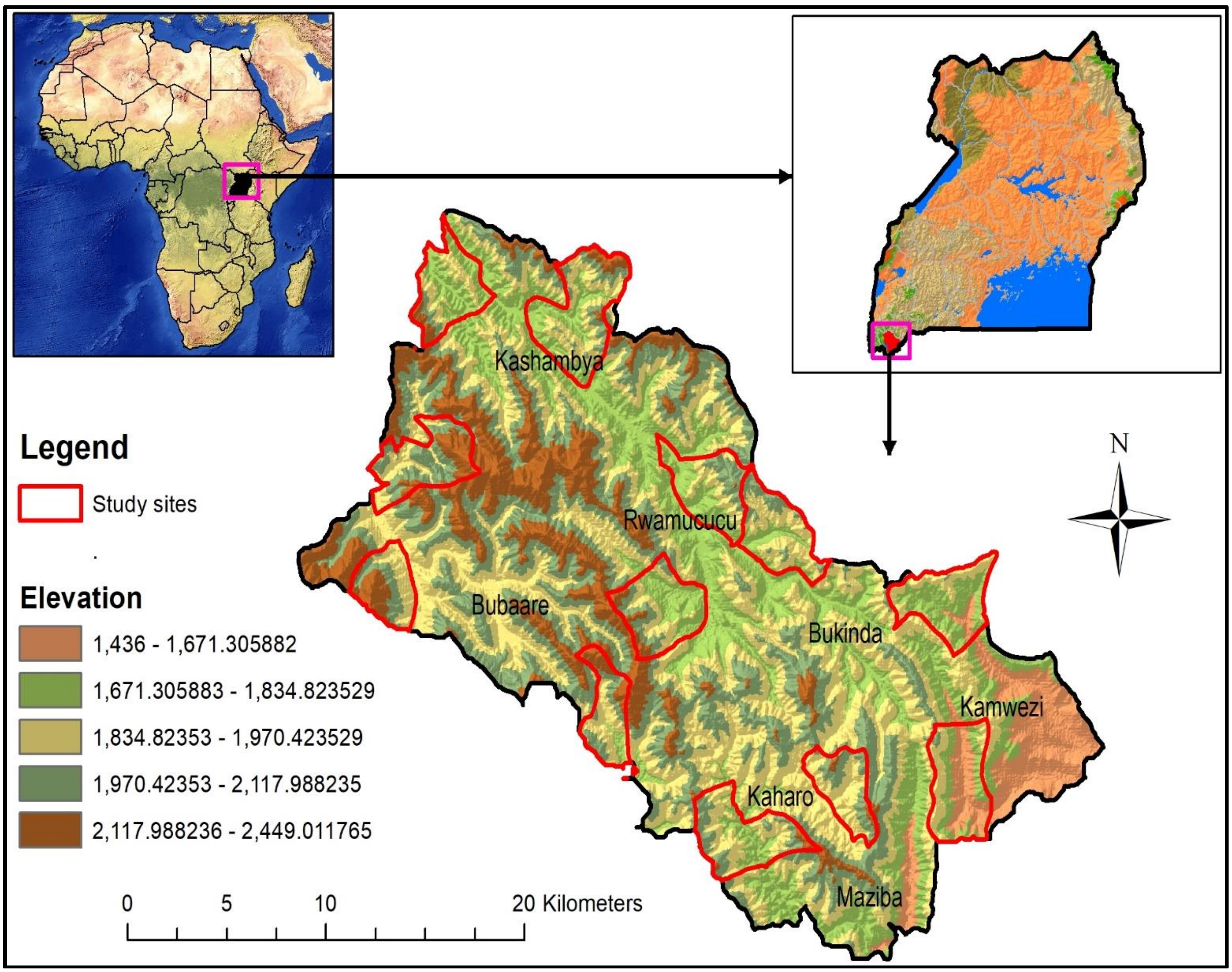

Fig. 1.4: The location of the study area.

In order to decide on the study area, a landslide inventory was carried out with the guidance of local people. During this inventory, landslide scars were identified mapped and characterized. The study area was demarcated using water delineation tools in ArcGIS. The drainage network and topography served as baseline information for establishment of catchment area and micro watersheds in the region. The drainage network and watershed used in this study were delineated automatically from a 10m DEM using the Arc Hydro tools in ArcGIS 10.1.

The area was also 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 different from what is observed elsewhere in the region. 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 landslide scars are easily concealed by soil materials mobilized from the hilltops and spur slopes into topographic hollows where landslide occurrence is dominant. The soils accumulating within the landslide scars encourage high vegetation regeneration owing to the high rainfall amounts in the study area.

1.5.2 Relief and Topography

The relief of Kigezi highlands varies from about 1,436 meters above sea level in Kamwezi sub-county in the extreme South-East, to about 2,264 m a.s.l. in Rwemucucu sub-county (e.g. Rurengere ridge), 2,352 m a.s.l in Kashambya sub-county (e.g. Buchundura ridge), 2,606 m a.s.l. in Bwindi ridges (e.g. Rwamunyonyi ridge) and 2,529 m a.s.l. in Echuya ridges (e.g., Karengyere ridge) for the non-volcanic highlands (Fig. 1.5).

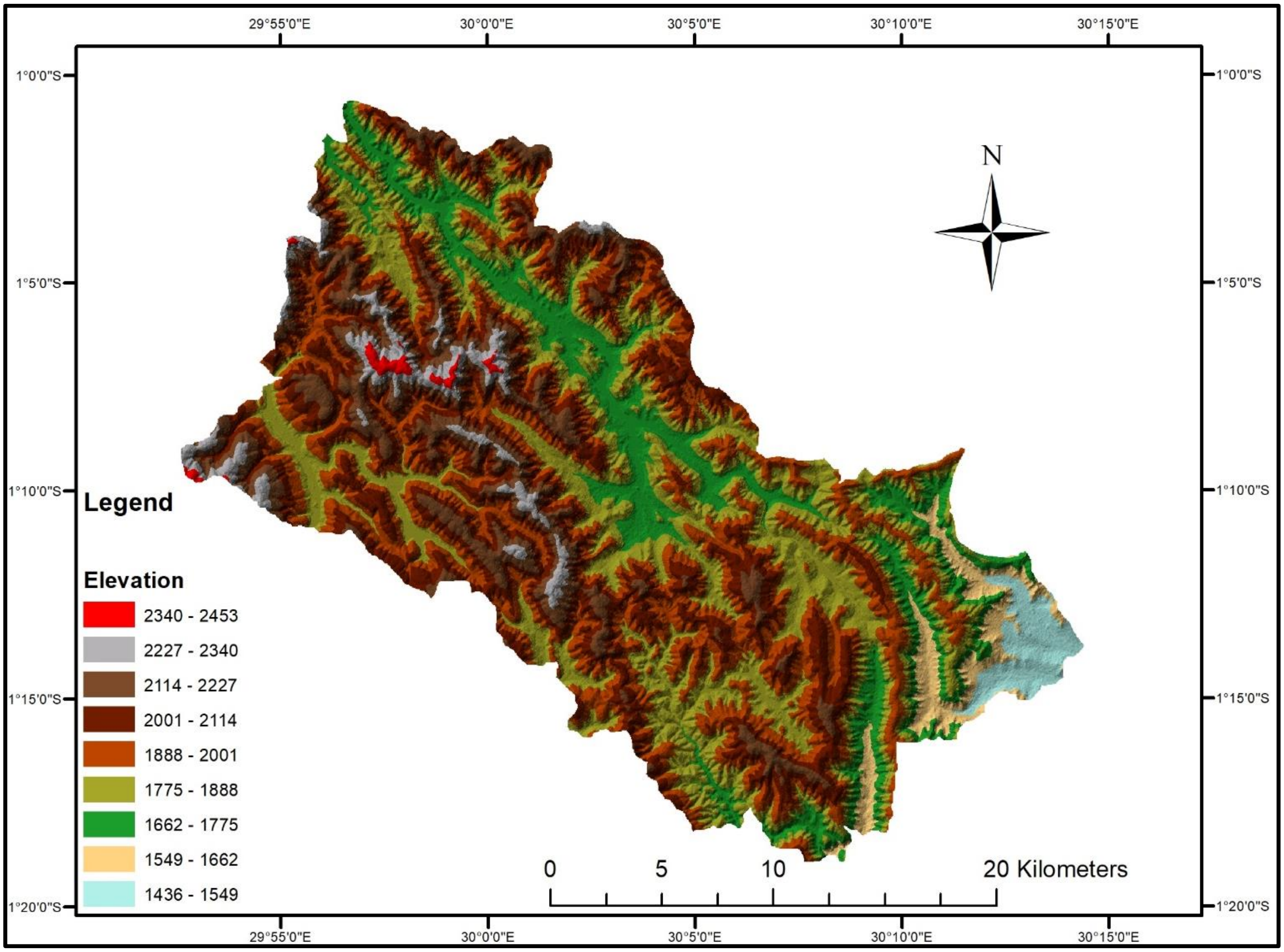

Fig. 1.5: Relief and topography of the study area.

The topography comprises mainly extensive flat-topped ridges and hills (Fig. 1.6), broken by short numerous steep-sided deep subsidiary strike valleys separated by fluted spurs, usually 3-6 km (Bagoora, 1993). The topography is extremely rugged, consisting of narrow steep convex (20-45˚) and gentle (10-15˚) slopes (Bagoora, 1998). The origin of this complex and intriguing landscape can generally be described in terms of both structure and climamorphogenesis (Bagoora, 1997). Due to the prevalent tropical humid conditions characterized by aggressive weathering, as well as fluvial and non-fluvial denudation processes, the highlands have been sculptured into deeply incised landscape (Tukahirwa, 1999). The landscape has steep slopes and deep narrow valleys (Fig. 1.5) with an amplitude of 600-700m or greater with very little arable land available on the less steep and gentle pediment slopes (NEMA, 2008).


Fig. 1.6: The landscape of the study area

1.5.3 Geomorphology

The geomorphology of Kigezi highlands was described by Wayland (1934), as belonging to the African surface formed during the Jurassic and Cretaceous times (180-135 and 135-65 million years B.P). By the mid-Tertiary (about 35 million years before present), most of the African surface including Uganda is thought to have been destroyed by denudation processes to a single peneplain. After mid-Tertiary, the peneplain suffered slow regional uplift and consequent dissection by erosion processes (Bagoora, 1997). In the south-west (location of the study area), this landscape was further up-warped to well above the general plateau by later structural movements associated with the formation of the western rift valley (Harrop, 1960; Bagoora, 1993). Therefore, the flat crested hills and ridges of Kigezi highlands belong to the remnants of the older upland erosion surfaces of the Jurassic and upper Cretaceous times (Bagoora, 1998). It is part of the Uganda plateau which Wayland (1934) classified as peneplain 1. The highlands are characterized by distinctive advanced erosion surfaces represented particularly by the main steeply rising slopes and flat- topped ridges that rise between 1,500 and 2,600 m above sea level (NEMA, 2010). Through time, some parts of the highlands have become lateralized, but the contribution of duricrust and near laterite conditions is very complex (Langlands, 1974).

1.5.4 Geology and soils

The geology of Kigezi highlands and other parts of South Western Uganda is composed of sedimentary rock system of the pre-Cambrian age (Ollier, 1969; Bagoora, 1993; 1997). Except for the Pleistocene (1-2 million years B.P) volcanic system which underlie the volcanic mountains in the extreme South-West, most of South-Western Uganda is founded on a sedimentary rock system. They are of the pre-Cambrian age dated between 390 and 3,235 million years old (Langlands, 1977). In some places, however, metamorphosed products such as schists and gneisses and later intrusions of granites exist (NEMA, 2004). The rocks have been collectively grouped into Karagwe-Ankolean system. The rock system is named after an extensive region in South-Western Uganda and North-Western Tanzania where it is dominantly underlain by this system (Bagoora, 1993). The Karagwe-Ankolean phyllite and shale’s with quartzite and occasional metamorphic-like schists form the characteristic long crested ridges. The ridges are characterized by steep middle slopes and gentle pediments that end in flat valley bottoms (Langlands, 1972). Phyllite and shale’s (mostly arenaceous) dominate the middle slopes. Towards the top, sediments normally become more arenaceous and are overlain by thick sandstones and sandy micaceous shale’s (Bagoora, 1988; 1989). Miscellaneous alluvia of sands and clay occupy many drowned water courses in the highlands, the texture of which depends on the rock in the area. For example sands and coarse sands occur in the warped valleys flanked or underlain by gneiss and granite intrusions, while shale’s and phyllite have given rise to clay deposits (NEMA, 2010).

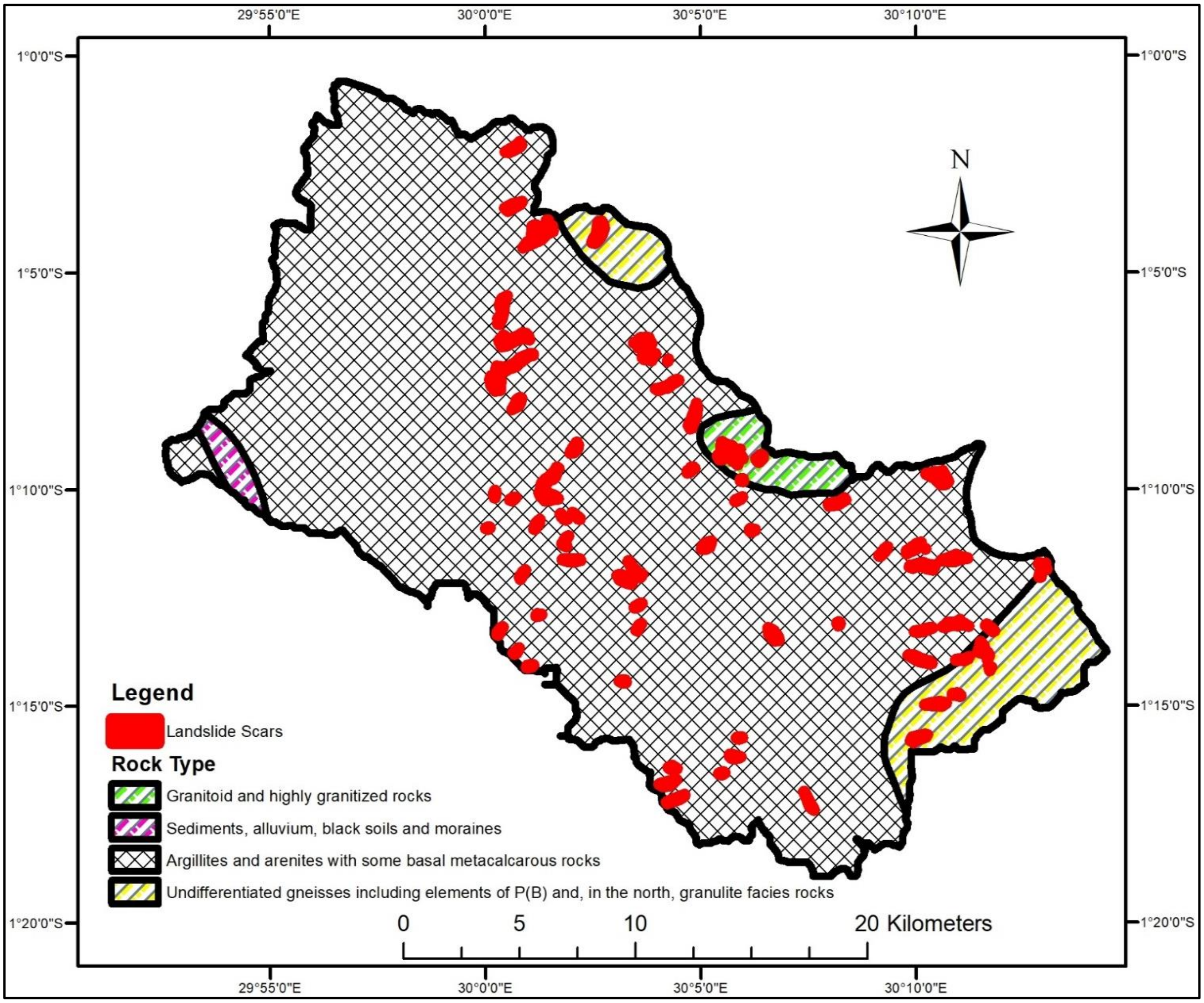

Fig. 1.7: Rock types and landslide distribution.
Source; Department of Geology for Uganda, 1965.

The rocks underlying slopes in the study area have been categorized by Ollier (1969) and Bagoora (1988; 1993), as phyllite, shale's, sandstones, quartzite, granites and gneisses of granitic composition. Other rock types include various grades of schists such as quartz-schists and fine textured mica-schists belonging to the Ankole-Karagwe rock system and the Achaean basement complex. Slopes which are dominantly underlined by quartzite and granitic intrusions are not very prone to landslide occurrence (Bagoora, 1988). During field investigations, it was observed that slope zones underlain by intrusive rocks experience only shallow soil slips. Landslides mainly occur on ridges underlined by shale and phyllite rocks (Fig. 1.7), but more dominant on sections underlined by shale than those with phyllite. A relationship exists between rock type and soil depth. Slope sections underlain by relatively weaker rocks like shale have deep soil profiles due to the high weathering rates (Bagoora, 1997). Slopes underlain by quartzite and granitic intrusions are covered by very shallow soils and in some cases bare rocks making them less prone to landslide occurrence. Soil development also determines the nature, magnitude and return period for landslide occurrence (Bagoora, 1993).

The rates of weathering of the rocks and the physical and chemical characteristics of most soils are strongly related to geology of the area (Bagoora, 1997). By implication, the local geology has an important influence on the problem of landslides in the highlands. Rock type also has direct influence on slope factors such as slope angle and slope stability. Slopes underlain by phyllite and shale’s are more affected by instability processes than those underlain by quarzitic and micaceous sandstones (Bagoora, 1988; 1989). The latter characterizes much of the steep slopes, and have more rock outcrops and cliffs. This indicates relatively greater stability with less landslide activity. The shallow sandy loam and sandy soils that have developed on such rocks are however highly susceptible to erosion and sliding (Bagoora, 1998). It has been observed elsewhere by Kirkby and Morgan (1980) that in humid areas sandstones are typically stable up to 30˚ whereas shale’s are stable up to only 8-15˚. This indicates a much lower angle of repose from the latter.

In the study area, relief and geology strongly influence soil characteristics (NEMA, 2012). The soils of the study area are associated with the geological system of the south-western highlands (Ollier, 1969). They are largely a product of climate, pedogenic and geomorphological processes that have taken place on an uplifted African surface (Langlands, 1972). The soil types in the study area include Luvisols, Histosols, Acric Ferralsols and Dystric Regosols (Bagoora, 1993). Acric Ferralsols and Luvisols are the most dominant soil types in the study area (Fig. 1.8). Other soil types include Latosols, which are either yellow or red. These are old soils representing almost the final stage of weathering and have little mineral reserves left (Langlands, 1974; Morgan, 1986). Landslide occurrence is dominant on Acric Ferralsols and Luvisols while Histosols and Dystric Regosols are not susceptible to slope failures (Fig. 1.8). Out of the 65 landslide scars investigated, 63% occurred on slope sections dominated by Acric Ferralsols while 37% occurred on zones covered by Luvisols.

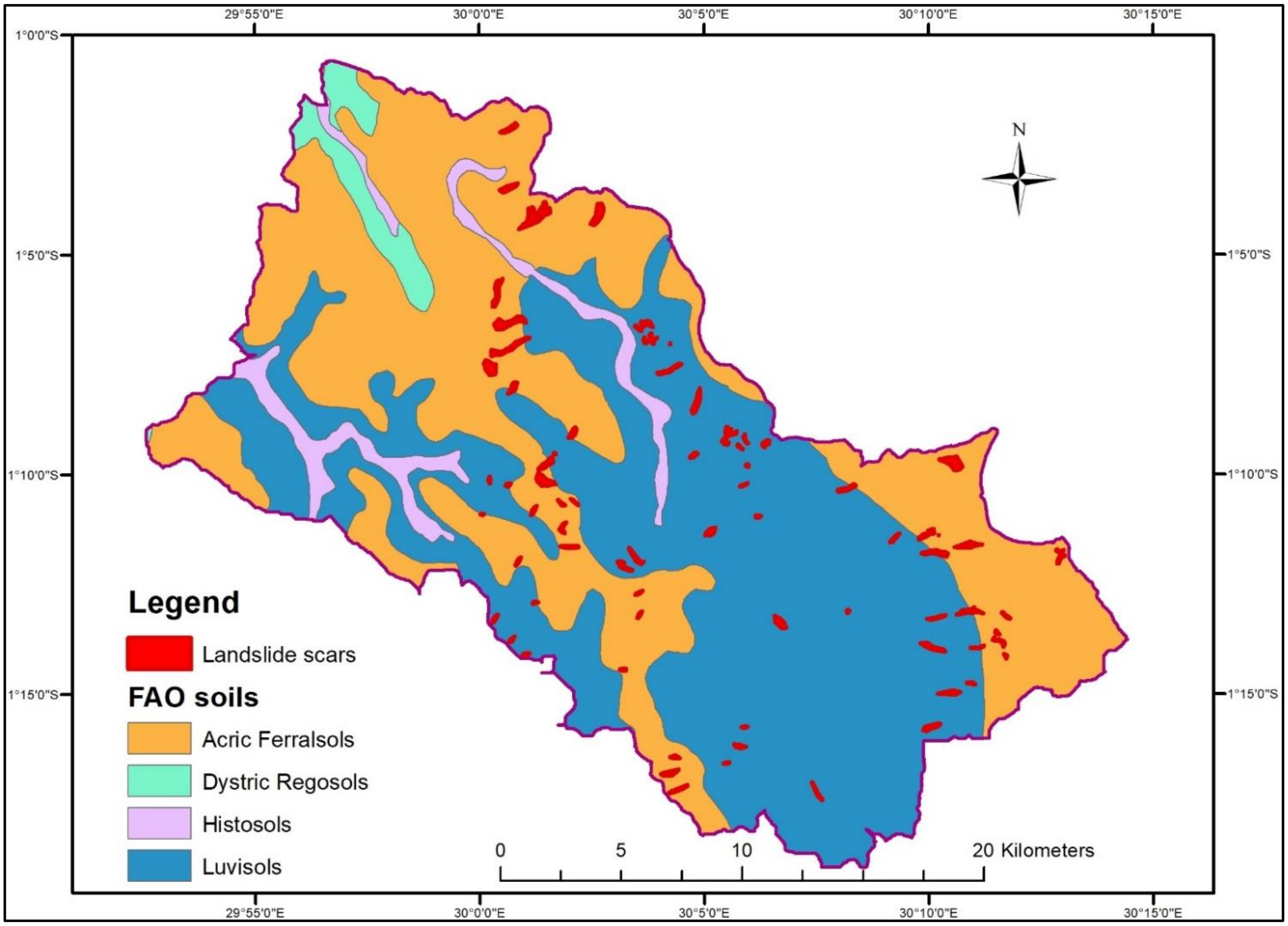

Fig. 1.8: Distribution of soil types and landslide occurrence
Source: National Agricultural Research Laboratories for Uganda, 2008.

In a high moisture environment like the Kigezi highlands with a short dry season (June to August), pedogenic processes have produced a variety of soil textures and structures (Bagoora, 1998). The soil texture range from sandy loams, loams to sandy clay loams. Furthermore, the soils possess a granular and porous structure, including sandy clays which are fairly fertile (Tukahirwa, 1995). The soils are deepest in the hollows, hill folds, and more importantly on the pediment slopes in the valleys (Bagoora, 1998), where the soils form prime agricultural land in the highland (Siriri *et al*., 2002).

1.5. 5 Climate and Hydrology

The Kigezi highlands have a warm to cool humid climate and the average annual rainfall at 01˚ 15’ south and 29˚ 59’east (1867m a.s.l) is 994mm (Kabale Meteorological Department, 2015), which can be classified as moderate. Rainfall however increases to 1250-1500mm or more in higher areas. The study area experiences a bimodal rainfall pattern. The main rainfall season is from mid-February to late May with a peak in March-April (NEMA, 2012). The second season is from September to December with a peak in October-November (Fig. 1.9). The rainfall received in most of the Kigezi highlands is orographic and it is influenced by incursion of westerlies (an air stream from Democratic Republic of Congo) but with much cloud cover (Tukahirwa, 2000). There is a marked dry season from June to August mainly characterized by dry winds and clear skies with small rainfall totals of about 16mm and a lesser one from December to February (Fig. 1.9). Landslide occurrence in the study area is experienced during the MAM and SON seasons especially during the months of April, May, October and November (NEMA, 2012).

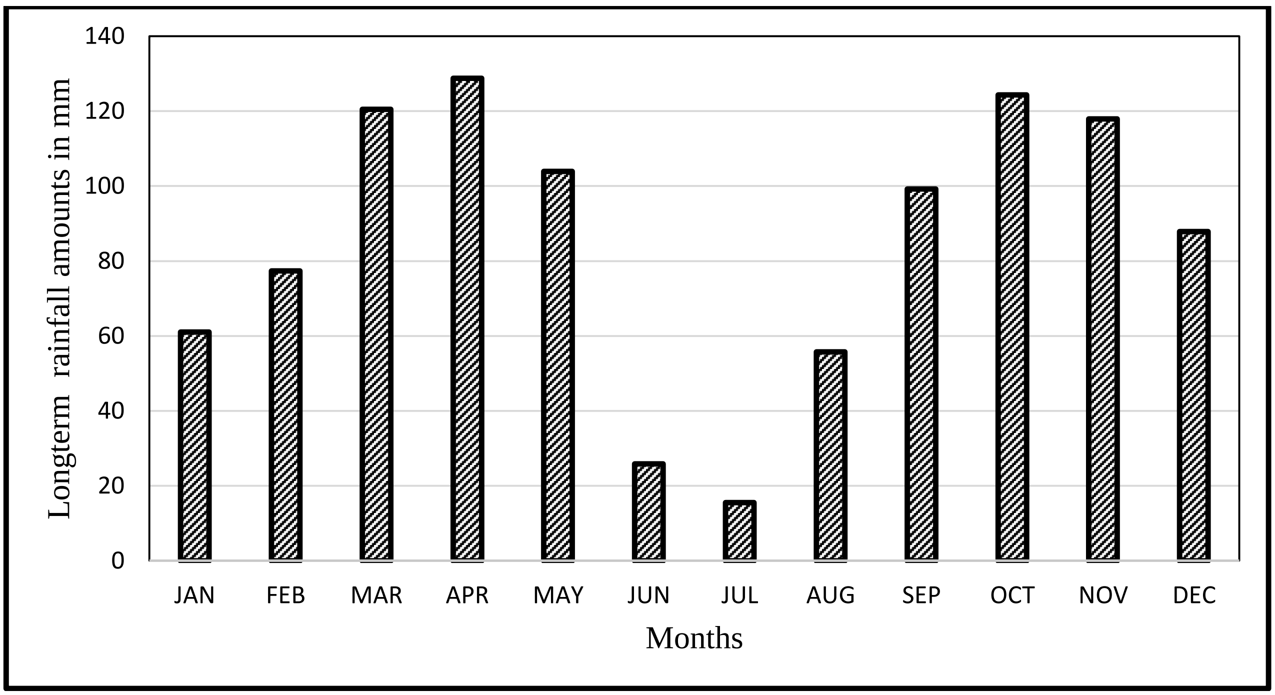

Fig. 1.9: Mean monthly rainfall distribution

Whereas rains in Kigezi highlands are generally of moderate to low intensity, occasional extreme rainfall events are also experienced, especially in the eastern part of the highlands (location of the study area). Events of over 25mm in 24 hours are not infrequent (Bagoora, 1997). Rainfall distribution has an implication on landslide occurrence especially its influence on the behaviour of soil characteristics. Although the nature and distribution of rainfall has an implication on landslide patterns (Knapen *et al.*, 2006; Broothaerts *et al.*, 2012), this has not hitherto been investigated empirically in the Kigezi highlands. The present study therefore also examined the impacts of rainfall patterns on landslide occurrence in these highlands. The mean maximum and minimum temperature values for Kigezi highlands are 23˚C and 10˚C respectively, while potential evaporation rates are also moderate (Kabale Meteorological Department, 2005; NEMA, 2010). These are enhanced by the commonly cloudy skies at such high altitudes. Thick morning fog and mist are characteristic of the highlands. Relative humidity is always high all the year round. The high humidity in combination with rainfall distribution nearly all year round leads to high soil moisture most of the year (NEMA, 2008). In turn, this has significant implications for the magnitude of slope failure, since antecedent moisture significantly influences hillslope response to rainfall (Bagoora, 1998).

The study area comprise numerous highland streams which drain valleys incised in the ridges and hills (NEMA, 2012). The dominant streams drain into south-east to north-west, following Kanyabaha-Nomuremu-Kyerero river valley, which is one of the major valleys in the highlands. The river flows through a major stretch of papyrus swamp that occupies these valleys (Bagoora, 1997). The dense network of the streams serves as an index of topography and morphology of the area and efficient drainage which is a characteristic of the highland environment (NEMA, 2008).

1.5.6 Vegetation cover and land use systems

It is believed that until about 500 years ago, the Kigezi highlands were covered by montane forests comprising mainly of mahogany and other hard woods, including species such as *Juniperus proce, Podcarpus milanjianus and Hagenia abyssinica* (Purseglove, 1946; NEMA, 2004). The area is now covered with grasslands dominated by *Hyparrhenia* species as well as a mountain grass known as *Pennisetum Clandestinum* (Bagoora, 1997). This grass gives rise to continuous sward of a meadow-like turf. The grass is sometimes called “kikuyu grass”, a name borrowed from Kenya due to its common occurrence in the Kenya highlands (NEMA, 2006). Above the forest altitude, a giant grass species known as bamboo (*Arundinaria Alpina*) existed (Siriri and Raussen, 2002). This type of vegetation does not usually form a continuous zone, and appears to be subject to fluctuations in extent due to its habit of dying off and regeneration (Morgan, 1986). Today, only remnants of the grass remain in areas like Echuya Forest Reserve (Carswell, 2002b). Centuries of human interference has led to serious degradation and in some cases depletion of vegetation cover

(Farley, 1996; Carswell, 1997). Today, except for the few natural vegetation patches surviving under legislative protection, the rest of the vegetation cover in the highlands is either very poor or long gone (NEMA, 2014; Kabale District Environmental Report, 2015). Most of the highlands now comprise poor vegetation cover with various human manipulated or impacted types (Fig. 1. 10) Many parts of the hillslopes are already bare due to degradation (Lindblade and Carswell, 1998). The degradation of the vegetation cover has increased the problem of increased runoff coefficients and landslides occurrence in the highlands (Bagoora, 1998; NEMA, 2012).

Fig. 1. 10: Intensively cultivated slopes with high vulnerability to landslide occurrence

Subsistence agriculture is the major economic activity for about 84% of population of the Kigezi highland region (UBOS, 2014). The farming system in the Kigezi highland was classified by

Parsons as the montane system (Siriri and Raussen, 2002) as medium-altitude-coffee system. It is essentially a peasant agricultural system based on annual food (Tukahirwa, 2000). Perennial bananas have been recently introduced and have gradually replaced the annuals on much of the prime agricultural land in the valley bottoms and valley sides, but rarely so on the ridge-tops, presumably due to lower moisture (Bagoora, 1993; 1997). A wide range of crops comprising of sorghum, maize, sweet potatoes, beans, peas, Irish potatoes and fresh vegetables are grown (Tukahirwa, 1999). There is no cash crop that has been found viable for the highland, except the introduced bananas (Bagoora, 1993). Most farmers herd and graze their animals on communal marginal hill slopes, valley bottoms, road sides and inter seasonal fallows (Tukahirwa, 1995).

1.5.7 Population

The conducive ecological condition of the Kigezi highlands have attracted a sedentary farming population (NEMA, 2008). The population density of the region is one of the highest in Uganda with 362 persons per km ^{2} for Kisoro, 296 for Kabale, 198 for Kanungu and 241 for Rukungiri districts. This is quite high compared to Uganda's population density of 174 persons per km^2 (UBOS, 2014). This high population density in the Kigezi highland region has put tremendous pressure on natural resources, especially land, leading to resource overuse and subsequently degradation (NEMA, 2014). Increasing hillslope degradation may be responsible for the occurrence of the observed landslides and associated processes. This is especially triggered by the humid and fragile environment which is dominated by steep slopes, degraded vegetation cover and poor farming practices. The situation has been further worsened by the impacts of climate change (NEMA, 2010).

1.6 Significance of the study

Growing population and expansion of settlements over hazardous areas largely increase the impact of natural disasters both in industrialized and developing countries. Developing countries like Uganda have difficulty in meeting the high costs of controlling natural hazards through major engineering works and national land use planning. Landslide costs include both direct and indirect losses affecting public and private properties. In addition to loss of lives, there are enormous economic losses due to the landslides with respect to direct losses. Indirect costs of landslides include loss of industrial, agricultural and forest areas, reduction in real estate values, loss of tax revenues, adverse effects on water quality in streams.

In Uganda, it is estimated that between 2010 and 2012 more than 500 people lost their lives due to landslide hazards (UBOS, 2012, NEMA, 2014). In addition, the socio-economic losses resulting from landslides are great and apparently are growing as human developments expand into unstable areas due to population growth. Reducing the damage from increasing occurrence of landslide hazards was one of the main objective of this study. The information generated by this study will help to identify potential landslide zones. The study will help in land management by identifying safer zones for community activities. This is important in reducing vulnerability of communities to landslide hazards in the region and the country at large. The information from the study, will help reduce the damage resulting from landslide hazards.

The conceptual model developed during this study should help in landslide disaster risk reduction in the country if extrapolated to other regions with similar conditions. Research results will be helpful in developing new regulations on protecting the design of new buildings, infrastructures, and facilities as well as support land management in the country. Development of a multi-factor analysis method of interaction between complex natural systems and various anthropogenic activities will enable planners to make well-founded predictions and take more informed decisions. Therefore, the results of this research could produce practical benefits for environment protection and more importantly preserve human life.

1.7 Structure of the thesis

The thesis contain seven chapters which discuss landslide occurrence in the Kigezi highlands of south western Uganda.

1.7.1 Chapter One: General Introduction

Chapter one provides a general background to the study and the problem statement, research objectives and questions. The chapter also gives a description of the study area in terms of location, relief and topography, geomorphology, geology and soils, climate and hydrology, vegetation and land use as well as the population. The chapter concludes by giving the significance of the study.

1.7.2 Chapter Two: Literature Review

This chapter presents a review of literature on the biophysical factors identified to have implications for landslide occurrence including topographical, hydrological, geological, and pedological. It also reviews the implications of land use and cover changes on slope stability and hillslope hydrology. The chapter finally presents reviewed literature on conceptual models for landslide occurrence.

1.7.3 Chapter Three: Analysis of topographic parameters underpinning landslide occurrence in the Kigezi highlands of south western Uganda.

This chapter presents the topographic parameters of the Kigezi highlands and highlights on how they influence the spatial distribution of landslides. The selected topographic parameters are slope gradient, profile curvature, topographic wetness index (TWI), stream power index (SPI), and topographic position index (TPI). The chapter concludes by showing the convergence of all the topographic parameters and landslide occurrence.

1.7.4 Chapter Four: The influence of soil properties on landslide occurrence in Kigezi highlands of south western Uganda.

This chapter presents the soil characteristics and its role in landslide occurrence in the uplifted non-volcanic Kigezi highlands of South-Western Uganda. The chapter describes the soil physical and hydrological properties, depth, particle size distribution, shear strength, plasticity, dispersion, expansiveness, liquid limits, clay minerals and infiltration. The chapter describes how the understanding of soil characteristics is an important step in landslide hazard mitigation. The chapter concludes by showing how the soil properties are a major cause of landslide occurrence in the study area.

1.7.5 Chapter Five: Assessment of land use / cover changes and their implications for landslide occurrence in Kigezi highlands of South Western Uganda

The chapter presents the influence of land use/cover changes on landslide occurrence. The land use/ cover changes especially decimation of forests and grasslands, due to increased cultivation and settlements are presented in this chapter. The chapter shows how landslides are predominantly occurring in areas where forests and grasslands have been converted to cultivated land and settlements. The chapter also shows the dominance of landslides where there is an interaction between land use/cover changes and topography. The chapter finally describes the most significant land use/ cover change trends identified.

1.7.6 Chapter Six: A Conceptual model for landslide occurrence in Kigezi highlands of South Western Uganda.

This chapter presents an integration of all the landslide parameters identified in a conceptual framework. The chapter presents an equation developed showing the inherent topographic, soil and land use/ cover parameters and their convergence to cause landslide occurrence. The chapter describes landslide susceptibility in the highlands and explains how the model can be extrapolated to other regions with similar conditions. The chapter concludes by showing how landslide occurrence can be predicted in the study area.

1.7.7 Chapter Seven: The synthesis of landslide occurrence in Kigezi highlands

This chapter presents the synthesis of the major findings, conclusion and recommendations. The chapter synthesizes the findings presented from the analysis of landslide hazard occurrence. This chapter also presents recommendations that could reduce vulnerability while increasing resilience of communities to landslide hazards based on the key findings. It also proposes future research areas. A list of references and appendices as used in all the chapters is provided at the end of the thesis.

CHAPTER TWO

Literature Review

2.0 Introduction

This chapter presents insights into existing literature on landslide occurrence and the parameters influencing their spatial temporal distribution. The topographical, hydrological, geological, and pedological characteristics that influence landslide occurrence are explicitly discussed. The chapter also reviews the implications of land use and cover change for landslide occurrence. The chapter also reviewed literature on modelling landslide occurrence.

2.1 Landslide processes and typologies

Landsliding refers to downward movement of slope material that occurs when shear stresses overcome the slope strength (Cruden, 1991; Knapen *et al*., 2006). The process involves movement of soil and rock down slope under the influence of gravity (Reed, 1992). They are part of geomorphologic cycle of landform development (Huabin, *et al.,* 2005; Massimo and Lorenzo, 2008; Singh, 2010). Landslides constantly generate irregularities on the earth‘s surface in the form of varying relief and landforms (Selby, 1993; Jamali and Abdolkhali, 2009; Moayedi *et al*., 2011; Broothaerts *et al.*, 2012). They are considered among the most devastating natural disasters in the world (Petley, 2008; Broothaerts *et al*., 2012; Gorokhovich *et al.*, 2013; Pankow *et al*., 2014; Kirschbaum and Zhou, 2015). They are most prevalent in mountainous and highland terrains of tropical and sub-tropical regions (Guzzetti, 1999; Dai *et al*., 2002; Petley *et al*., 2005; Das *et al*., 2010; Mugagga *et al*., 2011; Pankow *et al*., 2014).

Understanding the characteristics of the specific landslide type in a region is an important factor to consider in disaster risk reduction (Singh, 2010; Broothaerts *et al.*, 2012). The type of landslide determines the potential speed of movement, likely volume of displacement, possible effects and the appropriate mitigative measures to be considered (Moayedi *et al.*, 2011). The various types of landslides can be differentiated by the kinds of material involved and the mode of movement (Selby, 1993). The material in a landslide mass is either rock or soil and sometimes both (Highland and Bobrowsky, 2008). Different types of landslides are evident in mountainous and highland areas (Hong *et al*., 2007; Yashar *et al*., 2013). Varnes, (1978) distinguishes five types of movements namely falls, topples, slides, flows and spreads as illustrated in Fig. 2.1.

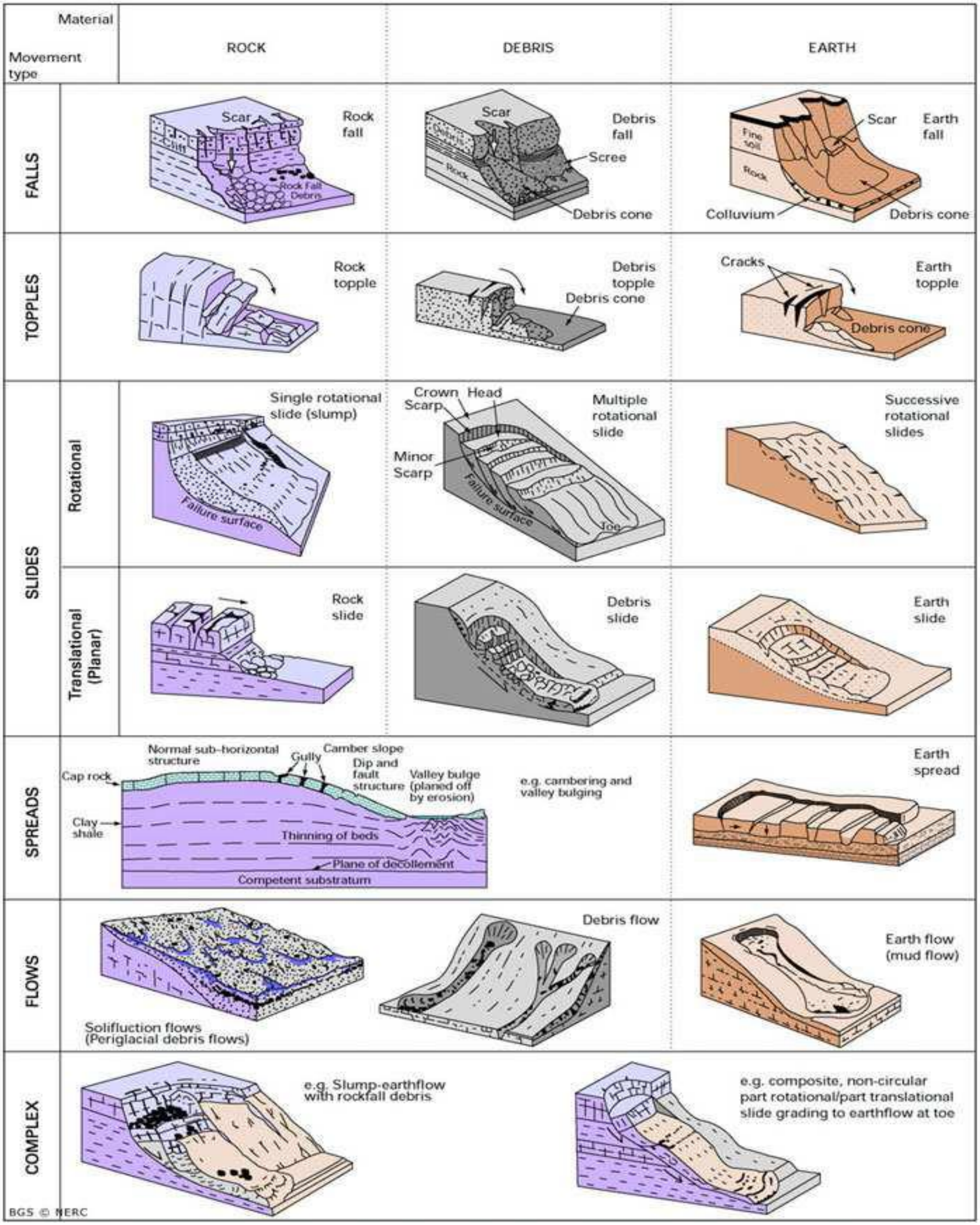

Fig. 2.1: Landslide processes and typologies
Source: http://pubs.usgs.gov/fs/2004/3072/images.

Falls

Falls are abrupt movements of masses of geologic materials, such as rocks and boulders that become detached from steep slopes or cliffs (Highland and Bobrowsky, 2008). Separation occurs along discontinuities such as fractures, joints, and bedding planes, and movement occurs by free-fall, bouncing, and rolling (Broothaerts *et al.*, 2012). Falls are strongly influenced by gravity, mechanical weathering, and the presence of interstitial water (Chen and Lee, 2004). Falls are usually due to undercutting of the toe or face of the slope by a river or by wave action (Selby 1993; Knapen *et al*., 2006). Falls are not common in the Kigezi highlands due to the deep soil profiles covering the slopes generated by humid tropical weathering processes and were therefore not considered for investigations in this study.

Topples

A topple failure is a process that has a distinct component of rotation and sliding before a fall takes place (Selby, 1993). Toppling is common in slates and schists but also occurs in thinly bedded sedimentary rocks and in columnar jointed igneous rocks such as basalt and dolerite (Kitutu *et al*., 2004). Toppling is also influenced by the structure of steeply jointed rocks (Cruden, 1991). Toppling failures are distinguished by the forward rotation of a unit about some pivotal point, below the unit, under the actions of gravity and forces exerted by adjacent units or by fluids in cracks (Sassa *et al*., 2005; Massimo and Lorenzo, 2008). Topples are not common in the Kigezi highlands and were thus not considered in this study.

Spreads

Spreads are distinctive and usually occur on very gentle slopes of flat terrain (Cruden, 1991). The dominant mode of movement is lateral extension accompanied by shear or tensile fractures (Hong *et al.,* 2007). The failure is caused by liquefaction whereby saturated, loose sediments (usually sands and silts) are transformed from a solid into a liquefied state (Uzielli *et al*., 2008). Failure is usually triggered by rapid ground motion, such as that experienced during an earthquake, but can also be artificially induced (Guzzett *et al.,* 2008). When coherent material rests on materials that liquefy, the upper units may undergo fracturing, liquefy and flow (Sassa *et al.,* 2005). Lateral spreading in fine-grained materials on shallow slopes is usually progressive (Kitutu *et al*., 2004). The failure starts suddenly in a small area and spreads rapidly (Selby, 1993). Such landslide processes are not common in the Kigezi highlands due to the steep topography which does not encourage spreading and were thus not considered for investigations in this study.

Slides

Slides are the down slope movements of rock and soil along a slip surface characterised by almost permanent contact between the moving mass and the slide surface (Bryant, 1991; Alexander, 1993). Slides occur where there is a distinct zone of weakness that separates the material from more stable underlying material (Selby, 1993; Yashar *et al.,* 2013). They form when a coherent mass of regolith or bedrock breaks free and then slides down slope along either a planar or curved surface (Moayedi *et al.,* 2011). The geometry of the detachment or rupture surface and the degree to which the sliding material remains coherent determines the type of slide (Sassa *et al*., 2005). The major types of slides are rotational and translational (Singh, 2010). Rotational slide is where the surface of rupture is curved concavely upward and the slide movement is roughly rotational about an axis that is parallel to the ground surface and transverse across the slide (Jamali and Abdolkhani, 2009).

Translational slide is where a landslide mass moves along a roughly planar surface with little rotation or backward tilting (Moayedi *et al.,* 2011). Transitional slides are also called planar slides (Corominas, 2014). The mass of material moves down-slope on a largely planar surface (Winter, 2014). Translational slides can have very different impacts to rotational slides (Corominas *et al*., 1996). Where the slope is sufficiently steep and the shearing resistance along the slip surface remains low, the movement can continue on for a long distance, being different to rotational slides (Msilimba and Holmes, 2005). Translational slides in rock usually occur along discontinuities such as bedding planes or joints (Eidsvig, 2014). In the case of debris slides, failure can occur on shallow shear surfaces at or near the base of the surface materials where there can be marked changes in strength and permeability (Alexander, 1993). Translational slides are common on Mt Elgon in Eastern Uganda and occur at a shorter distance from the water divide because they require a less increase in pore water pressure to occur on rectilinear slopes with shallow soils (Knapen *et. al*, 2006). In Kigezi highlands, the dominant type of movement is rotational sliding which mostly occurs on concave slopes and at a large distance from the water divide where runoff and subsurface water concentrates (Bagoora, 1993). In the present study therefore, rotational slides were considered in the characterization of landslide occurrence.

Flows

Flows are down slope movements of viscous masses composed of fluidized soil and other materials (Corominas, 2014). A flow is a spatially continuous movement in which the surfaces of shear are short-lived, closely spaced, and usually not preserved (Chen and Lee, 2004). In a flow, the structure of the material changes into quasi – fluid (Bryant, 1991). The component velocities in the displacing mass of a flow resemble those in a viscous liquid (Yashar *et al*., 2013). There is a gradation of change from slides to flows, depending on the water content, mobility and evolution of the movement (Broothaerts *et al.,* 2012). The most common type of flow landslide is the debris flow (Corominas, *et al.,* 1996; Knapen *et al*., 2006). A debris flow is a form of rapid mass movement in which a combination of loose soil, rock, organic matter and water mobilize as a slurry that flows downslope (Hong *et al.,* 2007). Debris flows are commonly caused by intense surface-water flow due to heavy precipitation that erodes and mobilizes loose soil or rock on steep slopes (Selby, 1993). Debris flows also commonly mobilize from other types of landslides that occur on steep slopes (Moayedi *et al*., 2011). Debris flows include less than 50% fine materials (Varnes, 1978). They are nearly saturated, and consist of a large proportion of silt and sand-sized material (Reed, 1992).

Debris flow source areas are often associated with steep gullies (Yashar *et al*., 2013). The debris-flow deposits are usually indicated by the presence of debris fans at the mouths of gullies (Kamp *et al*., 2008). Debris flows occur following intense torrential rainfall (Huabin *et al*., 2005). When lose regolith on steep slopes become saturated and unstable, they give way and flow down slope (Chen and Lee, 2004). Debris flows also commonly mobilize from other types of landslides that occur on steep slopes, are nearly saturated and consist of a large proportion of silt-and sand-sized material (Eidsvig, 2014). Debris – flow source areas are often associated with steep gullies usually indicated by the presence of debris fans at the mouths of gullies (Winter, 2014). Debris flows are one the most dangerous type of landslide because they often extend far from their sources. They move rapidly and their depositional areas often include inhabited sites. The 2010 Bududa landslide on the slopes of Mt Elgon in Eastern Uganda (Kitutu, *et al.*, 2011) is a typical example of debris flows and they are associated with extensive damage to property and life. Debris flows are also experienced in the Kigezi highlands especially after prolonged and continuous rainfall (Bagoora, 1989) and were therefore considered in the characterization of landslide occurrence during this study.

Landslides have become a major geomorphological processes in the East African highlands (Muwanga *et al.,* 2001). Several studies consider the East African highlands to be prone to landslides due to their humid climate, steep topography, and high weathering rates (Knapen, 2003; Knapen *et al*., 2006; Kitutu *et al.,* 2009; Mugagga *et al*., 2011**)**. Although landsliding has been recognized as a widespread phenomenon in the East African highlands, having a great social, economic and geomorphological impact, relatively little research and documentation can be found in the international literature (Ngecu and Mathu, 1999; Knapen *et al*., 2006). In many East African highlands, a clear insight into the local conditions for landsliding is lacking (Muwanga *et al.,* 2001; Knapen *et al.,* 2006). Therefore, the search for region-specific solutions is hampered. The subject of landsliding in the humid tropics like the Kigezi highlands of south western Uganda is still poorly served by the major literature on the subject. The topic is neglected in symposia on problems on land stability and instability (Brunsden and Ibsen, 1984, Anderson and Willebrand, 2003), engineering geomorphology (Foster *et al*., 2008), and tropical geomorphology (Chang, 1984). These omissions are odd, since some of the most landslide prone areas of the world are situated in the humid tropics including Kigezi region which has a long history of these landslide hazards (Bagoora, 1988, 1993 and 1997). The present study aimed at identifying and characterising the major parameters that influence the spatial distribution of landslide occurrence in the highlands.

Landslides can be triggered under multiple geological and morpho-hydrological conditions (Ngecu and Mathu 1999; Muwanga *et al.,* 2001; Breugelmans, 2003; Knapen 2003; Glade and Crozier 2004; Kitutu *et al.,* 2004; Knapen *et al.,* 2006). Identifying the factors controlling landslide distribution and determining their relative importance may be difficult (Guzzetti *et al*., 2006a; Jacob, 2010). Steep slopes, high annual rainfall, increasing population pressure and deforestation, have been identified as the major cause of landslides in the East African highlands (Glade and Crozier, 2005; Broothaerts *et al*., 2012). These factors can be divided into preparatory factors, which make the slope susceptible to movement without actually initializing it, and triggering factors, which finally initiate the movement (Broothaerts *et al.*, 2012). These parameters are elaborated in the subsequent subsections.

2.2 Topographic characteristics and landslides

The topographic characteristics of any region have a greater implication than any other parameter including soil and land cover on landslide occurrence (Fernandes e*t al*., 2004; Gao and Maro, 2010; Loos and Elsenbeer, 2011). According to Gao and Maro (2010), topography has been widely reported to bear a close association with landslides. The exact control of topographical settings over landsliding is usually analysed through spatial overlay of landslide affected areas with topographic parameter layers (Garcia-Mora *et al.,* 2012; Frazier *et al*., 2013). The major topographic parameters that influence landslide occurrence include slope gradient, aspect, curvature, roughness, distance from drainage network and discontinuities (Selby 1986; Bagoora 1988; Knapen *et al*., 2006). Other complex topographic parameters include topographic wetness index, stream power index, slope length and topographic position (Broothaerts *et al*., 2012; Frazier *et al.,* 2013). These topographic parameters have a spatial variation depending on the terrain and therefore regional or site-specific examination is important for the understanding of landslide occurrence.

2.2.1 Slope gradient

According to Garcia-Ruiz *et al*. (2010), slope gradient is one of the condition that influence landslide occurrence. Many studies have frequently used slope gradient in preparing landslide susceptibility maps due to its direct relationship to landslides (Lee and Talib, 2005; Sassa *et al*., 2005; Yalcin, 2011; Yashar, *et al*., 2013). Slope angle is an essential component of slope stability analysis. It has been reported that as the slope angle increases, the shear stress in soil or other unconsolidated material also increases (Garcia-Miro *et al.*, 2012). Yang *et al*. (2007) and Wati *et al.* (2010) note that steep slopes, particularly those at high elevations, are susceptible to failure. This is due to the increasing shear stress against reducing shear strength (Selby, 1993). Landslide occurrence is concentrated on relatively steeper slopes due to the high downslope component which pulls materials (Bagoora, 1988; Appolinaire *et al.*, 2007; Claessens *et al.*, 2007; Guan-Wei and Hongey, 2012; Mugagga *et al.,* 2012; López-Davalillo *et al.*, 2014). Gentle slopes are expected to have a low frequency of landslides because of the lower shear stresses associated with low gradients (López-Davalillo *et al.*, 2014). It has, however, been stated by Liesbet *et al*. (2015) that landslide occurrence increases with an increase in slope gradient up to a certain extent, and then decreases. Few landslides occur on very gentle and very steep slopes. Very steep slopes are less likely to develop a thick cover of superficial material conducive to certain types of landslides (Selby, 1986).

There is a close relationship between soil depth and slope gradient (Boehner, 2002; Boehner and Selige, 2006). Soil depth reduces with increase in slope gradient and elevation (Selby, 1993). According to Zung *et al.* (2008), the distribution of soil depth and topography are important parameters for the occurrence, magnitude and return period of landslides. More soil materials develop on lower and moderately steep slopes than very steep slope sections (Liang and Uchida, 2014). Landslide occurrence is, however, low on slopes with lower gradients due to the less downslope force required to move materials (Nath *et al.,* 2013). Landslide occurrence is dominant on moderately steep slopes due to the availability of materials and the downslope force to move materials (Sidle *et al*., 1985; Sidle and Terry, 1992). Precipitous slopes are generally stable because rapid erosion removes the erodible soils from them, leading to exposure of weathered rocks (Loos and Elsenbeer, 2011). The thin soils in such sections also means limited materials for downslope movement and hence low landslide incidences (Nath *et al.*, 2013). Although there is a high downslope force on very steep slopes, landslide occurrence is low due to thin soils (Bagoora, 1989).

Although slope gradient is a major topographic attribute affecting both the hydrological conditions and stability analysis, its importance seems to always be over-estimated in landslide hazard mapping (Fernandes *et al.,* 2004). Consequently, gentle hillslopes initially considered as having low landslide susceptibility are also affected by landslides (Bagoora, 1997; Hickey, 2000). This implies that other topographic parameters must be taken into consideration. Few studies have tried to consider the contribution of other topographical attributes on landslides. In Southern Brazil, hillslope form, although earlier suggested as an important parameter (Coelho and Fernandes, 1990) was not incorporated into stability analysis and landslide hazard mapping procedures. More recently, the role of concave forms has been intensively investigated, including their effects on surface and sub-surface hillslope hydrology and landsliding (Fernandes *et al*., 2004; Knapen *et al*., 2006; Liesbet *et al*., 2015). The present study, in addition to examining the role of slope gradient also analysed other topographic characteristics and their influence on landslide occurrence.

2.2.2 Slope curvature

Slope curvature has been reported by many authors as a major topographic parameter controlling landslide occurrence (Acharya *et al*., 2005; Mugagga, 2011; Liang and Uchida, 2014). Knapen *et al*. (2006) observe that plan concave slopes are more susceptible to landslides than convex slopes. Hillslopes with a convergent plan shape tend to concentrate subsurface water into small areas of the slope. This leads to a rapid increase in pore-water pressure during rainstorms (Montgomery and Bradon, 2002; Fernandes *et al*., 2004). An overlay of several landslide sites on a curvature surface by Mugagga (2011) on Mount Elgon in eastern Uganda also revealed a spatial correlation between landslide occurrence and topographical concavity. Concavities on slopes are areas where eroded soils and water collects (Knapen *et al*., 2006). The moisture collecting in concave sections induces weathering, which avails more materials for downslope movement (Liang and Uchida, 2014). The moisture in the concave sections also leads to saturation making the materials therein unstable and the result is downslope movements (Gao and Maro, 2010).

