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Investigation of sediment source and delivery dynamics in an east African hydropower reservoir using sediment tracing technology

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INVESTIGATION OF SEDIMENT SOURCE AND DELIVERY DYNAMICS IN AN EAST AFRICAN HYDROPOWER RESERVOIR USING SEDIMENT TRACING TECHNOLOGY

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A Thesis Submitted in Fulfilment of the Requirements for the Degree of Doctor of Philosophy in Environmental Science and Engineering of the Nelson Mandela African Institution of Science and Technology

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ABSTRACT

This study aimed to reconstruct the sedimentation rates over time and identify the changing sources of sediment in a major hydropower reservoir in Tanzania, the Nyumba ya Mungu (NYM). The study also aimed to evaluate the soil carbon as a proxy for erosion risk in the catchment. Fallout ^{210}Pb measurements were used to estimate age of sediment deposits and broad changes in sedimentation rates were reconstructed. Sedimentation peaks were cross referenced to geochemical profiles of allogenic and autogenic elemental constituents of the sediment column to confirm a causal link. Finally, geochemical fingerprinting of the sediment cores and potential sources were compared using a Bayesian mixing model (MixSIAR) to attribute the dominant riverine and land use sources to the reservoir. Reservoir sedimentation generally increased from 0.1 g cm^{-2} yr^{-1} in the lower sediment column to 1.7 g cm^{-2} yr^{-1} in the most recent deposits. These results correlated to changes in allogenic and autogenic tracers. The model outputs revealed that the Kikuletwa River with 60.3%, was the dominant contributing tributary to the total reservoir sediment and the Ruvu River 39.7%. However, downcore unmixing results indicated that the latest increases in sedimentation is moistly driven by an increased contribution from the Ruvu River. Cultivated land (CU) was shown to be the main land use source of riverine sediment, accounting for 38.4% and 44.6% in Kikuletwa and Ruvu rivers respectively. The “soil slake test” method for soil aggregate stability in water (WSA) indicated a significant decrease in soil aggregate stability in cultivated land in comparison to other land use types which indicates that the unsustainable land use changes can thus potentially increase the susceptibility of soils to erosion by water when soil organic matter (SOM) is reduced. This study has explicitly demonstrated that the integration of sediment tracing and -dating tools can be used for quantifying the dominant source of sediment infilling in East African hydropower reservoirs. The results underscore the necessity for catchment-wide management plans that target to limit soil erosion and reduce further impact to rivers and reservoirs to maintain and enhance food, water and energy security in Eastern Africa.

DECLARATION

I, Aloyce Isaya Mdimi Amasi, do hereby declare to the Senate of the Nelson Mandela African Institution of Science and Technology that this thesis titled "*Investigations of Sediment Source and Delivery Dynamics in East African Hydropower Reservoirs Using Natural Tracer Technology*" is my own original work and that it has neither been submitted nor being concurrently submitted for degree award in any other institution.

Aloyce Isaya Mdimi Amasi

Name and Signature of the Candidate Date

The above declaration is confirmed by:

Prof. Kelvin Mark Mtei

Name and Signature of Supervisor 1 Date

Prof. William Blake

Name and Signature of Supervisor 2 Date

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CERTIFICATION

The undersigned certify that they have read and hereby recommend for acceptance by the Senate of the Nelson Mandela African Institution of Science and Technology a thesis titled "*Investigations of Sediment Source and Delivery Dynamics in East African Hydropower Reservoirs Using Natural Tracer Technology*" in fulfilment of the requirements for the Degree of Doctor of Philosophy in Environmental Science and Engineering at Nelson Mandela African Institution of Science and Technology.

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DEDICATION

This work is dedicated to my first son Brian Aloyce who has been suffering from mental health since day one I registered for the PhD program and throughout my studies.