Landslides often occur in areas of convergent topography in which subsurface soil water flow paths give rise to excess pore-water pressures downslope (Yashar *et al.*, 2013). According to Gao and Maro (2010), areas of concave curvature tend to remain saturated between storms due to convergence of ground water flow. In concave forms, water flow is concentrated in hollows. This increases the moisture content of the soil and the amount of time soil will remain saturated (Gu and Wylie, 2016). Curvature therefore affects surface and subsurface hillslope hydrology in determining slope stability (Infascelli *et al.*, 2013; Raju and Nandagiri, 2015). Several studies indicate that landslides are mostly confined to medium and steep slopes along topographic hollows (Reneau *et al.*, 1987; Acharya *et al*., 2005; Gao and Maro, 2010). Some studies note that landslides may also occur on planar slopes (Van Den Eeckhaut *et al.*, 2013). Landslides in the present study area follow topographic hollows with profile concave forms along clearly defined lines. An investigation into the conditions of such sections is therefore important for the understanding of landslide occurrence in the region. It is worth to analyse the properties that confine landslides along topographic hollows which is a unique topographic attribute. Other complex topographic parameters including topographic wetness index, stream power index and topographic position index also need to be examined. This would give a proper understanding of the role of topographic characteristics on landslide occurrence in the study area.

2.2.3 Topographic wetness index (TWI)

According to Loos and Elsenbeer (2011), topography is the driving force for water movement. At any particular point on the landscape, Topographic Wetness Index (TWI) is the ratio between the catchment area and the slope at that point (Infascelli *et al*., 2013). Whereas high moisture is expected in areas with convergence topography, slope sections with divergence topography normally have low moisture (Grabs *et al.*, 2009). The high saturation rates in topographic hollows lead to instability of slope materials due to reduced cohesion (Agnew *et al.*, 2006). Steep convex areas on spur slopes and hilltops are not vulnerable to landslide occurrence due to the low saturation rates in such sections (Ali *et al*., 2014; Raju and Nandagiri, 2015). Convex areas remain dry and therefore stable. Some studies indicate that valley bottoms have the highest saturation rates due to low slope gradient which does not allow water to drain away fast enough (Gomi *et al.*, 2008; Grabs *et al.*, 2009). The high saturation however does not result in landslide occurrence in the valley bottoms. The valley bottoms lack a gradient that is steep enough to initiate movement of materials (Ali *et al.*, 2014). Landslides often occur in areas with moderate saturation which also have steep gradients and have conducive energy to initiate downslope movement of materials (Nath *et al.*, 2013).

2.2. 4 Stream power index

Stream power index (SPI) is an indicator of sediment transport capacity in the landscape (Ferreira *et al*., 2015). According to Moore *et al*. (2007), SPI is a measure of the erosive power of water flow based on the assumption that discharge is proportional to specific catchment area. Merino-Martín, *et al*. (2015) indciate that SPI is directly related to both slope and catchment area. SPI and slope erosion risk increases when the amount of water contributed by upslope areas and the velocity of water flow increase (Rousseau *et al.*, 2012; Ferreira *et al*., 2015). This is due to increase in the gradient and the specific catchment's area (Moore *et al.,* 1993b). It has also been reported by Ferguson (2005), that the potential erosive power of overland flow depends on SPI. Several studies observe that high SPI values indicate areas on the landscape that have a high potential for erosion during and after rainfall events (Gomi *et al.*, 2008; Buda, 2013). The ridge tops and spur slopes are not vulnerable to landslide occurrence due to low erosive power (Fernandes *et al.,* 2004; Gao and Maro, 2010). High landslide occurrence is expected along drainage lines with more power to transport materials (Buda, 2013).

2.2.5 Topographic position index (TPI)

According to Weiss (2001), Topographic Position Index (TPI) measures the differences between elevation at the central point and the average elevation around it within a predetermined radius. Many physical processes acting on the landscape, including landslides, are highly correlated with the topographic position of the landform (Tagil and Jenness, 2008; De Reu *et al*., 2013). The landforms include hilltop, valley bottom, exposed ridge, hollows, flat plain and upper or lower slope (Weiss, 2001). Characterized by topographic hollows, the middle slope positions are vulnerable to accumulation of the eroded materials from hilltops and spur slopes (Briggs and Knapp, 2008; Mokarram *et al.*, 2015). TPI is capable of predicting areas susceptible to saturation and those that have the potential to produce overland flow (Seif, 2014). A number of studies show that locations exhibiting long low-angled slopes are the most prone to the formation of variable saturation areas and thus susceptible to landslides (Reneau and Dietrich, 1987; Weiss, 2001). The ridge tops and spur slopes are relatively stable because of thin soils due to high erosion on them (Gao and Maro, 2010).

In the present study, an analysis of topographic characteristics was important in order to understand its influence on landslide distribution in the region. Topographic parameterisation involved analysing both primary and complex topographic variables and how they influence landslide occurrence in Kigezi highlands of south western Uganda.

2. 3 Hydrological parameters and landslides

According to Selby (1982), water is an important factor in slope stability just like gravity. Water is very critical in downslope movement of materials (Morgan, 1993; Selby, 1993). Heavy rainfall on a slope known to be near its angle of repose can spell trouble in populated areas like Kigezi highlands (Bagoora, 1997). According to Bhudu (2000), hydrological conditions in the soil are very crucial in the initiation of landslides. Water has the effect of adding significant weight and reducing friction (Sidle and Ochiai, 2006). Increased humidity can cause enhanced scouring, undercutting and higher pore-water pressure of structurally pre-conditioned mountain fronts and valley walls (Raju and Nandagiri, 2015). Several authors show that deep rotational landslides on concave slopes can be attributed to the concentration of runoff and sub-surface water which reduces slope shear resistance (Coker and Flores, 1999; Westerberg, 1999; Glade, 2002; and Knapen *et al*., 2006). The amount of water entering a slope is a function of vegetation cover, drainage network, soil type and rock structure (Bagoora, 1998 and Buda, 2013). The water entering is an important factor indicating the significant linkage between landslide hazards, the processes and conditions in the atmosphere and hydrosphere (Garcia Lopez-Davalillo *et al*., 2014). To represent the hydrologic factor in landslide hazard assessments, indirect measures are used and mapped to show the influence of the area's hydrology (Guzzetti and Tonelli, 2004).

The main landslide triggering mechanism is the loss of soil cohesion arising from an increase in pore water pressure during periods of intense and/ or prolonged rainfall (Kitutu *et al*., 2009). Rainfall influences slope stability in many ways including the wetting of the slope forming material (Larsen and Simon, 1993; Lin and Jeng, 2000). The infiltrating water increases the total weight of the slope material. It leads to a rise in the water table and increases the pore pressure in fine-grained sediments and can therefore trigger a landslide (Glade, 2003; Broothaerts *et al*., 2012). An association of landslide occurrence and heavy rainfall has been documented by many authors (UNESCO/UNEP, 1985; Christiansen and Westerberg, 1999; Knapen *et al*., 2006). Prolonged rains with a lower intensity result in higher and deeper infiltration and lower runoff in sloping areas (Bagoora, 1998; Kitutu *et al*., 2009). Studies by Guzzetti *et al.* (2006b), and Jiang *et al*. (2014), show that for rainfall induced landslides, hydrological conditions in the soil like moisture distribution are crucial. High rainfall leads to development of pore water pressure and is considered as an important trigger of landslides (Iverson, 1997). Landslides are triggered by both high intensity and short duration (Jiang *et al.*, 2013) as well as prolonged rainfall (Larsen and Simon, 1993). The initiation of shallow landslide is in theory attributed to the occurrence of extreme rainfall events (Polemio and Sdao, 1999; Glade, 2003). Generally, rainfall is an important input for landslide occurrence as it leads to pore-water pressure which affects soil strength (Wieczorek, 1987; Polloni *et al.*, 1992; Knapen *et al.*, 2006).

The present study was based on the assumption that, the actual mechanism leading to slope failure in the highlands is related to the loss of cohesion due to soil saturation during the rainy season. The trigger in itself may be due to the development and attenuation characteristics of pore water pressure. As landslide initiation in the Kigezi region is assumed to be driven by a sharp rise in the pore water pressure following net precipitation input, rainfall data was required. Rainfall being the trigger of landslide occurrence in the region, a detailed analysis of the spatial-temporal rainfall behaviour in the catchment was paramount. Rainfall patterns for Kigezi region have changed and intensity-distribution become more extreme during some months, seasons and years than others (Bagoora, 1997; NEMA 2012). High rainfall amounts are normally received in the months of March to May, and September to November (Bagoora, 1998).

2. 4 Geology and landslides

The characteristics of bedrock materials like shear strength, their permeability and degree of weathering are generally influential condition for landslide occurrence and can determine the styles of failure (Bagoora, 1988; Selby, 1993). Bedrock influences landslide occurrence in several ways; for example, weak- incompetent rock is more likely to fail than strong-competent rock (Muggaga, 2011). The strength of a rock mass depends on the type of rock and the presence and nature of discontinuities such as joints or other fractures (Sparks, 1986). Bagoora (1997) shows that the more discontinuities present in bedrock, the greater the likelihood of rock instability. Geology has an important influence on the problem of landslides in any region. Geology has direct influence on slope factors such as slope angle and stability (Bagoora, 1988; 1989). For example, Kirkby and Morgan (1980) report that in humid areas like the present study area, sandstones are typically stable up to 30˚ whereas shale's are stable up to only 8-15˚. This indicates a much lower angle of repose from the latter.

Rock type may exert control on landsliding by influencing the strength of surface material in the area. For example, soils derived from schist's or shale's contain high percentages of clay (Bagoora, 1988). Such soils have different strength characteristics than coarser-grained soils such as those derived from granitic bedrock (Selby, 1993). Bagoora (1998) observes that shale's are notorious for landslide activity. Slopes developed on phyllite and shale's are susceptible to landslide risk. Sparks (1986) further note that pyroclastic materials are also associated with greater and more frequent landslides than rocks such as well cemented sandstones and granites. The complex rock system in the Kigezi highlands has been affected by various pedogenic processes that give rise to complex soil types whose inherent differences in erodibility greatly determines the magnitude of landslide hazards.

2.5 Soil properties and landslides

Soil types and physical characteristics are a key determinant of slope stability (Sidle and Ochiai, 2006; Liesbet *et al*., 2015). Sidle *et al*. (1985) observe that soil properties that affect slope stability are those that influence the rate of water movement in the soils and the capacity of the soils to hold water. These properties include depth, particle and pore size distribution in the soil matrix, cohesion, organic matter content, strength, and friction of the materials making up the slope (Cruden and Varnes, 1996; Yalcin, 2007). Soil depth forms one of the most important factors for assessing the stability of the soil and landslide susceptibility of an area (Fernandes *et al*., 2004). Liang and Uchida (2014) suggested that the distribution of soil thickness and topography are important parameters for the occurrence of shallow landslides. The depth of soil profile and its moisture content determine how water can be stored in the soil before saturation is reached (Merino-Martin *et al.*, 2015). The response of soil moisture in deep layers differs from that in surface soil layers. In surface soil layers, moisture responds intensively and quickly, and then reaches peak values within a short period of time (Morgan, 1993). During a rainfall event, drainage of water through the soil profile is stopped at the boundary of clay pans, thereby causing water to accumulate (Yalcin, 2011). The accumulating water leads to the development of pore-water pressures in the soil material (Yilmaz and Karacan, 2002). This can cause the overlying materials on top of clay pans to loose strength hence leading to slope failure.

Soil texture is another major pedological parameter that can explain landslide susceptibility (Day, 1965; Gee and Bauder, 1986; Mugagga *et al.*, 2011). Studies by Jadda *et al*. (2009), and Wati *et* *al*. (2010) also report on the susceptibility to landslides by fine-textured clayey soils due to their small pores that release water gradually. Such properties make soils prone to landslides because of the high-water retention. Wati *et al.* (2010) observe that low permeability of fine textured clayey soils exacerbates the vulnerability to landslides. Kitutu *et al*. (2009) also report that areas where sandy clay loams are underlain by sandy clay soils are prone to landslide occurrence. Mugagga *et al*. (2011), also indicate that landslides mostly occur in unconsolidated sands that have lower internal cohesion than clays. According to Kitutu *et al*. (2011), sandy soils allow fast flow of water into the soil which is held in the deeper layers with high clay content, causing water saturation and slope failure. The clay content in soil influences its erodibility due to inherent strong inter-particle bonding, which is strong in clay soils and weaker in silts and sandy soils (Morgan, 1993). According to Yalcin (2007) and Broothaerts *et al.* (2012), the high clay content in the soil can be an important precondition for landslides. This is due to the chemical and physical properties of clay minerals (Keller and Dexter, 2012). Several studies have demonstrated the influence of high clay content on landslide occurrence (Knapen *et al*., 2006; Kitutu *et al*., 2009; Wati *et al*., 2010 and Mugagga *et al*., 2011). These studies show that high clay content can cause soil materials to have low cohesion, especially during a rainfall event leading to landslides.

The dominant clay mineral present in a soil also influences its expansion and sliding potential (Baynes, 2008). For example, Ohlmacher (2000) and Yalcin (2007) associated landslide occurrence to smectite and illite clay groups. Such clays have lower shear strength and higher swelling potentials (Fauziah *et al.,* 2006). They are more prone to landsliding than shale's composed of kaolinite and chlorite (Ohlmacher, 2000). Highly plastic inorganic soils are prone to sliding during rainfall events due to the reduction in shear resistance (Inganga *et al.*, 2001 and Dai *et al*., 2002). Several studies report on the role of liquid limits in characterising the problem nature of soils (Van Der Merwe, 1964; Mario *et al*. 1996; Msilimba and Holmes 2005; Fauziah *et al*. 2006; Baynes, 2008). Studies by Chartwin *et al*. (1994) and Isik and Keskin (2008) show positive correlations between high plasticity, fine-grained inorganic clay and landslide occurrence. Plasticity is due to the presence of clay minerals or organic material (Keller and Dexter, 2012). Plasticity is influenced by the void ratio and is high in inorganic clays (Fauziah *et al*., 2006; Baynes, 2008). Other important parameters in triggering landslides are swelling properties of clay and the rate at which water infiltrates into clay at depth (Inganga *et al.*, 2001; Yang *et al*., 2007;

Jadda *et al*., 2009; Mukasa *et al.* 2016). Zung *et al*. (2009) observes that certain soil characteristics may be useful tools for assessing landslide frequency.

According to Fauziah *et al*. (2006), the most important cause of shallow landslides is the decrease of matric suction after a rainstorm and the development of positive pressures above the water table. In particular, soil shear strength decreases non-linearly with increasing soil matric suction (López-Davalillo *et al.*, 2014). When the suction becomes less negative as the soil approaches saturation, the soil becomes more susceptible to failure (Selby, 1993). Factors like stress level, material type, permeability and density, determine the soils shear strength characteristics (Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014). The development of positive pore water pressures can push particles apart (Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014), which acts against the normal stress, effectively reducing it. Since particles are pushed apart, cohesion and friction also reduce (Morgan, 1993).

Whereas negative pore water pressure will increase both cohesion and friction, and therefore strength, positive pore water will decrease both cohesion and friction, decreasing strength (Selby, 1993). According to Zezere *et al*. (1999), during rainstorms, majority of landslides occur due to pore water pressures. Increase in water pressure leads to a reduction in shear resistance and effective stress (Morgan, 1993; Palemio and Sdao, 1999) leading to landslide occurrence. All these pedological parameters have a significant spatial variation and require site-specific investigations to understand their contribution in landslide occurrence. The present study therefore examines specific soil properties including soil depth, angle of internal friction, clay mineralogy, particle size distribution, cohesion and bulk density. This information is required to determine landslide susceptibility in a particular region.

2. 6 Land-use /cover and landslides

Changes in land-use and vegetation cover can result in surface alterations (Van Western *et al*., 2003). Surface alterations have been identified as one of the condition for landslide occurrence (Promper *et al*., 2014). Alcantara *et al.* (2006) state that land use change drives land degradation and can make hillslopes more susceptible to mass movements. Human activities leading to changes in land cover patterns are the most rapid driver of global change (Van Western *et al.,* 2003; Slaymaker *et al*., 2000; Promper, *et al.,* 2014). Land cover changes cause large alterations in the hydro morphological functioning of hillslopes (Morgan, 1993; NEMA, 2007; Garcia-Mora *et al*., 2010; Mugagga *et al*., 2012). This can affect rainfall partitioning, infiltration characteristics, runoff production and even lower the shear strength of the soil leading to landslides. Many studies reveal a close correlation between slope stability, especially landslide occurrence and land-use/cover changes (Cruden and Miller, 2001; Breugelmans, 2003; Beguería, 2006; Breuer *et al*., 2009; Bauer *et al*., 2012).

The presence of vegetation cover substantially modifies parameters such as cohesion, internal friction angle, weight of the slope-forming material and pore-water pressure (Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014). Land degradation through deforestation affects material shear strength (Glade, 2003; Meusburger and Alewell, 2009; Wasowski *et al*., 2010; Mugagga *et al*., 2011). Destruction of vegetation cover reduces organic matter and humus, which is important for binding materials together. The loss of root networks reduces the cohesion of soil, while decreased evapotranspiration raises water levels (Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014). Slope failures often occur several years after logging, and after root systems decay away (Wasowski *et al*., 2010; Das *et al.*, 2011).

New land-cover patterns may occur as a result of anthropogenic activities such as economic developments, population growth or land abandonment (Carswell, 2000; Carswell, 2002b; Promper, *et al.,* 2014). These activities are capable of greatly altering slope form and ground water conditions (Swanson and Dyrness, 1975; Bagoora, 1998; Clark and Wilcock, 2000; Bamutaze, 2005; Barasa *et al*., 2010; Buda, 2013). The altered conditions may significantly increase the susceptibility to landslide occurrence of a given area (Meusburger and Alewell, 2008; Kato and Mutonyi, 2011; Mugagga *et al*., 2012). The expansion of settlements is increasing the impact of natural disasters both in the developed and developing countries (Ronsenfeld, 1994; Alexander, 1997; Guzzetti *et al*., 1999; Clark and Wilcock, 2000; Promper, *et al.,* 2014). Landslides have traditionally been regarded as key indicators of forest disturbance, particularly in association with logging activities (Montgomery *et al*., 2001; Promper and Glade, 2012; Roller *et al.*, 2012), land use and climatic change (Van Beek, 2004; Van Den Eeckhaut *et al*., 2013). They are also regarded as a response to human imposed changes such as road building (Larsson, 1986; Meusburger and Alewell, 2008). To address the spatial-temporal variability of landslide risk, one aspect is to analyse past land cover patterns as well as future development of the land use and cover which was one reason for this study.

According to the National State of Environment Reports for Uganda (2000, 2002, 2004, 2008, 2010, 2012 and 2014), land is becoming increasingly scarce as the country's population increases at a high rate. People are forced to exploit steep slopes for settlement and agriculture, causing land degradation. Land degradation may lead to increased landslides in extreme cases. Land use changes associated with techniques of land preparation, cause massive soil redistribution and slope morphology changes (Bagoora, 1998). This leads to increased risks of catastrophic mass movements (Muggaga *et al.,* 2010). Several studies elsewhere have quantitatively evaluated the beneficial effects of vegetation on slope stability. This is in terms of its role in net precipitation, inducing critical pore water pressure conditions (Dhakal and Sidle, 2004) and its mechanical effects, particularly root reinforcement (Jakob, 2000; Karlsi *et al.*, 2009). Most of such studies focus on Mediterranean and temperate areas (Ferrer and Ayala, 1997; Jakob, 2000; Foster *et al*., 2008; Karlsi *et al.*, 2009; Ferreira *et al*., 2015). Only a handful of studies (Collison *et al*., 1995; AGS, 2007b) have attempted to quantify the effects of plants on slope stability in the humid tropics (Sidle *et al.*, 2006). All these studies however, confirm that there exists a close correlation between land cover changes and landslide patterns. Land cover conversion from forest to pasture permanently reduces slope stability (Selby, 1993). Studies attempting to quantify the beneficial effects of vegetation on slope stability are important, especially where deforestation and infrastructure development in the landslide prone areas such as Kigezi highlands of south western Uganda are drastically increasing.

In the present study, it is common to observe cultivated fields on steep valley side slopes which have been massively washed by rains, especially early in the growing season. A pattern of extensive exposures of sub-soil and fine network of dendritic, or longitudinal rills can be seen on most of these hills (Bagoora, 1998). Many hillslopes in Kigezi highlands have been reduced to nearly bare rocky surface by poor methods of cultivation (NEMA, 2012). Steep upper slopes have been encroached upon, stripped of their vegetation and led to hipper soil erosion and landslides

(Bagoora, 1997). Almost every available space has been intensively cultivated and gardens appear like a continuous carpet for kilometres, in the valleys and on hill slopes (NEMA, 2010). The bulk of the area is a pocket of land congestion on a rugged and sometimes stony terrain (Bagoora, 1993). Worse still, the absence of vegetation cover and over-cultivation mostly using poor farming methods on the uplands, has led to severe accelerated erosion in the Kigezi highlands (Bagoora, 1998; NEMA, 2014).

Studies by Farley (1996) and Carswell (2000) indicate that the original natural ecosystems in Kigezi highlands were characterized by a forest-savannah mosaic in most of the medium altitude areas. Higher elevations had moist lower montane forests. More than 2000 years ago, cultivation, grazing and permanent settlements are reported to have started leading to the clearing of forests (Carswell, 2002a; Siriri and Raussen, 2002). According to Carswell (1997), there was already a very small forest cover in most of Kigezi region in the early 20th century as revealed by written accounts and photographs. A study by Breyer *et al*. (1997) shows that small scale farming covered the largest area while woodland and bush land was the second most important land cover type. Changes in land use and cover pose a risk to stability of Kigezi highlands, but the magnitude of the impact is not well understood. The present study considers the influence of dynamic human-induced land cover changes on the occurrence of landslides.

2.6 Modelling landslide occurrence

Characterization of the spatial-temporal distribution of landslides and their causative mechanism is important for geomorphological studies and for natural hazard evaluation (Guzzetti *et al*., 2006b; Biswajeet and Lee, 2010). One way to understand the controls of landslides and predict their spatial-temporal occurrence is given by phenomenological modelling (Van Beek and Van Asch, 2004). It is assumed that conditions which led to landslides in the past are likely to cause them in the future as well (Peter *et al*., 2010). A number of methods and techniques have been proposed to evaluate where or when landslides are most likely to occur (Carrara *et al.*, 1991; Montgomery and Dietrich, 1994; Ayalew and Yamagishi, 2005; Guzzetti *et al.*, 2005), including the use of Geographic Information Systems.

Many studies have been undertaken to assess slope susceptibility to landslides through heuristic, deterministic, conceptual and statistical approaches (Carrara *et al.*, 1991; Dai *et al.,* 2001; VanWesten *et al.,* 2003). A heuristic approach is a direct or qualitative approach completely based on field observations and an expert's priori knowledge (Dietrich *et al.*, 1995; Zêzere *et al.*, 1999; Guzzetti *et al.,* 2005). Deterministic approaches are based on slope stability analyses (Thornes and Alcantara-Ayala, 1998; Sidle and Ochiai, 2006; Isik and Keskin, 2008). They are applicable when the ground conditions across a study area are relatively homogeneous and the types of landslides are known and relatively simple (Jadda *et al.,* 2009).

Comparatively little work has been done on the systematic comparison of different techniques to determine landslide susceptibility, outlining advantages and limitations of the proposed methods (Carrara *et al.*, 1991; Van Westen, 1994; Glade and Crozier, 2005; Lee and Talib, 2005; Peter *et al*., 2010). Process-based landscape models use some variation on the factor of safety equation (Selby, 1982; 1993). Some focus on the geomorphic, hydrological, geological and vegetation data to estimate slope, cohesion and subsurface flow characteristics (Bevern and Kirkby, 1979; Brasington and Richards, 1998; Beven, 2000; Claessens *et al.,* 2006b; Sidle *et al.,* 2006). The fundamental aim of the present study is to develop a conceptual model which integrates all the inherent topographical, hydrological, pedological and anthropogenic parameters and evaluate their convergence to cause landslides. The present study therefore employs a conceptual model in which the convergence of all the landslide underpinnings in determined.

2.7 Conceptual framework

The interaction among different conditions that influence landslide occurrence is illustrated in Fig. 2.2. The spatial and temporal occurrence of landslides in the Kigezi highlands is on the increase due to a complex of interactions among a large number of partially interrelated factors which are both biophysical and anthropogenic (e.g., Meusburger and Alewell, 2008).

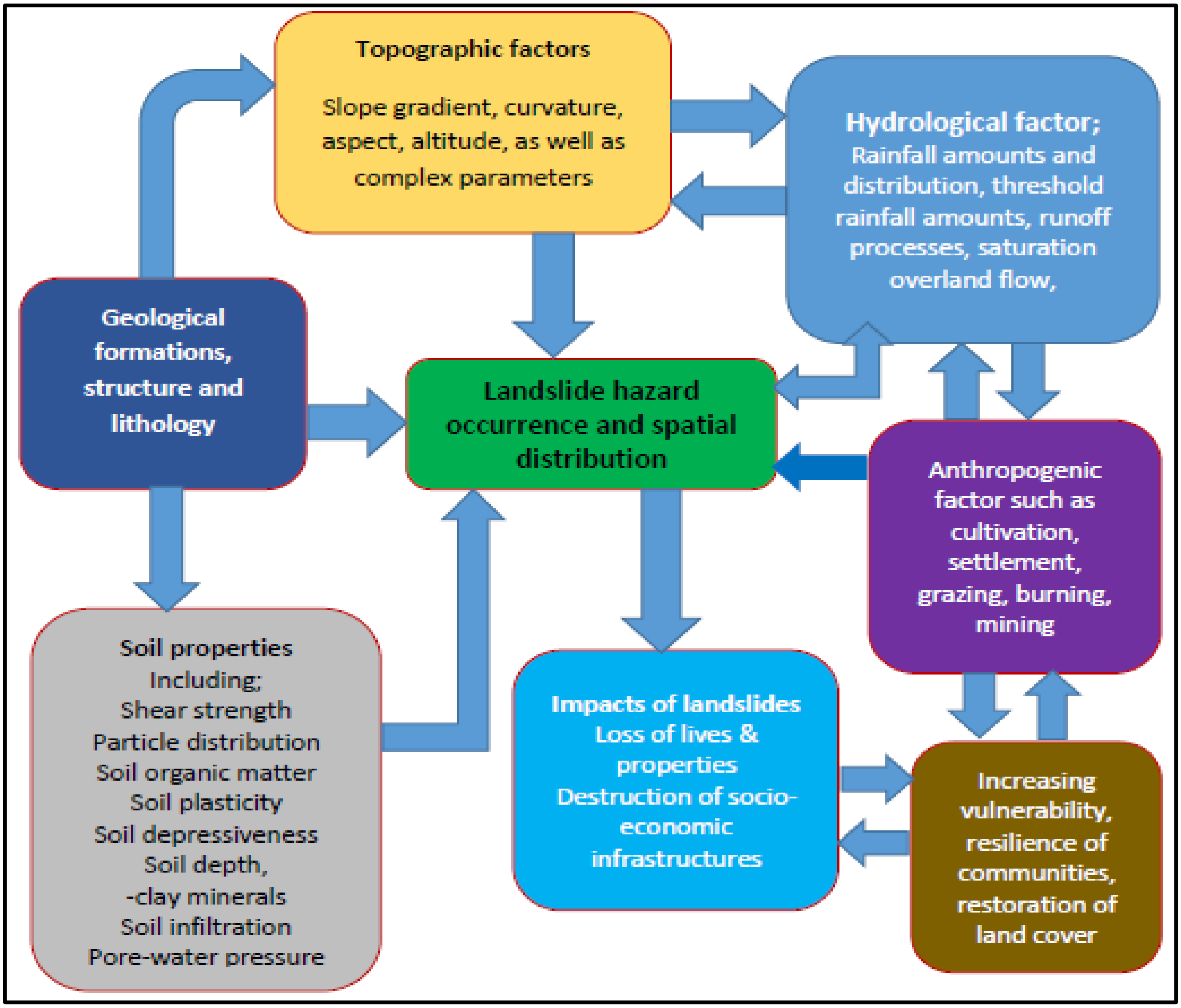

Fig. 2.2: Landslide hazard occurrence conceptual frame-work.

The main landslide occurrence determinants include; (i) the geological setting and formation (e.g., mudstone, phyllite, schists, shale, sandstone, limestone, greentuffes); (ii) topographical structure including; elevation, gradient, aspect, curvature, topographic wetness index, stream power index, topographic position index; (iii) hydrological processes especially; rainfall amounts and intensity, flow accumulation, flow direction, and, (iv) soil characteristics such as the types (e.g., clay, silt, loam, sand), texture, soil depth, cohesion, internal friction, organic matter content, soil moisture, infiltration, and mineralogy (Fig. 2.2). There is also the superimposition of the anthropogenic factors, such as the human impact on the land cover and landscape processes. There is clear evidence of an interaction between the human factor and the biophysical parameters. Most of the landslides have occurred where there is human impact in the highlands (Fig. 2.2).

The effects of landslides on people and structures can be lessened by total avoidance of landslide hazard areas. This can be done through restricting, prohibiting or imposing conditions through landslide hazard-zoning. Local governments can reduce landslide hazards through land use policies and regulations. Individuals can reduce their exposure to landslide by educating themselves on the past hazard history of a site. The responses however, need to be based on adequate scientific information that requires research designed to address both short and long-term issues. This study was therefore worth undertaking.

CHAPTER THREE

Analysis of topographic parameters underpinning landslide occurrence in Kigezi highlands of South Western Uganda

Abstract

The frequency and magnitude of landslide occurrence in the Kigezi highlands of south western Uganda has increased, but the topographic attributes underpinning their occurrence are not well understood. The present study focused on selected topographic parameters namely slope gradient, profile curvature, topographic wetness index (TWI), stream power index (SPI), and topographic position index (TPI). These factors were parameterized in the field and GIS environment using a 10m DEM. Sixty five landslide features were surveyed and mapped to produce landslide distribution maps. Landslides are concentrated along topographic hollows in the landscape. The occurrence is dominant in slope zones where slope gradient is between 25˚ and 35˚, profile curvature between 0.1 and 5, TWI between 8 and 18, SPI greater than 10 and TPI between -1 and 1. Landslides are less pronounced on slope zones where slope gradient is less than 15˚ and greater than 45˚, profile curvature less than 0, TWI less than 8 and greater than 18, SPI less than 10 and TPI greater than 1. The understanding of these topographic parameters and their influence on landslide occurrence is important in landslide hazard mitigation. It is now possible to identify and predict potential landslide zones in the landscape. Safer zones for community activities can also be demarcated using the topographic information. This knowledge generated about the area's topographic characteristics and landslide occurrence will help to reduce on vulnerability while increasing resilience of communities to landslide hazards in this fragile highland ecosystem. This can be done through designating zones for community activities while avoiding the potential landslide zones.

Keywords: Topographic parameters, landslide occurrence, non-volcanic Kigezi highlands.

3.1 Introduction

Landslides are hydro-geotechnical processes most prevalent in highland and mountainous terrains of tropical and sub-tropical regions (Guzzetti, 1999; Schuster and Highland, 2001; Dai *et al*., 2002; Petley *et al*., 2005; Highland and Bobrowsky, 2008; Petley, 2008; Das *et al*., 2010; Broothaerts *et al*., 2012; Mugagga *et al*., 2012; Kirschbaum and Zhou, 2015). Several studies consider the East African highlands to be prone to landslides due to their humid climate, steep topography, and high weathering rates (Knapen, 2003; Knapen *et al*., 2006; Kitutu *et al.,* 2009; Mugagga *et al*., 2011**)**. Although landslides are recognized as widespread phenomenon in the East African highlands, having a great social, economic and geomorphological impact, relatively little research and documentation can be found in the international literature (Ngecu and Mathu, 1999; Knapen *et al*., 2006). In many East African highlands, a clear insight into the local causes for landsliding is lacking (Muwanga *et al.,* 2001; Knapen *et al.,* 2006).

Landslides are triggered under multiple geological and morpho-hydrological conditions (Ngecu and Mathu 1999; Muwanga *et al.*, 2001; Breugelmans 2003; Knapen 2003; Glade and Crozier 2004; Kitutu *et al.,* 2004; Knapen *et al.*, 2006). Identifying the factors controlling landslide distribution and determining their impacts may be difficult (Guzzetti *et al*., 2008; Van Den Eeckhaut *et al*., 2013; Liesbet *et al.,* 2015). Topography is often considered as the most important parameter in landslide occurrence (Selby, 1993; Gao and Maro, 2010; Loos and Elsenbeer, 2011). Various scholars have inferred on the contribution of topographic parameters in landslide occurrence (Jakob, 2000: Zhou *et al.,* 2002; Begueria, 2006; Hong *et al.,* 2007). The topographic characteristics of any region have a greater implication than any other parameter including soil and land cover on landslide occurrence (Fernandes e*t al*., 2004; Gao and Maro, 2010; Loos and Elsenbeer, 2011). According to Gao and Maro (2010), topography has been widely reported to bear a close association with landslides. The exact control of topographical settings of a landslide is usually analysed through spatial overlay of landslide distribution maps on topographic layers (Lopez-Davalillo *et al.,* 2014). The major topographic parameters that influence landslide occurrence include slope gradient, aspect, curvature, roughness, distance from drainage network and discontinuities (Selby 1982; Bagoora 1988 and 1989; Knapen *et al*., 2006). Other complex topographic parameters important in landslide occurrence analysis include topographic wetness index, stream power index, slope length and topographic position (Grabs *et al*., 2009; Gao and Maro, 2010; Broothaerts *et al*., 2012; Infascelli *et al.*, 2013). These topographic parameters have a spatial variation depending on the area's terrain and therefore regional or site-specific examination is important for the understanding of landslide occurrence.

The Kigezi region of south western Uganda is one of the most remarkable landslide affected East African highlands due to its steep topography (Bagoora, 1988; NEMA, 2010). The disaster risk of communities to landslide hazards is poorly understood due to the paucity of information on the topographic characteristics of these highlands underpinning landslide occurrence. Although the frequency and magnitude of landslide occurrence in these highlands have increased (NEMA 2012; Kabale District Environmental Report, 2015), the topographic parameters underpinning their spatial distribution are not well understood. Information on topographic characteristics is important for predicting and/or identifying potential landslide zones. The information on topographic parameters and their influence on landslide occurrence would help reduce vulnerability and disaster risk of communities in these highlands. The present study therefore, examines the influence of topographical parameters on landslide occurrence, especially their influence on hillslope hydrology and soil development. Topographic parameterisation involved analysing both primary and complex topographic variables and how they influence landslide occurrence in the Kigezi highlands of south western Uganda.

3.2 Materials and methods

3.2.1 Mapping landslide dimensions and their spatial distribution

Detailed field investigations were carried out to establish the magnitude and spatial distribution of landslide occurrence in the Kigezi highlands. Analysis of the landslide dimensions (depth, width and length) and spatial distribution was done using both field surveys and GIS desktop techniques. A field inventory was undertaken to map the visible landslide scars using hand-held GPS receivers. Coordinates for the mapped landslide scars were imported to ArcGIS 10.1 software to produce landslide distribution map for the study area. By means of the spatial analysis in a GIS environment, the landslide and topographic parameter maps were compared to study their relationships. This relationship between landslide occurrence and topographic parameters was established at the failure zone or point of landslide origin. This was done by establishing the value of the topographic parameter at the failure zone.

Fig. 3.1: Field investigations to map and measure landslide scars and their distribution.

Landslides were surveyed by focusing on relief features such as concave and convex forms, slope gradient variations, shape of slope meso-forms, fractures, and other non-anthropogenic linear elements. Field surveys and measurements of landslide sites was done in relation to slope position and other topographic parameters (Fig. 3.1). The dimensions of landslide scars were established by carrying out measurements of their average width, depth and overall length from head to toe using tape measures. Landslide morphological analysis was carried out from the valley bottoms where deposition landslide debris occurred to the failure zones. Landslide scars were divided into several segments based on their morphology and location (e.g., head, middle and toe). Repeated measurements were conducted in each of the segments and averages computed. Landslide dimensions were used to establish the area of the scars and estimate the volume of materials removed by each occurrence. Historical landslide data and records were obtained from Kabale District Environmental Reports (2008, 2012 and 2015), local government records as well as interaction with the local communities.

3.2.2 Topographical parameterization

The topographic parameters analysed in this study were derived in a TOPMODEL framework. According to Quinn *et al.* (1995), the TOPMODEL is a modelling framework in which the role of topography in catchment hydrology can be explored. The TOPMODEL is a physically based watershed model for simulating runoff generation (Beven and Kirkby, 1979). The topographic index has been developed to describe the tendency of water to accumulate and to move down slope by gravity (Moore *et al*., 1991). It is assumed that within the TOPMODEL, areas with similar properties can be grouped together for computational purposes (Brasington and Richards, 1998; Beven, 2000). The TOPMODEL was written explicitly to model the role of topography in catchment hydrology (Beven, 1989; Quinn *et al*., 1991). Over the last decade, geographical information system and digital elevation models have been increasingly used to automatically classify landforms (Wilson and Gallant, 2002; De Reu *et al.,* 2013). The topographic parameters used in this study were topographically generated using a Digital Elevation Model (DEM).

A 10 m DEM for the region was used to extract major topographic attributes and characteristics of the highlands. The DEM was generated by first digitizing 10 m interval contours from 1:20,000 topographic maps obtained from the Uganda’s Lands and Survey Department. A 10m DEM was interpolated from contour lines using the Topo to Raster tool in ArcGIS 10.1, which converts vector topographic values to a raster DEM surface using the ANUDEM algorithm. Topographic parameterization was performed using ArcGIS 10.1 and SAGA GIS 2.3.1(Systems for Automated Geoscientific Analyses) software. SAGA GIS 2.3.1 (See Conrad *et al.,* 2015) was used for DEM pre-processing, including import, projection and merging of data, as well as gap filling, curvature calculation and cluster analysis. DEM preparation included morphological filtering and surface depression filling. The process of filtering helped to smoothen the data to create elevation averages for better interpretation.

Topographic analyses involved computation and establishment of selected topographic parameters namely slope gradient, slope shape (profile form curvature), topographic wetness index, stream power index and topographic position index. This was done using a 10m DEM in ArcGIS 10.1 as well as field measurements using clinometers and handheld GPS receivers. These parameters were prioritized in the present study due to their direct implication for soil development and hydrological aspects that influence landslide occurrence. Preliminary investigations indicated that parameters such as aspect and altitude do not have a direct influence on landslide distribution in the study area. Their effect on, for instance hydrology and moisture dynamics is in a relative sense. These parameters did not show a relationship with landslide occurrence and were thus not considered during the present study.

Slope gradient and curvature

Slope gradient and curvature values were generated directly from the DEM in ArcGIS 10.1. Curvature values for the highlands were established in the direction of slope gradient (profile curvature). In the present study, surface profile curvature was focused on, as it was identified during field surveys as the dominant curvature form closely related with landslide occurrence in the study area. According to Carson (1985), profile curvature affects the driving and resisting stresses within a landslide in the direction of motion. It therefore affects the flow velocity of water draining the surface and influences landslide occurrence (Gao and Maro, 2010). Curvature values were used to identify peaks, ridges, spur slopes, planar regions, topographic hollows and pits. A profile curvature map for Rukiga catchment was generated and overlain with the landslide scars to establish the influence of curvature on landslide occurrence. Curvature affects the rest of the topographic parameters, as discussed in subsequent paragraphs.

Topographic Wetness Index

The Topographic Wetness Index (TWI) for the highlands was established using the formula;

TWI = In (α / tanβ) where, *α* is the specific catchment area, and tanβ is the slope angle (degrees).

The value of TWI is to determine the zones of saturation (Grabs *et al*., 2009). It was used to estimate the accumulated water flow at any point in the watershed. In the present study, SAGA wetness index was computed because it is based on a modified catchment area calculation which does not consider flow as a very thin film (see Boehner *et al.,* 2006). Consequently, it was used to predict the cells situation in valley floors with a small vertical distance to a channel. SAGA wetness index is a more realistic and higher potential soil moisture determinant than the standard TWI calculation (Boehner *et al.,* 2002; Boehner and Selige 2006). Generally, high TWI values occur in concavities with the requisite upslope contributing area. Low TWI values indicate relatively little upslope contributing area and steep slopes typical of ridges and hilltops (Grabs *et al*., 2009). TWI is thus a surrogate for saturation levels and susceptibility to landslide occurrence. The generated TWI map for the catchment was overlain with the landslide distribution map and the influence of TWI on landslide occurrence was ascertained. TWI is closely related with SPI, which is discussed in the next sub-section.

Stream Power Index

Stream Power Index (SPI) is a measure of erosive power of water flow and is closely related to the topographic wetness index. It was used to identify the erosive effects of concentrated surface runoff (see Gartner *et al.*, 2015a). It is proportional to the specific catchment area (Quin *et al*. 1995). Values were calculated based on the formula;

SPI = *ln* (α* tanβ), where α = specific catchment area and tanβ = slope of the cell.

SPI was used to provide a correlation between water flow paths, flow accumulation and slope that together define the energy potential surface water has for erosion (see Moore *et al*., 1991; Moore *et al*., 2007). This eroding power of flowing water was used to identify potential and actual sites for landslide occurrence. Areas with high SPI values were considered to be more prone to landslide occurrence than areas with low SPI values. A map of the SPI of the catchment was generated and was overlain with landslide features and their relationship was ascertained. In order to remove spurious features, the resulting TWI and SPI maps were filtered using the "majority filter" routine (3×3 scanning window). TWI and SPI were used to quantify flow intensity and accumulation potential for the highlands in the study area. SPI is also closely related with the TPI, which is explored in the subsequent paragraph.

Topographic Position Index

The Topographic Position Index (TPI) measures the differences between elevation at the central point (ZO) and the average elevation (Z) around it within a predetermined radius (R) (Weiss, 2001; Gallant and Wilson, 2000). Calculation for TPI was based on the formulae; TPI=ZO-Z and Weiss (2001) method of landform classification was used. The method classifies the landscape into discrete slope position classes using the standard deviation of TPI. Using TPI, six discrete classes were defined. Following Seif (2014), TPI was used to measure the topographic slope positions, automate landform classification and to analyse the topographic location of landslide hazard zones. TPI values provided a simple and powerful means to classify the landscape into morphological classes (e.g., Jenness, 2006). The generated TPI map of the study area was overlain with the landslide distribution map and their relationship was ascertained.

3.2.3. Statistical analysis

Data analysis was aimed at detecting spatial trends in landslide-topographic relationships. Statistical data analysis was performed using Statistica software package. Inferential statistics, hypothesis tests and confidence intervals coefficient of variation were used to describe and compare topographic parameters and 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 determined. The topographic variables were correlated and analysis of variance was conducted on all the parameters and their influence on landslide occurrence. Linear regressions were used to determine the coefficients of determination (R^2) for topographic parameters and landslide occurrence.

3. 3 Results

3.3. 1 Landslide characteristics and distribution

Based on the field surveys carried out in these highlands, as well as reports from the local governments and communities, there is an increase in landslide occurrence in the region over the past 30 years. The spatial distribution of landslides in the study area is illustrated in Fig. 3.2.

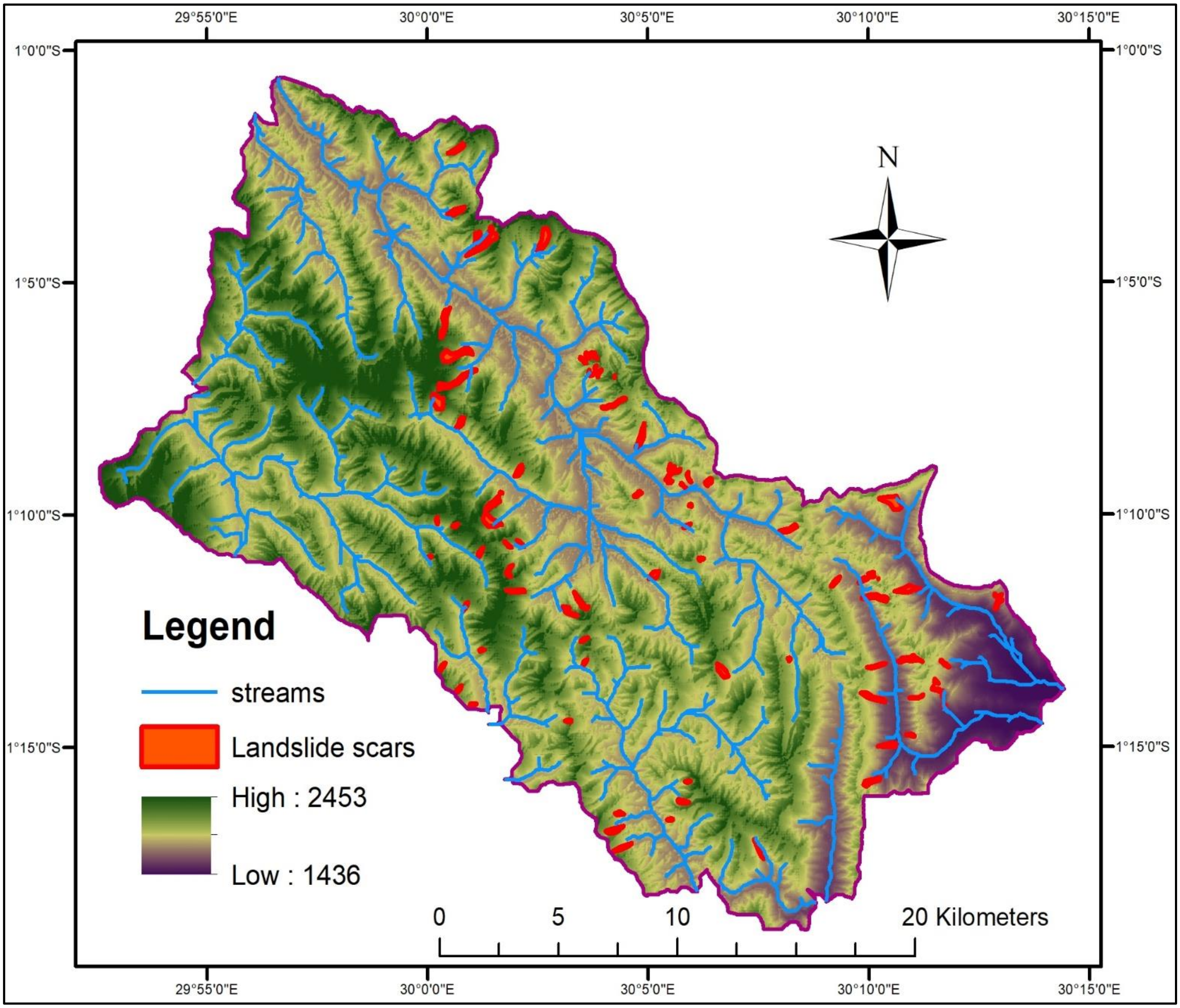

Fig. 3.2: A landslide distribution map for Rukiga catchment.

The landslide scars in the catchment varied from small 12.5m slides to longer, complex flows that extended in some cases to more than 890m. 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 of the landslide scars ranged between 0.9 meters for small scars to 17.5 meters for complex occurrences. The average depth ranged between 0.5 meters for shallow landslides to 5.3 meters for deep seated scars (Table 3.1). The mean area covered by each landslide occurrence varied from 125m^2 for smallest landslides to about 6000m^2 for large scars. The estimated volume of hillslope materials displaced by individual landslides varied widely. Small landslides displaced an approximate 62.5m^3, while large occurrences displaced close to 30 000 m^3 of materials (Table 3.1).

In the present study, out of the 65 visible landslide scars identified and mapped in the catchment, 92% are larger at failure zones than the toe which is a unique attribute. It is unique because elsewhere for example Mt Elgon region in Eastern Uganda, landslide scars are smaller at failure zones and larger towards the toe. 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 unravelled 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 cm 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.

Table 3.1: Landslide scar geometric characteristics

Landslide scars	landslide scar dimensions in meters			Area of the landslide scar in m^2	Volume of the scar in m^3	Gradient (in degrees) 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	30000	33-35
7	10	5	400	4000	20000	32-34
8	10	5.3	498	4980	26394	31-33
9	16.6	4.3	315	5229	22600	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	10508	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	14496	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
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

About 51% of landslide scars have a number of tongues which converged together following topographic hollows. Convergence of the landslide arms occurred in zones where two or more topographic hollows from upslope joined into one wider hollow.


Fig.3.3: Landslide occurrence in Kigezi highlands.

There are signs of landslide recurrence in some areas where there was remobilization of the materials around the former failure zones (Fig.3.3). Some of the landslide scars remain active with different phases of activity especially during the rainy season. In this study, landslide scars were considered as those features which resulted from past failures but which are still visible. Recent landslides are those failures which have occurred within the past 3 years and are still active. Out of the 65 visible landslides in the study area, more than 75% are older landslide scars where margins and head scarp have been degraded. Less than 25% are recent slides with well-defined margins, head scarp, with no or partially developed drainage channels. The recent landslides show signs of future activity.

3.3. 2 Topographic characteristics and landslide occurrence

Field investigations verified that landslides in these highlands occur within clearly defined channels along topographic hollows. Most of the landslides originate from the mid slopes, where materials had accumulated, then widened and deepened as the process continued downslope in zones of deeper soils. Others are shallow and longitudinal following shallow hollows between slopes, while in wider-deeper hollows, deep-seated translation landslides occurred.

3.3.2.1 Slope gradient and landslide occurrence

Slope gradient values for the highlands ranged between 0˚ in the valley bottoms and 63˚ in the upper slopes (Fig. 3.4).

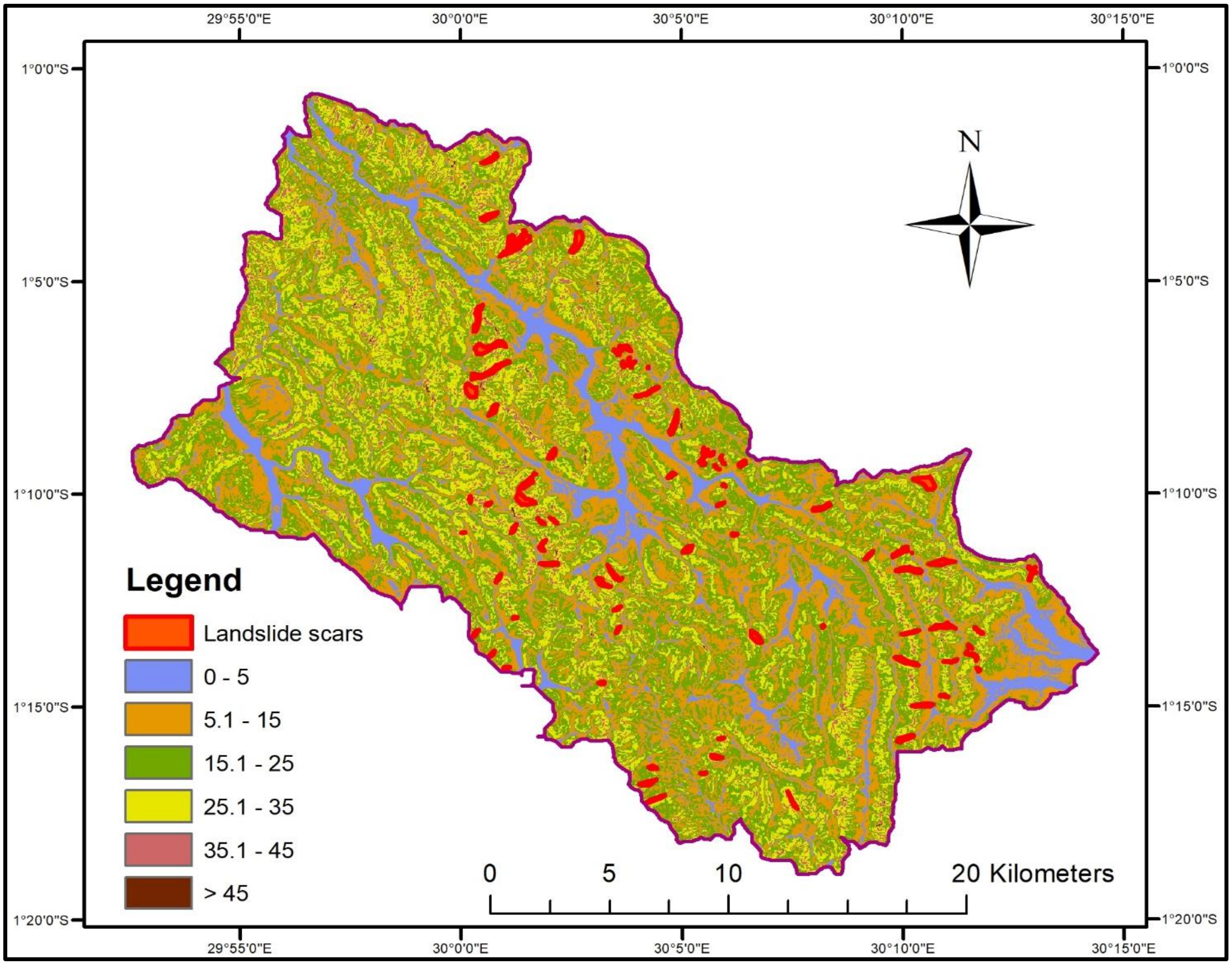

Fig. 3.4: A map of slope gradient and landslide distribution

Hillslope gradient plays a key role in determining landslide susceptibility in the study area. This relationship between landslide distribution and hillslope gradient was explored by binning hillslope gradient values into 10˚ intervals and plotting the frequency of hillslope gradients for the landslides. Using the FAO (2006) slope classification thresholds, slope gradient was categorized into five groups including moderate (<15˚), moderately steep (15˚–25˚), steep (25˚–35˚), very steep (35˚–45˚), and precipitous (> 45˚) slopes. In the study area, more than 55% of the landscape has hillslope gradients lower than 15˚ while less than 30% has gradients of greater than 25˚ (Fig. 3.4). More than 60% of the landslides in the highlands occur on hillslope zones with slope gradients between 25˚ and 35˚ (Fig. 3.5), yet this gradient category accounts for only 28% of the area's topography.

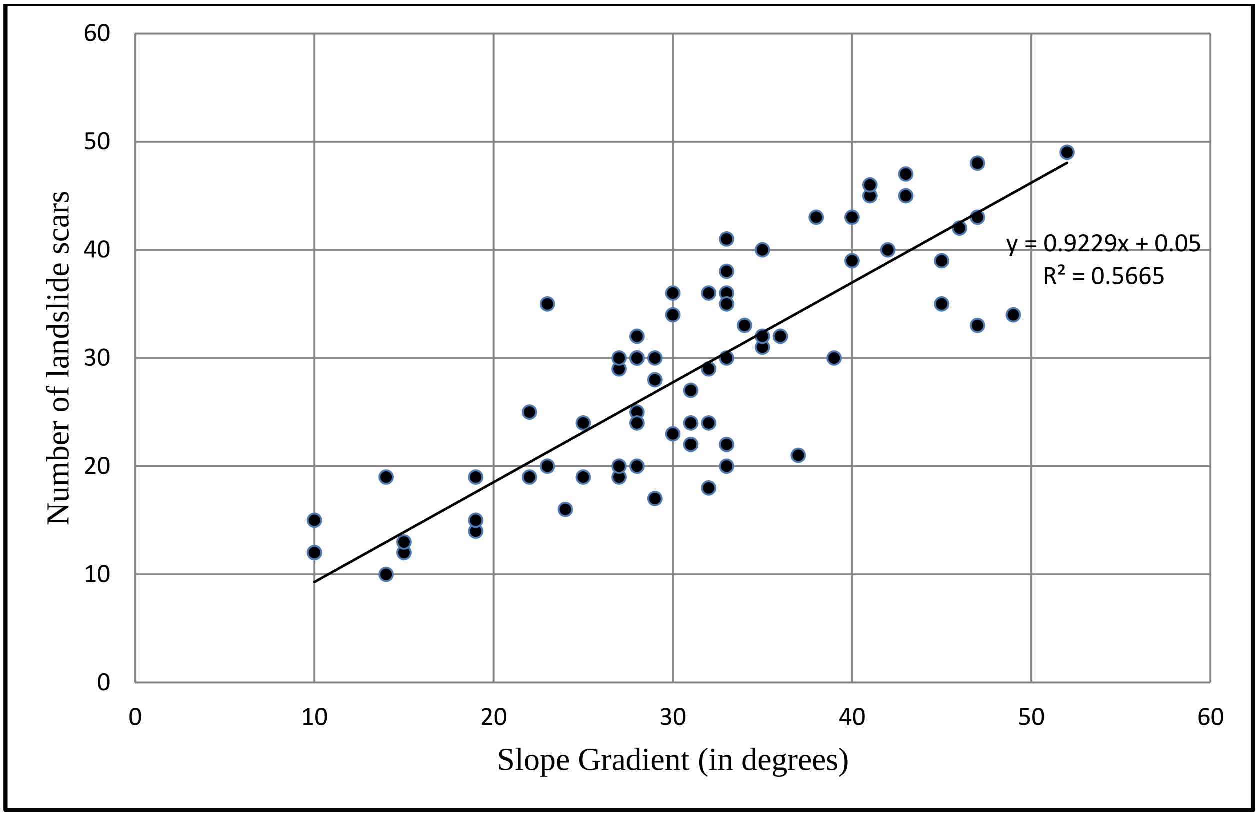

Fig. 3.5: Distribution of landslides with slope gradient

Landslide occurrence was found to be less pronounced within slope zones with slope gradients <15˚ and >45˚. Out of 65 landslide scars surveyed, 60 % occurred on zones with gradient between 25˚ and 35˚ (Table 3.1 and Fig. 3.5). Landslide occurrence increases from 4% on moderate slopes to 22% moderately steep slopes and then sharply increases to 60% on steep slopes, but fall sharply to 10% on very steep slopes and <4% on precipitous slopes (Figs. 3.4 and 3.5). This distribution is an indicator of the control that hillslope gradients exert on landslide occurrence. The correlation analysis show that the observed relationship between slope gradient and landslide occurrence is significant (R^2= 0.5665, P-value= 0.057).