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LIST OF ABBREVIATIONS AND SYMBOLS

AAR	Average Annual Rainfall
AfDB	African Development Bank
ASCE	American Society of Civil Engineers
BMM	Bayesian Mixing Model
BS	Bushland
Bq/kg	Becquerel per Kilogram
CB	Channel Banks
CoRiF	Consolidated Radioisotope Facility
CRS	Constant Rate of Supply
CU	Cultivated/Agricultural land
DL	Detection Limit
D-MixSIAR	Deconvolutional MixSIAR
DS	Riverbed sediment samples
DW	Dry Weight
EDXRF	Energy Dispersive X-ray Fluorescence
ENSO	El Nino Southern Oscillation
EU	European Union
EUROSEM	European Soil Erosion Model
FAO	Food and Agriculture Organization
FRNs	Fallout Radionuclides
GDP	Gross Domestic Product
GIS	Geographical Information System
GW	Gigawatt
g/cm^2/yr	Gram per Square Centimeter per Year
IEA	International Energy Agency
ICE	Intergovernmental Committee of Experts
ICWRER	International Conference on Water Resources and Environment Research
IAEA	International Atomic Energy Agency
IUCN	International Union for Conservation of Nature
KL	Kikuletwa River/Tributary
LOI	Loss on Ignition

MAR	Mass Accumulation Rate
mm/year	Millimeter per Year
m.a.s.l	Mean Above Sea Level
MCMC	Markov Chain Monte Carlo
MW	Megawatts
NBS	National Bureau of Statistics
NIR	Near Infrared
NYM	Nyumba ya Mungu
PBWO	Pangani Basin Water Office
PCA	Principal Component Analysis
pH	Potential Hydrogen
PRB	Pangani River Basin
PVC	Polyvinyl Chloride
SAR	Sediment Accumulation Rate
SDR	Sediment Delivery Ratio
SOC	Soil Organic Carbon
SOM	Soil Organic Matter
TAEC	Tanzania Atomic Energy Commission
t ha h $ha^{-1}MJ^{-1}mm^{-1}$	Tonnes Hectar Hour per Megajoule per Hectare per Millimeter of Rainfall
$t/km^2/yr$	Tonnes per Square Kilometers per Year
UN-ESA	United Nations-Economic and Social Affairs
USGS	United States Geological Survey
UTM	Universal Transverse Mercator
RB	Mainstem River Bank
RUSLE	Reversed Universal Soil Loss Equation
RV	Ruvu River/Tributary
WEPP	Water Erosion Prediction Project
WD-XRF	Wave Length Dispersive X-ray Fluorescence
WSA	Water Stability Aggregate

CHAPTER ONE

INTRODUCTION

1.1 Background of the Problem

Hydropower reservoirs are essential for producing climate-neutral energy (Hauer *et al.*, 2018) and ensuring long-term energy stability for economic growth in developing countries (European Union [EU], 2009). In addition, they provide other essential economic and ecological resources, such as irrigation and drinking water sources for agriculture and livestock, recreational spaces and fishing habitats (EU, 2009; Vörösmarty *et al.*, 2010; Creţan & Vesalon, 2017; Hauer *et al.*, 2018; Llamosas & Sovacool, 2020). The hydropower industry and its share of power production in East Africa are expanding linearly, while the East African population and its energy demands are growing exponentially (Palmieri *et al.*, 2003). Despite the key socio-economic services they offer, hydropower reservoirs are currently threatened by changing water supply and sediment transport dynamics in wider catchments (Palmieri *et al.*, 2003; Kondolf *et al.*, 2014). Unsustainable land use and climate changes increase soil erosion and sediment delivery rates, resulting in accelerating reservoir sedimentation (Yasir *et al.*, 2014). Consequently, water storage capacity is decreasing, and energy production capacity is declining (Yasir *et al.*, 2014). Moreover, increased sedimentation can cause flooding that may disrupt the local infrastructure. Among their longer-term negative impact, Mega-projects such as hydropower constructions could also often cause loss of life and property and involuntary resettlement which could further lead to poverty (Gutman, 1994; Vãran & Creţan, 2018). By confining sediments to reservoirs, dams also hinder sediment transfer to the downstream river system, which subsequently lacks the sediment input required for maintaining channel shape and preserving the aquatic habitats (Kondolf *et al.*, 2014). In addition, sedimentation in reservoirs can add compressional forces to the dam structure, thereby exceeding the normal hydrostatic design, while clogging of water intake also hinders the production of energy (Annandale *et al.*, 2016).