3.3.2.2 Slope curvature and landslide occurrence

Profile curvature values computed for the catchment ranged from -3 to +5 (Fig. 3.6).

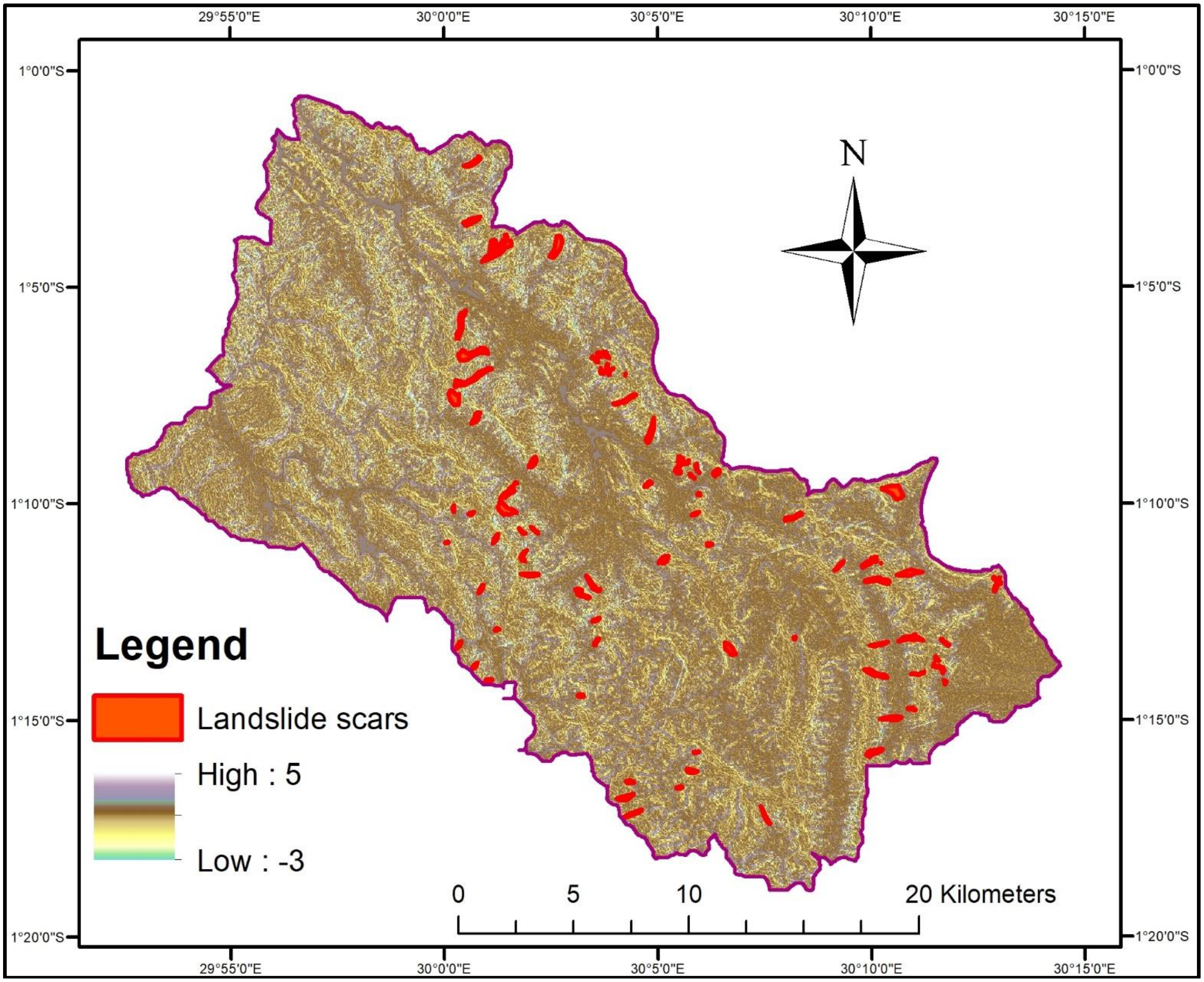

Fig. 3.6: A map of profile curvature and landslide distribution.

Landslide occurrence is dominant in zones where profile curvature values are between +0.1 to +5. Landslides are less pronounced in zones where profile curvature values are lower than 0 (Fig. 3.7). Out of the 65 landslide scars surveyed, 95% occurred in zones with profile concave curvature forms with values between 0.1 and 5 along topographic hollows. Only 5% of the landslides occurred in zones with profile curvature values between -0.2 and 0.1, which were conspicuously concave zones.

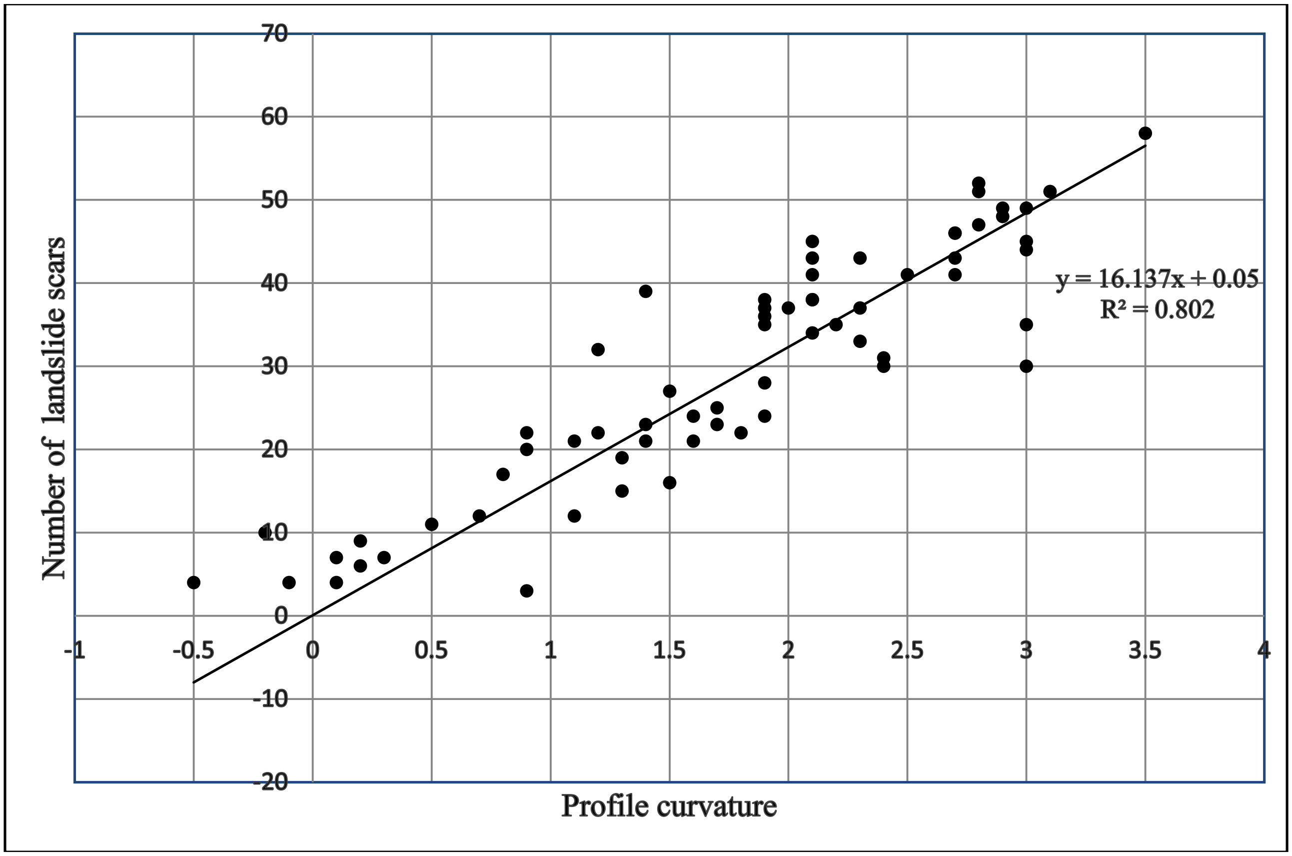

Fig. 3.7: Distribution of landslides with profile curvature.

Slope zones with profile convex curvature forms with values lower than -0.2, were mainly spur slopes and hilltops did not experience any landslide occurrence (Figs. 3.6 and 3.7). It was therefore noted that more than 95% of the landslides in the study area had occurred on profile concave forms in topographic hollows. A relationship between profile curvature with TWI and SPI is presented in the subsequent sub-sections. The correlation analysis show that the observed relationship between profile curvature and landslide occurrence is highly significant (R^2= 0.802, P-value= 0.088).

3.3.2.3 Topographic Wetness index and landslide occurrence

The Topographic Wetness Index (TWI) values computed for the highlands ranged between 2 and 24 (Fig. 3.8). Hydrologically active areas for the catchment were delineated using the TWI values.

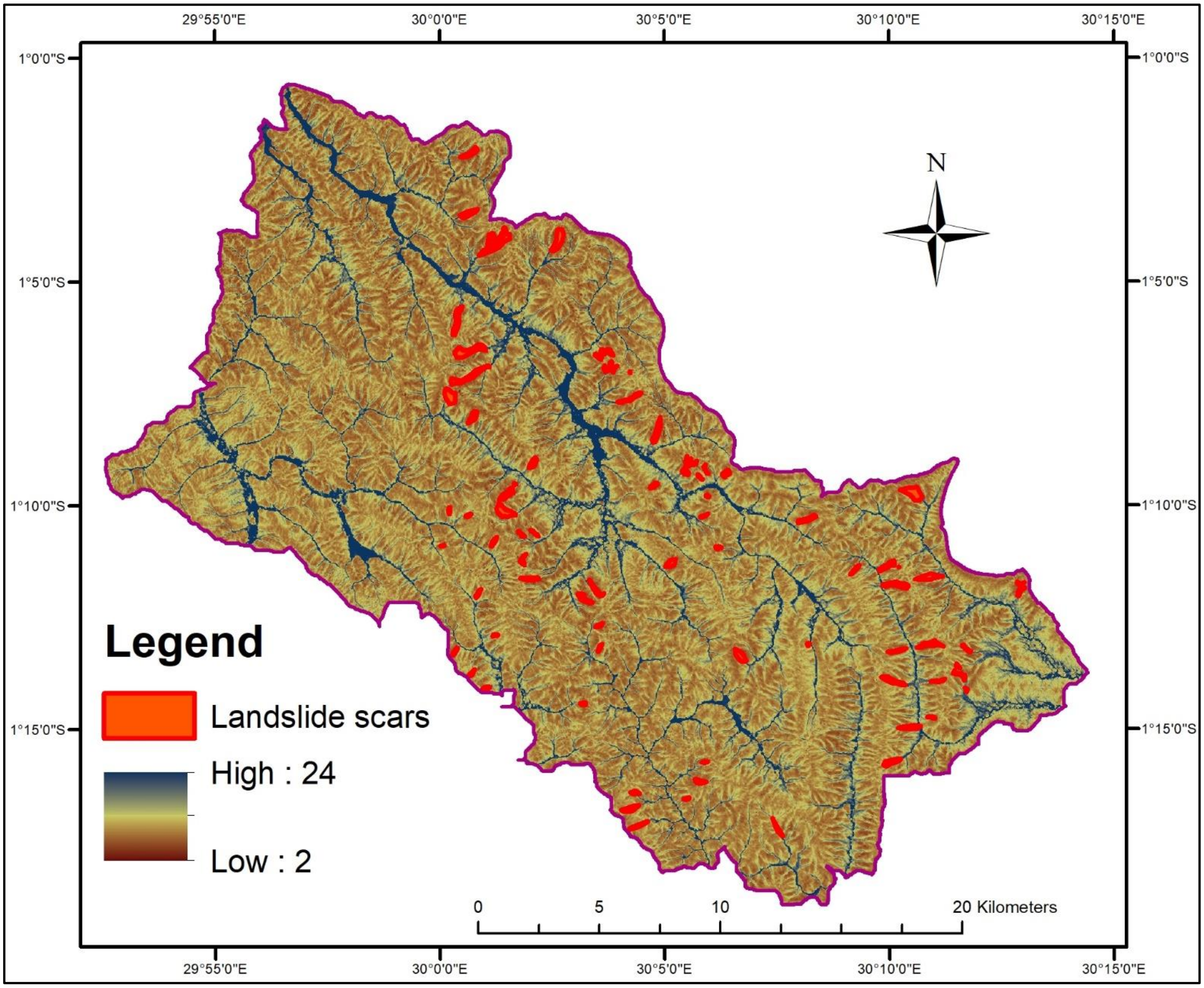

Fig. 3.8: A map of Topographic Wetness Index and landslide distribution

The TWI values were used to classify the landscape into three saturation zones or levels. The saturation zones classified using TWI included; low (< 8), high (between 8 and18), and very high (>18). Low TWI values were found on hilltops, spur slopes and steep slopes. High TWI values were mainly present along topographic hollows while very high TWI values were associated with valley bottoms (Fig. 3.8). High TWI values also correspond with profile concave curvature forms mainly along topographic hollows. Low TWI values correspond with profile convex forms dominant along hilltops and spur slopes. This relationship and its implications will be explored in the discussion section.

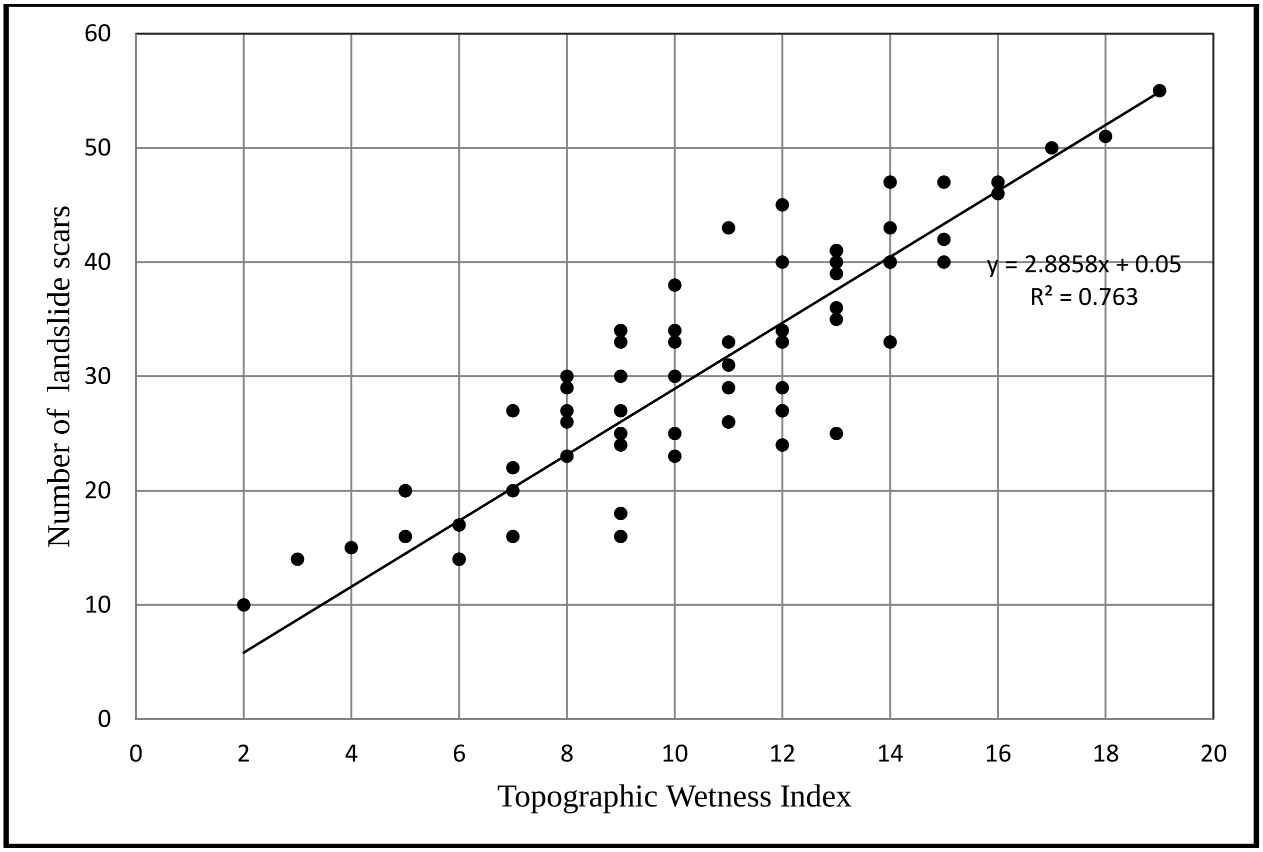

Fig. 3.9: Distribution of landslides with TWI.

Landslide occurrence was found to be dominant in zones with high TWI values ranging between 8 and 18 dominant along topographic hollows (Fig. 3.9). Zones with TWI <8 especially uppermost zones and spur slopes and >18 in the valley bottoms are not vulnerable to landslide occurrence. The valley bottoms are deposition zones for the landslide debris (Fig. 3.3). In the study area therefore, landslides significantly occur in zones with high TWI values, but reduce in zones with low and very high TWI values. The correlation analysis show that the observed relationship between TWI and landslide occurrence is highly significant (R^2= 0.763, P-value= 0.077).

3.3.2.4 Stream Power Index and landslide occurrence

The Stream Power Index (SPI) values computed for the highlands ranged between 0 and 43. Higher SPI values ranged from 10 to 43 and were along topographic hollows. Low SPI values (<10) were found along hilltops, side slopes and spur slopes (Fig. 3.10).

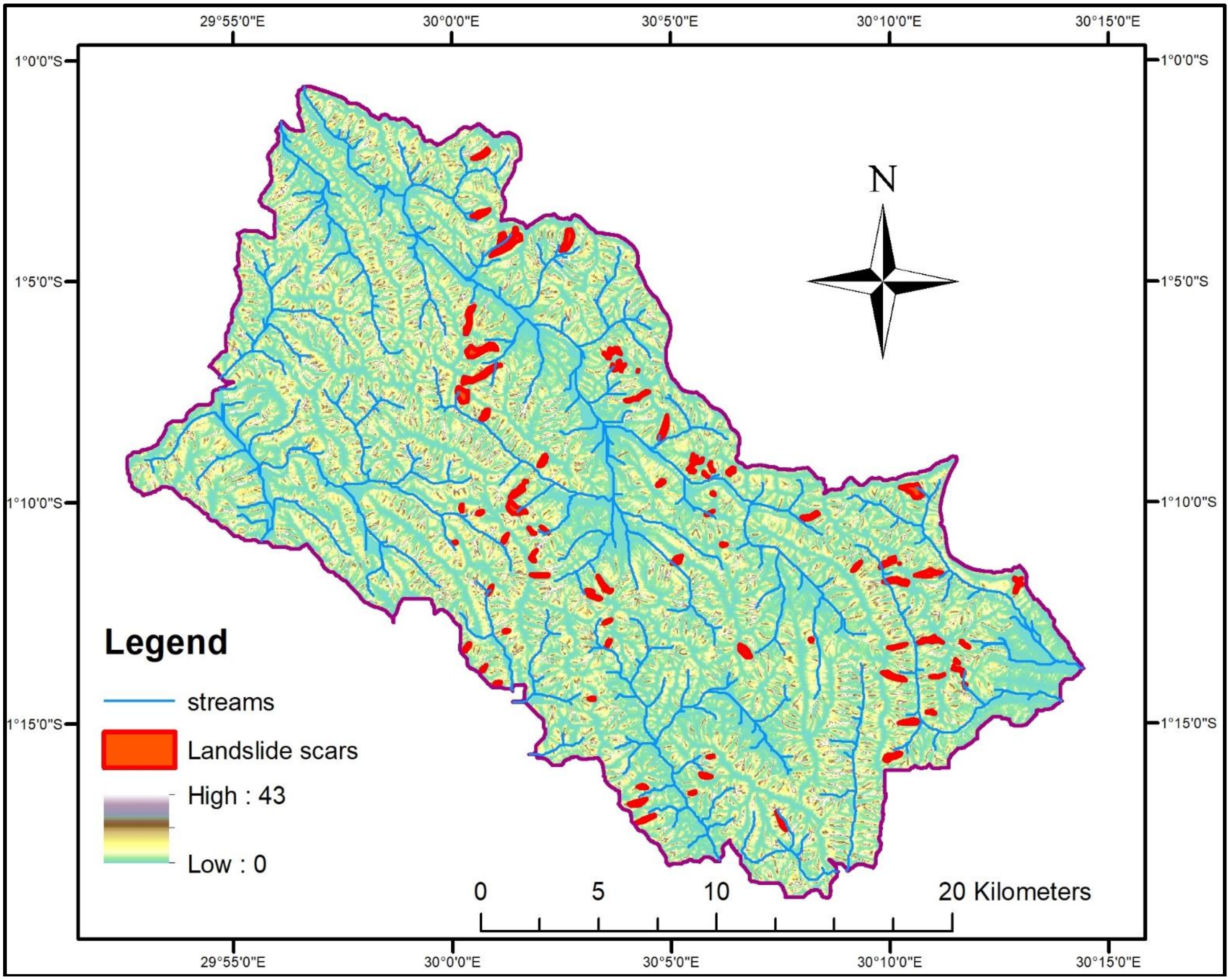

Fig. 3.10: A Stream Power Index map and landslide distribution

SPI values also corresponded with slope curvature of the landscape. Whereas high SPI values were mainly found along profile concave zones within topographic hollows, low SPI values were found along convex zones on ridge tops and spur slopes (Fig. 3.10). An overlay of SPI and landslide distribution maps indicated that landslide occurrence was dominant in zones where SPI values were >10 (Fig. 3.11). Such zones are mainly along topographic hollows.

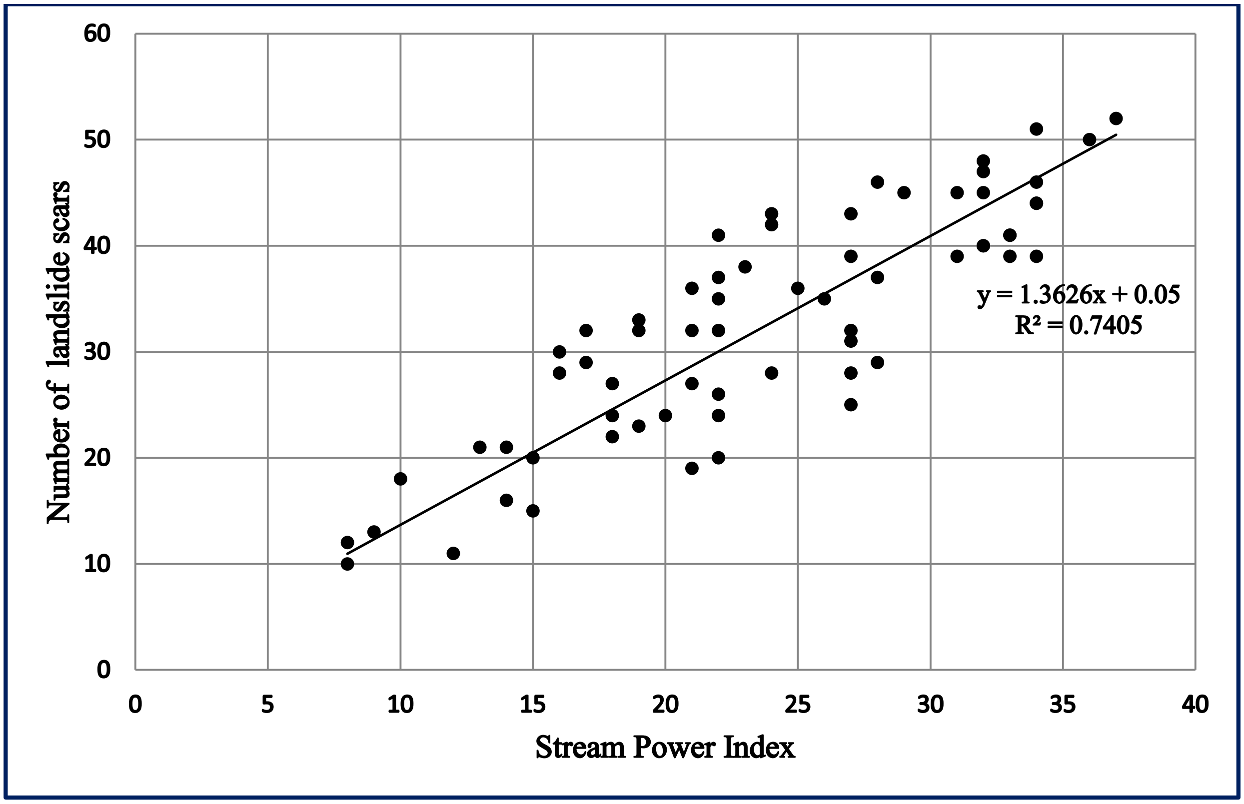

Fig. 3.11: Distribution of landslide features with SPI**.**

Slope sections with SPI values <10 (Fig. 3.11), mainly along hilltops and spur slopes had no landslide incidences. SPI and TWI were used to quantify flow intensity and accumulation potential. An overlay of SPI and TWI shows a relationship with landslide occurrence. High SPI (>10) and TWI (>8) values corresponded with profile concave forms mainly found along topographic hollows. These are also the zones with the highest incidence of landslide occurrence in the study area. Hilltops, uppermost and spur slopes have low SPI (<10) and TWI (<8) values and do not experience landslide occurrence. The correlation analysis show that the observed relationship between SPI and landslide occurrence is highly significant (R^2= 0.741, P-value= 0.0671).

3.3.2.5 Topographic Position Index and landslide occurrence

The Topographic Position Index (TPI) values computed for the highlands ranged between -7 and 5 (Fig. 3.12).

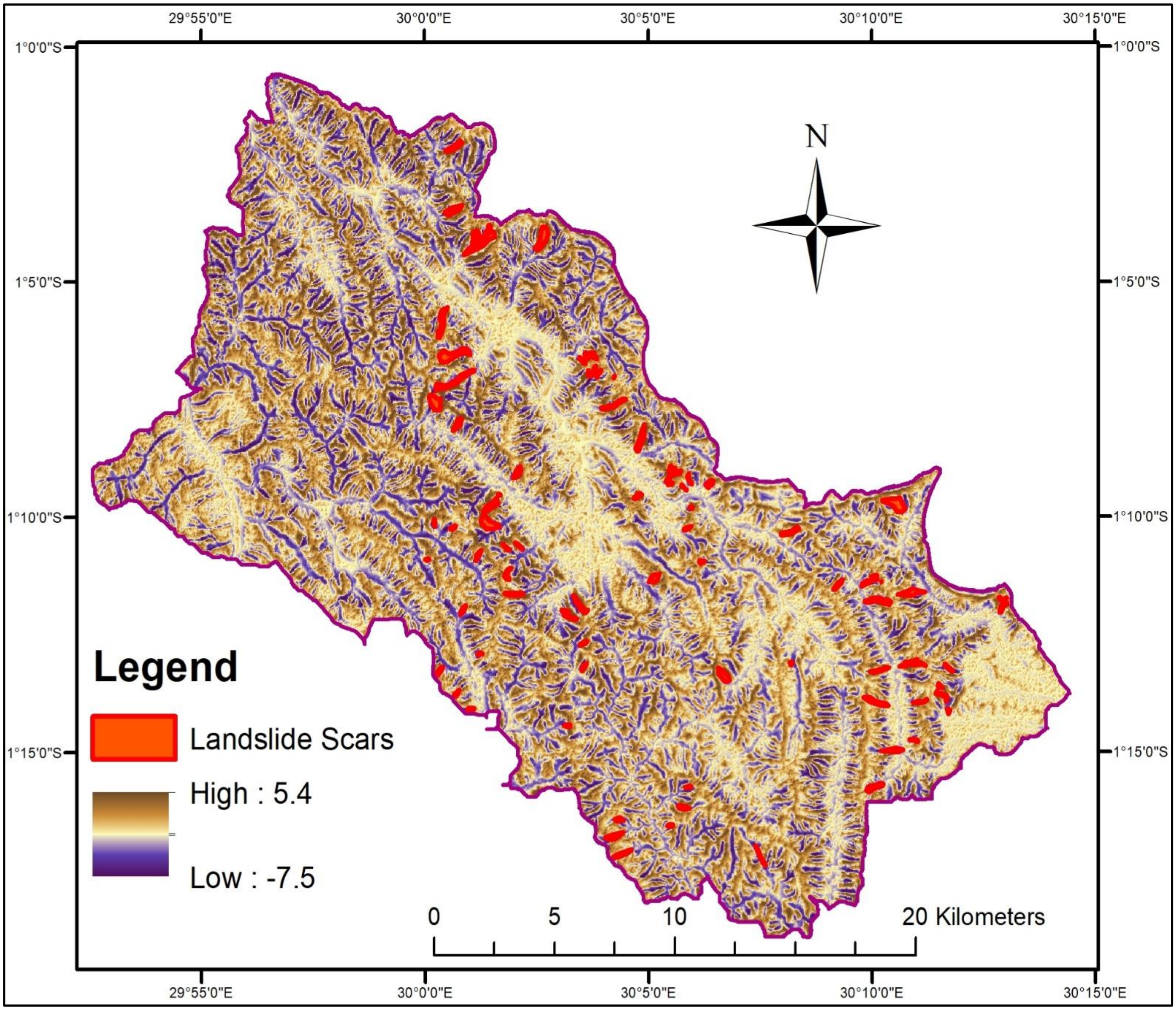

Fig. 3.12: A map of Topographic Position Index and landslide distribution

TPI values were used to compare the elevation of each cell in a DEM to the mean elevation of a specified neighbourhood around that cell. High TPI values (>+1) represented ridge tops, uppermost and spur slopes. Medium TPI values (between -2 and +1) represented middle slopes, while low TPI values (<-2) represented valleys bottom (Figs. 3.12 and 3.13). Positive TPI values also corresponded with profile convex zones, while negative TPI values mostly corresponded with profile concave forms. TPI was therefore used to determine whether a point is on a hilltop, in valley bottom or on an exposed ridge.

TPI values were used to classify the landscape into landform categories. The landform categories identified included; gentle valleys, open slopes, mesas, valley bottoms, hollows, hilltops, mid-slope ridges and spur slopes (Fig. 3.13).

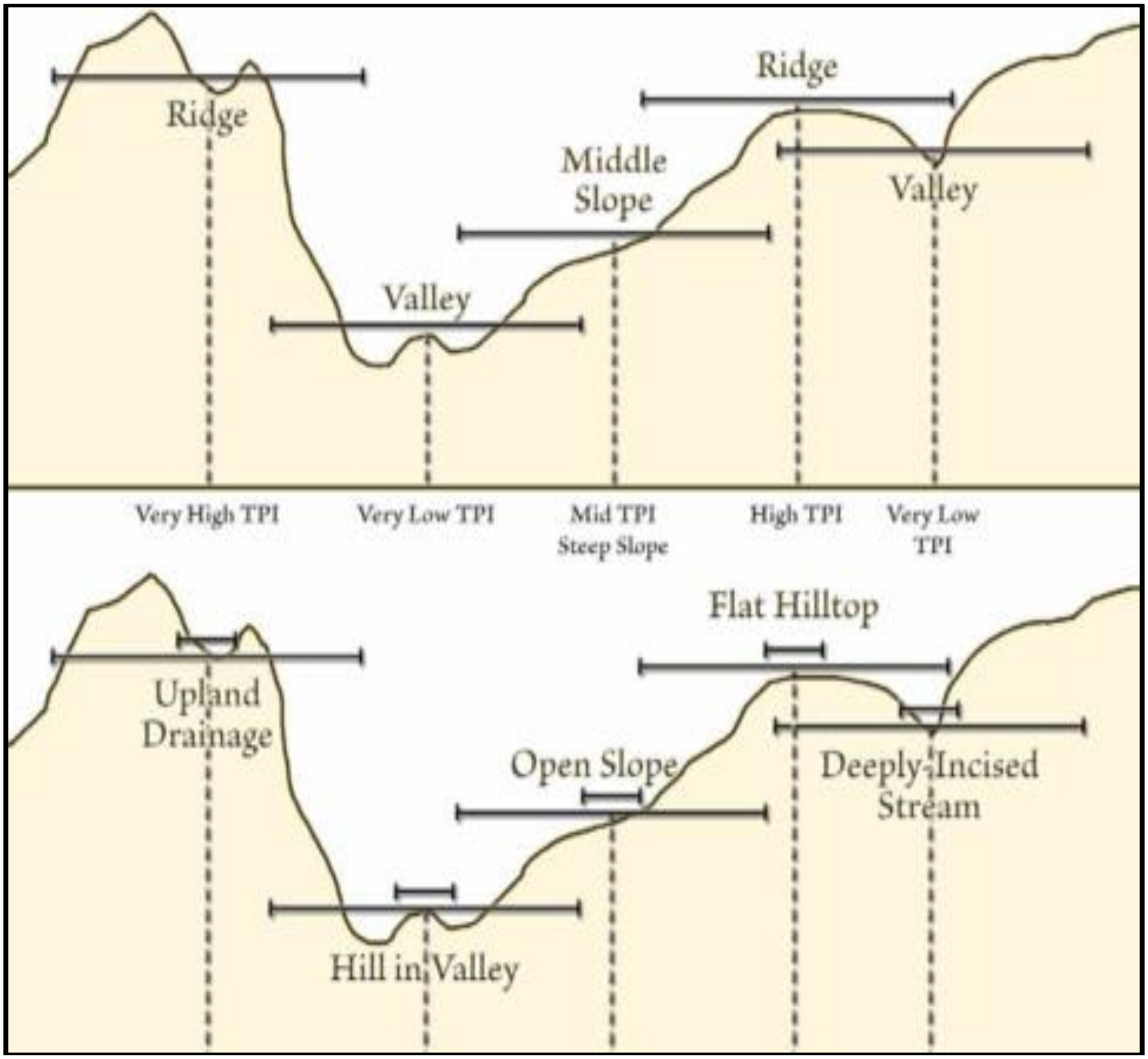

Fig. 3.13: Slope positions and landforms based on TPI values

The valley bottoms tended to have strongly negative TPI values (<-2) while topographic hollows corresponded with medium values (-2 to 1). The hilltops, mid-slope ridges, small hills in plains and high ridges all tended to have strongly positive TPI values (>1) (Fig. 3.12). TPI values were also used to classify the landscape into slope positions. The slope positions identified included lower slope positions, lower-middle, middle-middle, upper-middle and uppermost. Lower slope positions tended to have strongly negative TPI values while upper positions had strongly positive values. Slope position classification shows that 69% of the catchment is classified as middle-slope, 14% valley, 6% hilltop and 11% as flat surface.

Landslides are dominant in slope zones with TPI values between -1 and 1 mainly along topographic hollows. There was no landslide occurrence observed in zones with high TPI values (>1) mainly on hilltops, uppermost and spur slopes (Fig. 3.14).

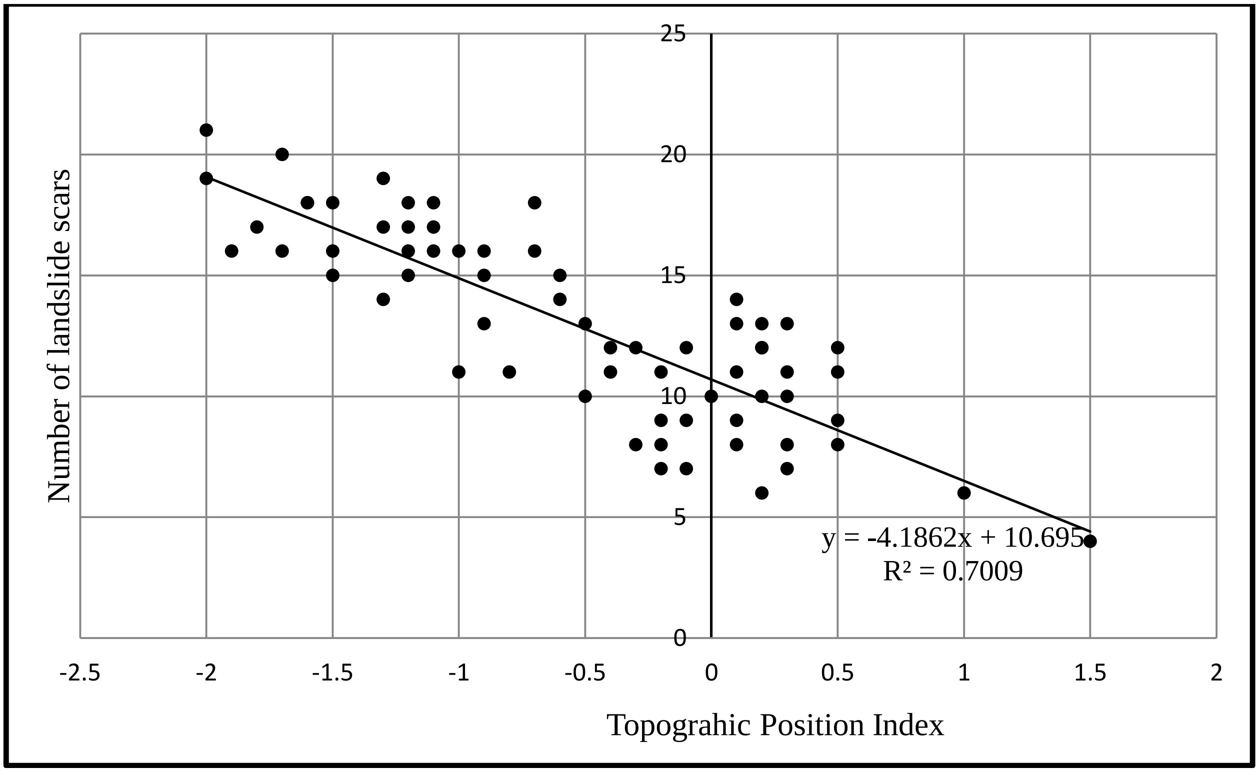

Fig. 3.14: Distribution of landslide scars with TPI**.**

The distribution of landslide occurrence in the study area exhibits a considerable variability with slope position. It was established that landslide concentration increases from 4% in the lower slope sections to 58% and 34% in the middle-middle and upper-middle elements respectively and then, declines rapidly to 4% in the uppermost sections. Thus, the most significantly affected slope position in these highlands is the middle slope section. The correlation analysis show that the observed relationship between TPI and landslide occurrence is significant (R^2= 0.7009, P-value= 0.065).

The different topographic parameters do interact to influence landslide occurrence. There is a convergence among these topographic parameters within the landscape to cause landslide occurrence. It was established that landslide occurrence is dominant in slope zones with slope gradient between 25˚ and 35˚, profile curvature between 0.1 and 5, TWI between 8 and 18, SPI >10 and TPI between -1 and 1. Landslide occurrence is less pronounced on slope zones with slope gradient <15˚ and >45˚, profile curvature <0, TWI <8 and >18, SPI <10 and TPI >1. This interaction among topographic parameters and landslide occurrence is explored in the discussion section. The coefficient of regression (R^2) and probability (P-value) reveal that whereas profile curvature is the most significant topographic parameter slope gradient is the least in influencing landslide occurrence in the study area (Table 3.2).

Table 3.2: Coefficient of Regression (R^2) and Probability (P-value)

Topographic parameter	Coefficient of regression (R^2)	Probability (P-value)
Slope gradient	0.567	0.057
Profile curvature	0.802	0.088
Topographic Wetness Index	0.763	0.077
Stream Power Index	0.741	0.067
Topographic Position Index	0.701	0.065

3.4 Discussion

3. 4.1 Landslide characteristics and distribution

The morphology of landslide scars in the study area denotes the presence of a simple or composite slide plane surface. The morphometric characteristics of the landslide scars diagnosed included the overall length, average width and depth, scar area as well as the volume of materials removed by each occurrence (Table 3.1). These dimensions are important in establishing the magnitude and type of the landslide occurrence. According to the landslide classification scheme of 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 movement is roughly rotational about an axis that is parallel to the ground surface and transverse across the slide. In the study area, rotational sliding mostly occurs on profile concave slopes and at a large distance from the water divide where runoff and subsurface water concentrates. In the study area shallow translational and rotational slides account for 85 % while debris flows, complex slides and rock fall make up less than 15 % of the landslides experienced.

The geometry of the slide depletion zones vary widely throughout the study area. This is similar to what has been observed by Knapen *et al.* (2006) on the foot slopes of Mt Elgon in Eastern Uganda. Whereas the volume of generated debris on Mt Elgon is dependent on the landslide depth (Knapen *et al.*, 2006), in the Kigezi highlands it is related to the length and width of the depletion zone. The depths of the landslide scars in the study area depends on soil profile characteristics especially soil depth as well as presence and location of clay pans (presented in Chapter 4 on soil properties). Whereas slope sections with profiles of less than 2.5m experience shallow slides, soil slips and debris flows, deep seated translation slides are experienced on slope elements with deep soil profiles of greater than 4m. More than 80% of the landslide scars were narrow and shallow because they occurred in small topographic hollows (discussed in the subsequent section on topographic parameters). Such small hollows gather limited materials from the surrounding hilltops and spur slopes and therefore have shallow soil profiles (Reneau and Dietrich, 1987; Grabs *et al.*, 2009; Raju and Nandagiri, 2015). Shallow landslides occur in such hollows due to limited materials available for downslope movement. About 10% of the analysed landslide scars were deep seated and wide. They occurred in deeper and wider topographic hollows. The wide hollows have a capacity to gather more materials from the surrounding spur slopes and hilltops and are associated with deep soil profiles (Gao and Maro, 2010). The occurrence in such hollows is deep seated due to availability of materials for undercutting. Whereas deep seated landslides had a depth greater than 2.5m, shallow landslides had a depth of less than 2m (Table 3.1).

The length of the landslide scars were observed to vary greatly across the landscape (Table 3.1). The length depends on the structure and size of the topographic hollow. Whereas longer topographic hollows experience elongated landslide processes of greater than 850m, shorter hollows are associated with small landslide processes of less than 12.5m. Therefore, the size and magnitude of the landslide processes depend on the characteristics of the hollows which determine soil material depth. The size of the hollow determines it ability to assemble moisture and soil from the surrounding spur slopes and hilltops (Reneau and Dietrich, 1987). Therefore there is a relationship between the landslide scar characteristics and the local topography. This relationship between landslides and topographic parameters is explored in the subsequent section.

In the study area, more than 75% of the landslides resulted from old failures of more than 3 years ago and are undergoing recovery. They are quite old, inactive and covered by vegetation, with a still visible semi-circular and concave head scarp and a deposit having an irregular morphology. Others are characterized by a complex activity, with a slow evolution and periods of acceleration. The landslide scars are being concealed by soil materials mobilized from the hilltops and spur slopes into topographic hollows where landslide occurrence is dominant. The soils accumulating within the landslide scars encourages rapid vegetation regeneration, owing to the high rainfall amounts in the study area (NEMA, 2012). These landslide scars are disappearing from the landscape due to the high rates of vegetation regeneration. Most of the shallow landslide scars are no longer visible on the landscape. It is only the deep seated landslides that are still visible premised to the fact that they require large quantities of materials to accumulate within them before they are completely concealed. The high rates of landslide scar recovery may lead to underestimation of landslide occurrence in the study area. About 25% of the analysed landslides have resulted from recent failures which occurred within the last 3 years and are still active. Often they suffer a reactivation, in the form of a shallow debris slide or a small debris flow in the head scarp or at the toe. This is due to remobilization of materials around the landslide scar which makes it active and susceptible to further failures.

3.4.2 Topographic parameters and landslide occurrence

The topographic characteristics of any region have a greater implication than any other parameter including soil and land cover on landslide occurrence (Fernandes e*t al.*, 2004; Gao and Maro, 2010; Loos and Elsenbeer, 2011). This is due to the fact that topography affects both hydrological processes and soil development (Selby, 1993). Field investigations revealed that most landslide occurrence in the study area is concentrated along topographic hollows than other slope elements. This relationship between different topographic parameters and landslide occurrence is explored in the subsequent sub-sections.

3.4.2.1 Slope gradient and landslide occurrence

Landslide occurrence is concentrated in sections with slope gradients ranging between 25˚ and 35˚. The lowest slope gradient where landslides occurred was 15˚ while the highest was 45˚. This distribution in landslide occurrence is due to the difference in soil depth and moisture. Soil depth is important because it represents the amount of materials potentially mobile and available for downslope movement. Moisture on the other hand leads to saturation resulting into reduced shear strength of materials. There is a close relationship between soil depth and slope gradient (Schmidt and Hewitt, 2004; Liang and Uchida, 2014). Soil depth reduces with increase in slope gradient and elevation (Sparks, 1986 and Selby, 1993). In the present study, it was established through soil profile analysis (presented in the subsequent chapter on soil properties) that sections with lower and moderately steep slope gradients had deeper soil profiles than those with very steep gradients.

According to Breugelmans (2003), soil depth and topography are important parameters for the return period of landslides. In the study area, slope sections with gradients ranging between 25˚ and 35˚ have deep well-developed soil profiles. The deep soils represent the requisite amount of material for downslope movement. The concentration of landslide occurrence on such slope sections is explained by the accumulation of displaced materials from uppermost very steep and precipitous slopes (Capitani *et al.*, 2013). The deep soil profiles in such slope sections also lead to high saturation rates as indicated by high TWI (>8) values. The high downslope force on saturated deep soils leads to landslide occurrence in such slope sections. 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 slopes with gradients <15˚ due to the less downslope force required to move materials (Nath *et al.*, 2013; Lopez-Davalillo *et al.*, 2014). Such slope sections are only depositional areas for the landslide debris.

Landslide occurrence was found to be less pronounced on slope zones with gradients greater than 45˚. Such slope elements which were classified as precipitous slopes have shallow skeletal soils and the percentage area coverage is relatively small. Erosion processes are more intensive on such slopes and hence soils are shallower. Precipitous slopes are generally stable because rapid erosion has removed the erodible soils from them. 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). More so, the rainfall intensity required to initiate landslides for weathered rocks is usually higher than that for soils because of their higher repose angle (Guan-Wei and Hongey, 2012). The thin soils in such sections also means limited materials for downslope movement and hence low landslide incidences (Nath *et al.*, 2013). Such slope sections are also moisture divergence zones with very low TWI values. They are stable with high angle of repose despite the steep slope gradients. Although there is a high downslope force on such zones, they remain relatively stable due to the absence of materials hence low landslide occurrence (Bagoora, 1989). This explains why landslide occurrence in Kigezi highlands is low on slope elements with very steep and precipitous gradients.

According to Lee *et al*. (2002), slope gradient is an important topographic parameter in slope stability analysis. Many studies have frequently used slope gradient (Lee *et al*., 2002; Lee and David, 2004; Lee and Talib, 2005; Sassa *et al*., 2005; Yalcin and Bulut, 2007) in preparing landslide susceptibility maps because it is directly related to landslides. Several authors report that the typical slope gradient for landslide occurrence is between 27˚ and 38˚ (Hutchinson, 1988; Hungr *et al.*, 2005; Hosseini *et al*., 2011; Lopez-Davalillo *et al.*, 2014). Likewise, the slope gradient of the source areas in the study area is between 15˚ and 45˚ but with a concentration between 25˚ and 35˚ which are relatively steep slopes. Landslide occurrence is concentrated on relatively steeper slopes due to the high downslope component which pulls materials (e.g., Bagoora, 1988; Appolinaire *et al.*, 2007; Claessens *et al.*, 2007; Guan-Wei and Hongey, 2012; Mugagga *et al.*, 2012; López-Davalillo *et al.*, 2014). It was revealed that landslide occurrence increases with increasing slope gradient but then reduce as the slope gradient increases further. As already indicated, this distribution is due to the variation in soil depth with slope gradient. Slope gradient also corresponds with profile curvature in influencing landslide occurrence in the study area. This relationship is explored in the subsequent sub-section. From the statistical analysis, it clear that slope gradient is the least significant topographic parameter in landslide occurrence in the study area as indicated by the coefficient of regression (R^2= 0.5665) and the P-values (0.057).

3.4.2.2 Slope curvature and landslide distribution

It was noted that landslides are concentrated in slope sections where profile curvature is concave with values ranging from 0.1 to 5. Landslides are less pronounced in sections where profile curvature is convex with values lower than 0 (Figs. 3.5 and 3.6). Profile concave forms are dominant along topographic hollows while convex forms appear on hilltops and spur slopes within the landscape. More than 95% landslide features surveyed occurred in topographic hollows with perfectly profile concave curvature forms. Less than 5% of landslides occurred in sections which were not perfectly profile concave forms. The concentration of landslides in topographic hollows with profile concave forms 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). During field investigations, it was revealed that there is an indication of water being mobilized from upslope and collecting in the hollows. The incoming water from upslope leads to rapid built-up pore water pressure in the soil mantle (Morgan, 1993). This results into reduced shear strength within hollows as will be discussed in the subsequent chapter on pedological parameters.

Topographic hollows have high TWI (>8) and SPI (>10) signifying high saturation rates and erosive power. The high saturation rates and confinement of flow results into reduced shear strength leading to slope failures. Profile convex forms however, have low TWI (<8) and SPI (<10), are water shedding sections and dry (Figs 3.8 and 3.10). This explains why profile convex forms in these highlands are not prone to landslide occurrence. It was also established through soil profile analysis (presented in the subsequent chapter on soil properties) that topographic hollows consist of deep soils. The thick soils are mobilized from immediate convex topography and have propensity to collect much infiltrated water during and after rainfall events. The moisture collecting in the hollows is not redundant, it induces weathering which avails more materials for downslope movement (Liang and Uchida, 2014). The availability of deep soils in hollows provides materials that are potentially mobile and susceptible to downslope movement (Fernandes *et al.*, 2004). Saturation also makes soil materials in hollows unstable and the result is downslope movements (Gao and Maro, 2010). Convex forms on the other hand, have shallow thin soils due to rapid erosion on them. They lack the soil materials for downslope movement and therefore, landslide occurrence on profile convex sections is low.

The findings are in line with what has been reported in literature that landslides often occur in areas of convergent topography in which, subsurface soil water flow paths give rise to excess pore-water pressures downslope (Corominas *et al.*, 1992; Montgomery *et al.*, 2002; Infascelli *et al.*, 2013). According to Gao and Maro (2010), areas of concave curvature tend to remain saturated between storms due to convergence of ground water flow. In concave forms, water flow is concentrated in topographic hollows (Grabs *et al*., 2009). This increases the moisture content of the soil and the amount of time soil remain saturated (Gu and Wylie, 2016). Curvature therefore, affects surface and subsurface hillslope hydrology and hence slope stability (Infascelli *et al.*, 2013; Lopez-Davalillo *et al.*, 2014). Topographic hollows 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. Several studies indicate that landslides are mostly confined to medium and steep slopes of topographic hollows (Reneau and Dietrich, 1987; Loos and Elsenbeer, 2011; Hosseini *et al.,* 2011; Buda, 2013). Fernandes *et al*. (2004) and Buckley *et al*. (2010) however, report that landslides may also occur on planar slopes. In the study area, topographic hollows in association with other topographic parameters to be explored in the subsequent sub-sections are prone to landslides. This is due to reduced shear strength resulting from the rapid build-up of pore water pressure in the soil mantle. From the statistical analysis, it was inferred that profile curvature is the most significant topographic parameter in landslide occurrence in the study area as indicated by the coefficient of regression (R^2= 0.802) and P-value (0.088). This is explained by the influence curvature has on the rest of the topographic parameters, as will be explored in the subsequent sections.

3.4.2.3 Topographic Wetness Index and landslide occurrence

Landslide occurrence is concentrated in sections with high TWI values ranging between 8 and 18. Such high TWI values are also found along topographic hollows with profile concave forms. Landslides were less pronounced in slope sections with TWI values <8 which have profile convex forms mainly hilltops and spur slopes (Fig. 3.10). As it has been explained in the previous sub-section, topographic hollows have high TWI due to the moisture converging in them from spurs and hilltops. Profile convex forms however, have low TWI as a result of divergence of water from them. The high TWI leads to saturation which reduces the shear strength of the materials leading to slope failures. The convex zones are however dry and stable due to the low saturation rates. The valley bottoms have the highest TWI values >18. This is due to greater amounts of water received from upslope areas which does not easily drain away leading to high saturation rates (Grabs *et al*., 2009). The high saturation, however, does not result into landslide occurrence. This is because such lower slope sections lack a gradient steep enough to initiate movement of materials. They are therefore deposition sections for the landslide debris.

According to Liesbet *et al*. (2015), topography is the driving force for water movement. At any particular point on the landscape, TWI is the ratio between the catchment area and the slope at that point (Wilson and Gallant, 2000). High TWI leads to saturation in the section. The high saturation rates along topographic hollows lead to instability of slope materials due to reduced cohesion (Reneau and Dietrich, 1987). The unstable materials can easily move downslope especially with a conducive gradient. Steep convex areas associated with spur slopes and hilltops are not vulnerable to landslide occurrence due to the low saturation rates (Ali *et al.*, 2014; Raju and Nandagiri, 2015). Materials in such sections are consolidated, have high cohesion since they are dry and therefore less susceptible to landslide occurrence. Studies by Chi *et al*. (2009) and Grabs *et al*. (2009), indicate that valley bottoms have the highest saturation rates. This is due to low slope gradient which does not allow water to easily drain away. The high saturation rates, however, do not result in landslide occurrence within the valley bottoms (Nath *et al.*, 2013). Likewise, it was established in the present study that the 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 a steep gradient to initiate the downslope movement of saturated materials. Topographic hollows are also sections with high SPI signifying greater erosive power. A combination of high TWI and SPI (to be explored in the subsequent sub-section) in topographic hollows explains the concentration of landslides in such slope sections.

3. 4.2 4 Stream Power Index and landslide occurrence

Landslide occurrence is dominant in sections with high SPI values >10 mainly along topographic hollows associated with drainage lines (Fig. 3.10). Slope sections with SPI values <10 especially hilltops and spur slopes had no landslide incidence (Fig. 3.11). There is a close relationship between SPI and TWI. SPI is an indicator of sediment transport capacity, while TWI is an alternative indicator of potential for landslide occurrence (Ferreira *et al*., 2015). The combination of both indices is important in predicting potential landslide zones. It was revealed that slope sections with high SPI (>10) also have high TWI (>8). Such sections correspond with profile concave forms mainly along topographic hollows within the landscape. As indicated in the previous sub-section, high TWI and SPI signify high saturation rates and erosive power dominant along topographic hollows. Such slope sections are associated with the highest incidence of landslide occurrence in the landscape. The results also indicated that slope sections with low SPI (<10) also have lower TWI (<8). Such sections correspond with profile convex forms (<0) mainly on hilltops and spur slopes. They have low saturation rates keeping them dry, with lower erosive power and therefore less incidence of landslide occurrence.