Dynamics of sediment availability in a catchment are complex in time and space and depend mainly on the climate, geology, topography, soil types, land cover and land use (Marttila & Kløve, 2010). The rapid expansion of agricultural land area with respect to population increase in Eastern Africa has led to an increase in the rates of soil erosion from large areas (Hathaway, 2008; Kidane & Alemu, 2015). In upstream catchments, fluvial processes are susceptible to land use and land cover changes on the basin scale, resulting in robust landscape reactions by modifying processes of soil erosion, sediment transport and deposition (Liébault *et al.*, 2005).

Conversely, natural climate variability and climatic changes in East Africa affect the hydrological cycle and, in turn, production capacity (Lalika *et al.*, 2015). In addition, increased runoff and gully incision also lead to increase in sediment connectivity and sediment supply leading to rapid transport of eroded sediment to downstream sinks (Blake *et al.*, 2018a). Increased erosion following land use or climate change and rapid downstream transport of eroded sediment is thus the biggest threat for the sustainability of reservoirs (Garzanti *et al.*, 2006; Yasir *et al.*, 2014). Moreover, the impacts of land use and climate change influence soil organic matter (SOM) across a range of timeframes through the harvesting of the live and dead vegetation, cropping, applying manure or compost, plowing (Mukumbuta *et al.*, 2019), deforestation (Karhu *et al.*, 2011), and afforestation (Feng *et al.*, 2018). Changes in SOM subsequently have substantial impacts on soil aggregate stability (Feng *et al.*, 2018; Miller *et al.*, 2019; Mukumbuta *et al.*, 2019) which influences their stability and erodibility. All these factors ultimately influence downstream siltation and sedimentation problems in dams/reservoirs (Yasir *et al.*, 2014; Lumbroso *et al.*, 2015).

While unsustainable land use, climate change and natural climate variability influence sediment transport (Lumbroso *et al.*, 2015), the processes by which they change catchment hydrology are non-linear in semi-arid East Africa, where the spatial and temporal dynamics of sediment connectivity are not well understood (Saavedra, 2005). Such dynamics are often neglected in reservoir planning (Harrison & Whittington, 2001). Sediment budgets as a functional reservoir management tool have rarely been established at the catchment scale in East Africa (Nyssen *et al.*, 2007). In this context, some pressing questions remain regarding hydropower management now and in the future. Are dam and reservoir systems managed in the same way the planners and designers intended (Goodwin *et al.*, 2006) concerning managing sediment accumulation? Are there any consequences of the construction and operation of the dam that were not foreseen by the designers (Goodwin *et al.*, 2006)? What are the potential sediment sources that contribute to the reservoir sedimentation? What are the dominant land uses and tributaries that are contributing to the infilling of the reservoir? What are the processes and features controlling sediment connectivity and sediment supply to reservoir sink zones? What are the best techniques to assess reservoir sedimentation rates? What approaches can reduce the quantity of sediment incoming to the reservoirs from upstream? What degree of the induced climate change variations in rainfall and temperature affect sediment delivery dynamics and can these be mitigated? What is the relationship between soil erodibility and the top soil organic carbon and can this understanding help support sustainable land management for soil conservation? These unknowns need to be answered and integrated into decision making for endorsements at early planning stages of future hydropower dams. Informed policy decisions and innovative mitigation solutions are required to move hydropower towards sustainable practices and meet the rising energy demands while ensuring water availability in East Africa.

While some exploratory work has been performed on the potential impacts of climate change and land use in East African hydropower reservoirs (Ndomba, 2007; Msuya *et al.*, 2010; Notter, 2010; Tadross & Wolski, 2010; Ehsani *et al.*, 2017) and the estimations of the sedimentation rate (Ndomba, 2007; Valimba, 2007; Ndomba *et al.*, 2008), there is little quantitative evidence available on the changing sedimentation rates and sediment sources contributing to the infilling of reservoirs. Radiometric dating using fallout radionuclides and geochemical fingerprinting can fill in this caveat in empirical data. In addition, there's currently still a lack of understanding on the dynamics and role of SOM on soil erodibility in Tanzania's complex soil systems. Although erosion risk is controlled by many factors, there is a need for evaluation of erosion risk linked to soil quality, an approach that has widespread applicability in the resource-poor agro-pastoral communities of Tanzania. In this context, SOM seems promising due to its all-embracing influence on the physical, chemical and biological properties of soils (Krull *et al.*, 2004), which makes it very sensitive to management, among other attributes.

Reconstructing changes in rates of reservoir sedimentation is crucial for evaluating the magnitude of siltation problems and therefore, the durability of hydropower reservoirs. Sediment tracing techniques evaluate the (dis)similarities between the physical or chemical traits of downstream sediments and the catchment potential sediment sources (Collins & Walling, 2004; Pulley *et al.*, 2015b; Nosrati *et al.*, 2019). Consequently the geochemical composition of downstream reservoir sediments depends on the relative contributions and geochemical properties of different tributaries (Haddadchi *et al.*, 2013; Walling, 2013; Wynants *et al.*, 2020). The proportional attribution of the tributary sources to downstream sediment can therefore be obtained through integration of the multivariate source and mixture geochemical fingerprints within mixing models (Collins *et al.*, 2010; Collins *et al.*, 2017; Blake *et al.*, 2018a; Wynants *et al.*, 2020). This is however, only possible when the eroded soils of fine particle fraction (particle size < 63 μm) transported from different watershed areas behaves conservatively from detachment, transport, deposition and after deposition. The ^{210}Pb is a natural geogenic radionuclide its deposition is continuous and basically constant from year to year (Du & Walling, 2012). Generally, the rate of decrease of $^{210}Pb_{ex}$ activity with depth in a sediment core offers the foundation for developing an age–depth correlation and for estimating sediment accumulation rates (SAR) (Mabit *et al.*, 2014). From its activity profile, it is feasible to determine the sedimentation rate and in some conditions to reconstruct environmental changes (Du & Walling, 2012) through time using numerous models comprising a number of different assumptions (Krishnaswamy *et al.*, 1971; Appleby & Oldfield, 1978; Du & Walling, 2012; Sanchez-Cabeza & Ruiz-Fernández, 2012).

These empirical models can subsequently be coupled with information on changing land use and climate to make meaningful deductions on the driving processes of increased sediment delivery. The present study therefore, will reconstruct the sedimentation rate over time in the Nyumba ya Mungu (NYM) reservoir using nuclear techniques (^{137}Cs and ^{210}Pb dating techniques) and quantify the main riverine and land use sources to the sediment using geochemical fingerprinting. The study will also evaluate the soil carbon as a proxy for erosion risk in the catchment.

1.2 Statement of the Problem

The changing demographics in East Africa create demand for land, food and water, leading to changes in land- and water use. The observed high levels of deforestation and the loss of permanent vegetation through the fast expansion of agricultural land and growing urbanization has accelerated soil loss rates and downstream siltation (Mbonile *et al.*, 2003; Tadross & Wolski, 2010; National Bureau of Statistics [NBS], 2012; NBS, 2018), thus negatively impacted hydropower production in the NYM Dam, fish catches, and other ecosystem services (International Union for Conservation of Nature/ Pangani Basin Water Office [IUCN/PBWO], 2008). Apart from ecological consequences, reduced water level in NYM Dam affects the country's economy and the livelihoods of smallholder farmers across the Pangani River Basin (PRB) who depends on the outflow for irrigation activities (Msuya *et al.*, 2010; Notter, 2010). It has been felt that the land use change pattern and other anthropogenic activities are the contributing factors for the increased sedimentation rate that resulted in a reduction in the reservoir volume (Valimba, 2007; Ndomba *et al.*, 2008), however, there is little empirical evidence available on the changing sedimentation rates and on the relative contribution of various sediment sources contributing to the infilling of reservoirs. Lack of these data has meant that relatively little is presently known about the magnitude of the impact of unsustainable land use practices on sedimentation rate and sediment delivery at the reservoir.

1.3 Rationale of the Study

There is a dearth of data on sedimentation rates in East African hydropower reservoirs. This represents a key restriction for sustainable land use and reservoir management. Data scarcity and limited studies have posed a significant challenge for national and regional planning towards reducing soil erosion. It also likely impairs the willingness of international organizations and decision-makers to invest in measures that could help tackle soil erosion for basin-wide benefit. The sediment budget concept integrates sediment transfer processes across all possible sources to all or any potential sinks in a system across the soil-sediment continuum of detachment, transport and deposition through a combination of different techniques. The integration of techniques provides the necessary information for mobilization, redistribution, transport and storage of sediments within a catchment area, including field assessment measurements, remote sensing Geographical Information System (GIS) Models, sediment core dating techniques and sediment tracing. The sustainability of hydropower reservoirs can only be preserved through continued scientific monitoring on the dynamics of soil erosion and sediment transport in the wider catchment of the reservoirs. The summary of the major issues that make an annotated bibliography are discussed in the context of the framework depicted in Fig. 1.