According to Moore *et al*. (1993b), SPI is a measure of the erosive power of water flow based on the assumption that discharge is proportional to specific catchment area. Merino-Martín *et al*. (2015), also indicate that SPI is directly related to both slope and catchment area. It is not therefore surprising that SPI was strongest where either the small neighbourhood or large neighbourhood values were negative. Negative topographic position index values (to be explored in the subsequent sub-section) indicate that a section is lower than its neighbours (Seif, 2014). Such sections have a larger catchment area than sections that are higher than their neighbours (Rousseau *et al.*, 2012; Ferreira *et al*., 2015; Gartner *et al.*, 2015a). SPI and slope erosion risk increases when the amount of water contributed by upslope areas and the velocity of water flow increase. This is due to increase in the gradient and the specific catchment's area (Freeman, 1991; Moore *et al.*, 1991). It has also been reported by Moore *et al*. (1993b) that the potential erosive power of overland flow depends on SPI. High SPI values in topographic hollows indicate high erosive power. Low SPI values on ridge tops and spur slopes indicate lower erosive power in such slope sections. Studies elsewhere also show that high SPI values indicate areas in the landscape that have a high potential for erosion during and after rainfall events (Ferguson, 2005; Gomi *et al.*, 2008; Buda, 2013). This explains why landslide occurrence is concentrated along topographic hollows which are associated with high SPI values. The ridge tops and spur slopes are not vulnerable to landslide occurrence due to low erosive power as demonstrated by lower SPI values. SPI is also closely related with TPI. This relationship is explored in the proceeding sub-section on TPI.

3. 4.2 5 Topographic position index and landslide occurrence

Landslides were identified as dominant on slope positions with TPI values between -1 and 1, which also correspond with topographic hollows. There was no landslide occurrence observed on slope positions with TPI values >1 especially on hilltops, uppermost and spur slopes (Fig. 3.12). This distribution shows a relationship between TPI and other topographic parameters. For example, as already indicated landslides are concentrated in sections with low TPI values along topographic hollows. These are also sections with high TWI and SPI. High TWI and SPI signify high saturation rates and erosive power as already indicated in the previous sub-sections. The combination of high TWI and SPI in such slope positions with profile concave forms leads to high incidences of landslide occurrence. Landslides 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. Following Jenness (2010), TPI values were also used to classify the landscape into morphological classes. Many physical processes acting on the landscape, including landslides are highly correlated with the topographic position of the landform. The landforms include hilltop, valley bottom, exposed ridge, hollows, flat plain, upper or lower slope (e.g., Tagil and Jenness, 2008; De Reu *et al*., 2013).

The most vulnerable slope positions have been identified as the lower-middle and upper-middle slopes, while the lower slopes and the uppermost sections are moderately affected (Fig. 3. 12). The lower slope sections are relatively stable in comparison with all other slope sections due to its low slope gradient. It is also noticeable that landslide concentration increases from 4% in the lower slope sections to 58% and 34% in the middle-middle and upper-middle elements respectively and then, declines rapidly to 4% in the uppermost elements (Fig. 3.14). Characterized by topographic hollows, the middle slope positions are convergence zones for eroded materials from hilltops and spur slopes (Briggs and Knapp, 2008; Mokarram *et al.*, 2015). Such slope positions are also convergence zones for moisture from the upper slopes. This leads to saturation which increases susceptibility to landslide occurrence within such slope positions.

Topographic Position Index (TPI) is capable of predicting areas susceptible to saturation and with high potential for overland flow (Seif, 2014). Previous studies indicate that locations exhibiting long low-angled hillslopes and low saturated hydraulic conductivity values are the most prone to the formation of variable saturation areas (Band, 1986; Grabs *et al.*, 2009; De Reu *et al*., 2013). It was established that valley bottoms and topographic hollows get saturated earlier than other sections due to high convergence. TPI was also used to subdivide the landscape into morphological classes based on topography. This is important for slope stability analysis aimed at identifying preferential zones for landslide occurrence (Weiss, 2001; Tagil and Jenness, 2008). In these highlands, topographic hollows are associated with high saturation rates, confinement of flow and deep soils. This leads to reduced shear strength and thus landslide occurrence (Selby, 1993). This explains why topographic hollows are the most affected landform classes by landslide occurrence. The ridge tops and uppermost slopes appear to be relatively stable due to thin soils resulting from high erosion on them (e.g., Gao and Maro, 2010; Liang and Uchida, 2014). They also have low TWI and SPI signifying low saturation and erosive power. This leads to stability of slope materials hence low landslide incidences. Therefore, TPI combines with other topographic parameters to cause landslides in the study area.

As earlier noted in the results section, all the topographic parameters analysed in the different sub-sections interact to induce landslides in the study area. It is established that there is a convergence zone within the landscape among all these topographic parameters to cause landslides. Profile curvature is the most significant topographic parameter influencing landslide occurrence in the study area. This is because it affects the flow of water and soil development, which in turn affects all the other topographic parameters. The study also established that slope gradient is the least significant topographic variable influencing landslide occurrence as revealed by the coefficient of regression (R^2) and probability (P-value) calculated.

Compared to other regions in the East African highlands, landslides in Kigezi highlands occur along clearly defined lines mainly the topographic hollows within the landscape. The landslide path can be easily identified and demarcated within the landscape. This is due to the fact that topographic hollows have moderately steep slopes, high TWI and SPI, low TPI as well as profile concave forms. Compared to other slope elements in the landscape, topographic hollows are conducive for landslide processes. On 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 (Knapen, 2003; Mugagga *et al*., 2012). This points to the differences in topographic parameters that underpin landslide occurrence in different mountain environments of the country.

3. 5 Conclusion

The topographic characteristics of Kigezi highlands of South Western Uganda have a major influence on landslide occurrence. This is due to its effect on hillslope hydrology and soil development. It was established that landslides predominantly occur along topographic hollows associated with profile concave forms, high TWI and SPI, low TPI and moderately steep slopes. 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 the materials. Combined with a steep gradient and deep soils, landslide susceptibility is enhanced. Hilltops, uppermost and spur slopes are, however, less susceptible to landslide occurrence. This is because such slope sections are associated with profile convex forms, low TWI and SPI, high TPI and very steep slopes. They are divergence zones for both soil and moisture. Such slope zones are dry, with thin soil profiles and are therefore relatively stable. An understanding of the interaction of these topographic parameters and their influence on landslide occurrence is important in landslide hazard mitigation. It would be possible to identify and predict potential landslide zones, and also demarcate safer zones for community activities. The knowledge generated about the area’s topographic characteristics and landslide occurrence will help mitigate community vulnerability and reduce disaster risk to landslide hazards in this fragile highland ecosystem.

CHAPTER FOUR

The influence of soil properties on landslide occurrence in Kigezi highlands of South Western Uganda.

Abstract

Analysis of soil properties is a prerequisite to understanding the spatial distribution of landslides in a given region. To evaluate the soil- landslide relationships, field investigations were undertaken and soil samples taken at different depths and points along slope profiles and positions. Onsite soil property analysis was conducted within the landslide scars, auger holes, and full profile representative sites. In order to measure soil-water infiltration rates, in situ infiltration tests were performed. A range of physical soil properties including porosity, dispersion, grain size distribution, Atterberg limits and shear strength tests were analysed in the laboratory. In order to determine clay mineralogical composition of the soil, XRD analysis was carried out. Results show that deep soil profiles ranging between 2.5 and 7 m are a major characteristic of the study area. Soils are characterized by clay pans at a depth ranging between 0.75 and 3 meters within the profiles. The study area is dominated by clay texture, except for the uppermost surface horizons, which are loamy sand. All surface horizons analysed had the percentage of sand, silt and clay ranging from 33 to 55%, 22 to 40% and 10 to 30% respectively. In the deeper horizons, sand was observed to reduce drastically to less than 23%, while clay increased to greater than 50%. The clay content is very high in most of the tested samples, exceeding 35% especially in the sub soil, implying the vertic nature of the soils, which is a very important factor in the occurrence of landslides. In terms of clay mineralogy, the upper soil layers predominantly contain quartz, while subsurface horizons have considerable amounts of illite/muscovite as the dominant clay minerals, ranging from 43% to 47 %. The average liquid limit was 58.43%, while the average plasticity index was 33.3%. The average computed weighted plasticity index was (PI_w) 28.4% and average expansiveness (ε_{ex}) was 38.6. All the soil samples tested were in the CH group and were classified as inorganic clay of high plasticity. Most of the samples have dispersion values greater than 30%. The average soil cohesion (C) was 8.2 kPa while angle of internal friction 5.4° signifying weak soils. Infiltration rates in the catchment are generally rapid to very rapid, varying between 12.2 and 88.5 cm h^{-1}.

Such soil characteristics have a great implication on landslide occurrence especially the timing and the nature of the processes. Landslides are not normally experienced either during or immediately after rainfall events but occur later in the rainfall season. This time lag in landslide occurrence and rainfall distribution, is due to the initial infiltration through quartz dominated upper soil layers, before illite/muscovite clays in the lower soil horizons get saturated. An understanding of pedological characteristics is an important step in landslide hazard mitigation in Kigezi highlands. Potential landslide zones can be identified based on the distribution of weak clay soil materials which are vulnerable to slope failure.

Key Words: soil properties, landslides, Kigezi highlands

4.1 Introduction

Landslides are among the most devastating natural disasters in the world occurring almost across all terrains with steep slopes (Guzzetti *et al*., 1999; Dai *et al*., 2002; Schuster and Wieczorek, 2002; Das *et al*., 2010; Mugagga *et al*., 2011; Kirschbaum and Zhou, 2015). The proximate and underlying causes of landslides are widely covered in literature (Ngecu and Mathu 1999; Muwanga *et al.,* 2001; Breugelmans, 2003; Knapen, 2003; Glade and Crozier, 2004; Kitutu *et al.,* 2004; Knapen *et al.,* 2006; Mugagga *et al.,* 2011). No individual factor can be singled out as the cause of landslides; a number of conditions usually interact to make rock or soil susceptible to land sliding (Selby, 1993). Several studies observe that the susceptibility of a slope to failure is dependent on many factors, including the topographic, geotechnical properties of material involved and the presence of discontinuities (Guzzetti *et al*., 2006b; Lopez-Davalillo *et al*., 2014). Most of these variables are specific to a particular area and therefore site-specific studies are important (Lee and Talib, 2005; Gonghui *et al.,* 2010; Kirschbaum and Zhou, 2015).

Several authors reveal that landslides are triggered under multiple geomorphological and geological conditions and therefore identifying the factors controlling landslide occurrence and distribution is difficult (Zhou *et al*., 2002 Guzzetti *et al*., 2005; Van Den Eeckhaut *et al*., 2013). Landslides are always associated with disturbance of the equilibrium relationship that exists between stress and strength in material resting on a slope. Instability arises when the shear strength is exceeded by downslope stress (Montgomery and Dietrich, 1994; Petley, 2008; Gemitzi *et al*., 2010; Regmi *et al*., 2010; Keller and Dexter, 2012). High clay content in the soil, slope curvature, steep gradients, land use and high rainfall amounts are considered influential conditions for landslide occurrence in East African highlands (Kitutu *et al.,* 2004; Knapen *et al.,* 2006; Kitutu *et al*., 2009; Mugagga *et al.,* 2011).

According to Hong *et al*.(2007), soil types and soil texture are primary-level parameters, while elevation, land cover types, and drainage density are secondary in terms of landslide inducement. As demonstrated by Yalcin (2007) and Jadda *et al.* (2009), fine textured clayey soils have small pores and release water gradually. Such properties make soils prone to landslides because of the high-water retention. Vagen (2010), observe that the low permeability of fine textured clayey soils exacerbates the vulnerability to landslides. This is due to increased saturation and pore water pressure which reduces soil shear strength leading to landslides (Mugagga, 2011). Kitutu *et al*. (2009), also observe that landslide occurrence is common in areas where sandy clay loams are underlain by sandy clay soils. This is because sandy clay loams are lighter and promote faster flow of water into the subsurface soil layers richer in clay and stagnates water flow through the soil. This leads to water accumulation which increases hydrostatic pressure, resulting in landslide occurrence. It has been reported by Sidle *et al*. (1985) that slope stability is influenced by the distribution of pores within the soil matrix and particle size. Several authors indicate that the stability of any slope is affected by specific soil parameters including bulk density, shear strength and cohesiveness (Sidle *et al*., 1985; Zezere *et al*., 1999; Kitutu *et al*., 2004; Kitutu *et al.*, 2009; Zung *et al*., 2009). To characterize the behaviour and the problem nature of soils, similar properties have been used by a number of authors (Bell and Walker, 2000; Schuster and Highland, 2001; Bell and Culshaw, 2001; Boehner, 2002; Bell, 2004; Mugagga *et al*., 2011).

Studies by Knapen *et al. (*2006), Yalcin (2007), Kitutu *et al. (*2009) and Mugagga *et al*. (2011) have also revealed that landslide occurrence is mainly as a result of high clay content in the soil materials. Several studies further confirm that landslide occurrence is linked to accumulation of clay in relict joints as well as specific clay minerals (Breugelmans, 2003; Fauziah *et al.,* 2006; Baron *et al.,* 2011). The susceptibility and stability of slope materials is influenced by the presence of swelling clay minerals (Morgan, 1993; Selby, 1993; Iganga *et al*., 2001; Zung *et al*., 2008; Zung *et al.,* 2009; Zinck, 2013). Soils with high clay content are considered as the leading cause of landslides and other mass movement process in most East African highlands (Kitutu *et al.,* 2009). Soil materials with clay content exceeding 30%, classified as vertic soils (Mukasa-Tebandeke *et al.,* 2015) are poorly drained. Such soils have swell-shrink characteristics with change in water content and are highly vulnerable to slope failure (Fauziah *et al.,* 2006*;* Baynes, 2008; Mugagga *et al*., 2011; Yalcin, 2011).

Li *et al.* (2006) state that clay mineralogy affects the shear and frictional resistance of the soils. Among the clay minerals, smectite, particularly montmorillonite and illite, decrease the soil residual strength, particularly due to their peculiar colloid-chemical characteristic (Mukasa-Tebandeke *et al.,* 2016) and contribute to landslide occurrence (Inganga *et al*., 2001). According to Bell (2004), montmorillonite as a swelling clay mineral has a negative behaviour, very strong attraction for water and can induce soil collapse due to susceptibility to volume change. The dispersive, collapsible and expansive nature of clay soils constitutes problem soils (Bell and Culshaw, 2001). William *et al.* (1985), note that problem soils are widespread around the world, notwithstanding the little attention they receive in many landslide studies. Some studies reveal that at various moisture contents, landslides may be induced by problem soils due to their distinct shrink-swell properties (Bell, 2004; Yilmaz and Karacan, 2002; Yalcin, 2007; Zung *et al.,* 2009; Yalcin, 2011).

According to Zinck (2013), soil physical properties in the unsaturated zone affect a large variety of processes determining the occurrence of shallow landslides. Unsaturated soil hydrology has been identified as a key factor for shallow landslide initiation and dynamics (Li *et al*., 2006; Kitutu *et al*., 2009; Li *et al*., 2010). According to Inganga *et al*. (2001) and Gonghui *et al*. (2010), the most important cause of shallow landslides is the decrease of matric suction after a rainstorm and the development of positive pore pressures above the water table. In particular, soil shear strength decreases non-linearly with increasing soil matric suction (Selby, 1993). The suction becomes less negative as the soil approaches saturation and therefore becomes more susceptible to failure (Kitutu *et al*., 2009). Such soil properties vary significantly in space and require site-specific investigations to understand their contribution to landslide occurrence. Area specific information on soil properties including soil depth, angle of internal friction, clay mineralogy, particle size distribution, cohesion and bulk density is required to determine landslide susceptibility (Gonghui *et al*., 2010; Mugagga *et al.*,2011; Roller *et al.*, 2012). Whereas previous landslide studies in Uganda have focused on the volcanic soils of the Mount Elgon Region in Eastern Uganda (e.g., Inganga and Ucakuwun, 2001; Muwanga *et al.*, 2001; Knapen, 2003; Kitutu *et al.*, 2004; Knapen *et al.*, 2006; Kitutu *et al.*, 2009; Kitutu *et al.*, 2011; Mugagga *et al.,* 2011), the present study examines the influence of the non-volcanic soils on landslide occurrence in Kigezi highlands, south western Uganda. This will provide a comparative understanding of the influence of soil properties on landslide occurrence in both mountain and highland environments of the country.

4.2 Materials and methods

4.2.1 Soil morphological analysis

To evaluate the soil-landslide relationships, field investigations were undertaken and soil samples taken at different depths and points along the slope profile as well as slope positions. Soil description was done according to FAO guidelines for soil descriptions (FAO, 1990 and 2006). In order to determine soil sampling sites, landslide scars were categorized in groups based on their morphological characteristics including depth, width, length, geology and structure. After analysing the profiles, the characteristics and similarities of soils in the landslide scars were identified. Landslide scars with similar visible morphological and soil characteristics were grouped together. After grouping the landslides, representative scars were sampled in each of the groups for soil analysis. It is therefore the landslide sites that determined the sample sites for soil analysis. The 65 visible landslide scars identified in the study area were categorized into 10 groups based on their characteristics. Only 10 landslide scars representing each of the 10 groups were sampled for analysis of soil profiles and particle size distribution. Four soil profiles were exposed on each landslide site in the valley bottoms, lower-middle, upper-middle and uppermost slope elements. In total, 40 soil profiles were used to characterize the in-situ soil properties. Three soil samples were collected from each of the profiles in the A, B and C horizons. In total, 120 soil samples were collected and used in the analysis of soil morphological characteristics (Table 4.2).

Onsite soil property analysis was conducted within the landslide scars, auger holes, and full profile representative sites. Profiles in the upper slope sections were dug up to a depth of 1 to 1.5m while those in the middle slopes ranged between 2 to 4m and the lower slope soil profiles were greater than 5m. Soil depth was classified into three broad groups of shallow, medium and deep. Soil samples were taken at a landslide site level within and outside the landside scars. At each landslide scar, were dug from the summit to the valley bottom to evaluate the pedological variations along soil profiles. The profiles were exposed for determination of soil textural and structural properties. Onsite soil property analysis was done within the mini-pits, auger holes, and full profile representative sites (Fig. 4.1). Soil horizons were analysed in detail with a specific focus on depth, colour, texture, presence and location of clay pans, and structure. Soil colour characteristics for each horizon were obtained using the Munsell soil colour chart. Site analyses sought to characterise among other things, soil matrices and bedrock within each landform.

4.2.2 Soil-water infiltration

Soil infiltration properties were analysed to gain an understanding of the hydrological characteristics of slope materials in the study area. According to Bagoora (1997), increased rainfall is the main factor triggering landslides in Kigezi highlands due to increased pore water pressure in voids. In order to understand the occurrence of landslides in Kigezi highlands, characterization of infiltration is of utmost importance. Quantification of infiltration rates is important to understand the mechanism of slope failure at hillslope and watershed scales. Infiltration rate is the amount of water which penetrates the soil per surface area and unit time (Bamutaze *et al*., 2010). Measuring infiltration rates involves a number of methods (e.g. ASTM D3385, 2003; Ahmed *et al*., 2011), including laboratory analysis as well as in situ measurements (Stolte, 2003; Wenck Associates, 2008; Rousseau *et al*., 2012).

In order to measure water infiltration rates of the soil, in situ infiltration tests were performed. Due to variations between soil characteristics with slope position and topographic configurations, infiltration tests were performed at 4 slope positions namely valley bottom, lower middle, upper middle and uppermost, as well as along spurs and topographic hollows. At each landslide site, eight infiltration tests were conducted, four along the topographic hollow and four on the spur slopes totalling to 32 experiments on the 4 landslide sites. Land use and cover categories considered in infiltration experiments included natural forests, shrub and thickets, grassland and bush land, fallow land, tree plantations, bananas, potatoes, sorghum, vegetables, beans, and intercropped gardens.

Field measurements of soil water infiltration were done using the double ring infiltrometer method (ASTM D3385-03, 2003), which consists of two concentric metal rings (Wenck Associates, 2008; Bamutaze *et al.*, 2010; Philips and Kitch, 2011; Ahmed *et al*., 2011). The purpose of the outer ring was to reduce edge effects and produce one-dimensional, vertical water flow in the inner ring. The rings were driven into the ground and filled with water, which was poured into both the outer and inner rings. Following Walsh and McDonnell (2012), water in both rings was kept at approximately the same level in order to promote vertical water movement and prevent lateral flow. A ruler was firmly placed vertically into the inner ring and water was poured continuously in both the inner and outer rings (Fig. 4.1). The time taken to reach steady state water infiltration rate at all sites varied between 1 and 2 hours. The drop-in water level or volume in the inner ring was used to calculate an infiltration rate.

Fig. 4.1: Field analysis of soil properties in the mini-pits, profiles and infiltration tests.

Given the importance of pore water pressure and antecedent moisture in landslide occurrence in the study area (Bagoora 1997 and NEMA 2010), rainfall data are important in the analysis of landslide occurrence (Bagoora, 1998). Rainfall data for the region were obtained from Kabale Meteorology Station, weather data 2015: WMO No. 63726, National No. 91290000, station name Kabale, at Elevation 1867m, Latitude 01˚ 15', Longitude 29˚ 59', and Kabale District local government environmental reports (2008, 2012 and 2015). The data obtained included the mean annual precipitation, the average number of rainy days, monthly and seasonal rainfall amounts and distribution, the precise or approximate period of the failures. Rainfall data were compared with landslide occurrence periods to ascertain the influence of rainfall distribution and amounts on landslide distribution.

4.2.3 Laboratory analyses

Soil samples for laboratory examination were taken vertically for each 0.4m in colluvium material. The samples consisted of undisturbed blocks and disturbed bag samples, all obtained from shallow trench pits. Scope of the laboratory testing comprised shear box, Atterberg limits, sieve and hydrometer analysis, specific gravity using test standards shown in Table 4.1 and XRD analysis. These tests were used in the estimation of porosity, dispersion, saturated specific weight, and saturation humidity. These properties are considered relevant in understanding and characterizing site-specific landslide occurrence (Van Der Merwe, 1976; Sidle *et al.*, 1985; Bell and Maud, 1994; Zezere *et al.*, 1999; Bell and Walker, 2000; Bell and Culshaw, 2001; Kitutu *et al.*, 2004; Kitutu *et al.*, 2009; Zung *et al*., 2009; Mugagga *et al*., 2011). A series of laboratory tests were conducted in order to define the main physical and mechanical characteristics of the soil. The tests performed included the natural water content, unit weight, degree of saturation, grain size distribution, Atterberg limits (liquid, plastic and shrinkage limits) and strength tests. The tests were performed at the Soil, Plant and Water Analytical Laboratory, Department of Agricultural and Environmental Sciences, Makerere University and TECLAB Limited (The Engineers laboratory) Nalukolongo, Kampala Uganda, in accordance with the British Standard BS 1377 (British Standards Institution 1990) procedures listed in Table 4.1.

Table 4.1: Tests and standards used on disturbed samples

Test Description	**Test Standard**
Particle size distribution	BS 1377: Part 2, Sub cl. 9.2: 1990
Liquid Limit	BS 1377: Part 2, Clause 4:1990
Plastic Limit and Plasticity Index	BS 1377: Part 2, Clause 5:1990
Linear Shrinkage	BS 1377: Part 2, Clause 6:1990
Shear Box Test	BS 1377: Part 7, :1990
Double Hydrometer	ASTM D4221 – 11

Mechanical analysis was used in determining the size of particles present in a soil, expressed as a percentage of the total dry weight. Two methods were used to determine the particle- size distribution of the soil; (1) *Sieve analysis* used for particle sizes larger than 0.075 mm in diameter (gravel and sand size particles), and (2) *Hydrometer analysis* used for particle sizes smaller than 0.075 mm in diameter (silt and clay size particles). Sieve analysis involved shaking of the soil sample through a set of sieves (ranging from 75.00mm to 0.075 mm). The percentage passed of the finest size and the percentage weights retained on each of a series of standard sieves of decreasing size were used to infer particle size distribution.

Atterberg Limits

Atterberg limits are a standard measure of the consistency of fine-grained soils depending on its moisture content (Kitutu *et al.,* 2009). These limits were used to determine soil plasticity, which provides a clue to the type of mass movement that would characterise a given area (e.g. Moussadek *et al.,* 2017). The susceptibility of slopes to landslide processes is explained by the problem nature of the soil, particularly its expansion properties at different moisture and clay content which is determined by Atterberg Limits (Fauziah *et al*., 2006; Isik and Keskin, 2008 and Mugagga *et al*., 2011). In order to carry out the Atterberg limit tests, 40 soil samples were collected from 10 representative landslide sites. On each of the representative landslide site, 4 samples were collected from the valley bottoms to the uppermost slope elements. Atterberg limit tests were carried out in the geotechnical laboratory on a number of soil samples obtained from the landslide sites. They were conducted to describe the plasticity of clay and to measure the critical water contents of a fine-grained soil. The tests gave an indication of the levels of saturation and response of a soil material to landslide occurrence. Atterberg limits provided useful information regarding soil strength, behaviour, stability, type, and state of consolidation. The limits were determined as the Liquid Limit (LL) and the Plastic Limit (PL) using the procedures in ASTM Standard D 4318 and D431/84 and also CNR UNI 100141 (e.g., Gao and Maro, 2010). The Plastic Limit (PL) and Liquid Limit (LL) for each sample were determined using the drop cone penetrometer method at the soil mechanics laboratory. The results of grain-size distribution and Atterberg limits tests were used to classify the colluvium soils according to the Unified Soil Classification System which enabled further classification of the fine material.

Soil dispersion tests

Dispersive soils are those which behave as unconsolidated single grained fine particles rather than as a cohesive mass. They have almost no resistance to mass movement and have low shear strength (Fauziah *et al*., 2006; Zinck, 2013). This renders them highly dispersive and therefore susceptible to landslides (Zung *et al*., 2008). Four samples were collected from each of the 10 representative landslide sites, giving a total of 40 samples used for soil dispersion testes. A double hydrometer test was used based on Stoke's law of settling velocity as an indicative laboratory test for identification of dispersive soils using the following procedure;

(a) Determination of the % passing 0.005mm with standard hydrometer testing using a chemical dispersing agent,
(b) Carrying out a separate hydrometer analysis using no dispersing agent and determining the percentage passing 0.005mm,
(c) The percentage dispersion was defined as; Dispersion = [(% passing 0.005mm from b)/ (% passing 0.005mm from a)] x100. Values greater than 30% were considered significant while values greater than 60% were considered critical.

Soil expansiveness

To determine whether the soils in the study area are susceptible to slope failure, investigations were done to determine the expansiveness of the soil. Expansive soils are those that exhibit particularly large volumetric changes (swell and shrinkage) following variations in their moisture content (Morgan, 1993; Selby, 1993; Bagoora, 1997). Three procedures were done to determine the expansiveness of the soil in the study area. The first procedure involved carrying out field surveys. During such field surveys, particular attention was paid to the swelling-shrinking characteristics of particularly clay soils (slickensides and shrinkage cracks), which play an important role in predisposing slope to failure. A description of the crack patterns on the soil surface was done at different sites. The second procedure entailed calculating the weighted Plasticity Index (PI_w) on the fraction <425µm and weighted for the sample's actual content of particles <425 µm. As defined by Isik and Keskin (2008), plasticity is the ability of a soil to undergo unrecoverable deformation at constant volume without cracking or crumbling. PI_w was determined using the formula:

PI_w = PI * (% passing 425µm)/100.
Where PI is the Plasticity index.

The third procedure involved calculating the expansiveness (ε_{ex}) using the formula:

$\varepsilon_{ex} = 2.4w_p - 3.9w_s + 32.5$,

Where; w_p = (Plastic Limit) * (%passing 425mm)/100 and w_s = (Shrinkage Limit) * (%passing 425mm)/100.

Clay mineralogy analyses

Several studies link landslide occurrence to specific clay properties (Yalcin, 2007; Kitutu *et al.*, 2009; Mugagga *et al.*, 2011; Yalcin, 2011). Properties of clays are determined by their mineralogical compositions (Mukasa-Tebandeke *et al.*, 2016). In the present study, clay mineralogical composition was analysed using X-ray diffraction. Due to financial consideration associated with budget constraints, only 16 samples were collected for clay mineralogical analysis. The samples were obtained from 8 representative landslide sites consisting of both the top soils and sub soils. Soil samples were prepared for XRD analysis using the back loading preparation method (e.g., Okalebo *et al*., 1993). It was analysed with a PANalytical Aeris diffractometer with PIXcel detector and fixed slits with Fe filtered Co-Kα radiation. The phases were identified using X'Pert Highscore plus software. The relative phase amounts (weight %) were estimated using the Rietveld method.

Shear strength tests

Shear strength parameters are also crucial for slope stability analyses (Bhudu, 2000; Li *et al.*, 2006; Kitutu *et al*., 2009; Gonghui *et al*., 2010; Mugagga *et al*., 2011). To determine the strength of materials, shear strength tests were executed on samples obtained from the landslide sites. These tests included Shear Box and Unconsolidated - Undrained (UU) tests. The tests were conducted to determine the mechanical soil properties which were used in slope stability analysis. Specimens from a relatively undisturbed soil samples were tested to determine the shear strength parameters of soil cohesion (c) and the angle of internal friction (ϕ).

Soil samples retrieved from the field were passed through a 2-mm sieve and the large fragments that could influence measurement were removed. Materials finer than 2mm were then used to prepare specimens for the direct shear tests. The shear strength test was carried out in accordance to British Standard (BS) 1377: Part 7: 1990 as given in Table 4.1, on the recovered undisturbed soil samples. Results of the tests on each specimen were plotted on a graph with the peak (or residual) stress on the y-axis and the confining stress on the x-axis to determine the cohesion (C) and the internal angle of friction (ϕ) values of the materials, which were then used to calculate the factor of safety of materials.

4.3 Results

4.3.1 Soil morphological properties

4.3.1.1 Soil profile 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 groups were 1.5 – 4 m and greater than 6 m, occupying midslopes along topographic hollows and lower slopes respectively (Table 4.2). Surface layers (A horizon) were covered by deposited black soils with an average depth of 0.5m to 1m. They are composed of fine particles and high humus content from decomposing litter (Fig. 4.2). The B and C horizons had depths ranging from 0.65m to 3.44m and 0.88 m and 5.7m, with a reddish brown color and coarse-textured materials concentrated in the lowest layer, respectively (Table 4.2 and Fig. 4.2). The unweathered material was buried at a depth of 1.3m along the spur slopes, 4.5m along topographic hollows and 7m in the valley bottoms.

Table 4.2: Soil morphological characteristics

landslide site	Soil Profile	Slope position	Slope Gradient in degrees	Soil depth in meters	Location of clay pan	Horizon A in metres	percentage distribution			Horizon B in metres	percentage distribution			Horizon C in metres	Percentage distribution		
							Sand	Silt	Clay		sand	silt	Clay		Sand	silt	clay
	P1	Upper slope	38	0.92	0.55	0.61	44	33	23	0.82	28	24	48	0.88	12	28	60
	P2	Upper-Middle	31	1.72	1.02	0.75	44	32	25	1.21	27	25	48	1.55	18	31	51
	P3	Lower middle	21	3.32	1.17	1.17	54	23	23	1.94	21	23	56	2.94	23	27	50
1	P4	Bottom valley	9	5.11	3.14	1.44	38	34	28	2.82	22	16	62	4.12	21	26	53
	P5	Upper slope	47	0.71	0.43	0.67	40	31	29	0.69	23	24	53	0.7	20	25	55
	P6	Upper-middle	26	2.12	1.22	0.82	33	40	27	1.44	29	25	46	2.02	18	19	63
	P7	Lower middle	18	3.83	2.16	0.93	38	32	30	2.11	30	24	46	3.21	22	26	52
2	P8	Bottom valley	10	6.21	3.11	1.11	42	30	28	3.21	21	17	62	4.23	17	22	61
	P9	Upper slope	34	0.62	0.72	0.62	39	36	25	0.92	24	21	55	1.22	23	26	51
	P10	Upper-middle	28	2.03	1.44	0.74	42	31	27	1.23	28	25	47	1.95	21	26	53
	P11	Lower middle	21	3.92	2.82	0.92	35	37	28	1.94	19	33	48	3.32	19	27	54
3	P12	Bottom valley	11	5.82	3.32	1.23	40	34	26	3.11	24	20	56	4.74	21	30	49
	P13	Upper slope	48	0.78	0.69	0.65	43	36	21	0.75	28	27	45	0.91	23	27	50
	P14	Upper-middle	30	1.83	1.12	0.82	44	30	26	1.22	22	14	64	1.73	19	37	44
	P15	Lower middle	16	3.41	1.92	0.93	43	32	25	1.88	21	29	50	3.11	22	23	55
4	P16	Bottom valley	8	6.34	3.17	1.13	40	34	26	3.11	31	18	51	5.44	20	27	53
	P17	Upper slope	43	0.65	0.73	0.61	41	33	26	0.82	25	28	47	1.11	21	22	57
	P18	Upper-middle	31	1.74	0.93	0.73	44	32	24	1.44	20	28	52	1.72	22	25	53
	P19	Lower middle	19	3.12	1.83	0.84	45	32	23	2.11	24	26	50	2.88	18	28	54
5	P20	Bottom valley	7	6.81	3.24	1.06	47	30	23	3.33	26	30	44	5.81	21	25	54
	P21	Upper slope	46	0.87	0.61	0.66	44	31	25	0.93	23	30	47	1.31	20	27	53
	P22	Upper-middle	29	2.31	1.22	0.81	50	32	18	1.71	25	27	48	2.11	19	20	61
	P23	Lower middle	15	4.12	2.12	1.22	42	22	36	2.34	22	36	42	3.88	19	31	50
6	P24	Bottom valley	11	7.12	3.11	1.54	40	28	32	3.33	21	18	61	4.78	22	29	49

	P25	Upper slope	44	0.93	0.82	0.77	52	32	16	0.87	27	23	50	0.93	23	19	58
	P26	Upper-middle	33	2.21	1.36	0.88	43	29	28	1.32	26	24	50	2.01	16	34	50
	P27	Lower middle	18	3.36	1.97	0.91	46	36	18	2.11	22	31	47	3.12	23	25	52
7	P28	Bottom valley	13	6.61	3.92	1.22	55	35	10	3.42	18	36	46	4.68	18	36	46
	P29	Upper slope	42	0.96	0.78	0.71	50	35	15	0.88	31	23	46	0.97	21	26	53
	P30	Upper-middle	25	2.42	1.44	0.92	44	37	19	1.94	28	23	49	2.11	22	28	50
	P31	Lower middle	20	4.21	2.33	0.95	48	30	22	2.88	30	25	45	3.89	18	21	61
8	P32	Bottom valley	12	7.21	3.12	1.23	47	37	16	3.44	22	36	42	5.88	20	27	53
	P33	Upper slope	48	0.83	0.48	0.66	49	33	18	0.84	30	21	49	0.95	23	26	51
	P34	Upper-middle	28	1.94	0.78	0.77	42	36	28	0.91	28	21	51	1.44	19	27	54
	P35	Lower middle	22	3.56	2.21	0.93	53	31	16	2.45	27	25	48	3.22	18	30	52
9	P36	Bottom valley	8	6.72	3.37	1.11	50	27	23	3.11	19	37	44	5.77	19	24	57
	P37	Upper slope	45	0.94	0.53	0.71	44	33	23	0.88	29	23	48	1.17	21	24	55
	P38	Upper-middle	34	2.44	1.52	0.93	40	33	27	1.98	23	27	50	2.23	23	20	57
	P39	Lower middle	17	4.23	2.17	1.03	48	33	19	2.44	17	20	63	4.11	22	21	57
10	P40	Bottom valley	11	6.83	3.13	1.33	45	30	25	3.12	18	35	47	5.88	22	18	60

Where P is the soil profile number

Soil profile analysis in the mini pits showed remarkable variations in colour and texture with depth. The soil profiles show visible colours and textural gradation within 3 distinct horizons. Most of the analysed profiles in the upper slope sections exhibit an abrupt change in colour from yellow (5Y8/2) to brown (7.5YR5/2) clay with γ 2.06 g/cm³. The subsurface horizons have a dark reddish brown (5YR 3/6) and orange (7YR 5/6) colour in moist and dry conditions respectively (Fig. 4.2). Profiles in the lower slope sections have surface horizons with brown (7.5YR 4/4) colour both in moist and dry conditions and a dark to dull reddish brown colour (7.5YR 4/6) in underlying horizons. The materials then gradually grade into a light grey (2.5Y7/1) silty clay with alternating grey and brown strips. The colour for the surface horizons in the profiles along the middle slope sections is dark (7.5YR 4/1) and dull brown (7.5YR 5/4) in moist and dry conditions respectively. The colour changes with depth to dark yellowish brown (10YR 5/6) (Fig. 4.2). Together with other physical properties, soil colour was used to differentiate between types of horizons of the same and different soil profiles. Soil colour as a physical property of soils was used in the present study to gain insights into some of its most important characteristics, such as mineral composition, age and soil processes (e.g. chemical alteration, carbonate accumulation and the presence of humified organic matter).

A relationship exists between soil depth and slope gradient, 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 soils (>5m) are found on slope sections with gradient lower than 10˚, moderately deep soils (2 to 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 than1m (Table 4.2). 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.

4.3.1.2 Location of clay pans

Field observations within mini pits and dug profiles indicated the presence of clay pans at a depth ranging between 0.75 and 3 m within the profiles (Table 4.2 and Fig. 4.2). A clay pan is an indurated and/or cemented dense-less permeable layer in the subsoil having a much higher clay content than the overlying material (Gillot, 1986). They are sticky when wet and hard when dry (Jiang *et al.*, 2007a). In the study area, there is variation in the location of clay pans within the soil profile. In some profiles, the clay pans were observed near the surface while in some horizons the pans were identified at greater depths. Whereas the upper slope soil profiles have clay pans near the surface at a depth of less than 0.85 m, lower slope soil profiles have pans at a depth greater than 2 m.

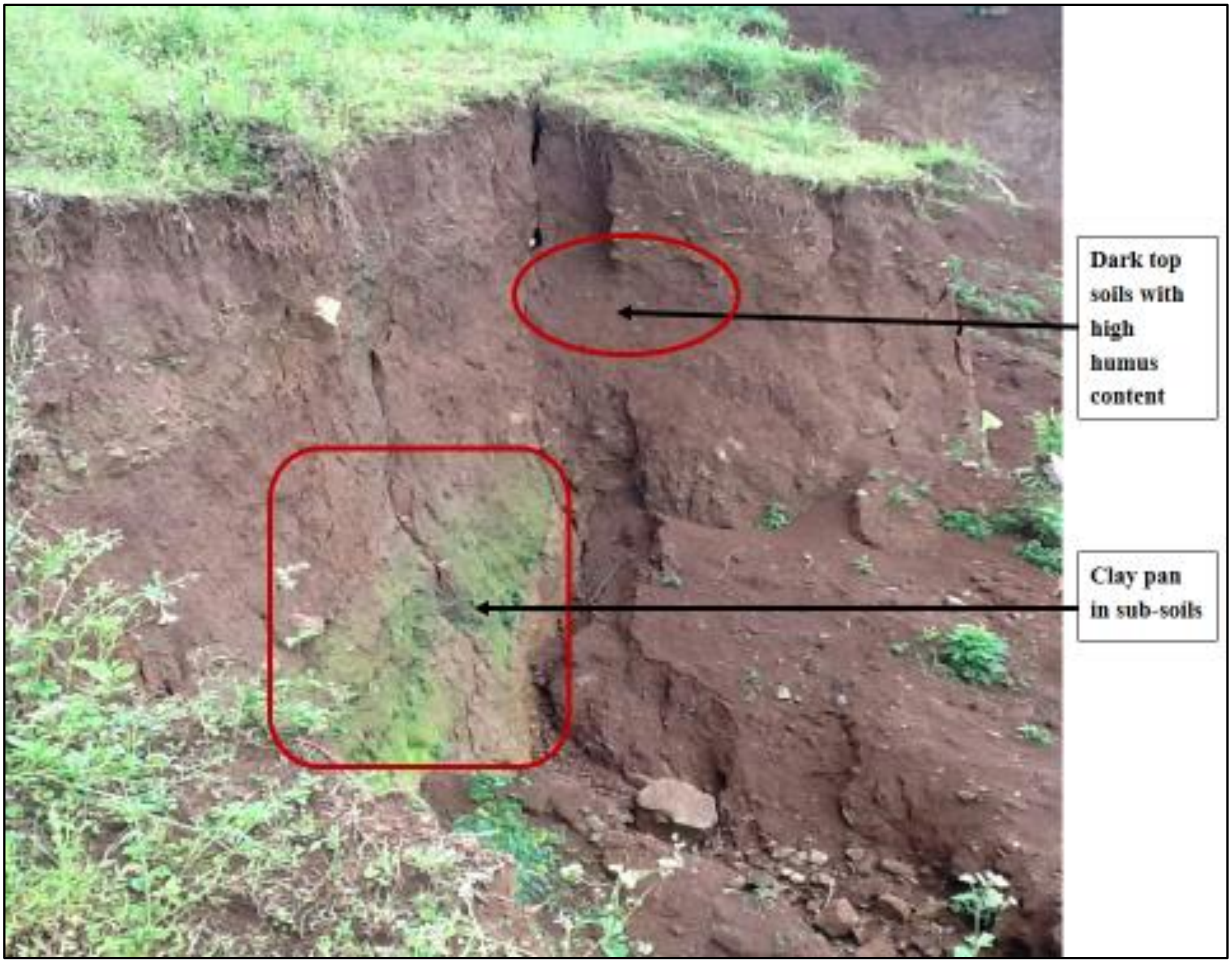

Fig. 4.2: Soil profiles dominated by clay pans in the sub-surface soils.

Clay pans in soil profiles along the middle slope sections are located at depth ranging from 1- 2 m (Table 4.2). Characterizing the variation in the depth of the clay pan horizon is a helpful step in describing other soil properties, as explored in the discussion section.

4.3.2 Effect of particle size on landslide occurrence

The study area is dominated by clay soils in the subsurface horizons. The particle size determination (Fig. 4.3) shows that the soils of the study area are dominated by clay presence, except for the uppermost surface horizons. The coarse-grained soils (sand and gravel) are represented by the soil particle proportion passing through sieve #200 and are less than 50% (Table 4.2). The fine-grained soils (clays and silts) are represented by the percentage of soil passing through sieve #200 and are greater than 50% of the soil in all the samples (Fig. 4.3 and Appendix 2).

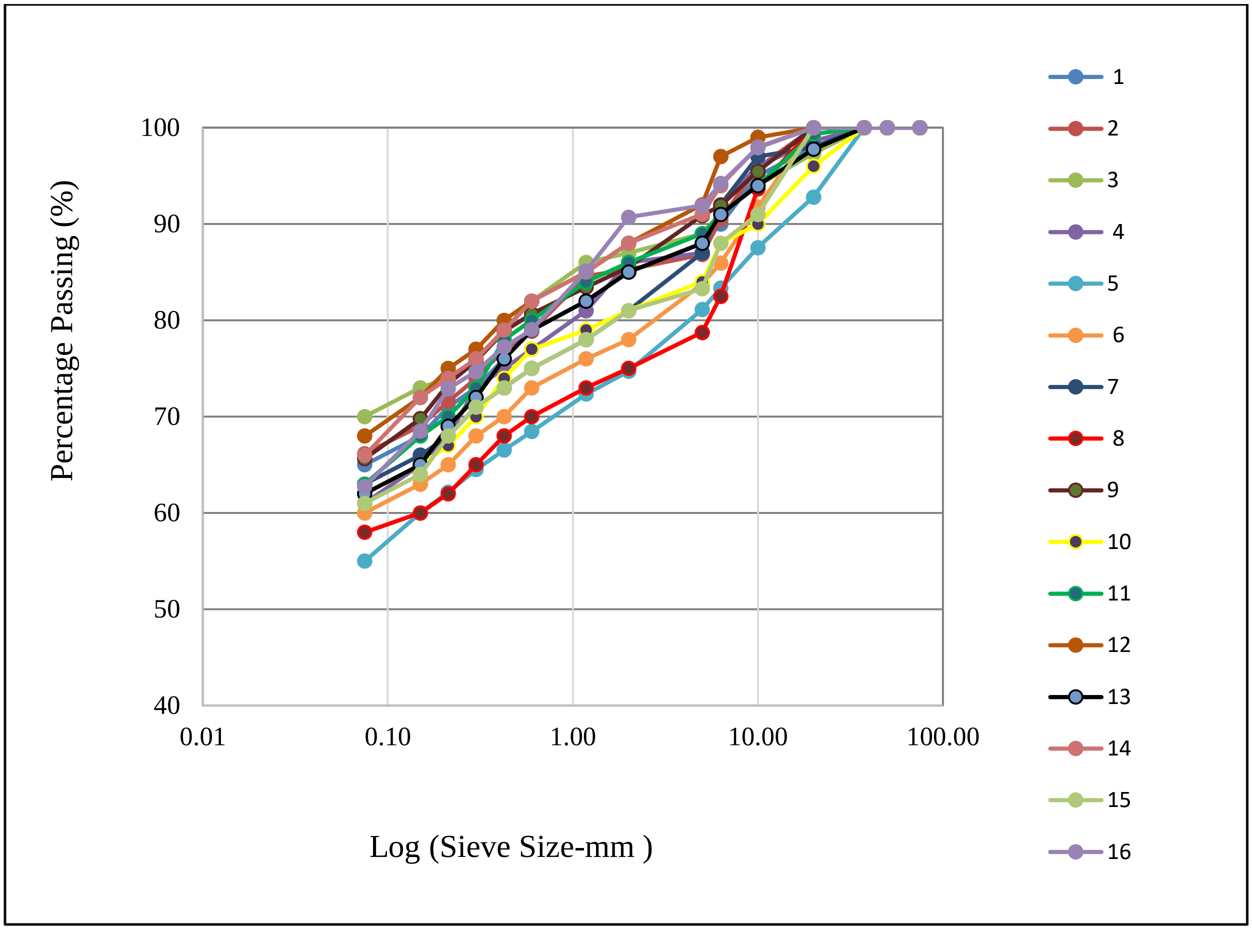

Fig. 4.3: Particle size distribution curves.

A predominance of fine grained materials of either silt or clay was identified. Sieve analyses indicated that soil texture varies with soil profile depth. Whereas sand particles dominate the top soils, finer silt and clay materials dominate the subsurface horizons. All surface horizons analysed had the percentage of sand, silt and clay ranging from 33 to 55%, 22 to 40% and 10 to 30% respectively. Whereas sand and silt percentages reduced from 18 to 30% and 14 to 36% in the subsurface horizons, the clay percentage increased from 42 to 64%. In the deeper horizons, sand was observed to reduce drastically to less than 23%, while clay increased to greater than 50% (Table 4.2).

The variation of clay content in the soil samples from the threshold values is presented in Figs 4.4 to 4.6.

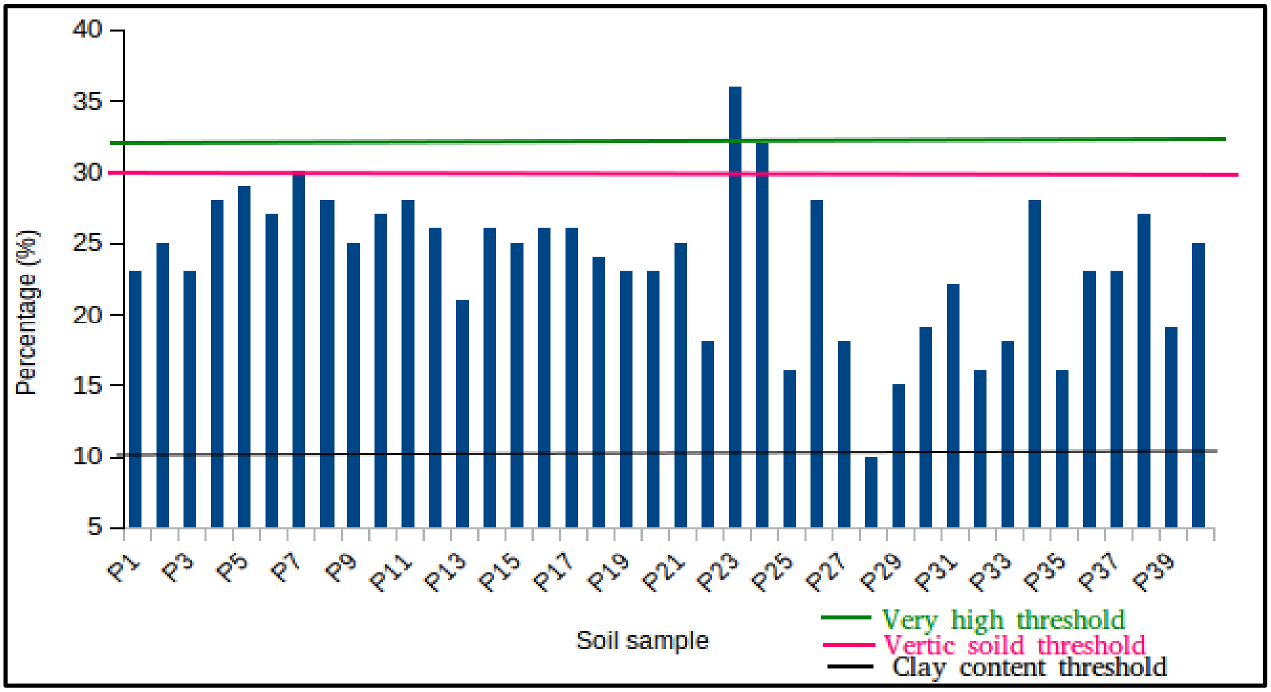

Fig. 4.4: Variation of clay content from the threshold values for the top soil horizons

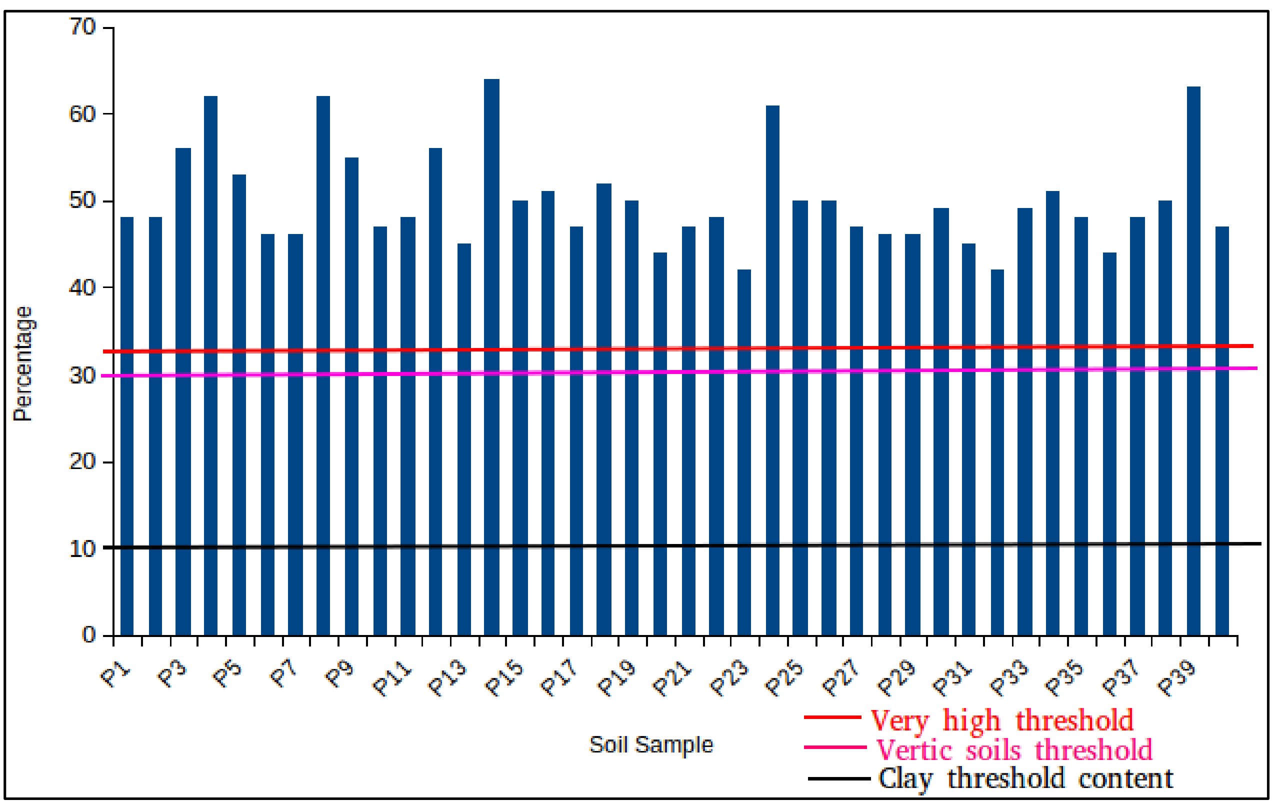

Fig. 4.5: Variation of clay content from the threshold in the subsoil horizons.

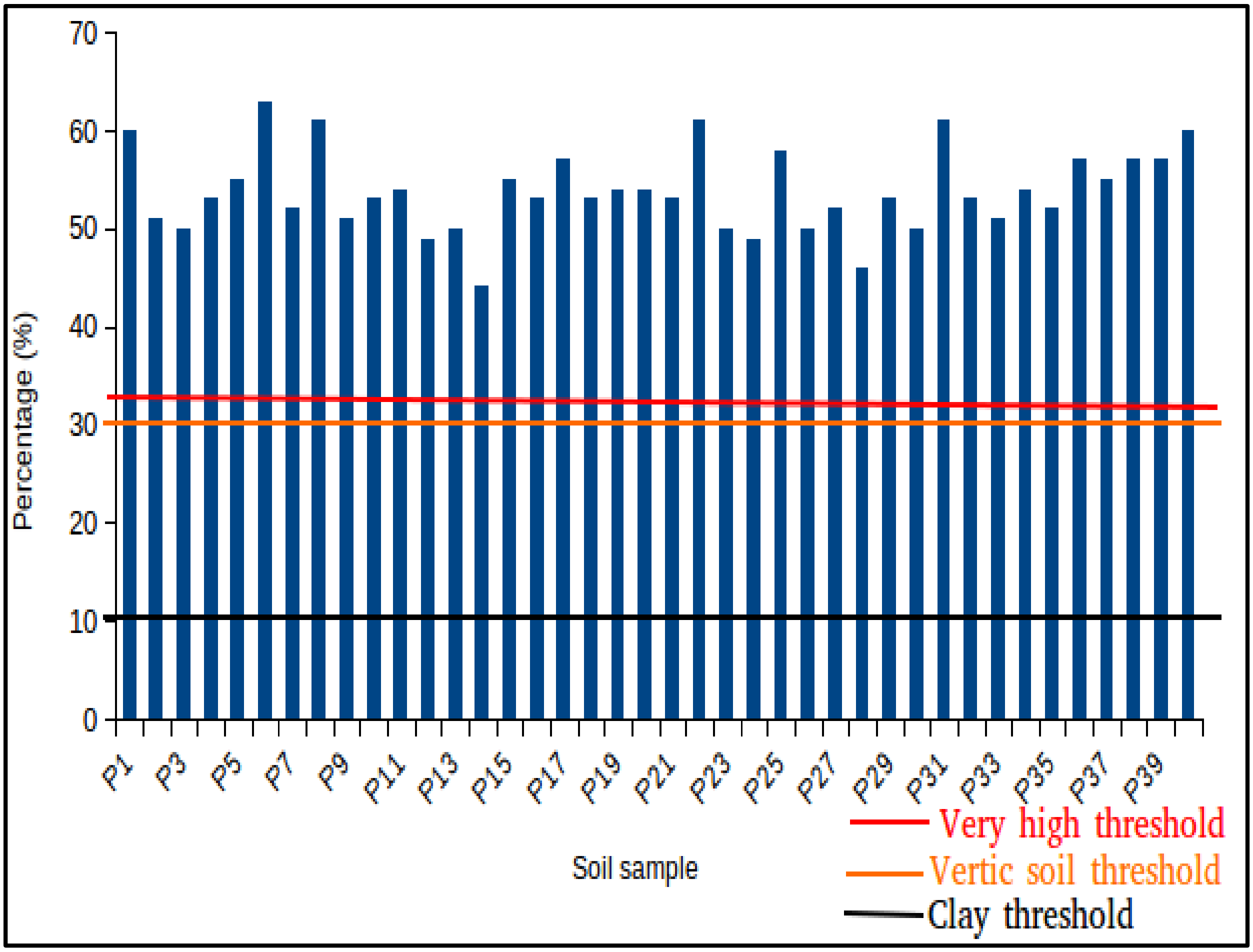

Fig. 4.6: Variation of clay content from the threshold values in the deeper soil horizons.