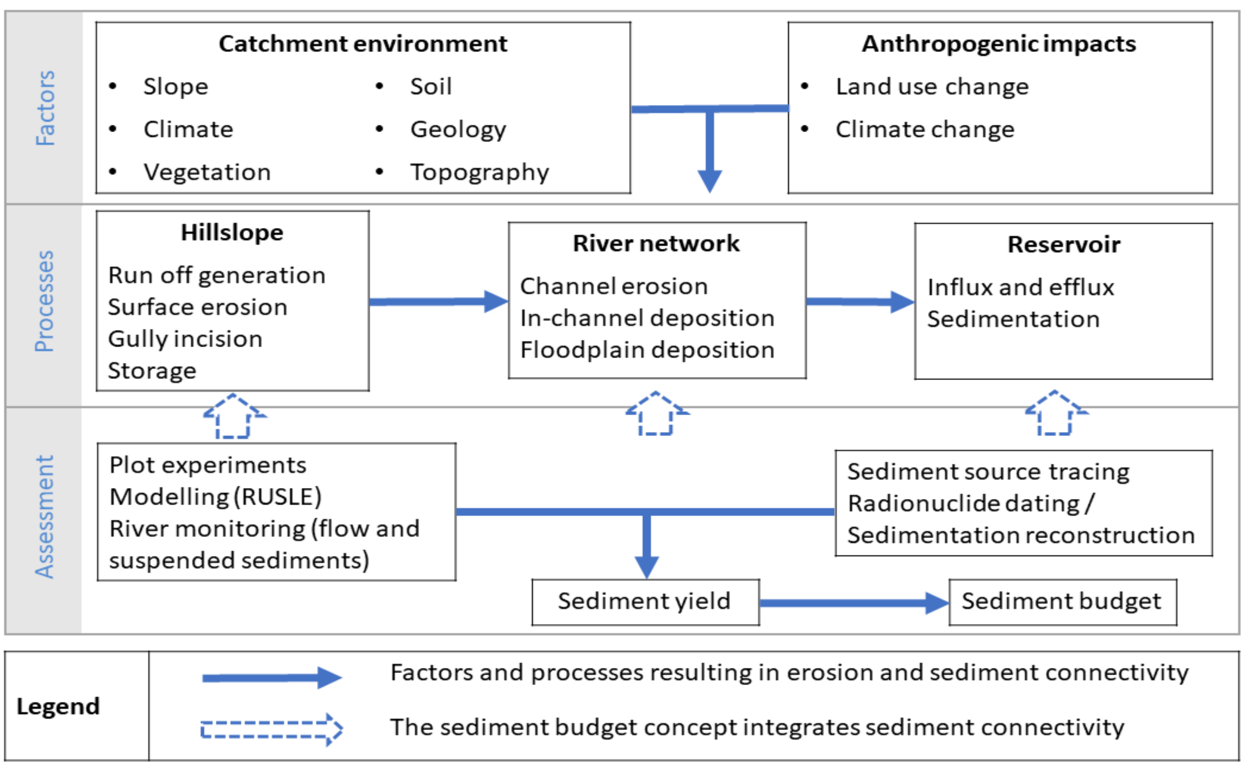


Figure 1: Sediment delivery in the complex catchment and the sediment budget processes

1.4 Objectives of the Study

1.4.1 General Objective

To reconstruct the sedimentation rate and to investigate sediment sources in the Nyumba ya Mungu hydropower reservoir using natural tracer techniques.

1.4.2 Specific Objectives

(i) To reconstruct the sedimentation rates over time in the Nyumba ya Mungu hydropower reservoir.

(ii) To establish the relative proportions of various catchment sediment sources.

(iii) To evaluate soil carbon as a proxy for soil erosion risk and sediment generation in the catchment.

1.5 Research Questions

(i) What are the potential sediment sources that contribute to the reservoir sedimentation?

(ii) What are the dominant land uses and tributaries that are contributing to the infilling of the reservoir?

(iii) What is the relationship between soil erodibility and the top soil organic carbon and can this understanding help support sustainable land management for soil conservation?

1.6 Significance of the Study

The findings from this study will improve understanding of different spatial sources of sediments and their relative contribution in the NYM reservoir as an East African exemplar catchment. The study will also promote understanding of the sedimentation rate and sediment dispersion pattern at the reservoir and other similar examples. Better understanding of the sediment dynamics will contribute to the effective sediment control strategies including remedial actions for mitigating the impacts of excessive sedimentation in reservoirs and for developing soil erosion management plans. Such knowledge can also help to reduce the costs of maintenance of water storage reservoirs and contribute to more specific plans for restoration of the reservoir, increasing capacity for water storage for irrigation and hydropower generation and increasing lifetime of the dam. This will also reduce the costs of the unnecessary dredging requirements and water treatment costs.

1.7 Delineation of the Study

This study aimed at investigating sediment sources and delivery dynamics in hydropower reservoirs using natural tracer technology. The radiometric analysis, geochemical fingerprints and the Bayesian mixing model (BMM) for source apportionment are discussed. It was observed that, differences in tracer concentrations between land uses replicated the processes distinguishing physical and chemical properties of sub-surface soils. The prevalence of deeply weathered soils in the study site and the entire sampling of the bank profile it might include a mixture of some surface and subsurface soil of which could have led to the overlap between channel banks (CB) and riverbanks (RB). Given the strong difference in signature between the Kikuletwa and Ruvu river sediment, it is assumed that intrasource variations will remain smaller than intercourse difference of which the model is strong enough to ascribe the proportional contribution of different land uses and tributaries. However, a major weakness of the used approach is that the riverine sediment fingerprint originates from samples spanning 1 year, while the cores integrate >60 years. This is the reason why the riverine sediment thus does not include potential variations in sediment fingerprint over time. This forms one of the limitations of this study and forms a gap for future studies of this kind. On another hand, the prospect of using water aggregate stability (WSA) as a rapid proxy for soil organic carbon (SOC) change by farmers where the advanced measurements are limited would offer agronomists with a new tool for monitoring soil health. However, more research is essential to initiate its potential in different soil types in a range of management scenarios.

CHAPTER TWO

LITERATURE REVIEW

2.1 Background

Soil erosion is responsible for the accelerated siltation and sedimentation in the reservoirs. Soil erosion occurs in two phases that leads to both on-site and off-site damages. First, is the detachment of discrete soil aggregates from the ‘in-situ’ soil, as a result of several processes such as precipitation events, water run-off, biochemical activity, geochemical and physical weathering, wind and other processes that disrupt the soil (Kirkby, 2008; Vercruysse *et al.*, 2017). Secondly, the detached soil particles are then transported away from their sources of origin by wind or water flow (Morgan, 2005; Vercruysse *et al.*, 2017). The mainly factors driving the soil erosion includes: (a) The erosivity of the eroding agent, (b) The erodibility of the soil (i.e. the susceptibility of the soil to detachment, entrainment and transport by the eroding agent), as determined by soil properties, (c) The topography (i.e. the slope length and steepness of the land), and (d) The nature of the surface cover, including land use and management practices (Renard *et al.*, 1997; Morgan, 2005; Vercruysse *et al.*, 2017). Apart from the erosion of soils, sediment can also originate from mass movements (such as landslides), riverbank erosion and/or anthropogenic activities and interventions in the landscape (Fryirs, 2013; Morgan, 2005; Vercruysse *et al.*, 2017). The eroded material in the catchment make a movement downstream between different zones whereby some are directly deposited to the nearest channel and others deposited before it reaches the channel and later remobilized by other more effective transporting agents and processes. The sequence of transport, deposition and remobilization has also been described as a sediment cascade (Collins & Walling, 2004; Fryirs, 2013; Vercruysse *et al.*, 2017).