From Figs 4.4 to 4.6, it is clear that the clay content in all the horizons is well above the 10% threshold values for expansive materials (e.g., Wati *et al.*, 2010; Mugagga *et al.*, 2011). A few samples in the top soil are above the 32% threshold values for vertic soils (see Van Der Merwe, 1964; Baynes, 2008). The clay content for all the samples in the sub soil and deep soil horizons is well above the 32% threshold for vertic soils. The extremely fine-textured nature of the material in sub soil horizons of the study area have strong implications for landslide occurrence, as will be explained in the discussion section. The clay fraction, which is well above the 10% threshold, explains the shrink-swell properties. The soils exhibit extreme expansion potential and are hence susceptible to landslides. The clay content is very high in most of the tested samples exceeding 35% (Table 4.2), especially in the sub and deeper soil horizons, implying vertic nature of the soils. The swelling and shrinkage characteristics of vertic soils are a very important factor in the localisation of landslides, as it will be explored in the subsequent section on clay mineralogy.

4.3.3 Clay mineralogy and landslide occurrence

The results of the X-ray diffraction tests are provided in Fig 4.7 and Table 4.3.

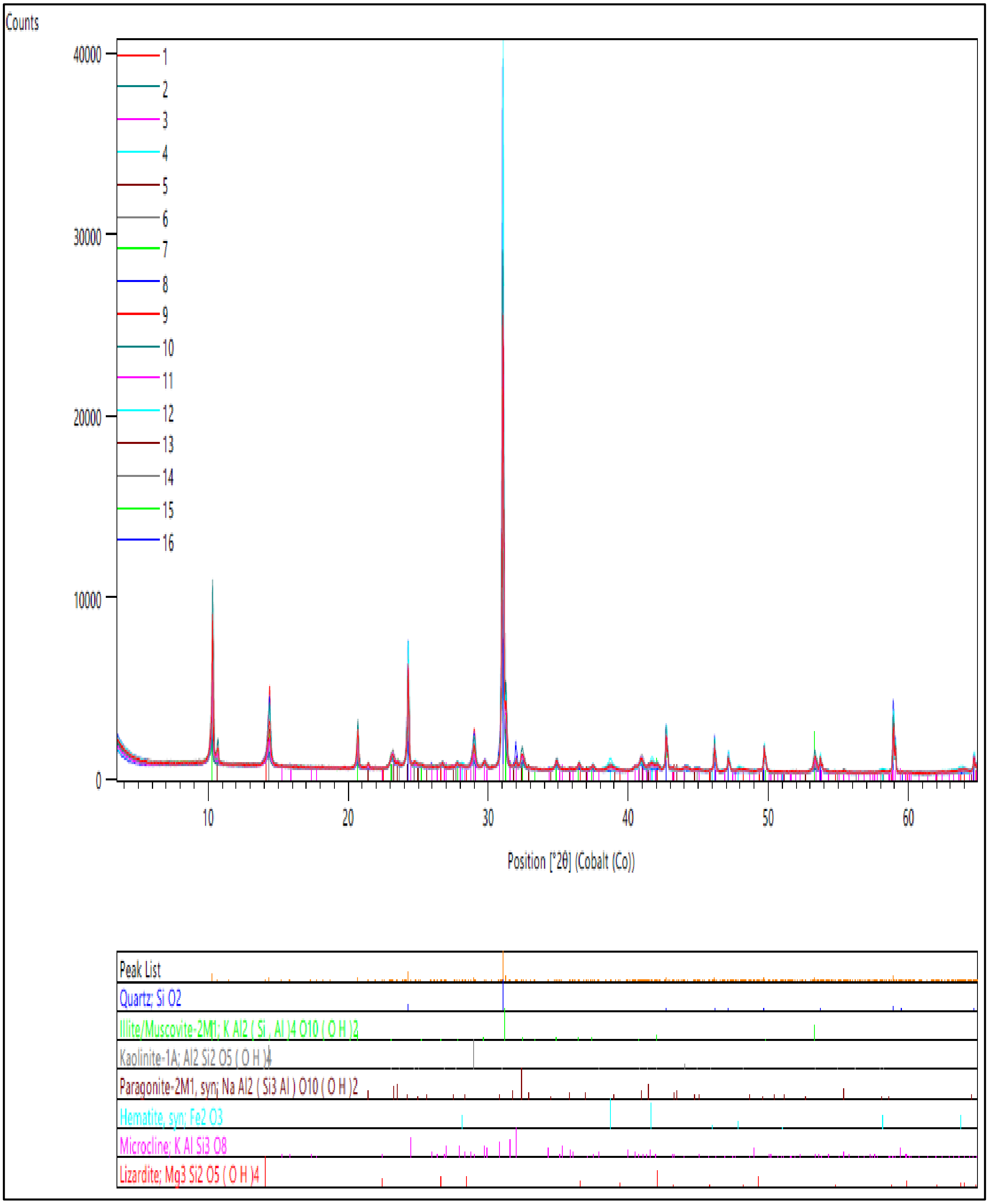

Fig. 4.7: XRD patterns of clay minerals

Table. 4.3: XRD mineral distribution in percentages

Samples	Soil horizon	Quartz	Illite/ muscovite	Kaolinite	Paragonite	Hematite	Microcline	Lizardite
1	Top soil	56	26	9	4	1	0	4
2	sub soil	46	43	7	2	1	0	1
3	Top soil	60	27	6	6	2	0	0
4	sub soil	39	43	8	4	3	1	1
5	sub soil	42	41	10	3	2	1	1
6	sub soil	43	45	8	2	1	1	0
7	sub soil	43	42	9	3	1	1	0
8	sub soil	36	47	11	3	1	1	1
9	sub soil	36	44	12	5	1	1	1
10	Top soil	75	18	5	0	2	0	0
11	Top soil	62	28	6	2	1	0	2
12	Top soil	76	16	6	1	0	1	0
13	Top soil	71	22	5	0	1	0	2
14	sub soil	41	46	7	4	0	1	1
15	Top soil	64	22	6	4	2	0	2
16	Top soil	82	12	3	2	0	2	0
0 = n.d. – not detected above the detection limit of 0.5 -3 weight percent								

From the X-ray diffraction patterns, it is clear that the main constituent minerals of the samples include quartz, illite/ muscovite and kaolinite. Samples from the top soil predominantly contain more quartz. The implication of such high amounts of quartz/sand in the top soil horizons for soil behaviour will be unravelled in the discussion section. Subsurface horizons have considerable amounts of illite/muscovite as the dominant clay minerals, ranging from 43% to 47 % (Table 4.3 and Fig 4.7). Notwithstanding the absence of smectite clays, the soils contain large amounts of moderately expansive clays, particularly illite/ muscovite.

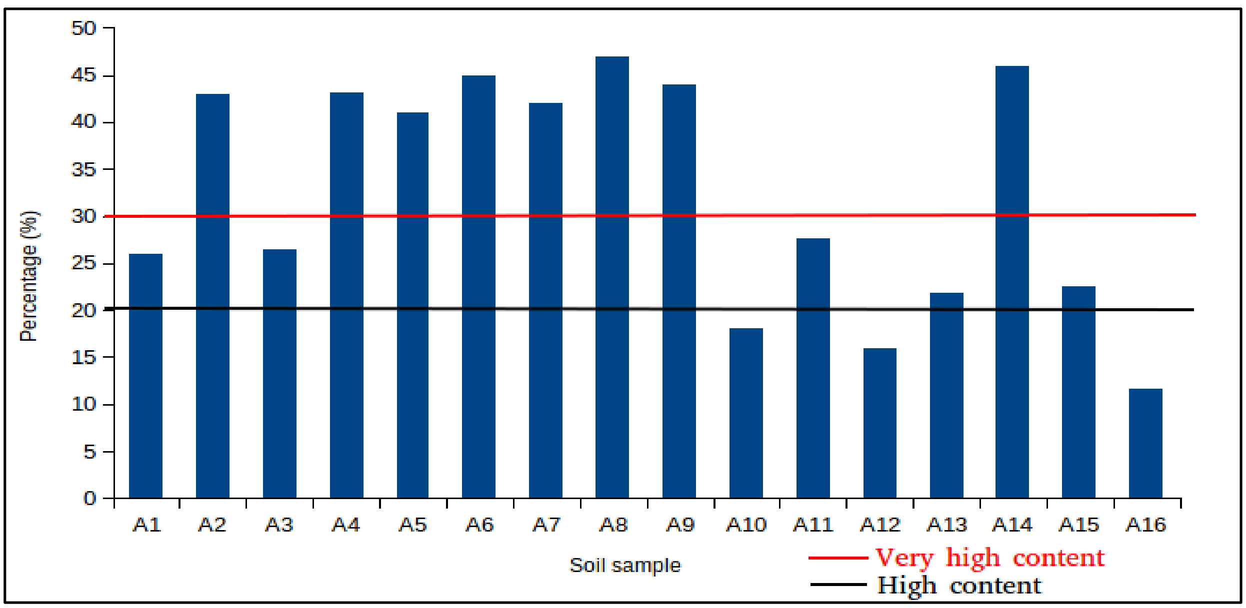

Fig. 4.8: Variation of illite/muscovite content from the threshold values.

From Fig. 4.8, it can be observed that more than 81% of the analysed samples had percentage of illite/muscovite minerals above 20% threshold value (see Ohlmacher, 2000; Wati *et al*., 2010) and are considered as highly expansive. More than 50% of the samples had values greater than 30% and are considered as very high (e.g., Husein *et al.*, 1999).

4.3.4 Soil dispersion and landslide occurrence

Atterberg Limit tests results are presented in Table. 4.4 and Fig. 4.9. The Liquid Limit and Plasticity Index for all the tested samples is greater than 50% and 30% respectively (Table 4.4). From the plasticity chart (Fig. 4.9), all the soil samples tested were inorganic clay of high plasticity belonging to the CH group. Such soils can easily move when saturated, leading to high incidence of landslide occurrence.

Table. 4.4: Plasticity index and shear strength parameters

Soil sample	Plasticity								Shear strength Parameters		Soil Class	USCS
									C '	φ '		
	Liquid Limit	Plastic Value	Plasticity Index	Linear Shrinkage	w_p	w_s	PI_w	ε_{ex}	kPa	(Degree)		
1	58	26	32	15	20	11	24	37	7	3	Clayey	CH
2	54	21	33	15	16	12	25	26	7	4	Clayey	CH
3	66	22	44	21	18	16	35	11	5	5	Clayey	CH
4	58	27	31	15	20	11	23	39	7	4	Clayey	CH
5	53	29	25	11	19	8	16	50	9	5	Fat Clay	CH
6	60	26	34	16	18	11	24	32	10	7	Clayey	CH
7	54	27	27	13	20	9	20	45	8	6	Clayey	CH
8	52	25	27	12	17	9	18	40	6	3	Clay Loam	CH
9	56	24	32	15	19	12	26	31	9	3	Clay Loam	CH
10	66	25	41	20	18	15	31	20	10	8	Clay Loam	CH
11	50	23	28	13	18	10	21	37	11	7	Clay Silt	CH
12	49	29	20	11	23	8	18	55	5	4	Clay Silt	CH
13	60	27	33	16	20	12	25	35	7	4	Sand Clay	CH
14	57	23	34	16	19	13	27	28	8	4	Sand Clay	CH
15	57	28	29	14	21	10	21	43	7	3	Clay Loam	CH
16	53	26	27	13	20	10	15	36	8	5	Clay Loam	CH
17	49	24	25	13	18	8	21	41	7	3	Clay Silt	CH
18	50	26	24	15	20	11	24	37	7	4	Silty Clay	CH
19	52	21	31	15	16	12	25	26	5	5	Silty Clay	CH
20	66	22	44	21	18	16	35	11	7	4	Clay Loam	CH
21	58	27	31	15	20	11	23	39	9	5	Clay Loam	CH
22	53	29	24	11	19	8	16	50	10	7	Clay Silt	CH
23	60	26	34	16	18	11	24	32	8	6	Clay Loam	CH
24	54	27	27	13	20	9	20	45	6	3	Sand Clay	CH

25	52	25	27	12	17	9	18	40	7	3	Sandy Loam	CH
26	56	24	32	15	19	12	26	31	7	4	Clayey	CH
27	66	25	41	20	18	15	31	20	5	5	Silty Clay	CH
28	50	23	27	13	18	10	21	37	7	4	Clay Loam	CH
29	49	29	20	11	23	8	18	55	9	5	Clayey	CH
30	60	27	33	16	20	12	25	35	10	7	Silty Clay	CH
31	57	23	34	16	19	13	27	28	8	6	Clay Loam	CH
32	57	28	29	14	21	10	21	43	6	3	Clay Loam	CH
33	53	26	27	13	20	10	21	41	9	4	Clay Loam	CH
34	49	24	25	13	17	9	20	39	9	6	Clayey	CH
35	51	25	26	15	20	11	23	39	7	4	Clay Loams	CH
36	62	28	34	11	19	8	16	50	9	5	Silty clay	CH
37	63	26	37	16	18	11	24	32	10	7	Silty clay	CH
38	55	24	31	13	20	9	20	45	8	6	Clayey	CH
39	58	29	29	12	17	9	18	40	6	3	Clayey	CH
40	52	25	27	15	19	12	26	31	9	3	Clay Loam	CH

The fine-grained soil materials tested had Liquid Limits (LL) with an average value of 58.43%, ranging between 50.43% and 66.43%, considered to be of high plasticity.

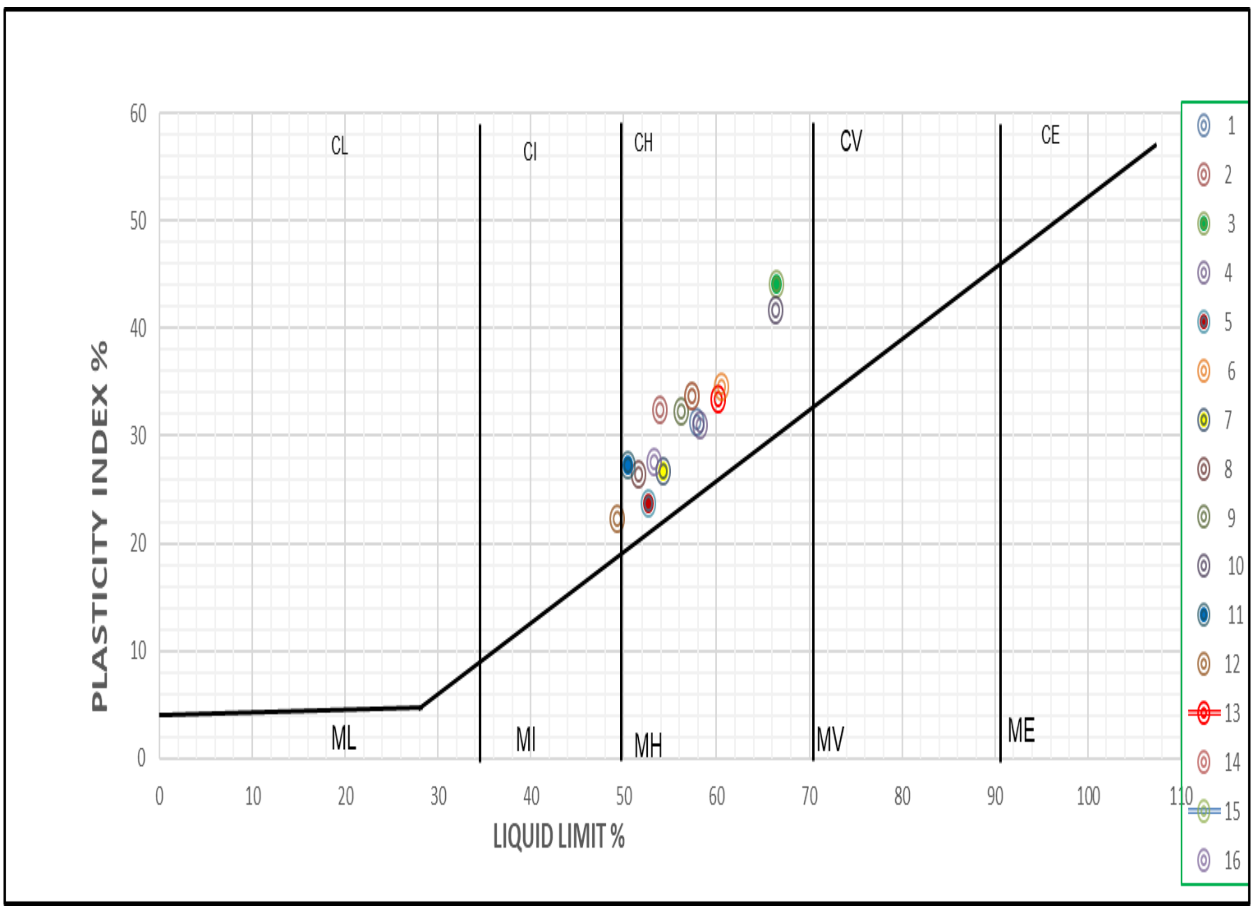

Fig. 4.9: The distribution of samples on the plasticity chart for the USCS.

Fig.4.9 shows the results of the Atterberg limits tests plotted on a plasticity chart in order to determine soil classification based on the Unified Soil Classification System ASTM D2487 (ASTM International, 2004a). It is evident that soils in the study area have high plasticity as already revealed by XRD results.

The variation of the Liquid Limits from the threshold values is illustrated in Fig. 4.10.

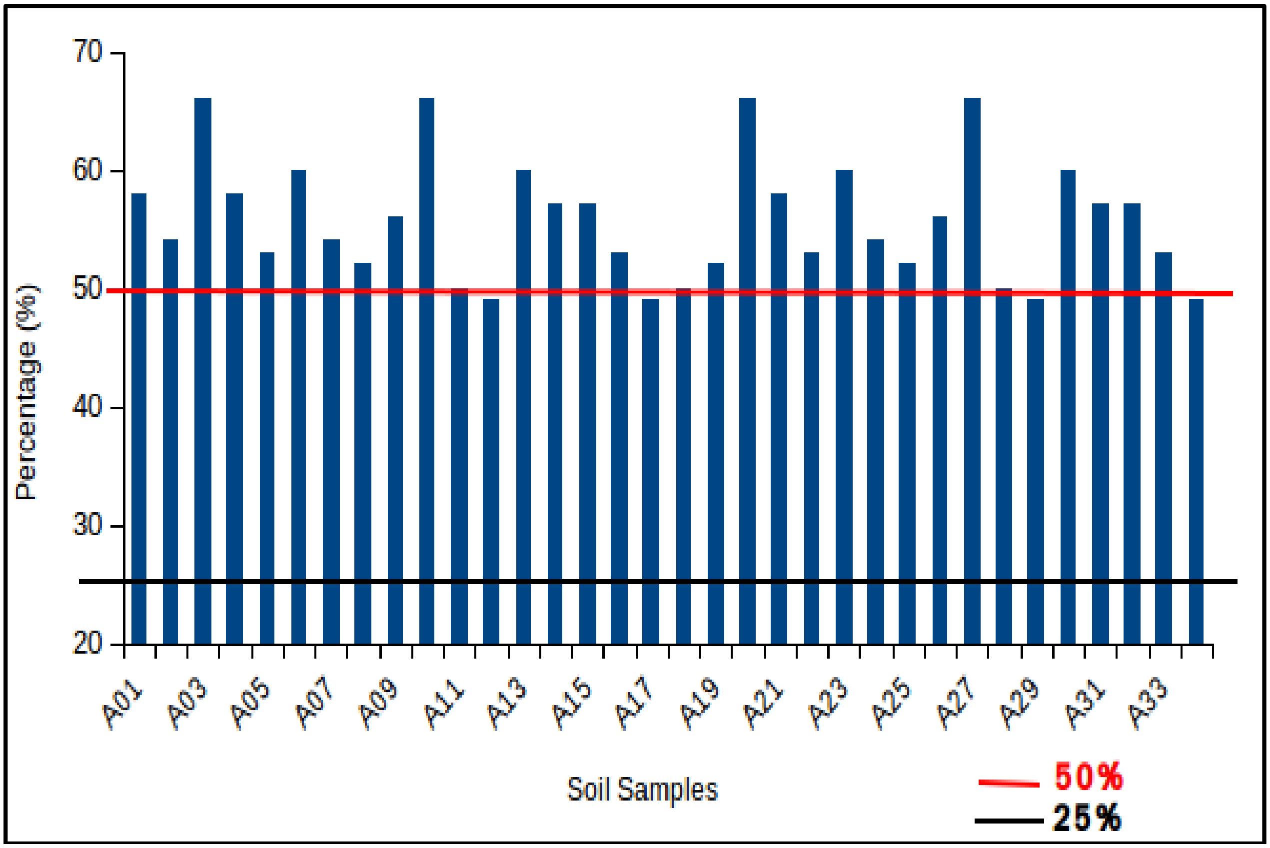

Fig. 4.10: Variation of Liquid Limits from the threshold values for expansive materials

From Fig. 4.10, it is clear that all the samples have Liquid Limits above the 25% threshold for expansive potential (See Baynes, 2008). More than 90% of the samples have Liquid Limits above 50% and are highly expansive.

The plastic value ranged from 21.3% to 28.9% with an average of 25.1%, while the average plasticity index was 33.3% ranging from 22.4% to 44.2% (Table 4.4). The clay materials have plasticity index greater than 30, hence considered as highly plastic and expansive. The variation in plasticity index values from the thresholds are illustrated in Fig. 4.11.

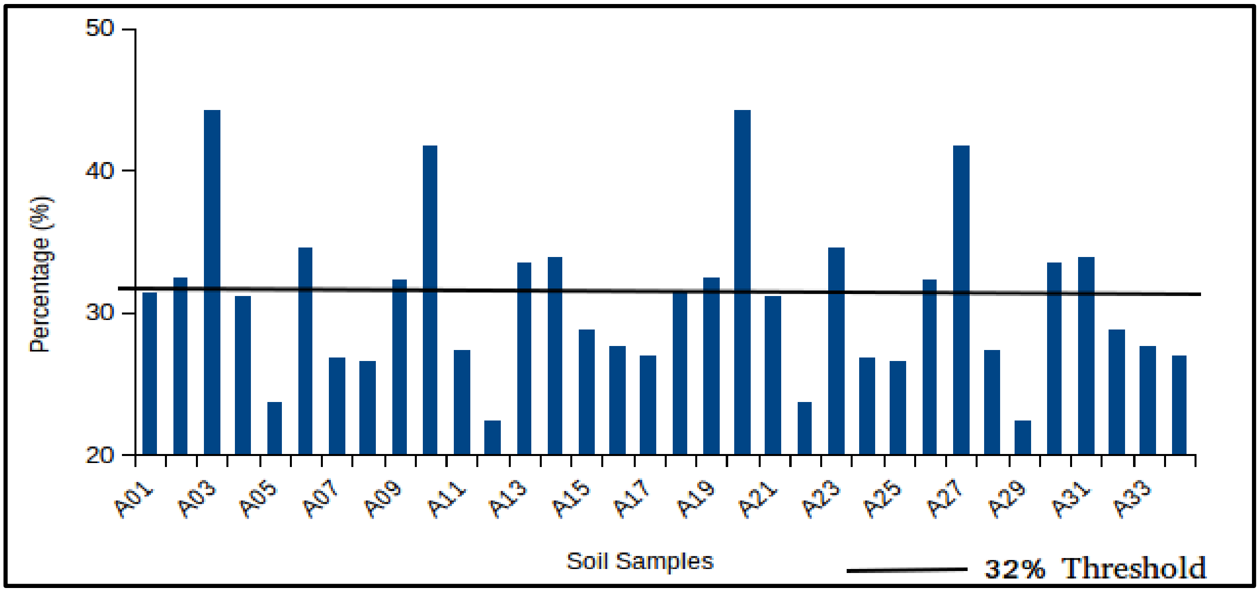

Fig. 4.11: Variation of Plasticity Index from the threshold values for expansive materials

From Fig. 4.11, it is clear that more than 40% of the analysed samples have a plasticity index above the 32% threshold for extremely high expansive potential (see Van Der Merwe *et al*., 2002; Baynes, 2008). The presence of many cracks, observed on the soil surface during field investigations confirmed this (Fig. 4.12). The presence of such cracks is also a characteristic of Vertisols with high expansive potential.

Fig. 4.12: Highly expansive materials with many surface cracks.

The computed Linear Shrinkage (LS) ranged between 10.53 and 20.76 (Table 4.4). The computed weighted Plasticity Index (PI_w) ranged between 17.92 % and 34.92% averaging 28.4% while the expansiveness (ε_{ex}) ranged between 10.7 and 54.8 averaging 38.6 for all the soil samples (Table 4.4). The computed weighted plasticity index for most of the tested soils is above the 20% threshold for expansive soils (See Gourley and Schreiner, 1993), signifying highly unstable soils. From Fig. 4.13, it can be observed that more than 65% of the analysed samples have a weighted plasticity index above the 20% threshold.

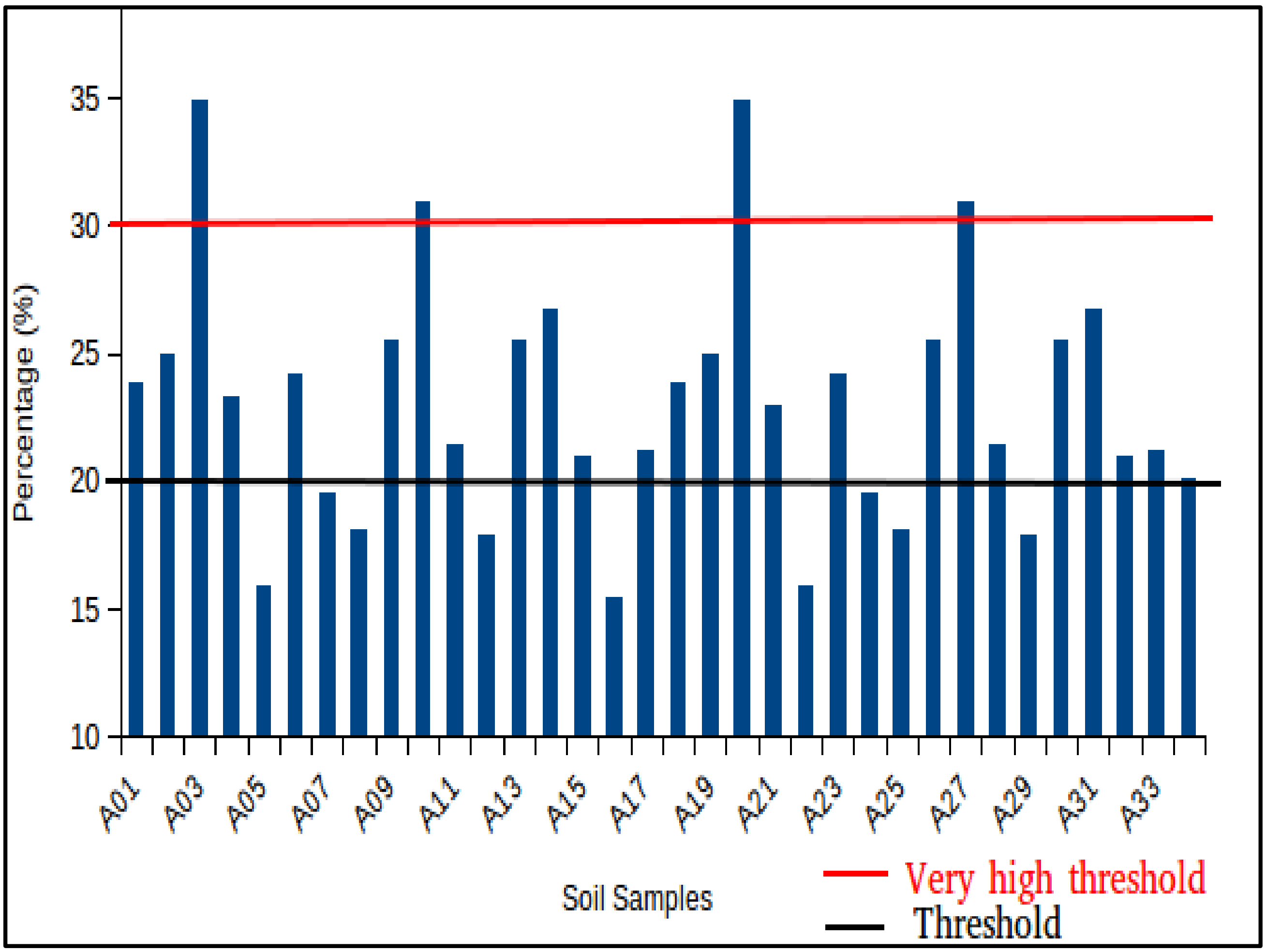

Fig. 4.13: Variation of computed weighted plasticity index from threshold values for expansive soils.

All tested samples had expansiveness (ε_{ex}) within the range of 20 to 50 (Table 4.4), indicating medium expansive soils susceptible to slope failure. More than 90% of the analysed samples have expansiveness above the 20% threshold (Elges, 1985; Gourley and Schreiner, 1994). More than 80% of the samples have values between 20 and 50% showing medium expansive soils (Fig. 4.14).

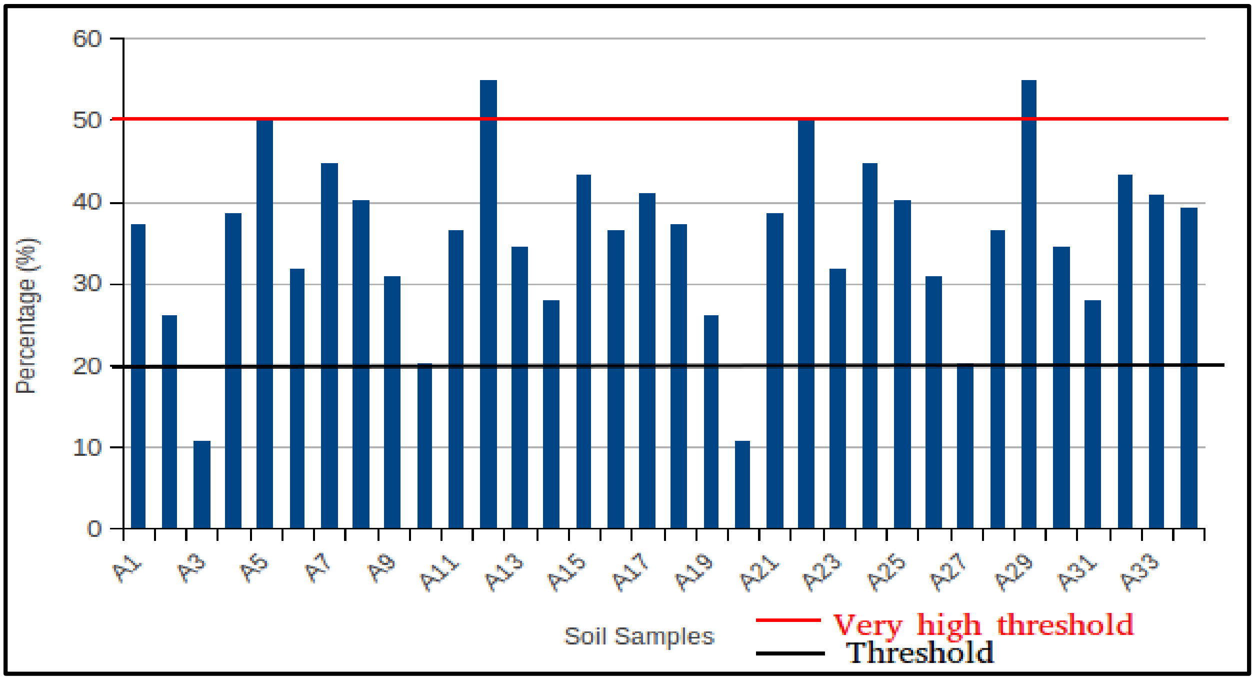

Fig. 4.14: Variation of expansiveness of the soil from the threshold values.

Results of double hydrometer test (Table 4.5) show that most of the samples have dispersion values greater than 30% (See Bell and Maud, 1994), implying that the majority of the soil materials are prone to slope failure. More than 90% of the soil samples have critical dispersion values greater than 50% (Table 4.5). This signifies high susceptibility to landslide occurrence. In the study area, dispersive soils are generally confined to the subsurface horizons with concentrated illite/ muscovite clay minerals.

Table 4.5: Double hydrometer test results

Soil samples	Hydrometer Test (proportion of particle sizes %)		
	% Passing 0.005mm using chemical (Sodium Hexametaphosphate solution)	% Passing 0.005mm without using chemical	%age Dispersion
A1	30	19	63
A2	30	17	57
A3	30	15	50
A4	31	14	45
A5	23	7	30
A6	21	7	33
A7	27	12	44
A8	21	8	38

A9	23	13	57
A10	31	10	32
B1	28	7	25
B2	30	11	37
B3	37	16	43
B4	29	12	41
B5	22	9	41
B6	30	8	27
B7	33	10	30
B8	36	17	47
B9	22	12	55
B10	27	14	52
C1	31	11	35
C2	31	13	42
C3	29	15	52
C4	33	16	48
C5	28	13	46
C6	31	18	58
C7	27	17	63
C8	22	14	64
C9	38	10	26
C10	34	12	35
D1	30	14	47
D2	29	18	62
D3	23	12	52
D4	24	9	38
D5	31	7	23
D6	37	13	35
D7	33	12	36
D8	40	17	43
D9	23	14	61
D10	39	10	26

Variation of soil dispersion values from the thresholds is illustrated in Fig. 4.15. It is clear that more than 80% of the analysed samples have dispersion values above the 30% threshold (see McCook, 1980; Elges, 1985) and are considered moderately dispersive materials. Five samples are above the 60% critical dispersion value (Bell and Maud, 1994) and are considered as highly dispersive.

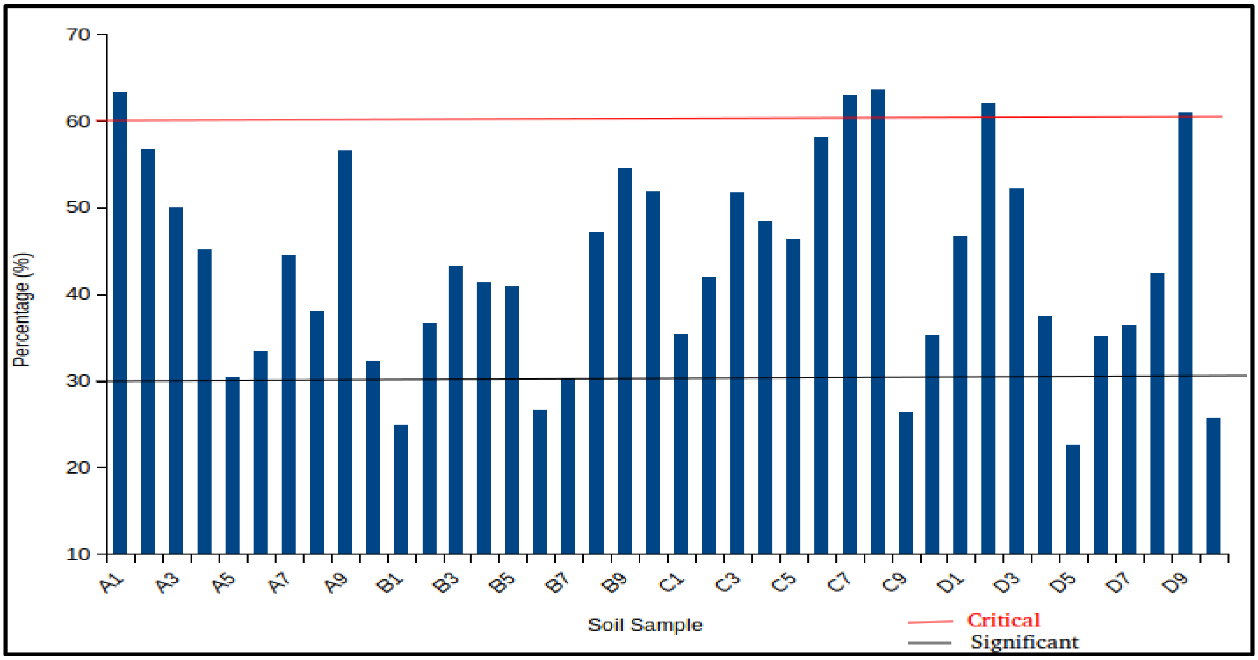

Fig. 4.15: Variation of soil dispersion from the threshold values.

4.3.5 Shear strength parameters

Shear strength parameter test results are presented in Table 4.4. The soils have cohesion (C) ranging from 5.2 kPa to 11.1 kPa with an average of 8.2 kPa. The angle of internal friction ranged from 2.6˚ to 8.1˚ with an average of 5.4˚. The lowest cohesion of 5.2 kPa was detected in soils completely saturated with water. The present study considered a minimal cohesion value of 8.2 kPa as the critical state equilibrium all over the area, because failure is expected to occur while soils are saturated with water. The soil materials have a very low angle of internal friction (<8.5˚) and are thus considered weak and susceptible to landslide occurrence.

4.3.6 Soil water infiltration

Soil water infiltration test results are presented in Fig. 4.16. High infiltration rates were observed in the top soils with depth ranging from 0.3 to 0.8m but drastically reduced in the sub soils. The high infiltration rates in the top soils are explained by the presence of loamy sandy soils. The subsoil is predominantly clay, with clay pans distinctly underlying the top soils, consequently limiting infiltration. This has an effect on the response of soil materials to incoming infiltrating water and consequently the timing of landslides, as will be explored in the discussion section.

Steady infiltration rates significantly varied with slope position and topographic configurations, as well as land use/ cover types. Lower slope zones tended to have higher infiltration rates than the upper elements (Fig. 4.16). The average soil-water infiltration was 24cm/h^{-1} in uppermost slope sections, 30cm/h^{-1} on the upper-middle sections, 70 cm/h^{-1} on the lower middle sections and greater than 80cm/h^{-1} in the bottom valleys. This signifies more saturation in the lower slope sections than the upper slopes. A relationship was also established between soil infiltration and topographic configuration. More water was observed to infiltrate along the topographic hollows than on spur slopes. Infiltration along the topographic hollows in the upper slopes was greater than 30cm/h^{-1}and less than 12cm/h^{-1} on the spur slopes. On the middle slope sections, infiltration rates along topographic hollows was greater than 70cm/h^{-1} and less than 45cm/h^{-1} on the spur slopes. Whereas soil infiltration along the topographic hollows was greater than 85cm/h^{-1} in the lower slope close to the valley bottoms, it was less than 55cm/h^{-1}on the spur slope sections (Fig. 4.16).

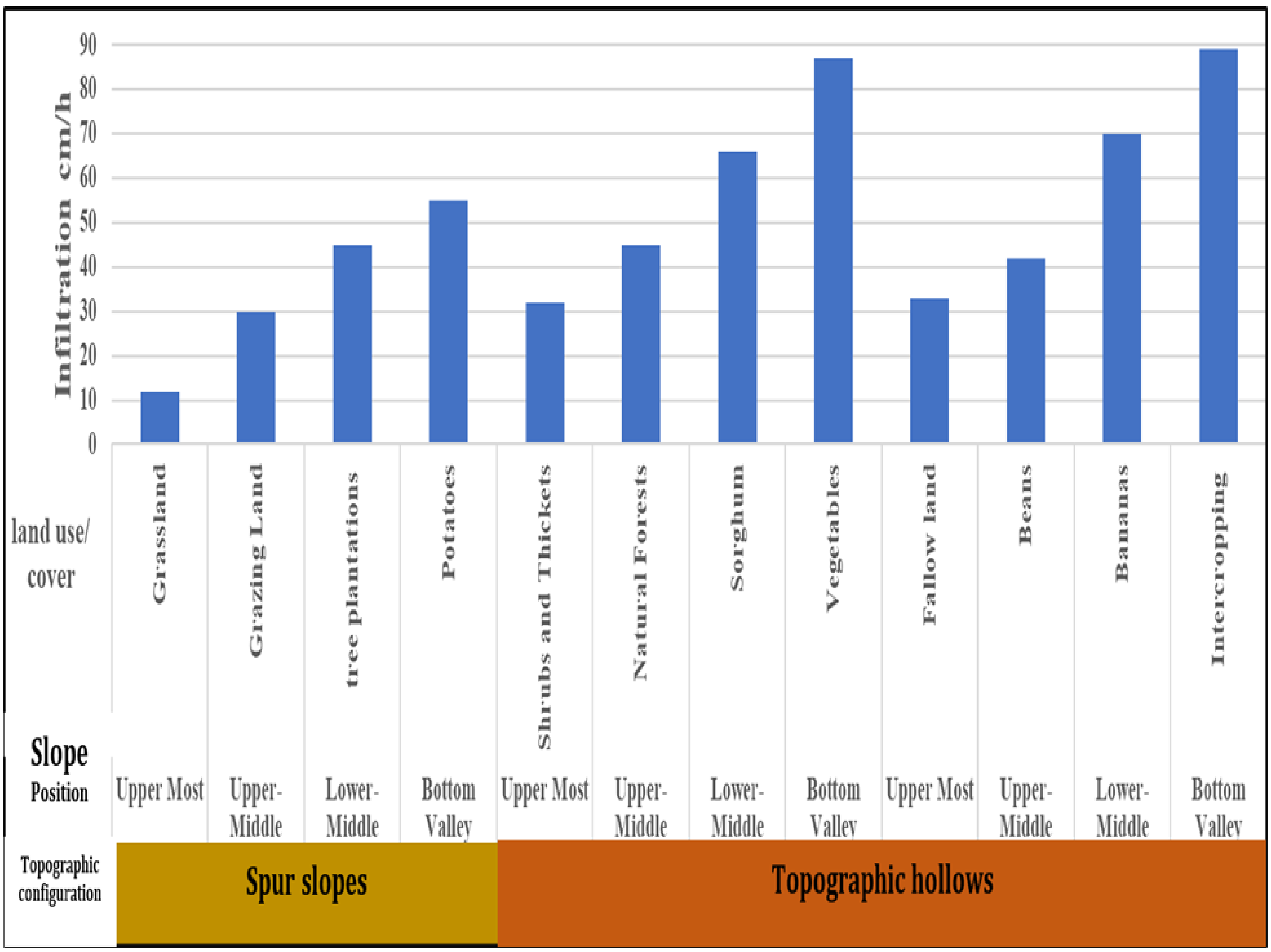

Fig. 4.16: Infiltration rates along slope positions, topographic configurations and land use/cover.

Soil water infiltration was observed to correspond to land use and cover types. The observed rates of soil infiltration were higher on the agricultural land use, especially cultivated areas than natural land cover types in the catchment. With the exception of bean covered areas which had infiltration rates lower than 42cm/h^{-1}, the rest of the agricultural land cover types had infiltration rates greater than 65cm/h^{-}1. In some cases, rates as high as 89cm/h-1 were recorded in sections with intercropping. On the other hand, with the exception of natural forest cover which had infiltration rates greater than 45cm/h^{-1}, the rest of the natural land cover types had infiltration rates less than 30cm/h^{-1} (Fig. 4.16). In some areas, rates lower than 12.2 cm h^{-1}were recorded on grassland areas. In the light of the above results, it was established that infiltration rates in the catchment vary from rapid to very rapid. They also vary with slope positions and topographic configurations as well as land use types. The steady state water infiltration rates vary between 12.2 cm h^{-1} and 88.5 cm h^{-1}.

4. 3.7 Rainfall distribution and soil behaviour

Rainfall amounts and distribution have implications for soil behaviour and hence landslide occurrence. The role of soil antecedent moisture and effects on soil pore water pressure are therefore important considerations in the analysis of landslide occurrence. An analysis of 35-year (1980-2014) rainfall records was undertaken. Rainfall amounts, distribution and implications to landslide occurrence is illustrated in Figs. 4.17, 4.18 and Table 4.6. Monthly distribution shows March, April, October and November as the wettest months. Seasonal rainfall distribution shows more rain is received during the MAM (352.5mm) and SON (327.8mm) seasons, while DJF (238.4 mm) and JJA (104.72mm) seasons receive less rainfall (Fig. 4.17, Table 4.6 and Appendix 3).

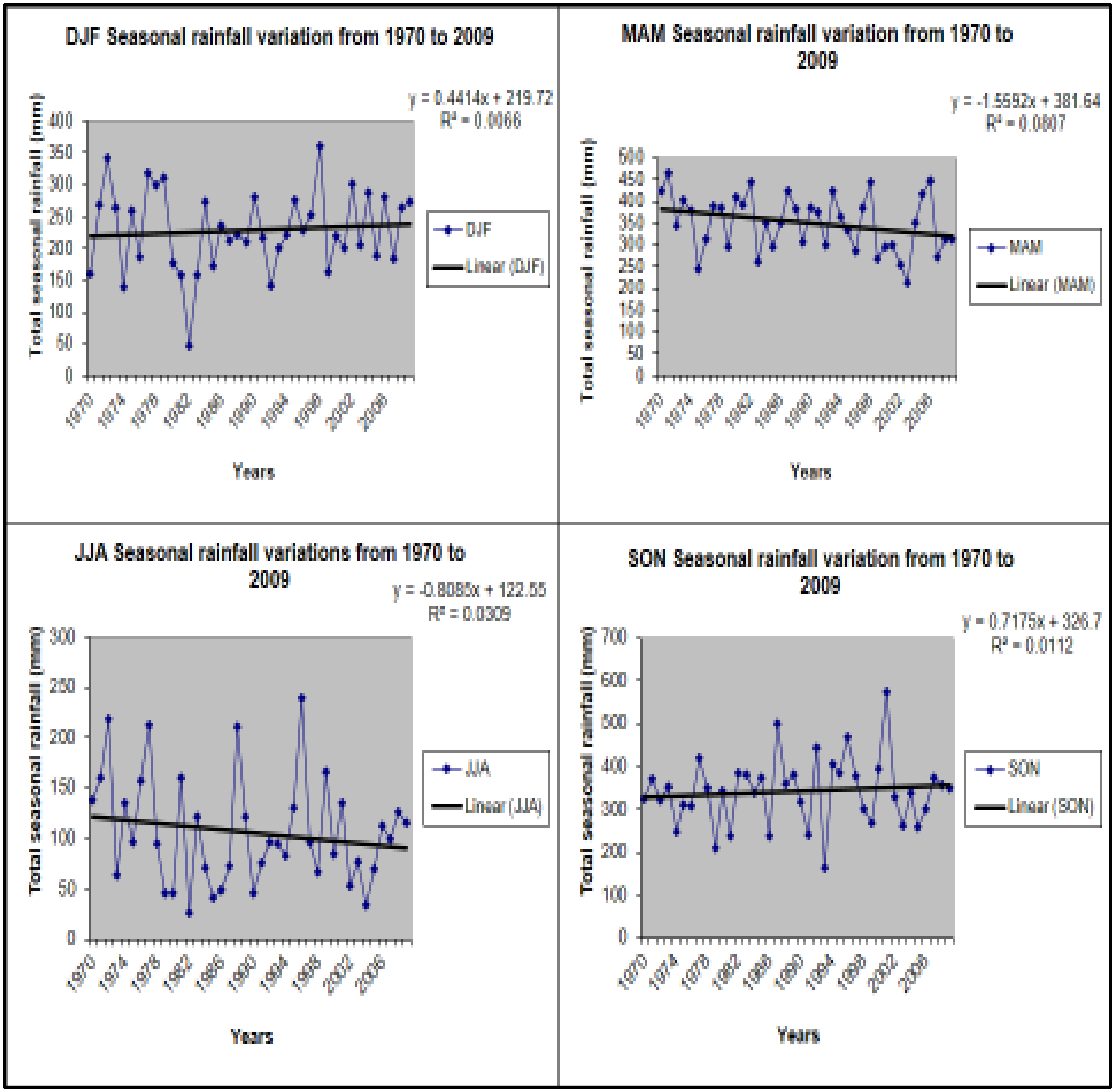

Fig. 4.17: Long-term seasonal rainfall distribution for Kigezi highlands (1970-2009). Source: Kabale Meteorology Station, weather data 2015: WMO No. 63726, National No. 91290000, station name KABALE, Elevation 1867m, Latitude 01˚ 15', Longitude 29˚ 59'.

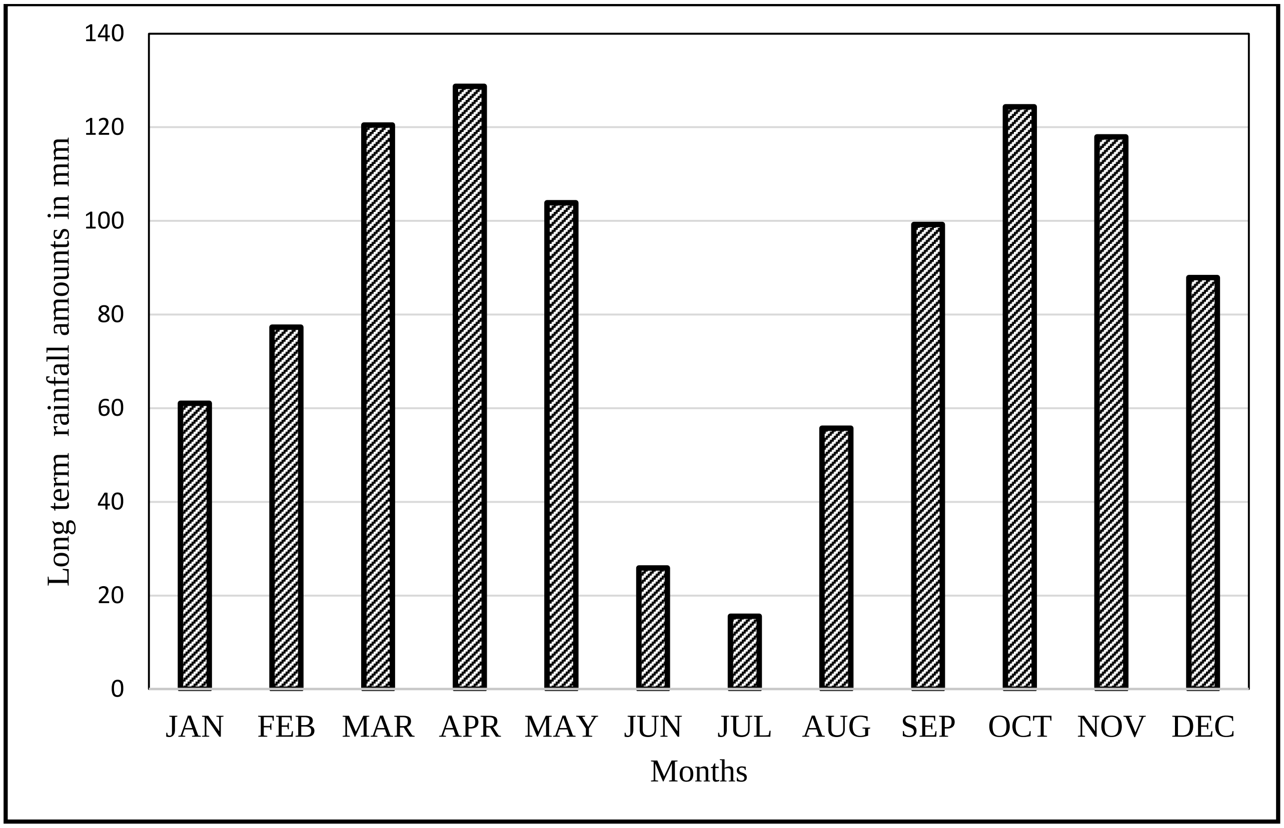

Fig. 4.18: Long term average monthly rainfall distribution for 1980 to 2014

It is noteworthy however, that landslide occurrence in the study area is not linked to individual rainfall events, but correspond with seasonal rainfall distribution. Landslides in the study area occur during the MAM and SON seasons and are not experienced in DJF and JJA seasons. There are more landslide occurrences during the MAM than the SON seasons (Table 4.6). The monthly distribution shows that more landslides are experienced during months of May and November than other months. These are, however, not the wettest months in the region. For example, in 2010, landslides occurred in the month of May despite receiving less rainfall (97.7mm) than the preceding months of March (149 mm) and April (133 mm). Similarly, in 2013, landslides occurred during the month of November which received less rainfall amounts (122. 2mm) than the preceding months of September (134.1mm) and October (154 mm) (Table 4.6, Fig. 4.18 and Appendix 3).

Table 4.6: Relationship between seasonal rainfall distribution and landslide occurrence.

YEAR	Season	Rainfall in mm	No of landslides	Month of occurrence
2001	SON	298.3	5	November
2002	MAM	253.7	6	May
2003	MAM	213.8	3	April
2004	SON	339.9	2	November
2005	MAM	415.6	5	May
2006	MAM	448.2	4	May
2007	MAM	271.8	4	April
2008	MAM	463.8	12	May
2009	SON	347.7	4	October
2010	MAM	456.7	32	May
2011	MAM	460.2	23	May
2012	MAM	454.4	9	May
2013	SON	303.1	4	November
2014	SON	354.2	2	October

Source: Kabale Meteorology Station weather data: WMO No. 63726, National No. 91290000, station name KABALE, Elevation 1867m, Latitude 01˚ 15', Longitude 29˚ 59', and Kabale District local government environmental reports (2008, 2010, and 2015).

It is therefore noteworthy that landslides in the study area do not necessarily occur in the wettest months of the year. The implications of this phenomenon will be unravelled in the discussion section. Analysed trends also reveal that landslide occurrence has been concentrated in years when there has been greater rainfall amounts. For example, more landslides were experienced during the years 2008 (12), 2010 (32), 2011 (23) and 2012 (9), which were also wetter years than the less wet 2005 (5), 2006 (4), 2007 (4) and 2009 (4) years (Table 4.6). The implications for landslide occurrence from surface soils high infiltration rates and the subsequent saturation of the clay dominant subsurface materials alluded to earlier, warrant deeper reflection, as will be discussed in the ensuing section.

4.4 Discussion

4.4.1 Soil profile characteristics

Deep soil profiles ranging between 2.5 and 7 m are a major characteristic of the study area. Deeper profiles are more pronounced along topographic hollows and valley bottoms (Table 4.2). Soil depth forms one of the conditions for assessing the stability of the soil and landslide susceptibility of the landscape (Sidle *et al*., 1985; Selby, 1993; Breugelman, 2003; Liang and Uchida, 2014). As observed by Iida (1999), the distribution of soil thickness and topography are the most important parameters for the return period of shallow landslides. The depth of soil profile and its moisture content determine how water can be stored in the soil before saturation is reached (Merino-Martín *et al.*, 2015). The response of soil moisture in deep layers differs from that in surface soil layers where moisture responds intensively and quickly, and then reaches peak values within a short period of time (Rousseau *et al.*, 2012).

Notwithstanding the deep soils (mostly >6m) on most slopes in the study area, most landslide features are shallow, occurring within less than half of the profile (1 to 3m). It is not common to find shallow landslides in areas covered by deep soils. As indicated by Kitutu *et al*. (2009), most sections covered by deep soils on the slopes of Mt Elgon in Eastern Uganda experience deep seated landslides. Shallow landslides on deep soil covered slopes in the Kigezi highlands was therefore considered as an anomaly and necessitated and investigation. Field analysis to ascertain this anomaly revealed the presence of 0.9 -3m thick clay pans within the profiles (Table 4.2). Infiltration tests revealed that clay pans reduce infiltration within the soil profile. The restriction of vertical flow of water through the soil profile by clay pans was also confirmed by Jiang *et al*. (2014). During the infiltration tests, water was observed to accumulate within the clay pans. The accumulating water leads to saturation of clay pans sandwiched between more stable materials. Saturated clay pans can act as a sliding surface for the overlying materials, consequently inducing landslides.

According to Li *et al*. (2010), understanding the spatial characteristics of clay pan soil properties is important in characterising soil behaviour. This is because clay pans create unique hydrological conditions characterized by poor drainage and very slow permeability in the soil matrix (Rawls *et al*., 1992; Yilmaz and Karacan, 2002). Reynolds *et al*. (2002), indicate that soil saturation is fast in clay pan soils due to slow permeability. This can make the soils sensitive to the incoming precipitation which can lead to reduced shear strength and therefore susceptibility of overlying materials to landslides. During a rainfall event, drainage of water through the soil profile is stopped at the boundary of clay pan, thereby causing water to accumulate (Jiang *et al*., 2014). The accumulating water leads to the increase in pore-water pressures in the soil material (Gillot, 1986; Brady and Weil, 2010). This phenomenon results in a semi-solid subsurface soil material that easily flows or slump under pressure from the top soil and accelerated by the slope gradient (Day, 1994; Yalcin, 2007). This explains why, despite having deep soil profiles of greater than 6m on most ridges, most landslide scars are shallow with a depth range of 1 to 3m.

A variation in clay pan depth with slope position was identified. Whereas the upper slope sections have clay pan layers located at a depth between 0.9 and 1.1m, they range between 1.5 and 3 m (Table 4.2) on the lower slopes. This is in agreement with studies by Jiang *et al*. (2014) and Liang and Uchida (2014). Clay pan horizons in the soil profile depend on erosion and deposition processes in the landscape (Jamison *et al.*, 1968; Jiang *et al.*, 2014). This explains why clay pans are close to the surface in the upper slopes with more erosion processes. In the lower slope sections with more depositional processes, clay pans are located deep in the profile. Several studies show that soil water holding capacity is greatly influenced by variations of clay pan profile properties over the landscape (Yilmaz and Karacan, 2002; Xiang *et al*., 2007; Ahmed *et al.*, 2011). Other studies observe that clay pans can lead to low saturated hydraulic conductivity of the soil material (Zhou *et al*., 2010; Zinck, 2013). This can cause perched water in the surface horizon creating a higher susceptibility to surface compaction. It can be inferred that the occurrence and characteristics of landslide processes in Kigezi highlands is highly influenced by the presence and position of clay pan horizons in the soil profile.

4.4.2 Particle size distribution

Fine-grained silt and clay soils (Fig. 4.3) are predominant in the study area, since more than 50% of the analysed soil passed through the 0.075mm sieve. Clay dominates, with more than 40 %, while sand and silt are less than 35 % and 25 % respectively in most samples. Owing to the high clay content greater than 35% on average and plasticity index (PI) of 33.3 %, the soils in the study area are categorised as vertisols, which are known for inducing landslides (Van Der Merwe, 1976; Mugagga *et al*., 2011). Vertic soils characteristically expand when wet and shrink in dry conditions due to the high clay content in them (Mukasa-Tibandeke *et al*., 2015). Particle size distribution is a fundamental soil property that affects many processes in soils (Selby, 1993). Many empirical relationships have been developed to relate particle size distribution to other soil properties, such as hydraulic conductivity and water retention characteristics (Bhudu, 2000; Brady, 2010). The presence of large amounts of clay in the soils of the study area is a major factor in landslide occurrence, since it affects the stability of the soils when wet. Clay has a great water-holding capacity, both upon the surface of the particles and within the cellule of the colloid (Mukasa-Tibandeke *et al.*, 2003; Mukasa-Tibandeke *et al.*, 2016). The absorption of water results in an increase in soil volume, frequently approaching 20 percent (Bhudu, 2000). Volume changes give clay soil a high shrink/swell potential, leading to failure. Studies by Corominas and Moya (1999) and Guzzetti *et al*. (2006a), show that fine-grained soil slopes become susceptible to landslide occurrence even under unsaturated state. This is due to the loss of matric suction, leading to a decrease in shear strength, as explored in the subsequent sub-section on clay mineralogy. Brady and Weil (2010), observe that soil properties including water retention characteristics and hydraulic conductivity depend on particle size distribution. Likewise, Kitutu *et al.* (2009), also observe that the susceptibility to landslide occurrence in Bududa District on the slopes of Mount Elgon is due to the abundance of fine-grained materials in the subsurface.

According to Mukasa-Tibandeke *et al.* (2015), clay is an abundant, naturally occurring, fine-grained mixture of minerals composed predominately of hydrous aluminium silicates. Due to its complex nature, clay presents problems to geotechnical engineers and that is why its properties have been investigated for their effects on landslide occurrence (Day, 1965; Bell, 2004; Mugagga *et al*., 2011). A 10% clay threshold has been used as an indicator of the expansion potential, whilst 32% clay content exhibits extreme expansion potential (Van Der Merwe 1976; Yalcin, 2007; Yalcin, 2011). Soils with such high clay content are highly vulnerable to landslide occurrence. Several studies elsewhere have demonstrated the influence of high clay content on landslide occurrence (e.g., Knapen *et al*., 2006; Xiang *et al*., 2007; Kitutu *et al*., 2009; Wati *et al*., 2010 and Mugagga *et al*., 2011). Yalcin (2007) and Broothaerts *et al*. (2012), observe that high clay content in the soil is an important precondition for landslide occurrence, due to its chemical and physical properties. Studies by Jadda *et al*. (2009) and Yalcin (2011), also reveal the susceptibility to landslides by fine-textured clayey soils due to their small pores that release water gradually. According to Zung *et al.* (2009), the vulnerability of such fine-textured clayey soils to landslides is exacerbated by their low permeability.

4.4.3 Clay mineralogy

XRD Clay mineral analyses indicated the presence of moderately expansive clays, particularly illite/ muscovite. Despite the absence of extremely expansive smectite, the dominance of illite influences the stability and susceptibility of slope materials to sliding (Skempton, 1985; Suzuki *et al.*, 2007; Van Den Eeckhaut *et al*., 2013). Previous studies indicate that the presence of illite clays can lead to landslide occurrence due to their swelling potential and low shear strength (Ohlmacher, 2000; Yalcin, 2007; Yalcin, 2011). Dhruba (2000) observes that the increasing occurrence of slumps in the Himalayas was due to the dominancy of illite and kaolinite clay minerals in the soils. In the same vein, Kitutu *et al*. (2009), also noted that landslide occurrence on slopes of Mount Elgon in Eastern Uganda is due to the presence of kaolinite and illite clay minerals. The presence of considerable amounts of illite/muscovite clay minerals in the study area also reveals Vertisols with high shrink-swell characteristics.

It has been reported by Inganga *et al*. (2001) and Kitutu *et al*. (2009) that the shear strength and swelling properties of different clay minerals differ variably and have a great significance on landslide occurrence. According to Ohlmacher (2000), illite and montmorillonite clay minerals have lower shear strength and higher swelling potentials, and are more susceptible to slope failure than other clay types like kaolinite. Soils containing illite are more susceptible to landslides than those containing kaolinite and chlorite (Yalcin, 2011). In illite clays, the properties of plasticity, bond strength, and shrinkage vary from low to moderately high (Yilmaz and Karacan, 2002). Illite/muscovite has a basic structure consisting of a sheet of gibbsite between and combined with 2 sheets of silica. In the silica sheet there is partial substitution of silicon by aluminium (Mukasa-Tibandeke *et al.*, 2015). The combined sheets are linked together by relatively weak bonding due to non-exchangeable potassium ions held between them (Mukasa-Tibandeke *et al.*, 2016). For this reason, clay containing more illite is weaker and susceptible to landslide occurrence than clay containing kaolinite (Yalcin, 2011).

Clays composed of illite have attractive forces on their surfaces (Day, 1994; Husein *et al*., 1999; Mukasa-Tibandeke et al., 2003). Just like clays composed of smectite, conditions are proper for the development of water films surrounding the flakes and for the existence of the plastic state when the clay is worked with water (Li *et al*., 2006; Li *et al.*, 2010). Although illite clays do not possess very high plasticity like smectite clays (Mukasa-Tibandeke *et al.*, 2015), some types of illite clays however, have very much greater plastic properties and bonding strength approaching those of montmorillonite (Moore, 1991; Owliaie *et al*., 2006; Mukasa-Tibandeke *et al*., 2016). Illite clays therefore have low strength, can absorb water and therefore expand resulting into loss of cohesion which can lead to landslides.

Several studies point out the influence of clay mineralogy on shear and frictional resistance of the soils (Xiang *et al*., 2007; Li *et al.*, 2010; Yalcin, 2011). Soil resistance to deformation is lowered by the presence of swelling clay minerals which impose high plasticity on soils (Bagoora, 1998 Suzuki *et al.,* 2007; Vagen, 2010; Walsh and McDonnell, 2012). The timing of landslide events during rainfall seasons depends on the behaviour of the soil materials. At the beginning of the rainfall season, there is rapid water flow through the top soil dominated by quartz into the subsurface soil with clay abundancy. The abundant illite/muscovite clay minerals in the sub soil absorb the incoming water. This leads to accumulation of moisture as the rainfall season progresses, resulting into building up of pore water pressure in the voids (Morgan, 1993; Bagoora, 1997; Morgan, 2009). The accumulating water in the soil leads to change in the soil behaviour which can swell and loose cohesion (Morgan, 1993; Van Den Eeckhaut *et al*., 2013). This explains why landslides in the study area are not experienced at the beginning of the rainfall season or immediately after extreme rainfall events, as is the case with Mt Elgon region in Eastern Uganda. Landslides in the study area are normally experienced after continuous rainfall in the season resulting into antecedent soil moisture build-up and eventual saturation.

Most landslides in the study area are experienced during the months of May and November, despite the preceding months of April and October receiving more rainfall. This time lag in landslide occurrence in the region can be explained by the initial infiltration through quartz dominated upper soil layers, before illite/muscovite clays in the lower soil horizons get saturated. The illite/ muscovite clays present do not immediately change in character due to incoming water but respond with time after accumulation of moisture. The accumulating moisture leads to saturation of illite/ muscovite clay dominated materials. This lowers the soil shear strength, leading to landslide occurrence (Van Den Eeckhaut *et al*., 2013). In summary, the presence of high amounts of illite/muscovite clay minerals capable of accumulating water over time is one of the key factors influencing landslide occurrence in the study area. The abundancy of high clay content which is greater than 35% on average renders the soil vertic. Vertic soils are known for inducing landslides because of affecting both cohesion and permeability (Van Der Merwe, 1976; Mugagga *et al.*, 2011).

4.4.4 Soil dispersion

Soils of the study area have high plasticity and are inorganic in nature (CH), indicating weak soils with high saturation (Fig. 4.5). The plasticity index which is greater than 30% signify vertic soils. Such soils easily slide, especially during continuous rainfall. Plasticity is influenced by the void ratio and is high in inorganic clays and low in organic clays (Fauziah *et al*. 2006; Yalcin, 2007). An increase in the void ratio leads to an increase in liquid and plastic limits, as indicated by the Atterberg limits and index properties of the soil (Table 4.4). The role of Liquid Limits (LLs) in characterizing the problem nature of soils has been reported by various scholars (Fauziah *et al.*, 2006; Kitutu et al., 2009; Mugagga *et al.*, 2011; Zinck, 2013). LLs for all samples analysed was above 50%, signifying high plasticity. Studies by Isik and Keskin (2008) show positive correlations between high plasticity and fine-grained inorganic clay and silts. Properties of plasticity and cohesiveness are displayed by silts and clays where a lump of soil can have its shape remoulded without breaking up or the soil volume changing. A clay or silt based soil is described as being in a liquid state when the moisture content increases and it becomes stickier and softer until it can no longer retain its shape (Fauziah *et al*., 2006; Zinck, 2013).

The dominance of illite in the study area was discussed in the foregoing subsection. The resistance of soil to deformation is lowered due to high plasticity resulting from the presence of swelling clay minerals such as illite in the soils (Zung *et al.*, 2009). The shear strength of materials is lowered due to higher plasticity resulting from absorption of water in clayey soils during heavy rainfall events (Zung *et al.*, 2008 ;). It was identified that the average weighted plasticity index (PI_w) and expansiveness were 26.4% and 32.8% respectively. By implication, soils in the study area are highly dispersive. Dispersive soils behave as single grained, very fine particles rather than as a cohesive mass like clay (Braja, 2011). As single grained with very fine particles, these soils have almost no resistance to erosion and have low shear strength (Fauziah *et al*. 2006; Keller and Dexter, 2012). Dispersive soils are structurally unstable in water due to their chemistry and collapse or disperse to form dissolved slurry when in contact with water (Bell and Maud, 1994; Bell, and Culshaw, 2001). This can interfere with the structural stability of the soil, making it highly prone to slope failures (Zung *et al.,* 2009). They are therefore excessively susceptible to landslide occurrence (Li *et al*., 2010; Yalcin, 2011).

Landslide occurrence in the study area is therefore associated with expansive soils which shrink and swell leading to loss of soil strength. The expansive potential of the soils in the study area is influenced by the high clay content and type especially illite/muscovite. Williams *et al*. (1985) and Yalcin (2007), also report on the influence of clay percentage and minerals on the expansive potential of any particular soil. In summary, clay soils dominated by illite/ muscovite minerals are promoting landslides in Kigezi highlands. This is due to their absorption of great quantities of water, slow drainage, high expansion potential, great shrink/swell potential, high plasticity, and subsequent loss of shear strength.

4.4.5 Shear strength

Field investigations revealed that the study area is covered by colluvium materials mainly along topographic hollows, which characteristically have low shear strengths in terms of their cohesion and internal frictional angles. The average cohesion for all tested soil samples in the present study is 7.2 kPa, while the average angle of internal friction is 4.1˚ (Table 4.4). The lowest cohesion of 5.2 kPa and the highest of 10.1 kPa, were detected in soils completely saturated with water. Such weak soils are susceptible to sliding especially after disturbance by natural and/or anthropogenic processes.

Clay which is the dominant material in the study has very low permeability and leads to building up of pore water pressure. The development of pore water pressures can push particles apart (Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; Lopez-Davalillo *et al.*, 2014), which acts against the normal stress, effectively reducing it. Since particles are pushed apart, cohesion and friction are also reduced (Morgan, 2009). Whereas negative pore water pressure will increase both cohesion and friction and therefore strength, positive pore water will reduce both cohesion and friction, decreasing strength (Selby, 1993; Zinck, 2013). Increased water content during a rainfall season also increases the weight of the soil and raises the water tables. This phenomenon can increase the shear stress, consequently decreasing shear strength, inducing landslides (Lopez-Davalillo *et al.*, 2014). It has also been reported that soil cohesion increases with soil water content, reaching a peak value and thereafter, decreases with further increases in soil water content (Brady, 2010; Broothaerts *et al.*, 2012). The magnitude of this variation is determined by soil texture and soil bulk density (Selby, 1993; Inganga *et al*., 2001).

The shear strength of a soil is a function of cohesion and the angle of internal friction, which is low in clay soils (Mukasa-Tibandeke *et al*., 2016). Cohesion is a function of bulk density, clay content, clay mineralogy, and is inversely related to moisture content (Morgan, 1993; Zung *et al*., 2009). The dominance by clay soils in the study area, which have very low strength parameters when saturated as indicated by shear strength parameters (c and ϕ) in Table 4.5, has strong implications for landslide occurrence. According to Zezere *et al*. (2005), during rainstorms, the majority of landslides occur due to increased pore water pressure. Increase in water pressure leads to a reduction in shear resistance and effective stress (Kitutu *et al.*, 2009). When dry, clays can be very firm and stable but become much weaker when they absorb water (Yalcin, 2011). They are capable of slow internal deformation and then can flow like a viscous liquid. Prolonged saturation of clay layers can therefore cause a progressive reduction in shear strength. It can also be seen from the present study results that the friction angle (ϕ) decreases with an increase in plasticity index. The value of ϕ generally decreases from about 8.1˚ with a plasticity index of about 22.4, to about 2.5˚ or less with a plasticity index of about 44.2 (Table 4.4). The presence of water in the void space of fine-grained soils like illite clay can have a major impact on the behaviour of the soil (Yashar *et al*., 2013). As moisture content of a soil increases, the soil changes from a brittle solid to a plastic solid and eventually to a viscous liquid (Bagoora, 1998; Yalsin, 2007; Zinck, 2013). This behavioural change in the soil material due to moisture content is a major trigger of landslides in Kigezi highlands.

4.4.6 Soil water infiltration

Soil water infiltration was noted to vary across slope position and topographic configuration (Fig. 4.16). This variation is due to the differences in soil depth and location of clay pans within the slope profile. The upper slope sections and spur slopes experience low infiltration rates due to shallow soils in such sections. The lower slope sections and topographic hollows are associated with high and very high infiltration rates due to the deeper soil profiles. This variation in infiltration rates also signifies differences in soil saturation levels (Rousseau *et al.*, 2012). Whereas the upper slope sections and spur slopes are dry due to low infiltration rates, the lower slope sections and topographic hollows experience high saturation rates. This is due to increased infiltration in such sections, as indicated by the high topographic wetness index values presented in Chapter 3 on topographic parameters. High saturation rates in the lower slopes lead to saturation overland flow processes which move from the slope base incrementally along topographic hollows upslope (e.g., Reynolds *et al.*, 2002). In association with the high clay content dominated by illite/muscovite minerals, as discussed in the previous sub-sections, materials along topographic hollows remain saturated most of the time. This results into reduced shear strength within the topographic hollows, facilitating landslide occurrence.

A relationship was identified between soil water infiltration and topographic parameters in the landscape, including slope gradient, position and curvature. The rate of infiltration decreases with an increase in slope gradient. Slope sections with lower gradients are associated with high infiltration rates than steeper slopes. This is consistent with observations elsewhere by Stolte (2003). Other studies however, show an increase in infiltration rates with increase in slope gradient (Wenck Associates, 2008; Bamutaze *et al*., 2010; Philips and Kitch, 2011; Walsh and McDonnell, 2012). In the study area, infiltration increases with reduced slope gradient due to increase in soil depth as explained earlier. The deeper soils in the lower slope sections require more water to accumulate before reaching saturation, hence more infiltration. The variation in infiltration with slope gradient is also due to the location of clay pans within the soil profile. Whereas clay pans are near the surface (0.9 to 1.2m) on the steep upper slopes, the converse is true with the lower slopes (>2.5m). As indicated in the previous sub-section, clay pans restrict vertical water movement in the soil profile and therefore reduce on infiltration. This explains why steep upper slope sections experience low infiltration rates because they reach saturation faster due to the presence clay pans near the surface.

A relationship between soil water infiltration with slope position and topographic configuration was also identified. Lower slope elements and topographic hollows experience higher infiltration rates than the upper elements and spurs. Such differences in infiltration rates in relation to slope position and topographic configuration are related to variations in soil depth in the study area. Lower slope elements and topographic hollows are characterized by deeper soil profiles than the upper and spur slope counterparts, as noted earlier in the sub-section on soil profiles. The former are convergence zones for water and eroded materials from the latter. Materials accumulating in such zones lead to deep soil profiles. Such deeper slope profiles are capable of accumulating more water before saturation is reached, hence high infiltration rates. Merino-Martin *et al*. (2015), observe that soil properties such as pore space, texture, structure, soil depth and antecedent moisture condition are among the important factors that influence infiltration rate. The present study confirms that it is possible to estimate landslide occurrence using soil infiltration experiments along the slope profile.

4.4.7 Rainfall distribution and soil behaviour.

Landslide occurrence in the study area is not correlated with extreme rainfall events as is the case with Mt Elgon region in Eastern Uganda. An extreme rainfall event occurs whenever more than normal rainfall amount is received in a given area within a single event (Godi *et al*., 2006). In the Mt Elgon region landslides are usually experienced during or immediately after extreme rainfall events (Knapen, 2003; Knapen *et al*., 2006; Kitutu *et al*., 2009; NEMA 2010 and Mugagga *et al*., 2011). It was confirmed from local communities and district environmental reports that landslides in the Kigezi highland region are normally not experienced during or immediately after peak rainfall periods. Paradoxically, they occur during the less wet months of the rainfall season. This phenomenon can be explained by the unique infiltration dynamics through quartz dominated top soil layers and saturation of the clay pans, as discussed earlier. This leads to antecedent moisture building up in the sub soil materials as more rainfall is received. Soil saturation leads to loss of cohesion, hence landslide occurrence in the region. Several studies also observe that antecedent soil moisture condition prior to a rainfall event is the most significant factor in landslide occurrence (Bagoora, 1997; Polemio and Sdao, 1999; Panek *et al*., 2011; Guan-Wei and Hongey, 2012).

4.5 Conclusion

Deep soil profiles ranging between 2.5 and 7 m are a major characteristic of the study area. Notwithstanding the deep soils on most slope elements in the study area, most landslide features are shallow, occurring within less than half of the profile due to presence of clay pans which reduce infiltration within the profile, leading to saturation. The saturated clay pans act as a sliding surface for the overlying materials, inducing landslides. Fine-grained soils of silt and clay are predominant in the study area. Owing to the high clay content greater than 35% on average and PI of 33.3 %, the soils in the study area are categorised as vertisols, associated with landslide occurrence. The study area has moderately expansive clays, particularly illite/ muscovite which influences the stability and susceptibility of slope materials to sliding. The soils have high plasticity and are inorganic in nature (CH), indicating weak soils with high saturation rates and can slide, especially during continuous rainfall. They are highly dispersive and are therefore excessively susceptible to landslide occurrence. Soil water infiltration varies across slope position and topographic configuration. Lower slope elements and topographic hollows experience higher infiltration rates than the upper elements and spurs. In association with the high clay content dominated by illite/muscovite minerals, materials along topographic hollows remain saturated for longer periods. This results into reduced shear strength, facilitating landslide occurrence. The timing of landslide events during rainfall seasons depends on the behaviour of the soil materials. Landslides in the study area are not normally experienced during or immediately after extreme rainfall events but occur later in the rainfall season. This time lag in landslide occurrence is due the initial infiltration through quartz dominated upper soil layers, before illite/muscovite clays in the lower soil horizons get saturated. This leads to antecedent moisture building up in the sub soil materials as more rainfall is received resulting into loss of cohesion. This behavioural change in the soil material due to moisture content is therefore a major trigger of landslides in Kigezi highlands.

CHAPTER FIVE

Implications of land use and cover changes for landslide occurrence in Kigezi highlands of South Western Uganda.

Abstract

The present study considers the influence of human-induced changes on the occurrence of landslides in Kigezi highlands of South Western Uganda. An analysis of the spatial-temporal land-use and cover changes was undertaken using satellite images spanning 1985 to 2015. In this study, Landsat imagery data used for land-use and cover change analysis included; Landsat 5TM, 7ETM+ scenes, Landsat 8 OLI/TIR of 30m spatial resolution scene path 173 and row 061 which covers the area adequately. Post classification change detection technique was applied to identify the dynamic land cover elements from the successive years of satellite data. Five land-cover categories were identified namely; cultivation, forests, grasslands, wetlands and settlements. Whereas forest cover reduced from 40 % in 1985 to 8 % in 2015, cultivated land and settlements increased from 16% and 11% to 52% and 25% respectively during the same period. The distribution of cultivated land decreased in lower slope sections within gradient group < 15˚ by 59%. It however increased in upper sections within gradient cluster 25˚ to 35˚ by over 85% during the study period. There is a shift of cultivated lands to the steeper sensitive upper slope sections associated with landslides in the study area. Out of the 65 landslide features mapped, 54% occurred on cultivated areas, 26% on settlements, 14 % on grassland, and only 6% occurred on forests with degraded sections. A close spatial and temporal correlation between land-use/cover changes and landslide occurrence is therefore discernible. It is recommended that, tree cover restoration be done in the highlands and the farmers encouraged to re-establish terrace farming while avoiding cultivation of sensitive steep middle and upper slopes.

Keywords: land-use/ cover changes, landslides, Kigezi highlands

5.1 Introduction

Landslide occurrence is expected to increase in the near future given the current land pressure in tropical highland regions with increasing occupation of steep uplands (Broothaerts *et al*., 2012; Kirschbaum and Zhou, 2015; Gu and Wylie, 2016). Understanding the factors that control land use patterns in a region susceptible to landslide occurrence is therefore essential. Most landslide studies have focused on quasi-static factors such as geology, topography and soil strength properties which do not change in the considered time frame (Dai and Lee, 2002; Ohlmacher and Davis, 2003; Ayalew and Yamagishi, 2005; Claessens *et al*., 2007). Less attention has been focused on the effect of triggering variable factors including land-cover changes and their implications for hillslope hydrology (Ali *et al.,* 2014**)**.

Land cover change is one of the crucial factors that influence the spatial distribution of landslide processes (Ayalew *et al.,* 2004; Begueria, 2006; Mugagga *et al.,* 2012). The conversion of forests and natural grasslands to agriculture and pasture in highland environments is on the increase in developing countries (Stolte, 2003; Breuer *et al.,* 2009; Liesbet *et al.,* 2015). Several authors note that natural processes and anthropogenic activities that lead to land cover changes are continuously experienced in highland and mountainous regions (Breyer *et al*., 1997; Arinaitwe, 2004; Bamutaze, 2005; FAO, 2010). Studies by Prompter and Glade (2012), and Roller *et al*. (2012), point out that serious environmental impacts result from the modifications and conversion of land cover. Severe land sliding has been reported on Mt Kilimanjaro due to replacement of forests by agriculture and settlement (Breugelmans, 2003). Several studies have shown that the initiation and reactivation of landslides is due to the impact of human activities on the environment (Cruden and Miller, 2001; Meusburger and Alewell 2008; Van Den Eeckhaut *et al.,* 2013). According to Knapen *et al*. (2006), population pressure on Mount Elgon slopes forces people to cultivate unsuitable steep slopes, thus contributing to slope instability.

Human activities are a major driver of global environmental change (Glade 2003; Ives, 2004; Karsli *et al*., 2009; NEMA, 2014). According to Ferreira *et al*. (2015), forest logging, burning and cultivation on hillslopes are the most important contributing factors for landslide occurrence. Forest logging, burning and cultivation on hillslopes are the most important contributing factors for landslide occurrence (Jacob, 2000; Promper and Glade, 2012). Several studies point out that the effects of hydrology and mechanism of slope failure are influenced by land cover changes (Glade, 2003; Petley *et al*., 2005; Beguería, 2006; Petley, 2008). Changes in vegetation cover pattens often result in increased landslide occurrence due to its effects on soil behaviour (Promper and Glade, 2012; Liesbet *et al*., 2015). Some authors observe that changes in land use and cover are an important trigger for landslide occurrence (Glade, 2003; Beguería, 2006; Petley *et al*., 2005; Mugagga *et al*., 2012). Lambin *et al*. (2001), therefore recommends an understanding of locations and rates of land-cover changes.

Land-cover patterns are highly dynamic and rarely in a stable equilibrium (Glade 2003; Karsli *et al*., 2009). Several investigations of landslide occurrence as a response to land cover changes have been conducted (Glade, 2003; Van Beek and Van Ash, 2004; Alcantara-Ayala *et al.*, 2006; Beguería, 2006; Mugagga *et al*., 2012; Promper *et al*., 2012). A number of studies show that there is a rapid increase in landslide occurrence after land-cover change (Glade 2003; Alcantara-Ayala *et al*., 2006; Kato and Mutonyi, 2011; Liesbet *et al*., 2015). Few studies have, however, analysed the spatial-temporal correlation between land cover change and landslide occurrence (Dai *et al.*, 2001; Van Beek and Van Asch, 2004; Karsli *et al*., 2009). Changes in land- use and cover pose a risk to slope stability of the Kigezi highlands, but the magnitude of the impact is not well understood. The present study considers the influence of land-use/cover changes on the occurrence of landslides. The role of specific topographic parameters in landslide occurrence was explored earlier. Consequently, the relationship among land-use, specific topographic parameters and landslide occurrence is explored in this chapter. Land-use/cover changes in the form of cultivated land, grasslands, forests, wetlands and settlements in Kigezi highlands of south western Uganda are analysed. It also seeks to unravel the relationship between different land-use/cover change scenarios and landslide occurrence.

The specific objectives of the study were:

1. To analyse the spatial-temporal land-use and cover changes for the period between 1985 and 2015.
2. To establish the relationships between land-use/ cover changes and the spatial- temporal distribution of landslides.

5.2 Materials and methods

5.2.1 Datasets and sources

Land-use and cover change detection was undertaken using satellite images spanning 1985 to 2015. In this study, Landsat imagery used for land-use and cover classification of Rukiga catchment included Landsat 5TM, 7 ETM+ scenes, Landsat 8 OLI/TIR of 30m spatial resolution (Table 5.1). All these images were obtained from path 173 and row 061 as indicated in Table 5.1. The satellite imagery data were all sourced from United States Geological Survey Global Visualization Viewer (GLOVIS). In addition, available aerial photographs of 1:20,000 scale for 1985, 1995, 2005 (Table 5.1) which corresponds with the Landsat images were acquired from the Uganda Department of Mapping and Surveys for purposes of detecting past landslide features in the catchment. Landslide features are not detectable from Landsat imagery, owing to spatial resolution constraints.

Table 5.1 Data sources and specifications

Dataset type	Acquisition date	Band	Pixel resolution	Path/row	Source
Satellite Data					
Landsat 5 (TM)	08/07/1985	3,4,5	30m x 30m	173/061	USGS
Landsat 7 (ETM+)	03/22/1995	3,4,5	30m x 30m	173/061	USGS
Landsat 7 (ETM+)	09/14/2005	3,4,5	30m x 30m	173/061	USGS
Landsat 8 (OLI/TIR)	10/07/2015	3,4,5	30m x 30m	173/061	USGS
Aerial photograph					
Black & white	02/04/1985		1:20000		Lands and Survey Department for Uganda
Black & white	12/05/1995		1:20000		
Black & white	01/05/2005		1:20000		
Ancillary data					
Topographic maps			1:25000		Lands & Survey Department for Uganda
Field surveys			Conducted between June 2014 and August 2016		
Government records including National and Local Environmental Reports					
Google Earth image for 2006 (June) and 2015 (October)					

5.2.2 Image processing and classification

Landsat images of 1985, 1995, 2005, and 2015 were exported to ArcGIS 10.1, geo-registered to WGS 84 datum, and projected into the Universal Transverse Mercator (UTM) zone 36N of the coordinate system. Pre-processing of downloaded images was done to ensure that a higher classification accuracy is attained. According to Zhang *et al*. (2011), it is necessary to rectify geometric distortion from raw satellite data in order to enable correct measurement of area, precise localization and multi-source data integration. Following Lillesand and Keifer (2004), image to image rectification technique was applied given that it minimizes the residual rectification error. Geo-rectification accuracy was further improved using 25 Ground Control Points (GCPs) obtained during field verification despite image pre-processing by the suppliers. Atmospheric correction was not performed because the post-classification comparison technique adopted for land-use/cover change analysis, which also compensates for variations in atmospheric conditions and vegetation phenology between dates. This is because each land-use/cover classification is independently mapped (Coppin *et al*., 2004; Yuan *et al.*, 2005).

The supervised classification method was used to classify land-use and cover patterns in the study area after field verification. Supervised classification helps in processing and quantitative analysis of remotely sensed imagery data applied after defining the specific areas of interest called training sample classes (Lu and Weng, 2007). In the supervised classification process, maximum likelihood parametric rule was used to produce better accuracy maps (e.g., Jensen, 1996). Land-use and cover changes between 1985 and 2015 in the catchment were detected by means of pixel-based classification on Landsat images. All satellite data were analysed by assigning per-pixel signatures and distinguishing the land-use categories into five delineated classes namely forests, grassland, wetland, settlement, and cultivation (Table 5.3).

5.2.3 Field verification

Field verified data, historical black and white aerial photos, 1: 25,000 topographic map sheets, and Google Earth images were used to provide reference information for both the classification and the accuracy assessment. Field verification for land use/cover classification was conducted between July 2015 and April 2016 using draft classified map derived from satellite image for 2015 as a guide. The verification was conducted for land use and cover for 2015 Landsat image. Hand held GPS receivers were used in determining the five land use/cover classes as categorically indicated in Table 5.3. The verification involved identifying 100 locations as samples including main land use/cover categories indicated. The collected field verified data were loaded onto a GIS system for validation using ArcGIS 10.1 software. The land use and cover for the 1985 and 1995 images were validated using topographic map sheets (93/1, 93/2, 93/4 and 94/3) of scale 1:25000 taken in 1986 and 1994 respectively. The topographic map sheets were acquired from the Uganda

Department of Mapping and Surveys. The 2005 image was validated using Google Earth image taken in July 2006. The topographic map sheets were scanned and imported into GIS for geo-referencing, mosaic, processing, validation and boundary demarcation of the catchment using ArcGIS 10.1.

5.2.4 Classification accuracy assessment

Accuracy assessment is an essential requirement of image classification done using the confusion matrix (James and Randolph, 2011). Accuracy is essentially a measure of how many field verified pixels were classified correctly. Confusion matrices quantitatively compare information obtained by reference sites to that provided by classified images for a number of sample areas (Congalton and Green, 2009). Accordingly, overall accuracy, producer's and user's accuracies, and Kappa coefficient were calculated from the error matrix (Jensen, 2005; Lillesand *et al.*, 2007; Lu and Weng, 2007; James and Randolph, 2011). The accuracy of the classification was verified by randomly generated reference points using a stratified random algorithm (Jensen, 1996). The field verified data were utilized in the maximum likelihood report as the independent data set from which the classification accuracy was compared. Topographic map sheets (93/1, 93/2, 93/4 and 94/3) of scale 1:25000 taken in 1986 and 1994 and Google Earth image taken in July 2006 were also digitized in Arc GIS and used in accuracy assessment. The accuracy assessments were performed for classified images of 1985, 1995, 2005, and 2015. A minimum of 25 random points were generated per class using stratified random sampling approach for accuracy assessment (Skirvin *et al*., 2004; Congalton and Green, 2009). According to Lilesand *et al.* (2007), the minimum level of accuracy in the identification of land cover categories from remote sensor data should be at least 80%. Jensen (2005) indicates that classification accuracy is considered significant if Kappa coefficient is greater than 0.70. Higher classification accuracies were obtained for all the downloaded images due to improved sensors as illustrated in Table 5.2. An overall accuracy ranging between 82.37% and 90.28% was achieved for the Landsat images, while the Kappa coefficient accuracies ranged from 0.74 to 0.85.

Table 5.2: Classification accuracy

Landsat Images	User Accuracy (%) (Recall)	Producer Accuracy (%) (Precision)	Overall Accuracy (%) (OA)	Kappa coefficient
1985	83.32	81.42	82.4	0.74
1995	85.33	84.54	85.1	0.77
2005	87.62	85.32	86.5	0.81
2015	92.24	88.32	90.3	0.85

5.2.5 Land-use/cover change detection and analysis

Land-use and cover change analysis was conducted in three temporal periods including 1985-1995, 1995-2005, and 2005-2015. Various techniques are available for change detection. According to Lu and Weng (2007), image differencing, principal component analysis and post-classification are the most commonly used methods. In the present study, the post classification change detection technique was applied to identify the dynamic land-cover elements from the successive years of satellite data. As noted from Table 5.1, the respective imagery were captured using different Landsat sensors. With such spectral differences, post-classification is appropriate for change detection. The main advantage of post-classification is that images are individually classified (Zhang *et al.,* 2011). This technique avoids the intricacies in change detection associated with analysis of imagery acquired at different times (Coppin *et al*., 2004; Lu and Weng 2007). The land-cover change analysis was performed by automatic comparison of image sub-object hierarchies (e.g., Ozesmi and Bauer, 2002; Yang *et al.,* 2007). The change detection matrix was calculated to find the proportion of each class which has undergone change during the study period. Following Zhou *et al.* (2008), multiscale multiresolution segmentation was utilized to obtain image objects of every land-use/cover class throughout the whole study area. The major land-use/cover change trends were identified from maps so generated and the level of persistence established through a cross-tabulation matrix.

Table 5.3: Land-use and cover classification

Land-use/cover classes	**Description**
Forests	Tropical, deciduous, coniferous, and plantation forests

Grasslands	Short and tall grasses, thickets, shrubs
Wetlands	Seasonal and permanent wetlands, swamps, bog, streams
Settlements	Built-up areas, residential, commercials, rural & urban non-residential, roads and other structures
Cultivated lands	Cultivated gardens, fallow lands, plantations

5.2.6 Landslide surveying and mapping

Two methods were used to identify the spatial-temporal landslide distribution in the study area. These were aerial photograph interpretation, field surveying and mapping. Aerial photograph interpretation (API) was used to identify the past landslide scars which occurred before 2005. These landslide scars are no longer visible on the landscape due to the high rates of vegetation regeneration. The landslide scars are easily concealed by soil materials mobilized from the hilltops and spur slopes into topographic hollows where landslide occurrence is dominant. The soils accumulating within the landslide scars encourages rapid vegetation regeneration, owing to the high rainfall amounts in the study area. It is only aerial photographs taken during the period when the scars were still visible, that can be used to identify and analyse such landslides. The advantages of using API in landslide investigations include rapid definition of landslide boundaries, identification of large landslides, and an appreciation of slope conditions (Rembold *et al.*, 2000). Black and white aerial photographs for three different periods including 1985, 1995, and 2005 (Table 5.1) were used to assess the location of historical landslide scars in the study area. Following Zone (2007), a mirror stereoscope with dimensions 718 x 260 x 310 mm with an optical path of 50mm which allows the viewer to magnify on areas of interest was used to analyse the aerial photographs and identify the historical landslide scars within the study area. The landslide scar details from the aerial photographs were scanned, digitized and their locations were analysed using ArcGIS software. The scanned/ digitized landslide scar locations from the aerial photographs were incorporated with the rest of the field mapped landslide scars to generate landslide distribution maps for the study area in a GIS environment.


Fig. 5.1: Land-use/cover and landslide mapping

Field surveying was undertaken to identify and map visible landslide scars in the catchment as presented in Chapter 3 on topographic parameters. Field investigations were undertaken between June 2014 and August 2016 to identify visible landslide scars in the catchment (Fig. 5.1). The landslide scars were mapped using handheld GPS recievers. Coordinates for the mapped landslide scars were imported into ArcGIS 10.1 software to produce a landslide distribution map for the study area. Through spatial analysis in a GIS environment, the landslide and land use/cover maps were overlain to derive the requisite relationships.

5.2.7 Relationship between land-use/cover changes and topography

In order to establish the relationship between land-use, topographic parameters and landslide occurrence, topographic analyses were performed with the use of a 10m Digital Elevation Model (DEM). Topographic parameterization was performed using ArcGIS 10.1 and SAGA GIS 2.3.1 software. SAGA GIS 2.3.1 software was used to calculate slope gradient, curvature, slope position classes and other topographic characteristics as presented in Chapter 3, on topographic parameters. Following Kamusoko and Aniya (2009), the topographic surfaces calculated from the 10 m DEM were resampled to the 30m Landsat images for overlay purposes, using the nearest neighbour resampling technique.

By overlaying the classified land use and cover maps for the respective reference years onto the slope gradient and position maps, the relationship between spatial-temporal land-use/ cover changes with topography was extracted in ArcGIS. Identified as the most dynamic land-use form, special attention was paid to changes in cultivated land in relation to slope gradient and position. Boolean images for cultivated land, slope gradient and position classes as presented in Chapter 3 on topographic parameters were generated using the RECLASS module in ArcGIS. They were overlaid on the slope surfaces to highlight the extent to which cultivation had changed with topography between 1985 and 2015. A compound relationship among cultivation, topography and landslide occurrence was also established through the overlay procedure in ArcGIS.

5. 3 Results

5.3.1 Land use and cover changes between 1985 and 2015

Land-use and cover changes over the period 1985-2015 are presented in Figs 5.2 to 5.6 and Appendix 10. There are five land-uses/covers observed as shown in the figures. Whereas a decrease in forestland is noticeable from 1985, cultivated land and settlements have expanded drastically.

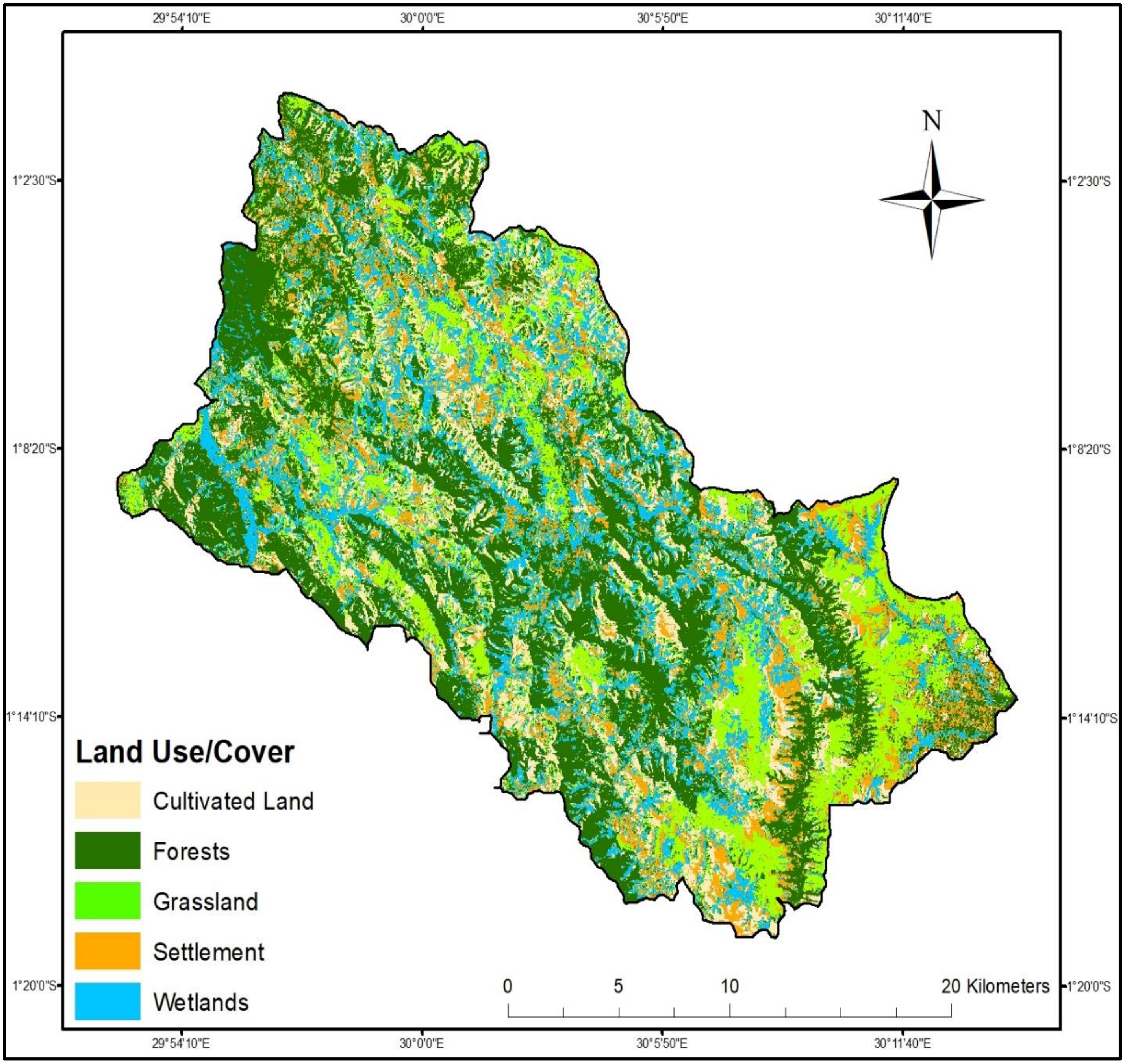

Fig.5.2: Land-use and cover distribution for 1985

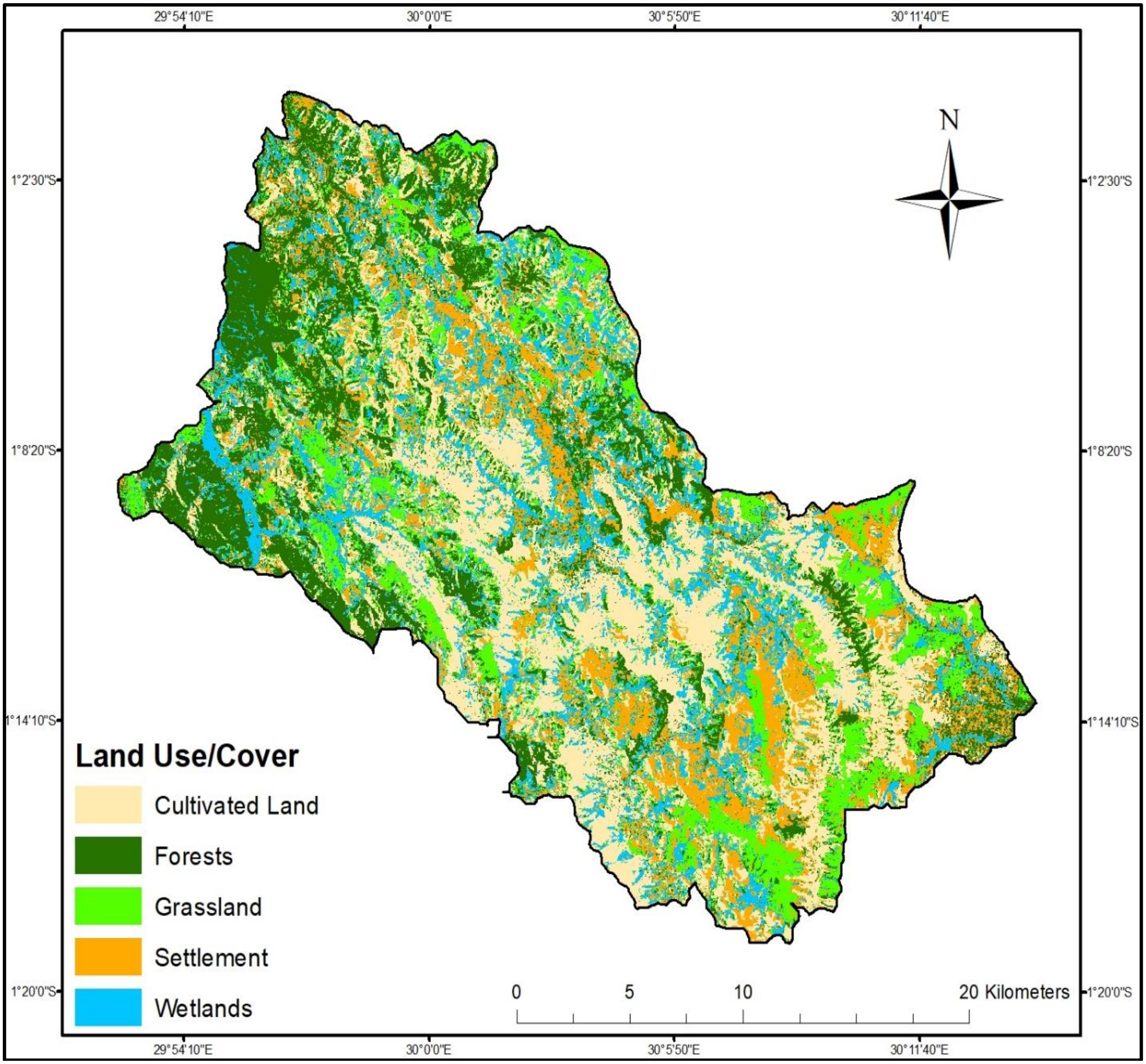

Fig.5.3: Land-use and cover distribution for 1995

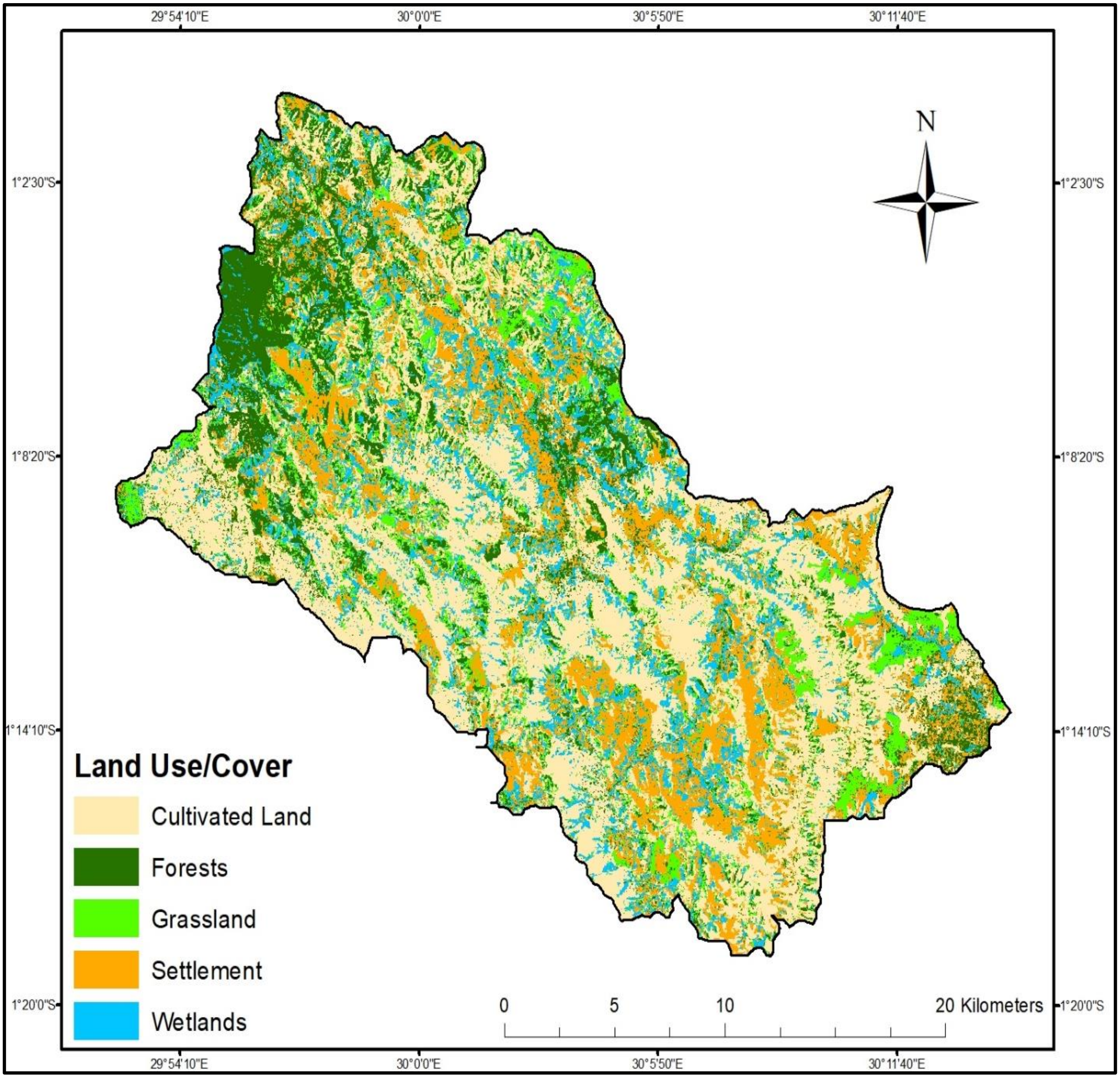

Fig.5.4: Land-use and cover distribution for 2005

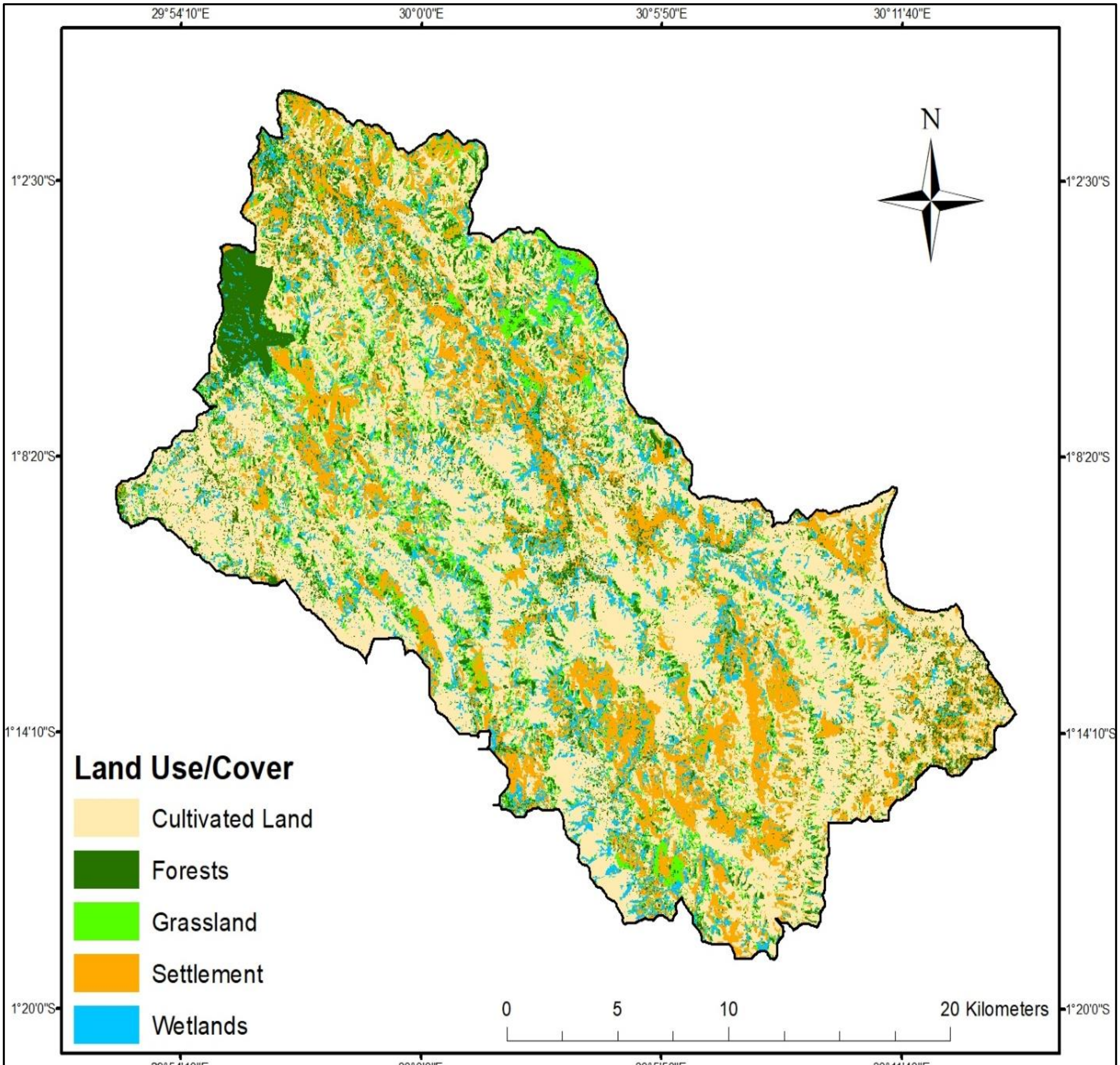

Fig.5.5: Land-use and cover distribution for 2015

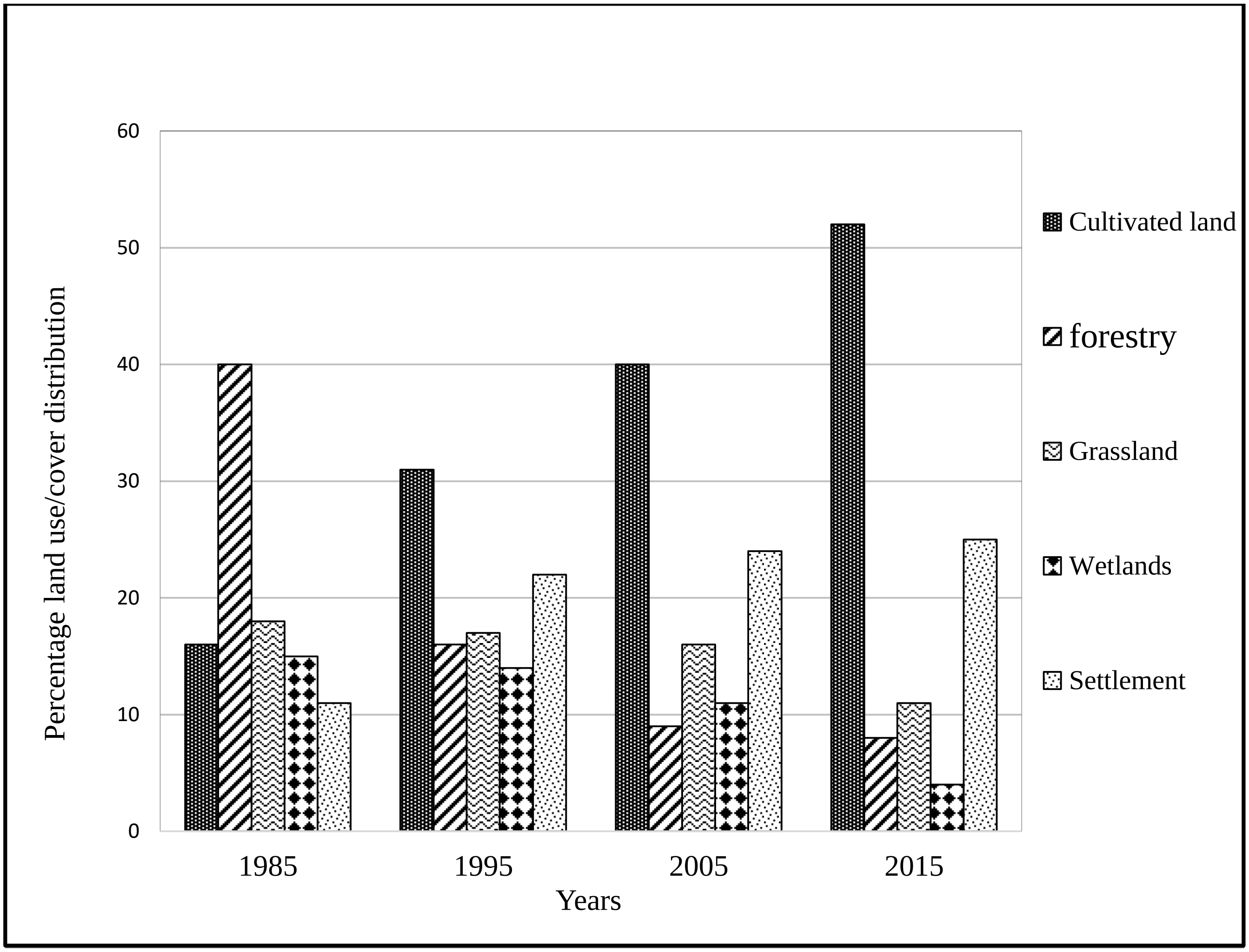

Fig. 5.6: Land-use/cover changes between 1985 and 2015

According to the land-cover change trends, forests as the dominant land-cover/ use type by 1985 spanned 40% of the total area followed by grassland at 18%. Cultivated land and settlements covered 16 % and 11% respectively (Fig. 5.6). Whereas forest land has diminished drastically by 73% between 1985 and 2015, cultivated land in particular has increased enormously by 218% over the same period. Likewise, settlements have expanded by 85% (Appendix 10). The hillslopes appear to be intensively cultivated with many settlements in the valley bottoms (See Google Earth image below, Fig. 5.7).

Fig. 5.7: Land cover distribution based on Google Earth image (October 2015).

5.3.2 Land use/cover changes and landslide distribution

The relationship between land use/cover and landslide distribution is illustrated by Figs. 5.8 and 5.9.

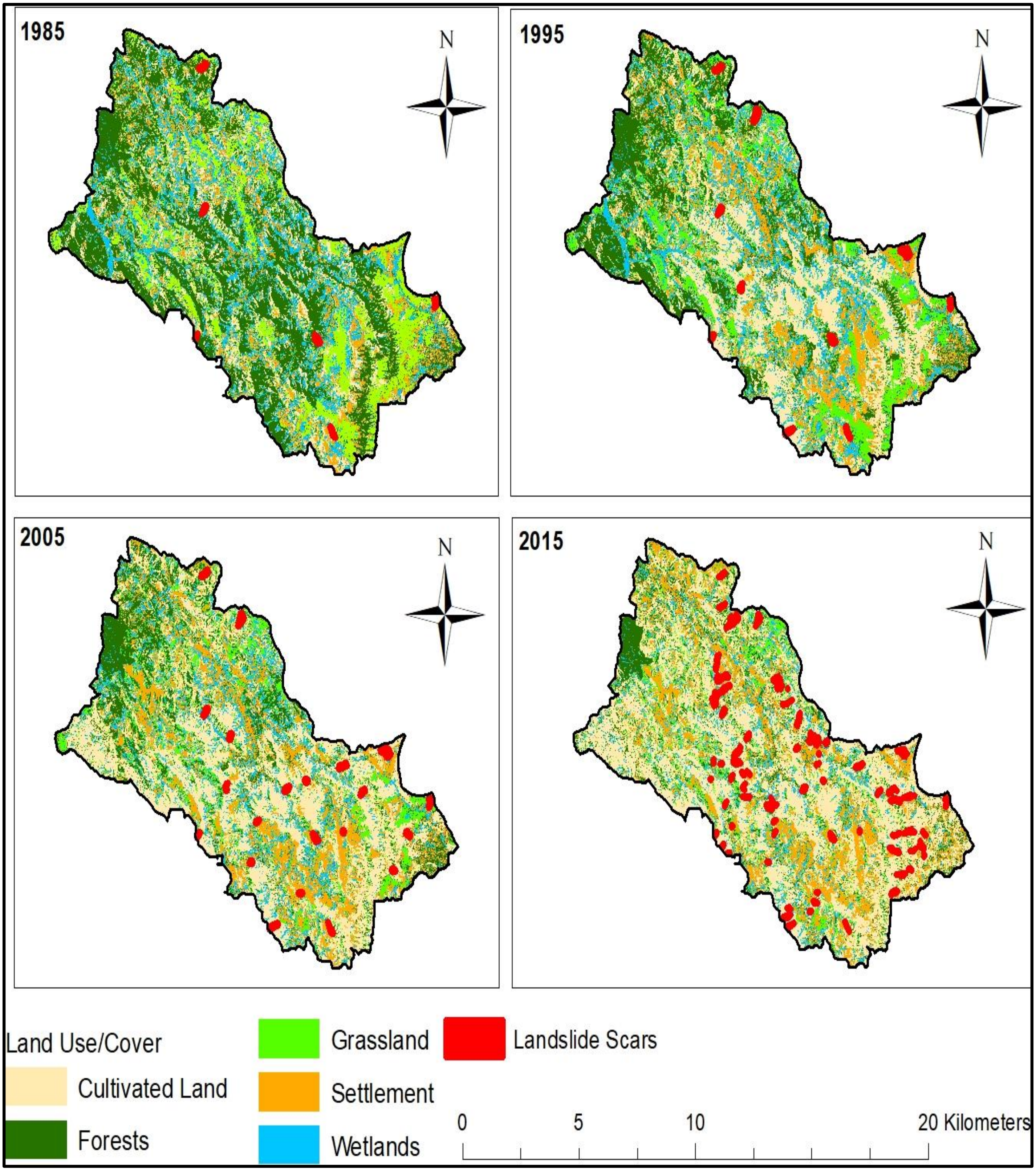

Fig. 5.8: Land use/cover and landslide occurrence distribution between 1985 and 2015.

The spatial distribution of landslides in the study area has changed (Fig. 5.8 and Appendix 13). The increasing occurrence of landslides in the study area is attributed to changes in the land use/cover distribution. This relationship is explored in the discussion section.

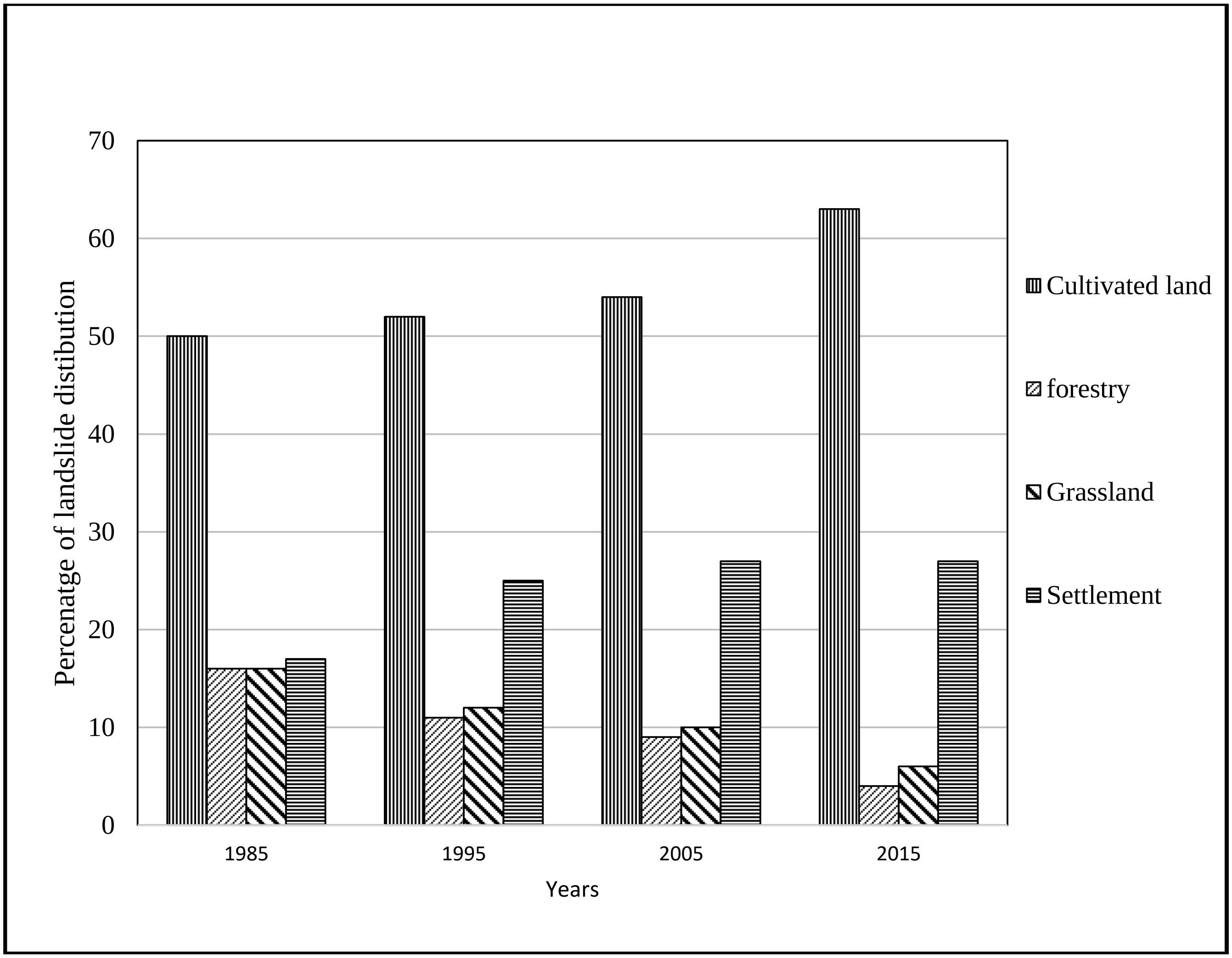

Fig. 5.9: Relationship between land use/ cover and landslide distribution.

A close spatial relationship between landslide distribution and land use/cover is discernible (Figs. 5.8 to 5.10). Most of the landslides have an association with degraded areas (Fig. 5.10). As can be noted from Figs. 5.8 and 5.9, cultivated land is the most affected land use/ cover category while forest areas dominated by woodlots experience the least landslide occurrence. More than 50% of the landslides are occurring on cultivated land, 20% on settlements while less than 15 % and 10% are occurring on grassland and forests with degraded areas respectively (Figs. 5.8 and 5.9).


Fig. 5.10: Landslide occurrence on degraded and intensively cultivated slopes.

The study area has experienced an increasing trend in landslide occurrence over the past 35 years. The temporal landslide occurrence pattern is shown in Fig. 5.11.

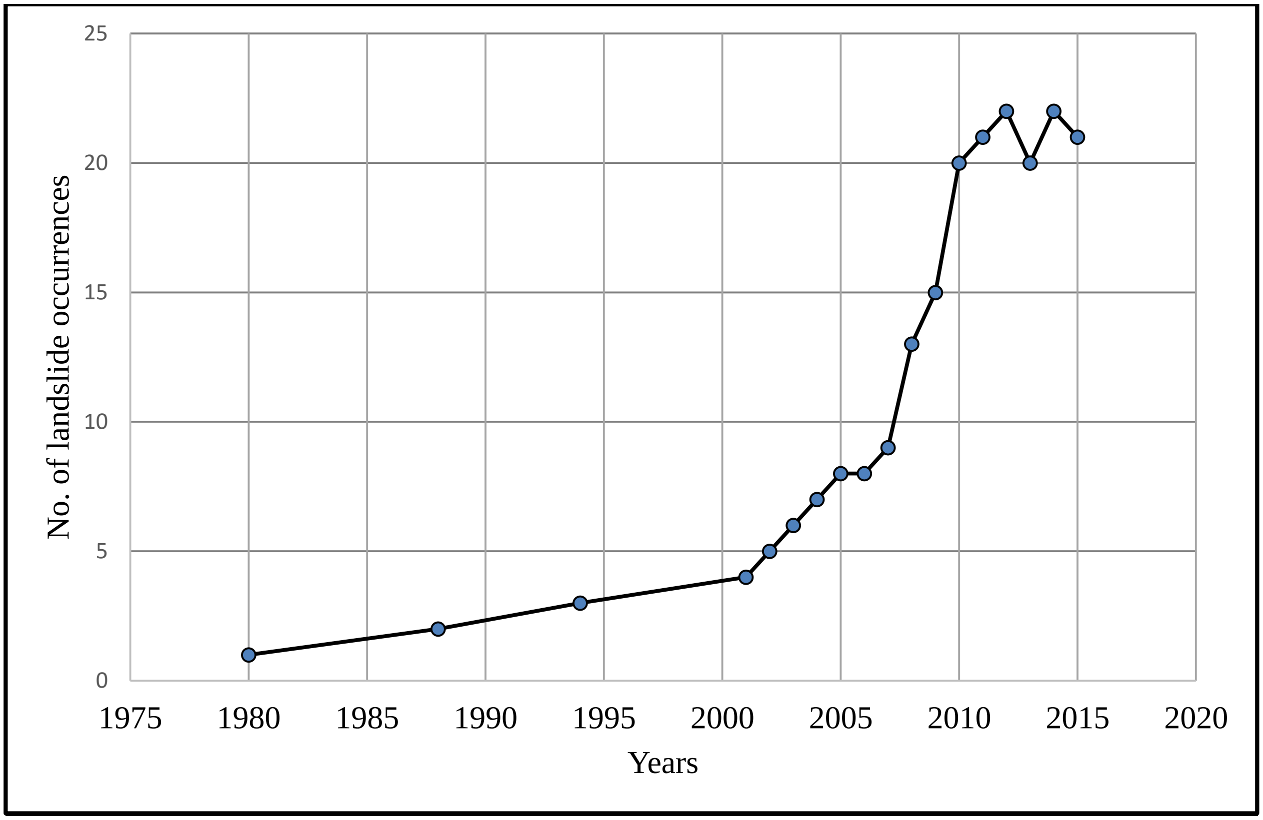

Fig. 5.11: Temporal landslide distribution in the study area.

Source; Kabale District Environmental Reports (2008, 2012 and 2015)

The overall trend shows that the study area is increasingly becoming vulnerable to landslides. Whereas the period between 1980 and 2004 experienced only 28 landslides, 180 occurrences were experienced between 2005 and 2015 (Fig. 5.11 and Appendix 14). A close relationship between temporal landslide distribution and land use/cover patterns is discernible. This relationship is explored in the discussion section.

5.3.3 Changes in cultivated land in relation to topography

A compound relationship among land use change, topography and landslide occurrence was established. The distribution of cultivated land in relation to topographic parameters is illustrated in Figs. 5.12 and 5.13. The major topographic characteristics in this relationship are slope gradient and slope position.

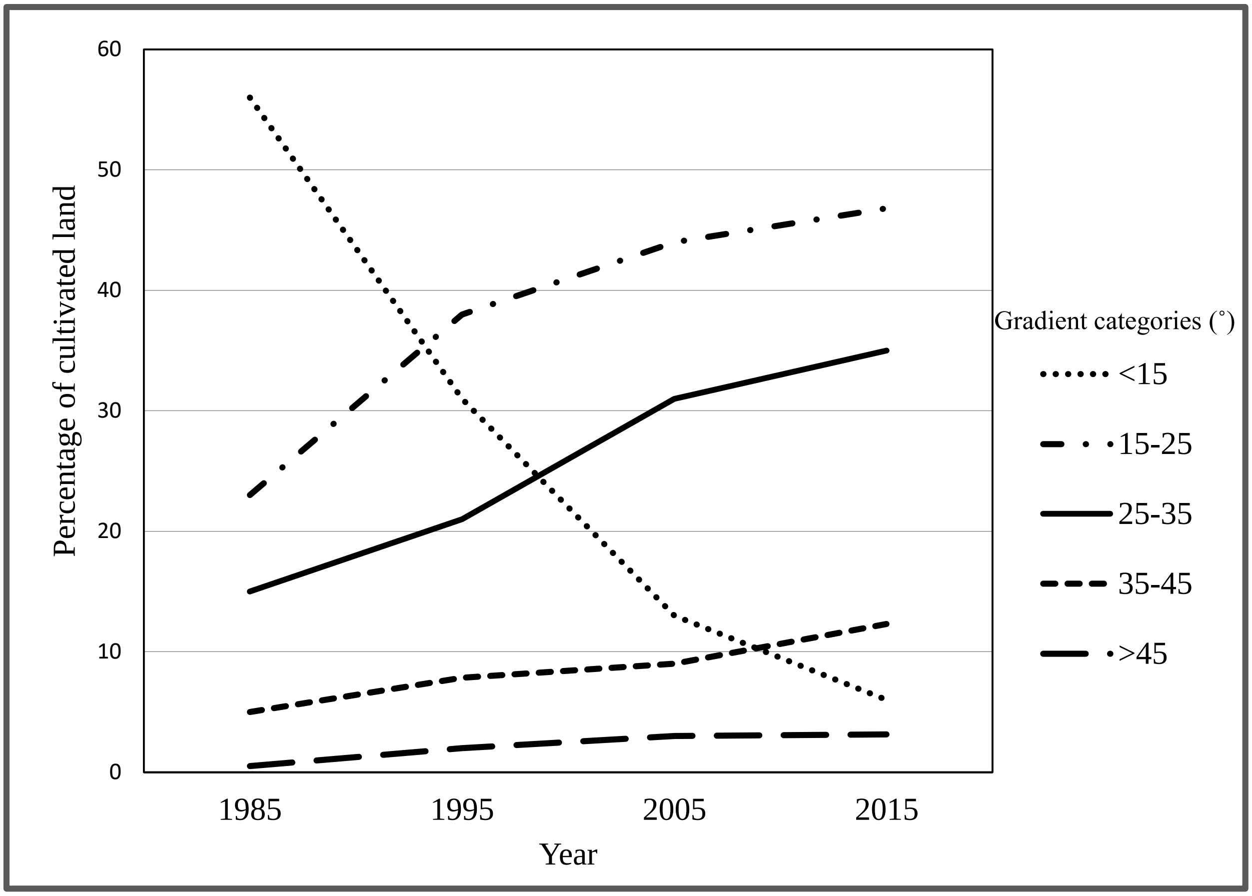

Fig. 5.12: Distribution of cultivated land along slope gradient clusters between 1985 and 2015.

The relationship between the distribution of cultivated land and hillslope gradient was explored by binning hillslope gradient values into 10˚ intervals and by plotting the frequency of hillslope gradients for the cultivated areas. As can be noted from Fig.5.12, there has been a significant shift in the distribution of cultivated areas along the different slope gradient categories. By 1985, 56 % of the cultivated land was in sections with gradients lower than 15˚. The overall trend shows a reduction in the distribution of cultivated land in sections with low gradient and an increase in steeper gradient zones. Whereas the distribution of cultivated land in sections with gradient cluster of less than 15˚ decreased by 88.6 %, it increased by 88.4% on 15˚ to 25˚ categories. Another significant shift in the distribution of cultivated land was experienced in slope gradient category of 25˚ to 35˚ where it increased by 91% (Appendix 11). The shift in the distribution of cultivated land is due to the conversion of most lowlands into settlements, as will be discussed in the relevant section. Landslides are concentrated in steep slope sections with intensively cultivated areas.

A relationship was also established between the distribution of cultivated land and slope positions. This relationship is illustrated in Figure 5.13.

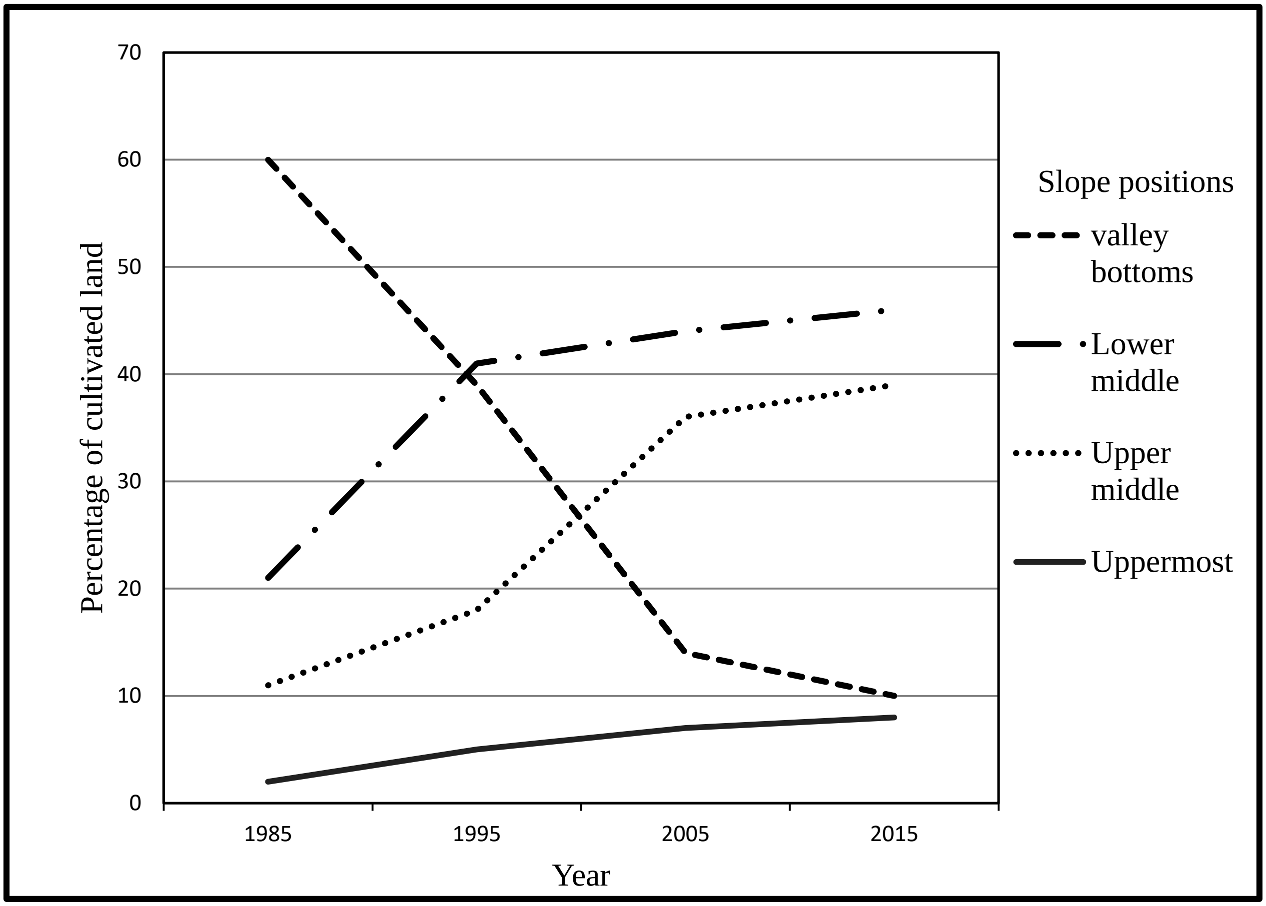

Fig. 5.13: Distributions of cultivated land along slope position from1985 to 2015

As can be noted from Fig. 5.13, there has been a significant shift in the distribution of cultivated areas along the different slope position categories. By 1985, most of the cultivated land was in the valley bottom sections covering 60% of the total cultivated land. The distribution of cultivated land has changed over the last 30 years with a shift from the lower slopes to upper slope sections. Whereas cultivated land decreased by 66% in the valley bottoms between 1985 and 2015, it has increased enormously by 187% and 128% in lower middle and upper middle sections respectively over the same period (Appendix 12). A relationship among cultivation, slope position and landslide occurrence shows that landslides are concentrated in the middle slope sections with intensively cultivated areas. This relationship is explored in the discussion section.

5.4 Discussion

5.4.1 Land use and cover changes

The land-use/cover change trends identified, reveal a drastic decimation of forests and grasslands due to increased cultivation and settlements. This is attributed to acute land shortages, resulting from the exponential population growth in this highland region (NEMA, 2010; UBOS, 2014). The high population density in the region has put tremendous pressure on the land cover, leading to resource overuse and subsequent degradation (Carswell, 2000; NEMA, 2014). The study area has experienced increased land clearance for cultivation, housing and commercial establishments. Most of the natural land cover is destroyed due to the increasing anthropogenic activities (Fig. 5.5). On most of the hillslopes in the study area, there has been complete depletion of vegetation cover (Farley, 1996; Carswell, 1997). Remnants of natural forest cover only exist in small patches surviving under legislative protection (Siriri and Rausen, 2001; Carswell, 2002b; Kabale District Environmental Report, 2015). Most of the highlands now comprise poor vegetation cover with degraded slopes. Many parts of the hillslopes are already bare due to degradation (Lindblade and Carswell, 1998; Carswell, 2002a).

During field investigations, it was observed that every available space has been intensively cultivated. Gardens appear like a continuous carpet for kilometres from the valley bottoms to uppermost slope sections (Fig. 5.7). Increased cultivation in the study area has led to uncontrolled clearance of forest cover (e.g., Farley, 1996, Carswell, 2000). Carswell (2002a) indicates that cultivation frequency in the study area was already very high in the 1930s. Plots were cropped twice a year for 4-5 years on average before being left to fallow, while fallow land was intensively grazed. Cultivation has therefore led to clearance of the natural forests leading to land degradation. The study area has a very long history of human impact on the environment which has resulted in bare slopes leading to landslides (Bagoora, 1997). This relationship between land use/cover changes and landslide occurrence is explored in the subsequent sub-section.

5.4.2 Land use/ cover changes and landslide relationships

Land use/cover changes in the study area have enhanced increased degradation of the hillslopes. This degradation is attributed to the high demand for firewood, charcoal, settlement and cultivation land. Land cover conversion from forest and grassland to cultivated fields permanently reduces slope stability (Selby, 1993). Land use/cover changes in the study area have affected slope stability and resulted into reduced soil cohesion. Such soils with reduced cohesion have become vulnerable to downslope processes, especially landslides. The unprotected soil loose cohesion due to saturation during heavy rainfall events and become susceptible to slope failure. Other studies also confirm that destruction of vegetation cover can reduce soil shear strength (Jakob, 2000; Sidle *et al*., 2006; Claessens *et al.*, 2007). Vegetation conversion results into soil property impairment and hydrological changes which can accelerate slope failures. This explains why landslides in the study area are concentrated in slope sections with degraded vegetation cover (Fig. 5.8). The presence of vegetation cover substantially modifies parameters such as cohesion, internal friction angle, weight of the slope-forming material and pore-water pressure (Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014). This explains why areas with intact vegetation are not susceptible to landslide occurrence in the study area (Fig. 5.8)

The findings of this study are in keeping with several previous studies which identified a close relationship between landslide occurrence and land-use/cover changes (Beguería, 2006; Mugagga *et al*., 2012; Promper and Glade, 2012; Roller *et al*., 2012; Gu and Wylie, 2016). Studies by Dai *et al. (*2001) and Glade (2002) show that barren and sparsely vegetated areas exhibit greater instabilities than forests. In Turkey, Karsli *et al*. (2009) reported an increase in landslide occurrence resulting from land cover changes on tea gardens. Van Beek and Van Asch (2004) also show that shallow landslides were a response to deforestation arising from forest logging and fires. All these studies confirm a close correlation between land-use/cover changes and landslide patterns. In the present study area, most of the landslides occur where there is an interaction between land-use/cover changes and topography. This relationship among land-use/cover changes, topography and landslide distribution is explored in the subsequent sub-section.

The temporal landslide distribution patterns in the study area show that most landslides (>85%) occurred during the period 2005 to 2015 than 1980 to 2005(Fig. 5.11). This distribution shows an increasing trend in landslide occurrence in the study area over the last 10 years. This implies that the disaster risk and vulnerability of communities to landslide occurrence in the study area is increasing. Studies elsewhere also reveal that the risks to landslide occurrence have increased (Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014).

5.4.3 Cultivated land, topography and landslide relationships

The most significant land-use/ cover change trend identified in the study area was the expansion of cultivated areas from valley bottoms and lower slopes to steep middle and upper slope sections. The shift in the distribution of cultivated land is attributed to conversion of most lowlands and valley bottoms into settlements over the past few decades. The valley bottoms and low lands favour construction of residential areas. The study area has experienced rapid settlement expansion due to the rapidly increasing population (NEMA, 2014; UBOS, 2014). There has been a significant increase in building density accompanied by large-scale construction of socio-economic infrastructure in low lands and valley bottoms. This has resulted into acute shortage of land for cultivation in the lowlands and valley bottoms, forcing people to encroach on the sensitive upper slope sections.

During field investigations, it was observed that most of the cultivation is taking place along topographic hollows. As indicated in Chapter 4 on pedological parameters, topographic hollows have deep soils. The deep soils along topographic hollows are due to convergence of materials from hilltops and spur slopes. The hollows are also associated with high moisture content as explored in Chapter 3 on topographic parameters. The deep saturated soils along topographic hollows have attracted heavy cultivation. Intensive cultivation is evident on the upper sensitive slope zones with gradients greater than 45˚. Due to the acute shortage of land in the study area, farmers have also continued to cultivate within and around the landslide features (Fig. 5.10). This has resulted into remobilization of materials (Fig. 5.10) around landslide features, rendering most of the already affected slopes vulnerable to further failures.

Increased cultivation on steep upper slopes is one of the major factors influencing landslide occurrence in the study area. This is due to the fact that cultivation affects the soil structure by weakening the internal cohesive forces (Selby, 1993). Furthermore, farmers in the study area have abandoned the practice of terracing (Bagoora, 1998; Nkonya, 2001; Siriri and Raussen, 2002). Terrace bans used to check on the speed of runoff and hold the soils on the intensively cultivated steep slopes. The remaining terrace bans have also been destroyed in order to get more land for cultivation. There is already evidence of land mismanagement as a result of poor methods of cultivation. As noted already, increased cultivation of steep concave slopes can affect soil material shear strength (e.g., Glade, 2003; Meusburger and Alewell, 2008; Mugagga *et al.*, 2012) leading to increased landslide occurrence. Wasowski *et al*. (2010) observed that higher frequency and susceptibility to landslides in south eastern Italy was a consequence of new ploughing on steep slopes for EU-sponsored wheat cultivation. Studies by Glade (2003) and Karlsi et al. (2009) also show that many shallow landslides are triggered by the expansion of cultivation on steep and marginal hillslopes. The spatial temporal patterns of landslide occurrence in the study area has changed due to increased cultivation of steep middle and upper slopes. Landslide hazards are likely to increase in this highland region given the current land pressure with increasing cultivation of steep uplands.

5. 5 Conclusion

The study area has experienced a drastic decimation of forests and grasslands, due to increased cultivation and settlements. Most of the highlands now comprise poor vegetation cover with degraded slopes. Landslides predominantly occur in areas where forests and grasslands have been converted to cultivated lands and settlements. They are also dominant where there is an interaction between land use/cover changes and topographic features especially slope gradient and position. The expansion of cultivated areas from valley bottoms and lower slopes to steep middle and upper slope elements is the most significant land use/ cover change trend identified. The shift in the distribution of cultivated land is attributed to conversion of most lowlands and valley bottoms into settlements over the past few decades. The spatial temporal patterns of landslide occurrence in the study area has changed due to increased cultivation of steep middle and upper slopes. A close spatial and temporal correlation between land use/cover changes and landslide occurrence is discernible.

5.6 Recommendations

This study has been established that slope sections with intact vegetation are not susceptible to landslide occurrence. It is therefore recommended that afforestation in the landslide prone and degraded areas be undertaken in the highlands, especially along the steep sensitive topographic hollows. Vegetation cover will increase the root density in the soils which can help to hold soil materials together. Root reinforcement will increase soil strength which can reduce on slope failures (Begueria 2006; Yalcin, 2007). The presence of vegetation cover will substantially modify parameters such as cohesion, internal friction angle, weight of the slope-forming material and pore-water pressure (e.g., Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014). The enhanced cohesion will derive from root matrix reinforcement and suction, through evapotranspiration and interception (Selby, 1993). The restored vegetation cover will therefore lead to stable hillslope water balance. Restoration of the vegetation cover will also increase humus and organic matter, which hold soil aggregates together and enhance the strength of materials (e.g., Morgan, 1993). Farmers should also be encouraged and helped to re-establish terrace farming, while avoiding cultivation of sensitive steep middle and upper slope elements. Terrace bans will help to retard the speed of runoff, while holding soil materials against downslope forces (Siriri and Raussen 2002).

Chapter Six

A conceptual model for landslide occurrence in Kigezi highlands of South Western Uganda

6.1 Introduction

On the basis of the findings of the present study, a conceptual model for landslide occurrence that can predict landslide susceptibility in the study area was developed. According to Brabb (1984) and Chao-Yuan *et al.* (2017), landslide susceptibility is the likelihood of an area to experience landslides, depending on the prevailing environmental conditions. Under specific geo-environmental conditions, landslide susceptibility is the probability for landslides to occur in a given area (Van Western *et al.*, 2003; Hungr *et al*., 2005). Many studies have been undertaken to assess landslide susceptibility and model their spatial-temporal occurrence (VanWesten *et al.*, 1999; Dai *et al.*, 2002; Sidle and Ochiai, 2006; Yalcin and Bulut, 2007; Peter *et al.*, 2010; Regmi *et al*., 2010). The conceptual model developed in this study does not consider how frequent and when landslides will occur. The model predicts where landslides are likely to be experienced in the landscape. It describes the potential landslide sites in the landscape. This is premised on the fact that the prediction of what will happen in the future depends on what happened in past and present (Lee and Talib, 2005; Chao-Yuan *et al*., 2017).

The characteristics of the past and present landslide sites were used in the determination of potential landslide zones. In this model, the inherent factors that caused the past landslides were used to predict where landslides are likely to occur in future within the landscape. This was premised on the fact that landslides are more likely expected to occur in areas where they have been experienced in the past (Glade and Crozier, 2005; Crozier, 2010). It is assumed that conditions which led to slope failure in the past are as well likely to cause landslides in future (Biswajeet and Lee, 2010; Susana *et al.,* 2017). Therefore, future landslide occurrence patterns depend on past and present failures. It has been suggested by some authors that landslide instability factors can be analysed and used to construct landslide occurrence predictive models (Hutchinson, 1988; Crozier, 1999; Dietrich *et al*., 1995; Crozier, 2010). Prediction of the susceptibility of an area to landslide occurrence depends on the identification and mapping of the inherent instability factors (AGS, 2007b; Guiseppe *et al.,* 2016). The factors underpinning landslide occurrence in the study area include topographic, soil, rainfall, as well as land use and cover changes (Fig. 6.1).

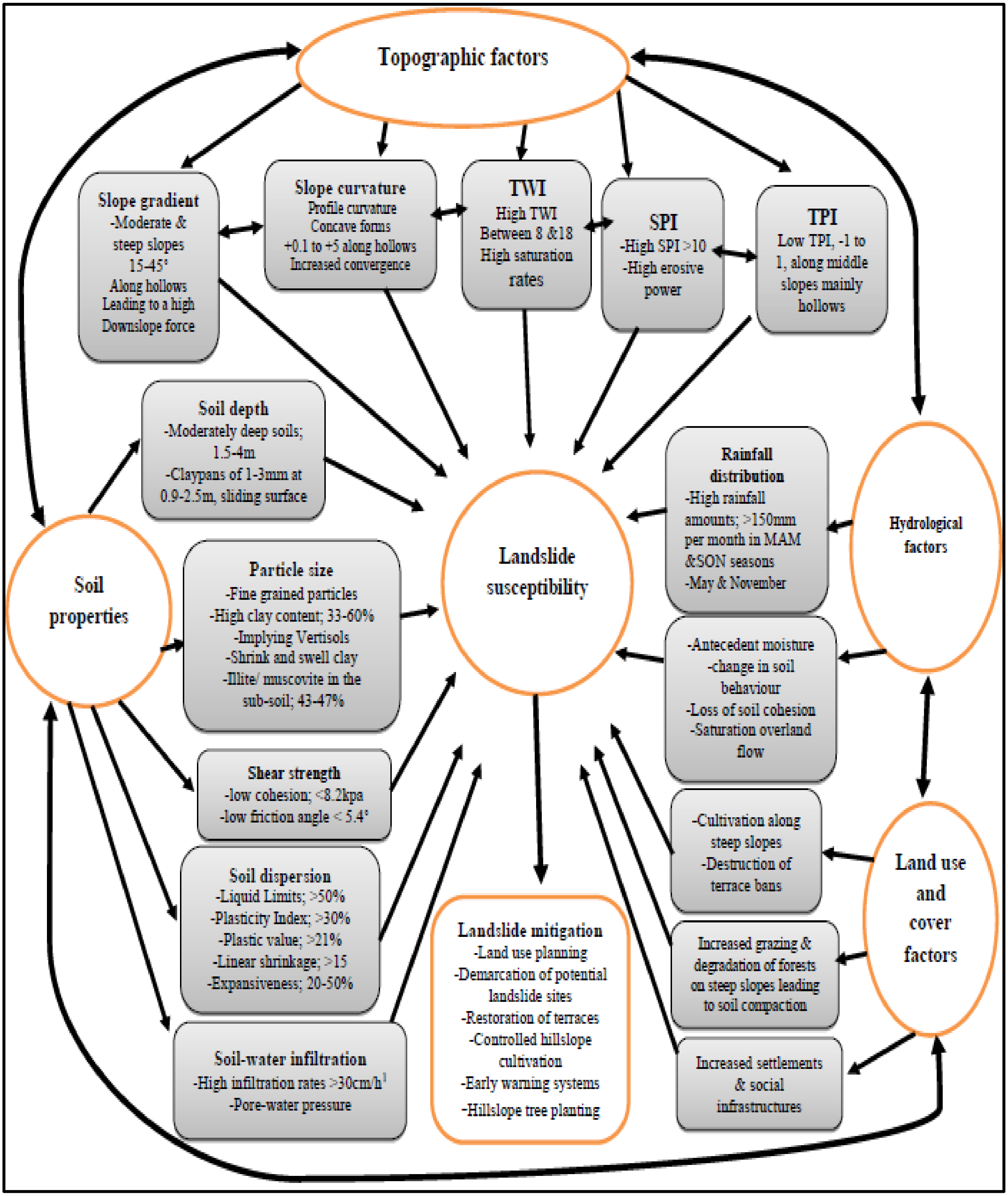

Fig. 6.1: A conceptual model for landslide occurrence for the study area.

Given the fact that landslide conditions are specific to a particular area, site-specific analysis of conditioning factors is important (Lee and Talib, 2005; Gonghui *et al.,* 2010; Kirschbaum and Zhou, 2015). The conceptual model developed in this study is dependent on the specific topographic, soil and land use/cover characteristics of Kigezi highlands. Several studies also indicate that in order to reduce the disaster risk to landslides, it is important to carry out landslide hazard assessment on local or regional scales (Pece *et al*., 2006; Guiseppe *et al.,* 2016; Chao-Yuan *et al.,* 2017). This is because landslide occurrence is a function of susceptibility and temporal frequency of landslide triggers (Corominas, 2014; Susana *et al.,* 2017). When assessing landslide hazard within a given area, recognition of the conditions that caused or may cause the slope to become unstable and the processes that triggered or may trigger the mass movement is of primary importance (Kirschbaum *et al.,* 2016). The factors that are responsible for creating a landslide in a particular area may be grouped into two categories: preparatory and triggering (Kirschbaum *et al.,* 2015; Chao-Yuan *et al.,* 2017). If triggering factors are not taken into account, the term 'susceptibility' may be employed to define the likelihood of landslide occurrence (Kirschbaum and Zhou, 2015).

The overall aim of this study was to integrate all the parameters underpinning landslide occurrence in Kigezi highlands in a conceptual framework and determine their convergence. The information provided by the model could help to mitigate vulnerability and enhance resilience of communities to landslide hazards. The conceptual model for landslide occurrence will help to inform policy, particularly in terms of developing early warning systems to landslide hazards in these highlands. Using this model, land use planning and management in the study area is possible.

6.2 The conceptual model

The parameters used to develop the conceptual model included topographic, soil properties, rainfall and land-use/ cover changes as illustrated in Fig. 6.1.

6.2.1 Topographic parameters

The topographic characteristics were used to identify potential landslide zones in the study area. The topographic parameters used in developing the conceptual model included; slope gradient, profile curvature, topographic wetness index, stream power index and topographic position index.

Slope gradient

Topographic analysis of the study area indicated that slope gradient has a major influence on landslide occurrence. This relationship is shown using equation A1:

$Ls = \varphi_g \tan\beta$ ………………………………………………………………………… A1,

Where Ls is landslide susceptibility and tanβ is the slope gradient.

Landslides are more likely to be experienced in sections with slope gradients ranging between 15˚ and 45˚. Such slope sections have deep well-developed soil profiles representing the amount of materials available and a high downslope gravitational component to pull materials leading to slope failures. This model is in line with what has been reported elsewhere about the importance of slope gradient in influencing landslide occurrence (Hickey, 2000; Lee and Talib, 2005; Saasa *et al*., 2005). Several authors report that the typical slope gradient for landslide occurrence is between 27˚ and 38˚ (Moeyersons, 2003; Hungr *et al.*, 2005; Hosseini *et al*., 2011; Lopez-Davalillo *et al.*, 2014). Likewise, the slope gradient of the source areas in the study area is between 15˚ and 45˚.

Slope curvature

Landslide occurrence in the study area is also influenced by profile curvature as illustrated by equation A2;

$Ls = \varphi_c \frac{\partial tan\beta}{\partial x}$ ………………………........……………………....……………………… A2

Where Ls is landslide susceptibility, φ_c is the profile curvature and $\partial tan\beta$ is the change in slope gradient.

Potential sites for landslide occurrence in the study area are likely to be in slope elements with profile concave curvature values ranging between 0.1 and 5, mainly along topographic hollows as indicated in the conceptual model (Fig. 6.1). This is due to the availability of high moisture content and deep soils converging in such zones as mobilized from upslope and spur slopes. The incoming water from upslope leads to rapid build-up pore water pressure in the soil mantle resulting in reduced shear strength within topographic hollows. Profile concave forms therefore remain saturated between storms due to convergence of ground water flow. The availability of deep soils within profile concave forms provides materials potentially mobile and susceptible to downslope movement. This model is in line with what has been reported elsewhere that landslides are mostly confined along topographic hollows with concave forms (Reneau *et al.*, 1987; Chen and Lee, 2004; Chen and Wang, 2010).

Topographic Wetness Index (TWI)

Landslide occurrence is also influenced by the distribution of TWI in the landscape as illustrated by equation A3:

$Ls = \varphi_w \left[In \left({}^{\alpha}/_{\tan\beta} \right) \right]$ …………………………………………………………………… A3,

Where φ_w is the wetness index, α is the catchment area and tanβ is the slope.

Landslides are likely to be experienced in slope sections with high TWI values ranging between 8 and 18. High TWI is dominant along topographic hollows due to the moisture converging in such slope elements from spur slopes and hilltops. The high TWI leads to saturation which reduces the shear strength of the materials and can encourage slope failures. The model is in keeping with other studies that report on the role of TWI in landslide occurrence (Reneau and Dietrich, 1987; Nath *et al.*, 2013; Ali *et al*., 2014; Raju and Nandagiri, 2015).

Stream Power Index (SPI)

Landslide occurrence is also influenced by SPI as shown in equation A4;

$Ls = \varphi_s \{ In\, [\alpha * \tan\beta] \}$ ………………………………..…………………………… A4

Where φ_s is the stream power index, α is the catchment area and tanβ is the slope

Landslide susceptibility is also likely to be high in slope sections with SPI values >10 especially along topographic hollows associated with drainage lines. High SPI values signify high erosive power along the topographic hollows rendering such sections vulnerable to landslide occurrence in the landscape. This model is also in line with other studies that have reported on the influence of SPI on slope instability (Freeman, 1991; Ferguson, 2005; Gomi *et al.*, 2008; Buda, 2013; Gartner *et al.*, 2015b).

Topographic Position Index (TPI)

Susceptibility to landslide occurrence is also influenced by TPI as illustrated in equation A5;

$Ls = \varphi_t \, [Zo - Z]$ ………………………………………………………………………… A5,

Where φ_t is the topographic position, Zo is the differences between elevation at the central point and Z is the average elevation around it.

Vulnerability to landslide occurrence is likely to increase on slope positions with low TPI values ranging between -1 and 1. They are likely to be dominant on slope positions characterized by topographic hollows within the middle slope sections as illustrated in the conceptual model (Fig. 6.1). This is due to accumulation of eroded materials from hilltops and spur slopes. Such slope positions are moisture and soil convergence zones. They are associated with high saturation rates, confinement of flow and deep soils. This leads to reduced shear strength and thus increasing susceptibility to landslide occurrence. The model is therefore in agreement with what has been observed elsewhere that high saturation rates along topographic hollows lead to instability of slope materials due to reduced cohesion (Agnew *et al.*, 2006; Ali *et al.*, 2014; Ferreira *et al.*, 2015; Raju and Nandagiri, 2015).

In summary, landslide occurrence (Ls) in Kigezi highlands is likely to be experienced where there is a convergence zone among all the topographic parameters as illustrated in equation A6 which is a summation of equations A1 to A5.

$$Ls = f\left[[(\varphi_g \tan\beta \,(15^\circ \; to \; 45^\circ)] + [\varphi_c \frac{\partial tan\beta}{\partial x}\,(0.1 \; to \; 5)] + [\varphi_w \left[In \; {}^{\alpha}/_{\tan\beta}\right](8 \; to \; 18)] + [\varphi_s In\,[\alpha * \tan\beta](> 10)] + [\,\varphi_t\,[Zo - Z](-1 \text{ to } 1)]\right] \text{……………A6}$$

The equation developed for the model is specific to Kigezi highlands where the topographic parameters integrated in the model prevail. This model can also be applied to other regions with similar topographic characteristics.

The dominant variables in the equation are $\tan\beta$ which is the slope angle, α catchment area, and Z, the slope position. The equation for the influence of topographic parameters on landslide occurrence in Kigezi highlands can therefore be summarised as;

$$Ls = f\,[\tan\beta, \alpha, z] \quad \text{...} \quad \text{A7}$$

The convergence zone for all topographic parameters to cause landslide occurrence in Kigezi highlands is on sections where slope gradient is between 15˚ and 45˚, profile concave forms between 0.1 and 5, TWI between 8 and 18, SPI >10, TPI between -1 and 1 mainly along topographic hollows as illustrated in the conceptual model (Fig. 6.1) and equation A6.

6.2.2 Soil properties

The analysed soil characteristics were also used to identify potential landslide sites in the landscape.

Soil depth and clay pans

An examination of soil properties revealed that landslides in the study area are likely to occur in areas with moderately deep soils >2.5m mainly along mid-slopes of topographic hollows. Such deep soils were associated with high TWI, signifying high saturation rates, which result in reduced shear strength of materials. Potential landslide zones in the study area are also those with clay pans located between 0.9 and 2.5m within the soil profile as illustrated in the conceptual model (Fig. 6.1). This is because clay pans reduce infiltration and lead to accumulation of water within the soil profile. The restriction of vertical flow of water through the soil profile by clay pans was also confirmed by Jiang *et al*. (2014) and Liang and Uchida (2014). The accumulating water leads to saturation of clay pans sandwiched between more stable materials. The saturated clay pans act as a sliding surface for the overlying materials, consequently inducing landslides.

Clay content

Landslide susceptibility (Ls) in the study area is also influenced by the clay content in the soil as illustrated in equation B1.

$Ls = \varphi_c[Cc]$ ……………………..……………………………………………………………B1,
Where Cc is clay content.

In Kigezi highlands, landslide susceptibility is high in areas with clay content greater than 30%. The soil materials in such areas with clay content above 30% have a great water-holding capacity, both upon the surface of the particles and within the cellule of the colloid. They have slow drainage, high expansion potential, great shrink/swell potential, high plasticity and subsequent loss of shear strength. Such soil materials, classified as vertic soils exhibit extreme expansive potential. Vertic soils swell when wet and shrink as they dry to form large cracks. The cracks can be deep and wide leading to reduced strength of materials and consequently landslides. Several studies elsewhere have also demonstrated the influence of high clay content on landslide occurrence (Knapen *et al*., 2006; Yalcin, 2007; Kitutu *et al.,* 2009; Wati *et al*., 2010; Mugagga *et al*.,2011; Broothaerts *et al*.,2012).

Clay mineralogy

Landslide occurrence in the study area is also influenced by the nature of clay minerals in the soil as illustrated in equation B2:

$Ls = \varphi_m[Cm]$ …………………………………………....……………………B2,
Where Cm is clay minerals.

In Kigezi highlands, landslide susceptibility is high on slope sections where the top soil is dominated by quartz while the sub soil has considerable amounts of illite/muscovite clay minerals ranging between 43% and 47 %. Such top soil materials encourage high infiltration. As the water flow gets into the sub soils with high clay content dominated by illite/ muscovite minerals, it stagnates. This leads to reduced shear strength and landslide occurrence. Previous studies also confirm that the presence of illite clays can lead to landslide occurrence due to their swelling potential and low shear strength (Ohlmacher, 2000; Yalcin, 2007; Kitutu *et al*., 2009).

Soil dispersion

Landslide occurrence in the study area is also influenced by other soil parameters such as liquid limits (LL), plastic limits (Pl), plasticity index (PI), weighted plasticity index (PI_w), expansivenes (ε_{ex}) and dispersion (D) as illustrated by equation B3:

$$Ls = \varphi_p \, f \, [LL, PL, PI, PI_w, \varepsilon_{ex}, D] \quad \ldots\ldots\ldots\ldots \text{B3}$$

Where φ_p is soil plasticity.

Landslide susceptibility is likely to be high in areas where soils have high plasticity and are inorganic in nature (CH) indicating weak soils. Such soils respond very fast to saturation and can easily move especially during prolonged rainfall events. Landslides in the study area are likely to occur where the LLs and plasticity index in the soil is above 50% and 30% respectively as illustrated in the conceptual model (Fig. 6.1). This signifies clays of high plasticity, which are weak and highly susceptible to landslides. The model is in line with what has been reported elsewhere on the role of pedological properties such as particle size distribution, claypans, clay content, vertic soils, plasticity and clay minerals in influencing landslide occurrence (Bell and Culshaw, 2001; Knapen *et al*., 2006; Zung *et al*., 2009; Mugagga *et al*., 2011; Yalcin, 2011).

In Kigezi highlands, the potential for landslide occurrence also increases in areas where soil materials have an average weighted plasticity index (PI_w) of 26.4% and expansiveness ranging between 20 and 50%. These are medium expansive soils which are susceptible to slope failures. Such soils are highly dispersive and responsive to changes in seasonal water distribution. They can shrink and swell, leading to loss of soil strength. Landslide occurrence is also likely to increase in areas with critical dispersion values of >50% (Fig. 6.1). Dispersive soils are structurally unstable in water due to their chemistry and collapse or disperse to form dissolved slurry when in contact with water (Bell and Maud, 1994; Bell and Culshaw, 2001). This can interfere with the structural stability of the soil, making it highly prone to slope failures.

Shear strength.

Landslide occurrence is also influenced by the shear strength parameters of cohesion ($C^{'}$)and angle of internal friction (φ') as illustrated by equation B4.

$$Ls = \frac{\varphi_{shear}}{f\ [C',\varphi']} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \text{B4}$$

Landslide occurrence in Kigezi highlands is also likely to be experienced on slope elements with soils having cohesion of less than 11.1 kPa and the angle of internal friction lower than 8.1˚. Such low cohesion and angle of internal friction signify weak soils which are susceptible to sliding especially after disturbance by natural and/or anthropogenic processes. This model is in keeping with several studies that have demonstrated the influence of soil shear strength on landslide occurrence (Yalsin, 2007; Zung *et al*., 2009; Mugagga *et al*., 2011; Yashar *et al*., 2013; Zinck, 2013)

Soil water infiltration

Landslide occurrence and distribution in the study area also depends on the variations in the soil water infiltration rates (I) as illustrated by equation B5.

$$Ls = \varphi_i[I] \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots \text{B5}$$

In the study area, landslides are also likely to be experienced in areas with high infiltration rates >30cm/h^{-1} mainly along mid-slopes of topographic hollows. High infiltration leads to increased saturation which lowers the soil shear strength. High saturation rates in the lower slopes lead to saturation overland flow processes, which move upslope from the slope base incrementally along topographic hollows. In association with the high clay content dominated by illite/muscovite minerals, materials along topographic hollows remain saturated most of the time. This results into reduced shear strength on such slope elements, facilitating landslide occurrence. Several studies also confirm the influence of soil saturation in determining soil stability and consequent slope failures (Duiker *et al*., 2001; Braja M. Das, 2011; Das *et al.*, 2011; Broothaerts *et al.*, 2012; Lopez-Davalillo *et al.*, 2014)

Rainfall distribution

Landslide occurrence in the study area is also influenced by the amount and distribution of rainfall (R) as given in equation B6:

$Ls = \varphi_{rf}[R]$ …………………………………………...…………….……………………………. B6

Susceptibility to landslide occurrence in the study area is likely to be high during the months of May and November, despite the preceding months of April and October receiving more rainfall. In the Kigezi highlands landslides are not likely to be experienced during or immediately after extreme rainfall events but will occur later as the rainfall season progresses. This time lag in landslide occurrence and rainfall distribution, is due the initial infiltration through quartz dominated upper soil layers, before illite/muscovite clays in the lower soil horizons get saturated. This leads to accumulation of antecedent moisture building in the soils during the months of April and October with heavy rainfall. Continuous rainfall during the successive months of May and November on already saturated soil materials is likely to result into loss of cohesion, leading to landslide occurrence in the region. Several studies also observe that antecedent soil moisture conditions prior to a rainfall event is a significant factor in landslide occurrence (Morgan, 1993; Bagoora, 1997; Rahardjo *et al*., 2007). Landslides in the study area are therefore likely to occur as a response to clay pan saturation during the rainfall season. Several studies also reveal a close relationship between landslide occurrence and rainfall distribution (Knapen *et al.*, 2006; Kitutu *et al*., 2009; Broothaerts *et al.*, 2012). It is illustrated in this model that landslide occurrence in the study area is due to the increase of pore water pressure in the soil due to continuous rainfall that reduce the soil strength. This model therefore shows that continuous rainfall is the main triggering factor for the initiation of landslides in the study area.

Landslide susceptibility and occurrence in the study area is function of different soil properties parameters. The soil properties converge to cause landslides in the study area as shown in equation B7 which is a summation of equations B1 to B6.

$$Ls = \varphi_c[Cc\ (30-60\%)] + \varphi_m[Cm\ (43-47\%)] + \varphi_p\ f\ [LL(>50\%), PL(25\%), PI(>30\%), pI_w(26.4\%), \varepsilon_{ex}(20-50\%), D(>50)] + \frac{\varphi_{shear}}{f\ [C'^{(11.1\ \mathrm{kPa})}, \varphi'(8.1°).]} + \varphi_i[I(>30\mathrm{cm/h}-1\ ] + \varphi_{rf}[R\ (>150mm/mth] \text{ ……………………………………………………….B7}$$

This equation is confined to Kigezi highlands characterized by the observed soil properties. This model can also be applicable to other regions with similar soil characteristics.

In summary, landslide occurrence in the study area is a function of soil properties and rainfall distribution. Rainfall distribution and amounts lead to changes in soil behaviour and consequently landslide occurrence. This can be summarised in equation B8;

$$Ls = f\ [Soil, Rainfall] \text{ ....…………………………………………………………….B8}$$

6.2.3 Land use and cover change

Landslide occurrence in Kigezi is also a function of land use and cover changes. This association is illustrated in equation C.

$$Ls = f\ [lucc] \text{ ……………………………………………………………………………C1,}$$

Where $lucc$ is land use and cover changes.

Landslide occurrence is likely to be experienced in slope sections associated with vegetation degradation. In the Kigezi highlands, landslide susceptibility is expected to be high in areas where forests and grasslands have been converted to cultivated and settlement lands. This is due to the fact land use/cover changes affect slope stability and result into reduced soil cohesion. Soil cohesion is lost once the tree roots which help to bind the soil materials together are removed due to destruction of vegetation cover. Vegetation conversion results into soil property impairment and hydrological changes which can accelerate slope failures. In the study area, landslides are likely to be experienced in sections where there is an interaction between land use/cover changes and topography. Landslide occurrence in the study area is likely to be experienced in sections with increased cultivation of steep middle and upper slopes as illustrated in the conceptual model (Fig. 6.1). This conceptual model is in keeping with several previous studies which identified a close relationship between landslide occurrence and land use/cover changes (Glade, 2002; Karlsi *et al*., 2009; Mugagga *et al*., 2012; Promper and Glade, 2012).

All these parameters converge to cause landslides in Kigezi highlands. This convergence is illustrated by summing up equations A7, B8 and C as shown blow;

$Ls = f\ [\tan\beta, \alpha, z]$ Topographical parameters……………………………………...…. A7

$Ls = f\ [Soil, Rainfall]$ Soil properties………………………………………….………..B8

$Ls = f\ [lucc]$ Land use and cover changes ……………………………………………….C

Therefore, susceptibility to landslide occurrence in Kigezi highlands can be illustrated by equation D where:

$Ls = f\ [\tan\beta, \alpha, z]\ + f\ [Soil, Rainfall]\ + f\ [luc]$……………………………………. D

Landslide occurrence in Kigezi highlands is therefore a function of topographical, soil and land-use/cover parameters. Landslide susceptibility is likely to be high on slope elements where all these parameters interact. This convergence zone within the landscape exist mainly along topographic hollows. The conceptual model illustrated in equation D is specific to Kigezi highlands characterized by the topographical, soil and land use/cover variables as shown in the various equations. The conceptual model developed can identify areas likely to be affected by future landslides. This could help to guide the decision makers for land use planning in the study area.

The model illustrates an interplay and convergence of parameters to cause landslides. The model can therefore help to conceptualize the landslide conditioning factors and susceptibility in Kigezi highlands. Although the model is unique to Kigezi highlands, it can also be applied to other areas with similar topographic, soil and land use/cover characteristics. Elsewhere, Mugagga (2011) also developed a conceptual model to explain landslide occurrence on the slopes of Mt Elgon in Eastern Uganda. In his model, steep concave slopes coupled with the 'problem nature of soils' and rainfall makes Mt Elgon slopes susceptible to landslides. Likewise, the conceptual model developed in this study shows that landslide occurrence in the Kigezi highlands is a function of an interaction of topographic, soil and land use/cover parameters. This model also indicate that continuous rainfall is the trigger of landslides in the study area due to its effect on pore-water pressure and change in soil behaviour. Landslide occurrence in the study area is therefore a function of the susceptibility factors including topography and soil properties as well as the triggering factors such as rainfall and the human factor. This model is in keeping with what has been observed elsewhere that landslide occurrence is an interplay of preparatory and triggering conditions (Kirschbaum *et al*., 2015; Chao-Yuan *et al*., 2017).

6.3 Conclusion

The conceptual model shows that topographic hollows are considered as potential landslide sites in Kigezi highlands. Hollows are susceptible to landslides because they are characterized by favourable gradient, profile concave forms, high TWI and SPI as well as low TPI. Hollows are also associated with moderately deep soils and experience high infiltration rates which lead to increased pore-pressure in the voids. Topographic hollows also experience saturation overland flow which moves incrementally from the slope base upslope leading to saturation and therefore reduced shear strength. The hollows also contain vertic soils with high clay content, plasticity index and have an extremely high expansive potential. The hollows are also intensively cultivated due to their deep soils. The convergence of all these parameters leads to landslide occurrence along the hollows. Due to the increasing landslide hazard occurrence in Kigezi highland region, developing an early warning system is imperative for disaster preparedness and management in vulnerable areas. This model can therefore be used in disaster risk reduction of communities to landslide hazards in this highland region.

Chapter Seven

The synthesis of landslide occurrence in Kigezi highlands

7.1 Introduction

This chapter presents a synthesis of the major findings, conclusions and provides recommendations. The parameters underpinning landslide occurrence in the non-volcanic Kigezi highlands are summarized. Recommendations are made to enhance disaster risk reduction of communities to landslide hazards. This is based on the key findings of the study. The chapter also proposes future research directions.

7.2 Topographic conditions and landslide occurrence

Slope gradient

This study establishes that landslide occurrence is concentrated in sections with moderately steep slope gradients, ranging between 25˚ and 35˚. The lowest slope gradient where landslides can occur in the study area is 15˚ while the highest is 45˚. It was established that the distribution in landslide occurrence with slope gradient is due to the difference in soil depth and moisture content. The study found out that soil depth reduces with increase in slope gradient. The concentration of landslides on moderately steep slope sections is explained by the accumulation of displaced materials from uppermost very steep and precipitous slopes (Capitani *et al.*, 2013). The role of slope gradient in inducing landslides has been explored and confirmed by a number of authors (Appolinaire *et al.*, 2007; Claessens *et al.*, 2007; Mugagga *et al.*, 2012; López-Davalillo *et al.*, 2014).

Slope curvature

Landslides in the study area were noted to concentrate in slope sections where profile curvature is concave, with values ranging between 0.1 and 5. Landslides are less pronounced in sections where profile curvature is convex with values lower than 0. This study found out that profile concave forms are associated with topographic hollows while convex forms appear along hilltops and spur slopes in the landscape. The study establishes that the concentration of landslides along topographic hollows with profile concave forms is due to, among other things, moisture concentration and deep soil profiles. This is because curvature affects surface and subsurface hillslope hydrology in determining slope stability (Infascelli *et al.*, 2013). The study results are in keeping with previous studies which identified landslides as occurring in areas of convergent topography, where subsurface soil water flow paths give rise to excess pore-water pressures downslope (Grabs *et al.*, 2009; Gao and Maro, 2010; Lopez-Davalillo *et al.*, 2014; Raju and Nandagiri, 2015).

TWI and SPI

This study found out that landslide occurrence was concentrating in sections with high TWI values, ranging between 8 and 18, mainly along topographic hollows with profile concave forms. They are less pronounced in slope sections with TWI values <8 associated with profile convex forms, mainly along hilltops and spur slopes. The study establishes that high TWI leads to saturation which reduces the shear strength of the materials, resulting into slope failures. The convex zones are, however, dry and stable due to the low saturation rates. Similar results have also been reported elsewhere on the influence of the TWI on slope failure (Agnew *et al*., 2006; Ali *et al*., 2014; Raju and Nandagiri, 2015). Landslide occurrence was also noted as dominant in sections where SPI values are >10, mainly along topographic hollows associated with drainage lines. High SPI values signify increased erosive power along the topographic hollows and therefore high susceptibility to landslide occurrence. Low SPI values on hilltops and spur slopes indicate reduced erosive power and therefore less incidence of landslide occurrence. Other studies also demonstrate that high SPI values indicate areas in the landscape that have a high potential for erosion and therefore landslide occurrence ((Ferguson, 2005; Gomi *et al.*, 2008; Buda, 2013).

Topographic position index

This study establishes that landslides in Kigezi highlands are dominant on slope positions with TPI values between -1 and 1 mainly along topographic hollows. Landslides are also concentrated in middle slopes with low TPI values, mainly along topographic hollows. Such slope elements were also found to have high TWI and SPI values, signifying increased saturation rates and erosive power. The middle slope sections characterized by topographic hollows are characterised by accumulation of the eroded materials from hilltops and spur slopes (Briggs and Knapp, 2008;

Mokarram *et al.*, 2015). The study establishes that the availability of materials increases the vulnerability of middle slope elements to landslides. Other studies have also associated landslide occurrence with the topographic position of the landform (Tagil and Jenness, 2008; De Reu *et al.*, 2013).

This study establishes a convergence zone among all the topographic parameters within the landscape to induce landslides. Landslides are common along topographic hollows with moderately steep gradient, profile concave forms, high TWI and SPI as well as low TPI.

7.3 Soil properties and landslide occurrence

Soil profile characteristics

This study found out that deep soil profiles ranging between 2.5 and 7m are a major characteristic of the study area. Deeper profiles are more pronounced along topographic hollows and valley bottoms. Notwithstanding the deep soils (mostly >6m) on most slopes in the study area, most landslide features are shallow, occurring within less than half of the profile (1 to 3m). This is explained by the presence of clay pans which form at a depth of 0.9 to 3m within the soil profiles. Clay pans reduce infiltration and lead to accumulation of water within the soil profile. The restriction of vertical flow of water through the soil profile by clay pans was also confirmed by Zhou *et al*. (2010) and Zinck (2013). The accumulating water lead to saturation of clay pans sandwiched between more stable materials. The saturated clay pans act as a sliding surface for the overlying materials, consequently inducing landslides. It can be inferred that the occurrence and characteristics of landslide processes in Kigezi highlands are highly influenced by the presence and position of clay pan horizons in the soil profile. Understanding the spatial characteristics of clay pan soil properties is important in characterising soil behaviour (Li *et al*., 2010). This is because clay pans create a unique hydrology, characterized by poor drainage and very slow permeability in the soil matrix (Rawls *et al*., 1992; Yilmaz and Karacan, 2002).

Particle size distribution

This study establishes that fine-grained soils of silt and clay are predominant in the study area. The study found out that clay predominates, with more than 40 %, while sand and silt are less than 35 % and 25 % respectively in most soil samples. Owing to the high clay content of greater than 35% and Plasticity Index (PI) of 33.3 %, the soils in the study area are categorised as Vertisols, which are known for inducing landslides (Mugagga *et al*., 2011). This study establishes that the presence of large amounts of clay in the soils of the study area is a major factor in landslide occurrence, since it affects the stability of the soils especially when wet. Clay has a great water-holding capacity, both upon the surface of the particles and within the cellule of the colloid (Baynes, 2008 Kitutu *et al.*, 2009; and Mukasa-Tibandeke *et al*., 2016). It can be inferred that soils with such high clay content are highly vulnerable to landslide occurrence. Similar results were also revealed by Kitutu *et al.* (2009) and Mugagga *et al*. (2011) on the slopes of Mount Elgon in Eastern Uganda. Several studies elsewhere have also demonstrated the influence of high clay content on landslide occurrence (Yalcin, 2007; Wati *et al*., 2010; Yalcin, 2011; Broothaerts *et al*., 2012).

Clay mineralogy

This study established the presence of moderately expansive clays, particularly illite/ muscovite, particularly in the sub soils. Despite the absence of extremely expansive smectite, the dominance of illite influences the stability and susceptibility of slope materials to sliding. Previous studies also indicate that the presence of illite clays can lead to landslide occurrence due to their swelling potential and low shear strength (Ohlmacher, 2000; Yalcin, 2007; Kitutu *et al*., 2009). Studies by Yilmaz and Karacan (2002) and Yalcin (2011) also report on the influence of clay mineralogy on shear and frictional resistance of the soils. A number of studies indicate that clay minerals significantly affect the physical, chemical and biological processes of soils (Day, 1994; Husein *et al*., 1999; Suzuki *et al*., 2007) including landslides (Selby, 1993; Inganga *et al*., 2001; Zinck, 2013). In summary, this study establishes that the presence of high amounts of illite/muscovite clay minerals capable of accumulating water over time is one of the key factors influencing landslide occurrence in the study area.

Soil dispersion

This study establishes that soils of the study area have high plasticity and are inorganic in nature (CH), indicating weak soils which respond very easily to saturation (Chapter 4). Such soils easily slide, especially during continuous rainfall. This study found out that the Liquid Limits for all samples analysed was above 50%, signifying soils with high plasticity. The average weighted plasticity index (PI_w) and expansiveness were 26.4% and 32.8% respectively. By implication, soils of the study area are highly dispersive. Dispersive soils are structurally unstable in water due to their chemistry and collapse or disperse to form dissolved slurry when in contact with water (Bell and Maud, 1994; Bell and Culshaw, 2001). This study establishes therefore that landslide occurrence in the study area is associated with expansive soils which shrink and swell, leading to loss of soil strength. The role of Liquid Limits in characterizing the problem nature of soils has been reported by various scholars (Msilimba and Holmes, 2005; Fauziah *et al.,* 2006; Baynes, 2008; Mugagga *et al.,* 2011).

Soil shear strength

This study established that colluviums cover most of the slopes mainly along hollows. The colluviums characteristically have low shear strengths in terms of their cohesion and internal frictional angles (Chapter 4). The study found out that the average cohesion for all tested soil samples in the study area is 7.2 kPa, while the average angle of internal friction is 4.1˚. Such weak soils are susceptible to sliding especially after disturbance. The dominance by clay soils in the study area, which have very low strength parameters when saturated as indicated by shear strength parameters (c and ϕ), has strong implications for landslide occurrence. Several studies also demonstrate the influence of shear strength parameters on landslide occurrence (Morgan, 1993; Selby, 1993; Lopez-Davalillo *et al.*, 2014).

Soil water infiltration

The study establishes that soil water infiltration varies across slope position and topographic configuration. This variation is due to the differences in soil depth and location of clay pans within the slope profile. This study observed that the upper slope sections and spur slopes experience low infiltration rates due to shallow soils in such sections which reach saturation very fast. The lower slope sections and topographic hollows are associated with high and very high infiltration rates due to the deep soil profiles in such zones. This variation in infiltration rates also signifies differences in soil saturation levels. The study establishes that high saturation rates in the lower slopes lead to saturation overland flow processes, which move from the slope base incrementally along topographic hollows upslope. In association with the high clay content dominated by illite/muscovite minerals, materials along topographic hollows remain saturated most of the time. This results into reduced shear strength within the topographic hollows, facilitating landslide occurrence.

This study also established a relationship between soil water infiltration and topographic parameters in the landscape, including slope gradient, position and curvature. Infiltration was observed to increase with reduction in slope gradient due to an increase in soil depth. The study found out that deeper soils in the lower slope sections require more water to accumulate before reaching saturation, hence more infiltration. This is consistent with observations elsewhere by Stolte (2003). This study observed that the variation in infiltration with slope gradient is also due to the location of clay pans within the soil profile. Clay pans restrict vertical water movement in the soil profile and therefore reduce infiltration. This explains why steep upper slope sections experience low infiltration rates because they reach saturation faster due to the presence clay pans near the surface. The study further observed that lower slope elements and topographic hollows experience higher infiltration rates than the upper slope elements and spurs due to variations in soil depth. Lower slope elements and topographic hollows are characterized by deeper soil profiles than the upper and spur slope counterparts (Chapter 4). The present study therefore confirms that it is possible to estimate landslide occurrence using soil infiltration experiments along the slope profile.

Rainfall distribution and landslide occurrence

The study establishes that landslide occurrence in the study area is not correlated with extreme rainfall events, as is the case with Mt Elgon region in Eastern Uganda. In the Kigezi highland region, landslides are not normally experienced during or immediately after extreme rainfall events, but occur later in the rainfall season. This is due to antecedent moisture building up in the sub soil materials as more rainfall is received. The study also found out that most landslides in the study area are concentrated in the months of May and November, despite the preceding months of April and October receiving more rainfall. This time lag in landslide occurrence and rainfall distribution, is due to the initial infiltration through quartz dominated upper soil layers, before illite/muscovite clays in the lower soil horizons get saturated. This leads to accumulation of antecedent moisture building in the soils during the months of April and October with heavy rainfall. This study establishes that continuous rainfall during the successive months of May and November on already saturated soil materials leads to loss of cohesion, hence landslide occurrence in the region. Several studies also observe that antecedent soil moisture conditions prior to a rainfall event are the most significant factor in landslide occurrence (Dai *et al*., 2003; Pedrozzi, 2004; Gonghui *et al*., 2010; Panek *et al.*, 2011; Qiang *et al*., 2011). This study therefore inferred that landslide occurrence in the Kigezi highlands is due to the initial rapid infiltration through quartz dominated top soils and subsequent saturation of the clay pan dominated sub soils.

7.4 Land-use/ cover changes and landslide occurrence

This study establishes a drastic decimation of natural vegetation cover due to increased cultivation and settlements was identified. This is attributed to acute land shortages resulting from the exponential population growth. Most of the natural land-cover is destroyed due to increasing anthropogenic activities. The study observed that on most of the hillslopes in the study area, there has been complete depletion of vegetation cover. The study found out that every available space in the study areas has been intensively cultivated. Gardens appear like a continuous carpet for kilometres from the valley bottoms to uppermost slope sections. This study established that land-use/cover changes affect slope stability and result into reduced soil cohesion. Several studies elsewhere also demonstrate that loss of root networks reduces soil cohesion (Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; Lopez-Davalillo *et al.*, 2014). The study established that most landslide occurrence is associated with vegetation degradation. This is in keeping with previous studies which observed the link between destruction of vegetation cover and reduction of soil shear strength and eventual landslide inducement (Glade, 2002; Begueria, 2006; Sidle *et al*., 2006; Claessens *et al.*, 2007; Karlsi *et al.*, 2009). Vegetation cover substantially modifies parameters such as cohesion, internal friction angle, weight of the slope-forming material and pore-water pressure (Van Beek and Van Asch, 2004; Promper and Glade, 2012). Similar studies have also identified a close relationship between land-use/cover changes and landslide occurrence (Meusburger and Alewell, 2008; Wasowski *et al*., 2010; Mugagga *et al*., 2012).

Most of the landslides were noted to occur where there is an interaction between land use/cover changes and topography. This study observed that the most significant land use/ cover change trend in the study area was the expansion of cultivated areas from valley bottoms and lower slopes to steep middle and upper slope sections. The shift in the distribution of cultivated land is attributed to conversion of most lowlands and valley bottoms into settlements over the past few decades. The study established that increased cultivation on steep middle and upper slopes is one of the major factors influencing landslide occurrence. Cultivation affects the soil structure by weakening the internal cohesive forces (Selby, 1993). Increased cultivation of steep concave slopes can therefore affect soil material shear strength, leading to landslide occurrence (Glade, 2003; Karlsi *et al*., 2009). This study therefore inferred that landslide occurrence is likely to increase in this highland region given the current land use pressure, which gives rise to increasing cultivation of steep uplands.

7.5 The uniqueness of landslide occurrence in Kigezi highlands

Compared to other regions in East African highlands, landslides in Kigezi highlands occur along clearly defined lines mainly the topographic hollows within the landscape. The landslide path can be easily identified and demarcated within the landscape. This is due to the fact that topographic hollows have moderately steep slopes, high TWI and SPI, low TPI, profile concave forms, deep soils and are intensively cultivated. Compared to other slope elements in the landscape, topographic hollows are conducive for landslide processes. On 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 (Knapen, 2003; Knapen *et al*., 2006; Mugagga *et al*., 2012). This points to the differences in topographic parameters that underpin landslide occurrence in different mountain environments of the country.

The landslide features in the study area are also predominantly shallow, despite the deep soil profiles covering the slopes. This is due to the presence of clay pans which form at a depth of 0.9 to 3m within the profiles. Clay pans reduce infiltration and lead to accumulation of water within the soil profile. The accumulating water leads to saturation of clay pans sandwiched between more stable materials. Saturated clay pans then act as a sliding surface for the overlying materials, consequently inducing landslides. It can be inferred that the occurrence and characteristics of landslide processes in Kigezi highlands is highly influenced by the presence and position of clay pan horizons in the soil profile. In other mountain environments, for example Mt Elgon region, landslide occurrence could occur up to the bedrock, removing all the overlying materials.

Landslides in the study area also have several tongues which converge together. Their appearance and morphology is associated with the convergence of several subsidiary hollows from upslope into one major hollow along the middle slope sections. This phenomenon is unique to the Kigezi highlands, denoting the influence of a unique topographic configuration. Furthermore, landslides in Kigezi highlands are not normally experienced during or immediately after extreme rainfall events as is the case elsewhere, for example the Mt Elgon region. They occur later in the rainfall season. They are concentrated in the months of May and November, despite the preceding months of April and October receiving more rainfall. This time lag in landslide occurrence and rainfall distribution, is due the initial infiltration through quartz dominated upper soil layers, before illite/muscovite clays in the lower soil horizons get saturated. This leads to accumulation of moisture in the soils during the months of April and October with heavy rainfall. Continuous rainfall during the successive months of May and November on already saturated soil materials lead to loss of cohesion, hence landslide occurrence in the region. In other regions such as Mt Elgon of Eastern Uganda, landslide occurrence immediately responds to extreme rainfall events.

7.6 Study recommendations

- An understanding of the topographic, pedological and anthropogenic parameters and their influence on landslide occurrence is important in landslide hazard management. It is now possible to identify and predict potential landslide zones, and also demarcate safer zones for community activities. This information and knowledge generated about the area’s topographic, pedological and land cover characteristics will serve to reduce the vulnerability and disaster risk of communities to landslide hazards in this fragile highland ecosystem. This can be done through designating appropriate zones for community activities, while avoiding the potential landslide zones. Potential landslide zones can be demarcated based on the distribution of susceptible soil materials along topographic hollows and degraded slope elements vulnerable to failures.

- Due to the increasing landslide hazards in Kigezi highland region, developing early warning systems in the light of landslides striking well after peak period rainfall months is very important. An early warning system will help in disaster preparedness and hazard management in vulnerable areas.
- To enhance the stability of materials within the topographic hollows, land-cover degradation should be strongly discouraged. It is important to embark on an upland-lowland ecosystem restoration and management undertaking in the study area. Healthy ecosystems can withstand the shocks of environmental change and lead to increased community resilience to landslide hazards.
- It is also recommended that afforestation in the landslide prone and degraded areas be done in the highlands, especially along the steep sensitive topographic hollows. Vegetation cover will improve the root density in the soil, which will help to hold soil materials together. Root reinforcement will increase soil strength, reducing slope failures (Beguería, 2006; Yalcin, 2011). The presence of vegetation cover will substantially modify parameters such as cohesion, internal friction angle, weight of the slope-forming material and pore-water pressure (Das *et al.*, 2011; Pánek *et al.*, 2011; Broothaerts *et al.*, 2012; López-Davalillo *et al.*, 2014). This is through increasing cohesion due to root matrix reinforcement and suction through evapotranspiration and interception (Selby, 1993; Morgan, 2009). The restored vegetation cover will therefore lead to a stable hillslope water balance. Restoration of the vegetation cover will also help to increase humus and organic matter which hold soil aggregates together, leading to increased strength of the materials.
- Farmers should also be encouraged and helped to re-establish terrace farming, while avoiding cultivation of sensitive steep middle and upper slope sections. Terrace bans will help to reduce runoff and hold soil materials against downslope forces (Siriri and Raussen, 2002).
- To reduce disaster risk of communities to landslide hazards, topographic hollows considered to be the landslide pathways should be avoided for human settlements. Human settlements must be relocated along the spur slopes, which according to this study are noted

as landslide-free zones. The conceptual model developed by this study can also be applied to other regions with similar topographic, pedologic and land use/cover change characteristics as the present study area.

7.7 Recommended research directions

- A study on vulnerability and disaster risk reduction of communities to landslide hazards in Kigezi highlands is recommended.
- A study on the hydrological characteristics and their influence on landslide occurrence in the highlands is also very important.
- A study of the efficacy of terracing in conserving and stabilizing the fragile soils to reduce landslide susceptibility is also recommended for the area.
- An investigation into land cover restoration and management and its role in landslide mitigation is also recommended in this fragile highland region.

7.8 General conclusion

In summary, landslides in Kigezi highlands are triggered by the complex interaction of multiple-factors, including dynamic triggers and ground condition variables. There is a convergence zone within the landscape where all the parameters interact to cause landslides in the study area. The topographic characteristics of the region have a major influence on landslide occurrence, owing to the effect on hillslope hydrology and soil development. Topographic hollows between spur slopes are the potential landslide sites in the highlands. The high clay content dominated by significant amounts of illite/muscovite minerals in the sub soils is a major factor promoting landslide occurrence. This is due to their ability to absorb large quantities of water, slow drainage, high expansive and shrink/swell potential, high plasticity, and subsequent loss of shear strength. The spatial-temporal patterns of landslide occurrence in the study area have changed due to increased cultivation of steep middle and upper slopes. A close spatial and temporal relationship between land use/cover changes and landslide occurrence is discernible.

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Appendices

Appendix 1: Landslide scar dimensions and characteristics

	Scar dimensions in meters				

Slide scar No.	Average. Width	Average depth	Length	Landslide scar area in m^2	Volume of the scar in m^3	Average Gradient at slide zone	Current land use on the slope
1	3.7	1.7	402	1487.4	2461.7	27	Sorghum, beans & potatoes
2	9.66	2.3	463.5	4477.4	6064.3	30	Maize and potatoes
3	17.5	0.74	350	6125	4426.5	34	Fallow land with grass
4	2.1	1.2	602	1264.2	3772.72	27	Grazing land
5	8.5	2.245	14.1	119.9	257.98	33	Sorghum. Potatoes, beans
6	10	5	600	6000	30000	35	Eucalyptus
7	10	5	400	4000	20000	34	shrubs and grasses
8	10	5.3	498	4980	26394	33	Grazing land
9	16.6	4.3	315	5229	22600	35	Maize, beans, bananas
10	10	0.5	12.5	125	62.5	25	Eucalyptus and pines
11	5.6	0.85	525.1	2940.6	1676.2	45	Potatoes and beans
12	5.8	2.3	530	3074	10508.3	42	Fallow land
13	2.7	1.8	885	2389.5	3669	36	Bananas,potatoes, sorghum
14	3.14	1.52	786	2468	4340.54	33	Potatoes,beans & sorghum
15	2.95	1.6	784	2312.8	4517.94	37	Sorghum, maize and beans
16	6.2	2.8	835	5177	14495.6	37	Grazing land
17	4.33	2.8	752	3256.2	9412.5	32	Sorghum and beans
18	5	2.5	600	3000	7500	35	Potatoes, bananas & beans
19	2.5	2.5	653	1632.5	4081.3	37	Sorghum and potatoes
20	1.7	1.9	268	455.6	865.64	31	Bananas, maize, potatoes
21	1.2	1.4	198	237.6	332.64	27	Maize and sorghum
22	0.9	2.1	213	191.7	402.57	33	Grazing land
23	2.4	2	201	482.4	964.8	30	Fallow land
24	1.7	1.4	341	579.7	811.58	19	Bananas, maize and beans
25	2.3	1.9	189	434.7	825.93	23	Potatoes, beans, sorghum
26	2.8	1.7	244	683.2	1,161.44	29	Grazing land
27	1.9	1.2	196	372.4	446.88	34	Shrubs and thickets
28	2.4	1.5	204	489.6	734.4	18	Beans and sorghum
29	2.8	1.9	302	845.6	1606.64	22	Maize and beans
30	1.9	1.6	194	368.6	589.76	32	Potatoes, sorghum & beans
31	2.1	2	219	459.9	919.8	29	Fallow land & grazing land
32	1.3	1.8	142	184.6	332.28	26	Maize, beans and sorghum
33	2.7	2.2	408	1101.6	2423.52	31	Potatoes and beans
34	2.5	2	386	965	1,930	37	Grazing land
35	1.6	0.9	125	200	180	29	Grazing land
36	1.7	2.1	184	312.8	656.88	23	Shrubs and thickets
37	2.4	2.2	296	710.4	1562.88	26	Potatoes and sorghum
38	1.8	1.2	202	363.6	436.32	31	Bananas, beans & sorghum
39	1.4	1.7	182	254.8	433.16	34	Fallow land
40	2.7	2.2	501	1352.7	2975.94	28	Grazing land
41	2.1	1.7	234	491.4	835.38	21	Potatoes, sorghum & beans
42	1.5	1.8	267	400.5	720.9	32	Maize and beans, sorghum
43	2.2	2.2	58	127.6	281	27	Bananas, cabbage, potatoes
44	1.9	2.7	135	256.5	693	25	Sorghum, beans, potatoes
45	2.8	1.2	196	548.8	658.6	33	Fallow land
46	2.3	1.9	243	558.9	1062	31	Fallow land
47	4.2	0.9	55	231	207.9	30	Grazing land
48	2.9	1.1	129	374.1	411.5	35	Eucalyptus
49	3.2	0.7	231	739.2	517.4	32	Maize beans potatoes
50	3.1	0.8	89	275.9	220.7	22	Sorghum, beans, potatoes

51	2.8	2.1	197	551.6	1158.4	21	Bananas, sorghum, beans,
52	3.4	1.7	238	809.2	1375.6	19	Potatoes, beans
53	3.2	1.2	345	1,104	1325	34	Fallow land
54	2.8	0.8	118	330.4	264.3	18	Grazing land
55	1.8	1.1	102	183.6	202	20	Grazing land
56	3.6	2.8	189	680.4	1905.1	32	Grazing land
57	3.9	3.1	213	830.7	2575.2	28	Fallow land
58	3.2	2.7	96	307.2	829.4	29	Fallow land
59	1.8	1.2	47	84.6	101.5	25	Grazing land
60	1.2	0.8	66	79.2	63.4	22	Maize beans potatoes
61	5.9	2.1	138	814.2	1709.8	29	Banana, potatoes, beans
62	3.6	1.9	123	442.8	841.3	34	Potatoes, beans, maize
63	6.2	3.2	84	520.8	1666.6	32	Sorghum, beans, banana
64	7.0	1.7	73	511	868.7	25	Grazing land
65	4.2	1.3	144	604.8	786.2	30	Grazing land

Appendix 2: Soil particle sizes passing through different sieve (mm) sizes

soil sample	% Passing Sieve (mm)														
	75.0	50.0	37.5	20.0	10.0	6.3	5.0	2.0	1.18	0.600	0.425	0.300	0.212	0.150	0.075
1	100	100	100	98	95	90	88	85	82	79	76	73	71	68	65
2	100	100	100	100	96	91	87	85	85	79	77	74	72	69	66
3	100	100	100	97	94	91	89	87	86	82	79	76	74	73	70
4	100	100	100	99	96	92	87	86	81	77	75	73	68	65	61
5	100	100	100	93	88	83	81	75	72	68	67	65	62	60	55
6	100	100	100	100	92	86	84	78	76	73	70	68	65	63	60
7	100	100	100	98	97	92	87	81	78	75	73	71	68	66	63
8	100	100	100	100	94	83	79	75	73	70	68	65	62	60	58
9	100	100	100	100	95	92	91	86	83	81	79	76	73	70	66
10	100	100	100	96	90	88	84	81	79	77	74	70	67	65	62
11	100	100	100	99	94	91	89	86	84	80	78	73	70	68	63
12	100	100	100	100	99	97	92	88	85	82	80	77	75	72	68
13	100	100	100	98	94	91	88	85	82	79	76	72	69	65	62
14	100	100	100	100	98	94	91	88	85	82	79	76	74	72	66
15	100	100	100	100	91	88	83	81	78	75	73	71	68	64	61
16	100	100	100	100	98	94	92	91	85	79	77	75	73	69	63
17	100	100	100	100	95	92	90	88	84	80	76	74	71	66	62
18	100	100	100	92	90	82	80	77	70	69	66	60	58	55	52
19	100	100	100	100	92	89	87	82	80	77	72	65	60	57	53
20	100	100	100	97	94	91	89	87	86	82	79	76	74	73	70
21	100	100	100	99	96	92	87	86	81	77	75	73	68	65	61
22	100	100	100	93	88	83	81	75	72	68	67	65	62	60	55
23	100	100	100	100	92	86	84	78	76	73	70	68	65	63	60
24	100	100	100	98	97	92	87	81	78	75	73	71	68	66	63
25	100	100	100	100	94	83	79	75	73	70	68	65	62	60	58
26	100	100	100	100	95	92	91	86	83	81	79	76	73	70	66
27	100	100	100	96	90	88	84	81	79	77	74	70	67	65	62
28	100	100	100	99	94	91	89	86	84	80	78	73	70	68	63
29	100	100	100	100	99	97	92	88	85	82	80	77	75	72	68
30	100	100	100	98	94	91	88	85	82	79	76	72	69	65	62
31	100	100	100	100	98	94	91	88	85	82	79	76	74	72	66
32	100	100	100	100	91	88	83	81	78	75	73	71	68	64	61
33	100	100	100	100	98	94	92	91	85	79	77	75	73	69	63
34	100	100	100	100	95	92	90	88	84	80	76	74	71	66	62

Appendix 3: Temporal distribution of rainfall in Kabale highlands

YEAR	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
1980	55.6	74.3	51.9	168	189	5.7	0	39.3	104.3	0	138.5	46.8
1981	66	17.7	148.8	139.2	103.8	22.6	7.4	129.8	59.2	121.4	57	73.7
1982	17.9	9.3	41.7	271.2	130.9	15.8	7.4	2.7	108.6	0	166.5	21.3
1983	12.4	50.8	87.7	127.3	45.4	12.7	19.1	88.5	70.6	238.5	73.3	94

1984	36.2	109.3	174.4	158.3	20.4	0	49.8	21.3	78.5	117.7	98.2	129.2
1985	31.4	34.8	104.4	149.5	40.7	2.9	12.3	23.9	90.6	120.4	71.5	36
1986	0	0	103.6	183	64.2	31.9	0	16.5	44.4	120.7	72.1	487.4
1987	82.7	99.2	113.5	131.1	178.4	44.6	3.9	23.3	108	129.1	261.5	31.3
1988	83.9	90.9	161.7	139.1	82.2	4	64.2	142.6	166	132.8	60.8	47
1989	34.7	99.8	82	89.6	134.9	35.6	10.1	74.7	180.8	128.5	72.3	74.3
1990	49	161.8	128.4	182.6	72.6	0	0	45	158.5	66.5	93.6	68.9
1991	72	73	158	101	117	39	18	18	54	135	51	71
1992	18	49	151	98	50	54	27	16	149	205	89	74
1993	96	28	176	87	160	34	0	61	10	59	95	77
1994	58	70	128	150	87	2	2	79	148	125	134	93
1995	45	128	105	82	147	123	1	6	114	166	105	102
1996	70	56	146	93	46	72	50	118	123	144	202	102
1997	101	0	114	122	149	33	27	37	25	155	196	151
1998	184	97	101	171	170	19	25	23	87	154	58	80
1999	77	37	145	72	51	0	0	167	65	87	116	49
2000	50.9	83.5	118.9	120	55.7	8.1	5.9	69.6	69.9	179.8	146.8	83.8
2001	86.3	51.2	83.9	135.7	77.8	22.4	46.7	65.7	231.1	201.9	139.5	63.9
2002	120.2	89.7	63.1	74.9	115.7	0	4.4	48.4	49.5	187.6	91.5	91
2003	66.9	80.6	74.7	139.1	96	29	22.8	25.5	82.6	86.5	94	57.3
2004	69.4	93.8	84.5	183.2	84.6	0	1.1	31.9	148.8	76.9	114.2	124.9
2005	25.7	121.8	170.1	123.4	122.1	40.5	0	29	84	107.6	66.1	41.1
2006	85.5	133.5	127.7	112.7	207.8	2.9	30.1	79.5	74.2	70.7	156.2	62.3
2007	55.2	102.5	80.3	103.6	87.9	34.1	42	23.6	99.5	112.1	162.9	25.2
2008	99	65.3	206.1	54.2	53.5	65.9	24	36.5	77.4	172.8	107.6	99.1
2009	61	114.2	122.7	99.5	90.7	19.8	1.1	94.6	87	86.6	174.1	98.8
2010	97.7	189	149.1	132.9	97.7	9.3	1.4	16.4	124.1	197	86.4	71
2011	33.5	64	139.2	88.2	63.9	62.5	12.1	103.8	71.7	73.9	157.5	54.7
2012	2.8	52.8	110.6	197	146.8	11.6	9.4	62.6	95.7	114.4	179.8	128
2013	30.1	101.1	142.2	98	192	21.1	3.2	74.5	134	154	122.2	77.1

Source: Kabale meteorology station, weather data: WMO No. 63726, National No. 91290000, station name KABALE, Elevation 1867m, Latitude 01° 15’, Longitude 29° 59’.

Appendix 4: Seasonal rainfall distribution for Kigezi highlands

YEAR	DJF	MAM	JJA	SON
1970	159	422	138	324
1971	269	464	161	369
1972	341	344	218	321
1973	263	402	63	353
1974	139	379	135	247
1975	260	245	97	311

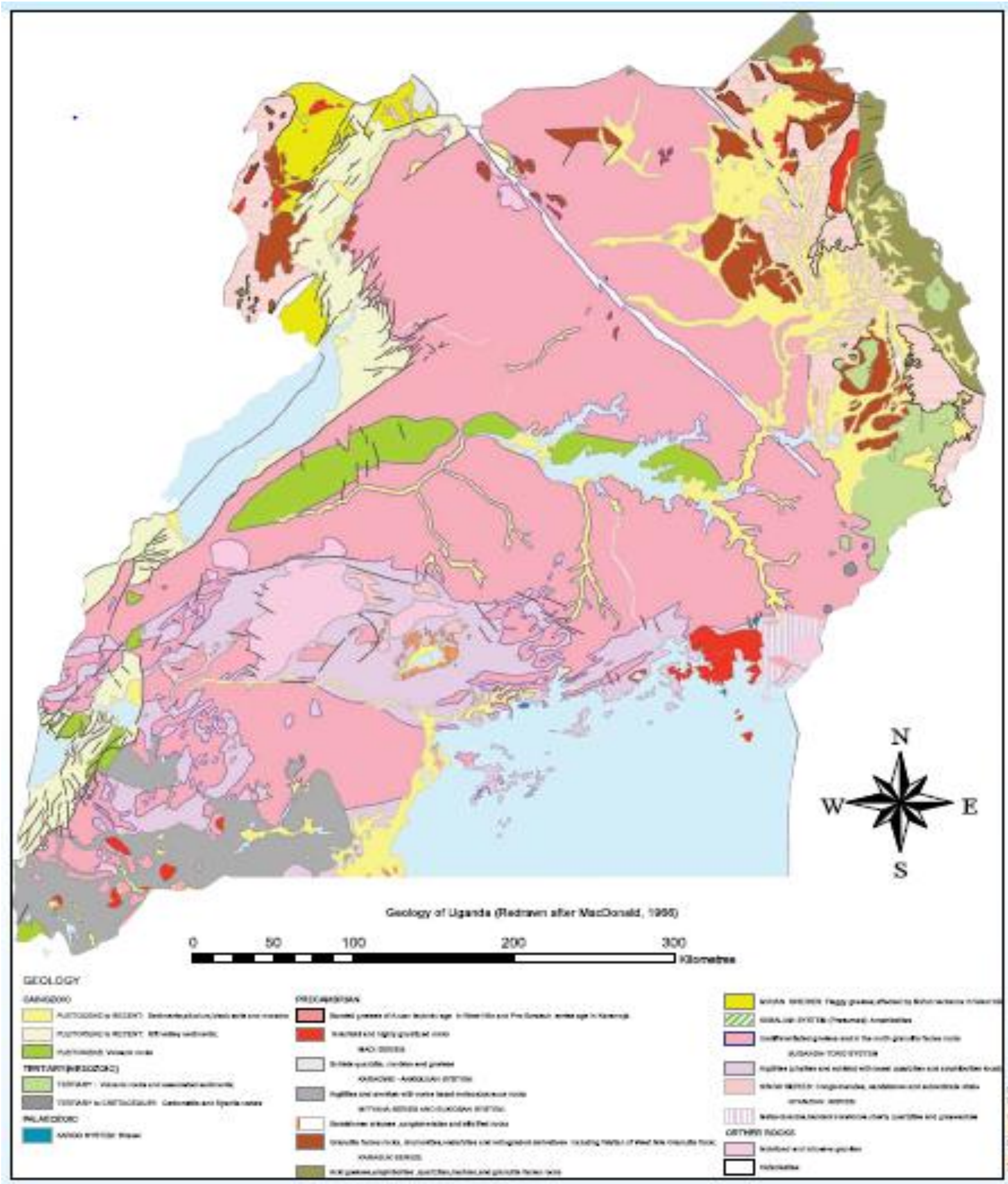
1976	185	313	157	308
1977	317	388	212	419
1978	300	384	95	349
1979	310	295	45	211
1980	177	409	45	341
1981	158	392	160	237
1982	48	444	26	386
1983	157	260	121	382
1984	274	352	71	339
1985	172	295	41	373
1986	236	351	49	237
1987	213	423	72	499
1988	222	383	211	360
1989	209	307	121	382
1990	280	384	45	318
1991	216	376	75	240
1992	141	299	97	443
1993	201	423	95	164
1994	221	365	83	407
1995	275	334	130	385
1996	228	285	240	469
1997	252	385	97	376
1998	361	442	67	299
1999	163	268	167	268
2000	218.2	294.6	83.6	396.5
2001	201.4	297.4	134.8	572.5
2002	300.9	253.7	52.8	328.6
2003	204.8	213.8	77.3	263.1
2004	288.1	352.3	33	339.9
2005	188.6	415.6	69.5	257.7
2006	281.3	448.2	112.5	301.1
2007	182.9	271.8	99.7	374.5
2008	263.4	313.8	126.4	357.8
2009	274	312.9	115.5	347.7

Source: Kabale meteorology station, weather data: WMO No. 63726, National No. 91290000, station name KABALE, Elevation 1867m, Latitude 01° 15’, Longitude 29° 59’.

Appendix 5: Location of study and sample sites

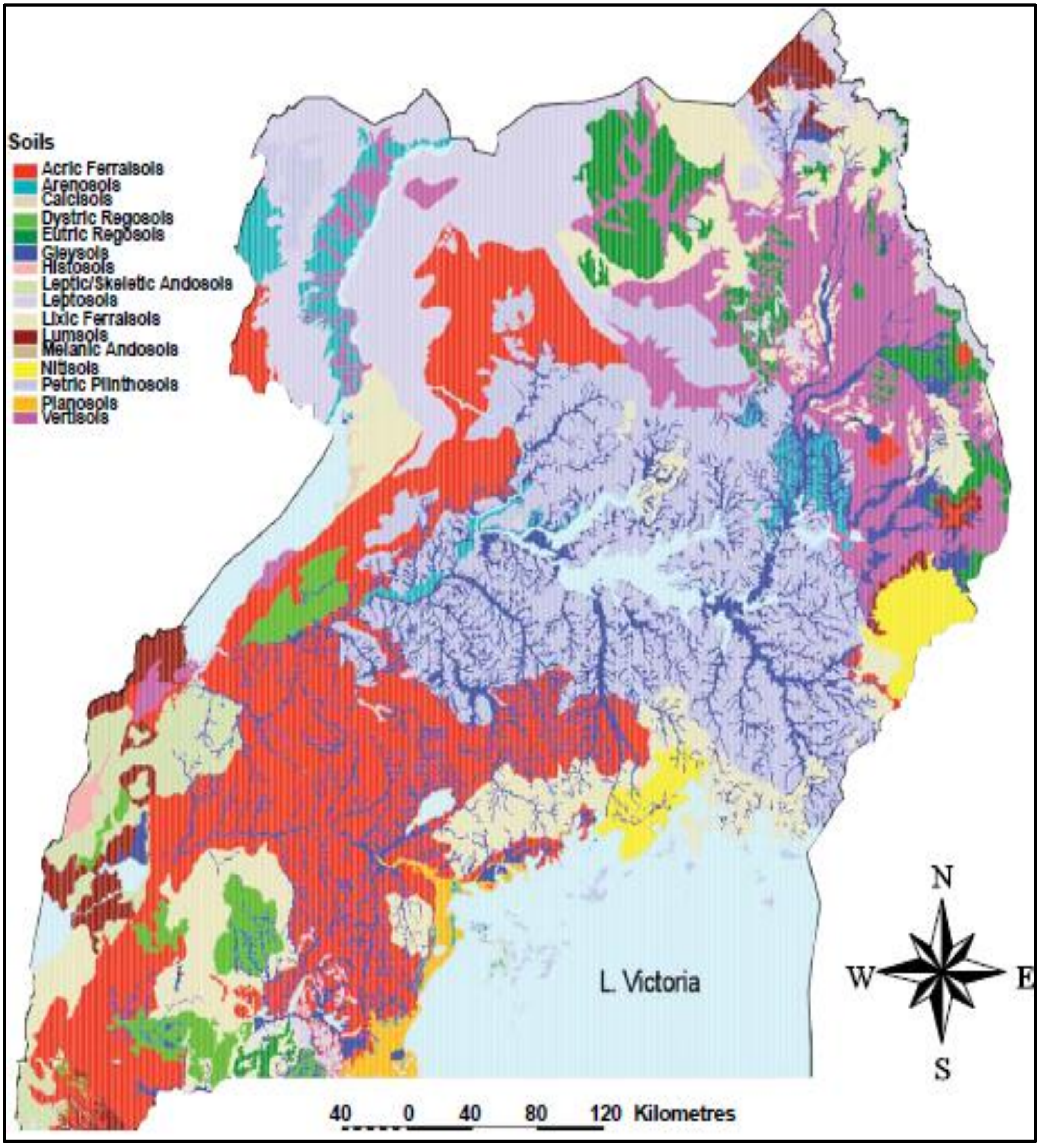
Sample-site code	Site name	Latitude	Longitude	Altitude
1	Rwenkyende	1° 09’ 18.868” S	30° 05’ 54.659” E	1780
2	Iragaza	1° 09’ 20.690” S	30° 05’ 56.369” E	1752
3	Kazindiro	1° 09’ 14.959” S	30° 05’ 55.800” E	1796
4	Kempunu	1° 09’ 12.603” S	30° 05’ 54.969” E	1817
5	Kandago	1° 09’ 11.816” S	30° 05’ 55.467” E	1815
6	Karorwa	1° 07’ 10.502” S	30° 05’ 54.854” E	1838
7	Ntungamo	1° 07’ 08.788” S	30° 05’ 53.653” E	1838
8	Kasambya A	1° 03’ 08.342” S	30° 05’ 52.590” E	1873
9	Kasambya B	1° 03’12.788” S	30° 05’ 27.354” E	1805
10	Kasambya C	1° 03’ 12.100” S	30° 05’ 27.742” E	1809
11	Rushebeya	1° 01’ 58.718” S	30° 05’ 28.267” E	1849
12	Kamwezi A	1° 08’ 57.425” S	30° 05’ 29.671” E	1843
13	Kamwezi B	1° 08’ 58.412” S	30° 05’ 35.341” E	1887
14	Kamwezi C	1° 08’ 02.596” S	30° 05’ 42.400” E	1974
15	Kyokyezo	1° 05’ 03.417” S	30° 05’ 42.320” E	1971
16	Maziba A	1° 12’05.591” S	29 35’ 37.302” E	1921
17	Maziba B	1° 12’ 05.568” S	29° 35’ 35.950” E	1928
18	Maziba C	1° 12’ 06.813” S	29° 35’ 34.823” E	2071
19	Rwamucucu A	1° 15’ 18.868” S	29° 25’ 54.659” E	1880
20	Rwamucucu B	1° 15’ 20.690” S	29° 25’ 56.369” E	1752
21	Rwamucucu C	1° 15’ 14.959” S	29° 25’ 55.800” E	1796
22	Muhanga A	0° 54’ 12.603” S	30° 02’ 54.969” E	1817
23	Muhanga B	0° 53’ 11.816” S	30° 02’ 55.467” E	1827
24	Muhanga C	0° 53’ 10.502” S	30° 02’ 54.854” E	1838
25	Nyangorogoro	0° 57’ 08.788” S	30° 00’ 53.653” E	1828
26	Nyamweru A	1° 18’ 08.342” S	29° 60’ 52.590” E	1873
27	Nyamweru B	1° 18’12.788” S	29° 60’ 27.354” E	1805
28	Nyamweru C	1° 18’ 12.100” S	29° 60’ 27.742” E	1815
29	Buhara A	1° 08’ 58.718” S	29° 55’ 28.267” E	1849
30	Buhara B	1° 08’ 57.425” S	29° 55’ 29.671” E	1843
31	Buhara C	1° 08’ 58.412” S	29° 55’ 35.341” E	1887
32	Bubare A	1° 13’ 02.596” S	29° 44’ 42.400” E	1934
33	Bubare B	1° 13’ 03.417” S	29° 44’ 42.320” E	1974
34	Bubare C	1° 13’05.591” S	29 44’ 37.302” E	1921
35	Kaharo A	1° 07’ 05.568” S	29° 65’ 35.950” E	1941
36	Kaharo B	1° 07’ 06.813” S	29° 65’ 34.823” E	2061
37	Kaharo C	1° 07’ 18.868” S	29° 65’ 54.659” E	1780
38	Bukinda	1° 10’ 20.690” S	29° 85’ 56.369” E	1752
39	Kyerero A	1° 10’ 14.959” S	29° 75’ 55.800” E	1796
40	Kyerero B	1° 10’ 12.603” S	29° 74’ 54.969” E	1857

Appendix 6: The geology map of Map



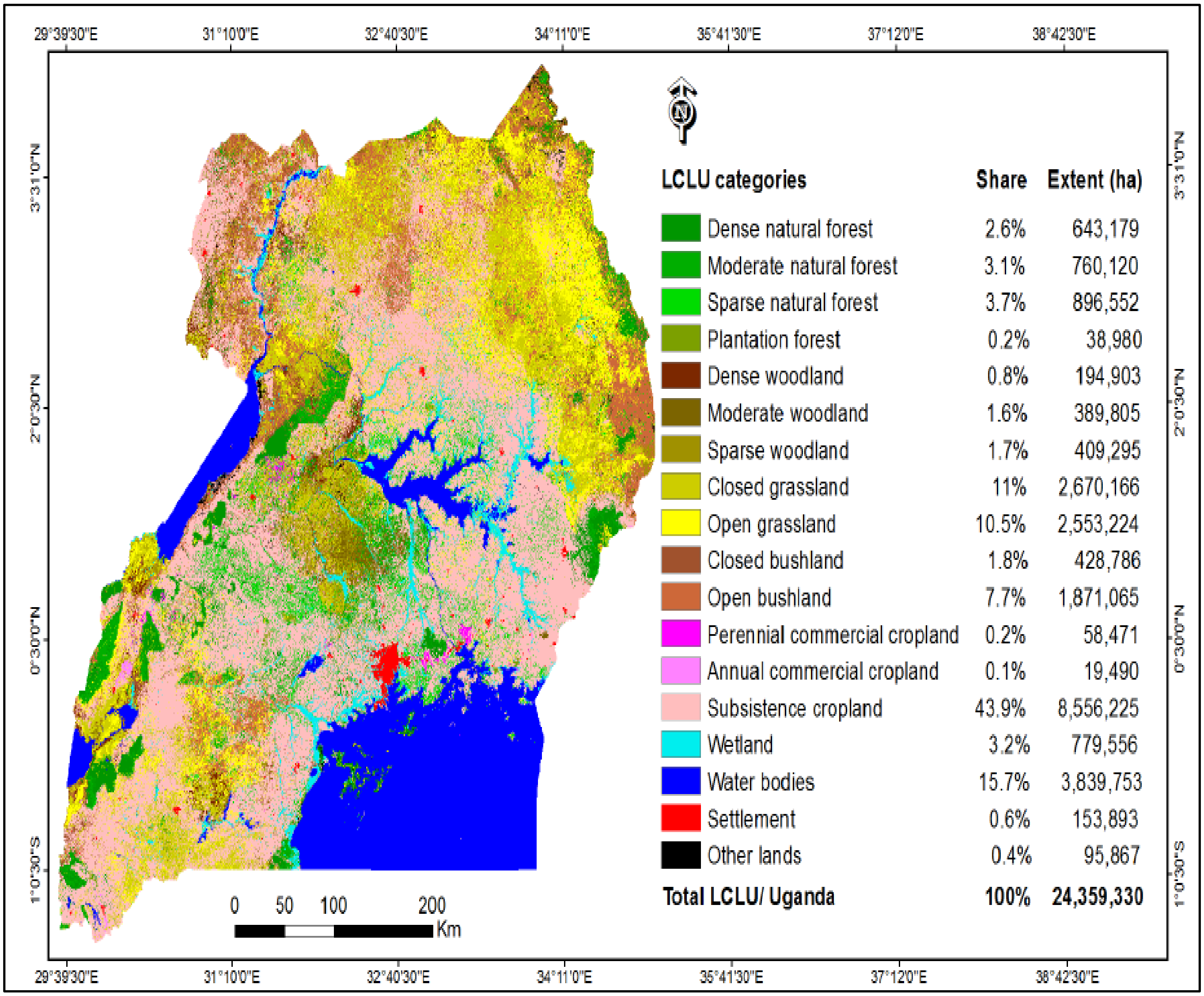
Source; Department of Geology for Uganda 1965

Appendix 7: Soil map for Uganda



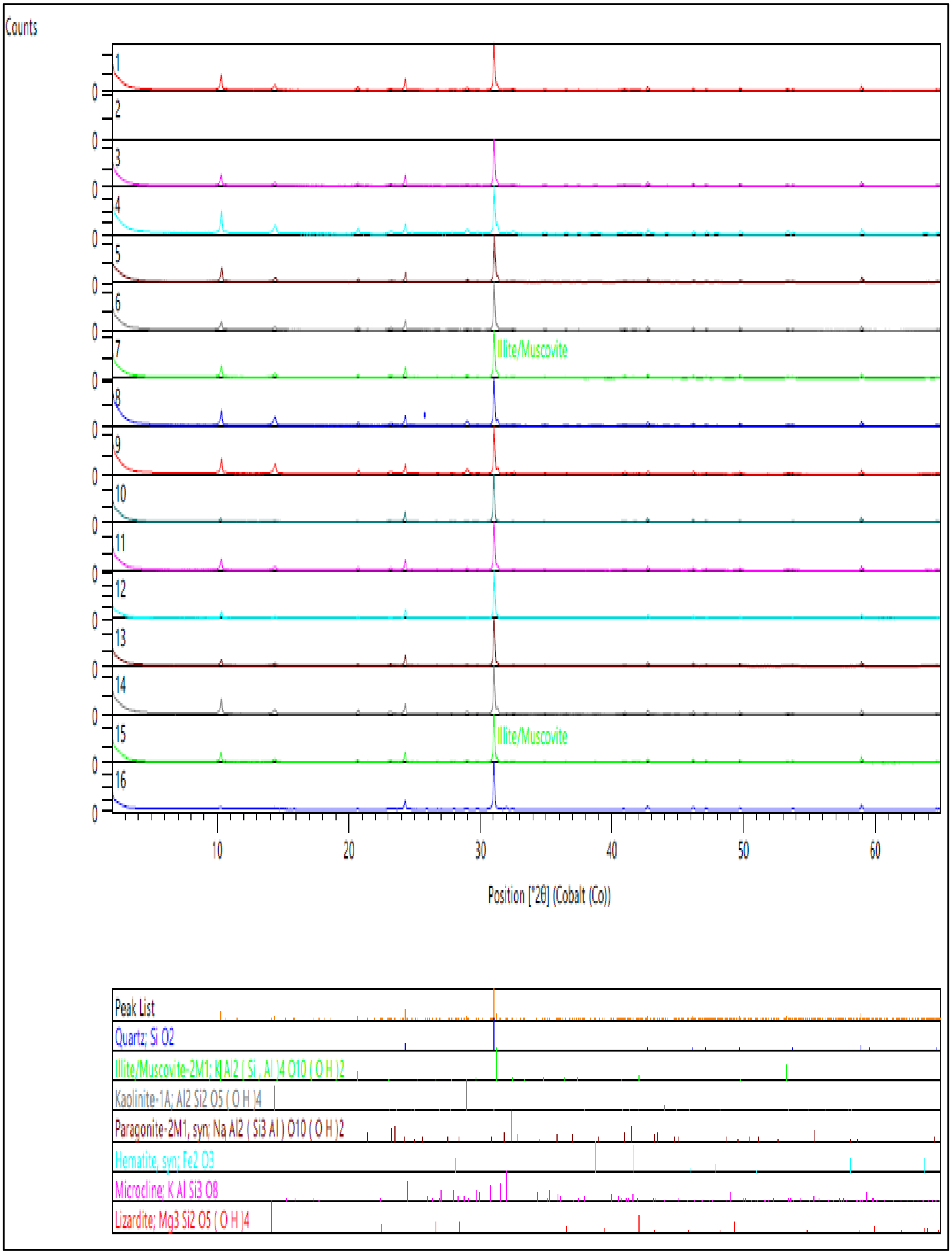
Source; **National Agricultural Research Laboratories 2008**

Appendix 8: Land cover classification for Uganda



Source: Uganda's Atlas for changing environment 2009.

Appendix 9: XRD analysis for clay mineralogy



Appendix 10: Land use and cover changes between 1985 and 2015

Years	1985		1995		2005		2015	
Landuse	Hectares	Percentage	Hectares	Percentage	Hectares	Percentage	Hectares	Percentage
Forestry	24599	40	9788	16	5315	9	4752	8
Grassland	10779	18	10265	17	9703	16	6828	11
Wetlands	9195	15	9001	14	7025	11	2641	4
Settlement	6926	11	13584	22	14677	24	15300	25
Cultivated land	10102	16	18963	31	24881	40	32082	52
Total	61602	100	61602	100	61602	100	61602	100

Appendix 11: Distribution of cultivated land along slope gradient clusters between 1985 and 2015.

	1985		1995		2005		2015	
Slope gradient	**Hectares**	**percentage**	**Hectares**	**percentage**	**Hectares**	**percentage**	**Hectares**	**percentage**
<15	5023	49.7	3490	34	4160	16.1	2108	6.6
15-25	3241	32.1	4006	39	10716	41.4	15017	46.8
25-35	1428	14.1	1850	18	8645	33.4	10008	31.2
35-45	358	3.5	804	7.8	2026	7.8	3944	12.3
>45	51	0.5	115	1.2	334	1.3	1004	3.1

Appendix 12: Distributions of cultivated land along slope positions from1985 to 2015

	1985		1995		2005		2015	
slope positions	**Hectares**	**percentage**	**Hectares**	**percentage**	**Hectares**	**percentage**	**Hectares**	**percentage**
Valley bottoms	6028	59.7	3947	38.5	3643	14.1	3234	10.1
Lower middle	2132	21.1	4682	45.6	10716	41.4	14017	43.7
Upper middle	1769	17.5	1124	11	9457	36.5	12533	39.1
Uppermost	172	1.7	512	5	2245	8.7	2297	7.2

Appendix 13: Relationship between land use/ cover and landslide distribution

	No. of landslide occurrences			
Land use/cover type	1985	1995	2005	2015
Cultivated land	3	5	11	41
Settlements	1	2	6	17
Forests with degraded areas	1	1	2	3
Grasslands	1	1	2	4
Total	6	9	21	65

Appendix 14: Temporal landslide distribution in the study area between 1980 and 2014

YEAR	No of landslides
1980	2
1988	3
1994	4
2001	3
2002	4
2003	3
2004	5
2005	5
2006	6
2007	4
2008	10
2009	13
2010	32
2011	23
2012	13
2013	10
2014	8