The on-site effects lead to a reduction of soil fertility, often less regarded (Evans *et al.*, 2005), however, more attention is given to the off-site impacts such as sedimentation of river channels or reservoirs (Evans *et al.*, 2017), flooding that may lead to loss of life and property, disruption of the local infrastructure and involuntary resettlement (Foley *et al.*, 2005; Byerlee *et al.*, 2014), deterioration of water quality through sediments and nutrients inputs that also cause eutrophication of surface water bodies (Evans *et al.*, 2017). Furthermore, limiting sediment transfer to downstream reaches lead to loss of biodiversity (Kondolf *et al.*, 2014).

2.2 Natural Factors Responsible for Soil Erosion Processes

The intensity and frequency of erosion processes depends on the location and characteristics of the catchment. Differences in geology, topography, soil types, climate and vegetation, influence the erosion and runoff dynamics and intensities (Morgan *et al.*, 1987). The dynamics of sediment availability in a catchment associated to these factors will create site- and catchment specific erosion and run-off dynamics (Jacobs *et al.*, 2018), thus, demonstrating the importance of considering particular catchment connectivity.

2.2.1 Topography and Geology

High geological activity such as earthquakes and volcanic vibrations generate landslides (Hovius *et al.*, 2011) and seismic weakening of rocks due to fracturing (Koons *et al.*, 2012) as a result leads to high levels of erosion and sediment transport. Consequently, these effects also cause a rise in slope erosion processes and river incision as a response to catchment uplift (Whittaker *et al.*, 2010). Slope influences erosion and run-off processes mostly through its gradient and length (Morgan, 2005; Kirkby, 2008). Gravity also has an influence in net downward movement as a response to neutral soil movement processes (Benedict, 1976; Moeyersons & Ploey, 1983; Roering *et al.*, 2002). More notably though, on the steep slopes, precipitated water infiltration is less giving rise to higher energy to erode the land due to the more rapid flow and higher amount of run-off and rapidly transport downstream (Govers, 1992; Poesen, 1992). Slope length on another hand influences the upslope contributing area and thus amount of run-off discharge. The longer the slope the higher amounts of run-off it generates and therefore the higher energy to erode the land and vice versa (Govers, 1992). As a result, a more substantially increase in rill erosion with increasing slope gradient and -length may be a prospect than increase in inter rill erosion (Govers & Poesen, 1988; Fox & Bryan, 2000). The meta-analysis of Montgomery (2007) has revealed that, globally, the sloped areas have higher erosion rates than flat regions with an increase in geological erosion rates in the following order: From gently sloping lowland landscapes ($<10^{-4}$ to 0.01 mm/yr), to moderate gradient hillslopes (0.001 to 1 mm/yr) and steep tectonically active alpine topography (0.1 to >10 mm/yr). Soil erosion experiences the diversity of features on different slope gradients, although with many exceptions consistent with local conditions and history. Although slope is significant to generates the forces for erosion, e.g. gravity and water flow, the interaction with other factors like soil, climate and vegetation controls the intensities and dynamics of erosion and run-off processes (Kirkby, 2008). As an example, Hudson and Jackson (1959) established the relation that the effect of slope is stronger under semi-arid conditions, where rainfall is more intense, suggesting the importance of climate. Moreover, Morgan and Smith (1980) and Lal (1976) found evidence that variations in vegetation cover have an exacerbated effect in sloped areas (McDonald *et al.*, 2002).

2.2.2 Soil Characteristics

The erosion vulnerability of an area is of course influenced by the differences in soil characteristics because soil characteristics defines the drainage capacity and resistance of the soil to both detachment and transport. The SOC influences soil erodibility through aggregate stability, soil structure and infiltration capacity (Morgan, 2005). Land use change can thus potentially increase the susceptibility of soils to erosion by water. However, soil erodibility is additionally also influenced by geology, pedology and other human activities. The steadiness of the soil aggregates is vital for soil structure and protection against erosion. The SOM plays a crucial role in the formation of the stable aggregates of the soil. The SOC stabilizes soil structure, improves soil physical properties and enhance nutrient recycling (Martens, 2000; Six *et al.*, 2000). Soil aggregates are the building blocks of soil structure and soil aggregate stability is therefore commonly used as an indicator of soil quality (Angers, 1992; Six *et al.*, 2000; Xie *et al.*, 2015; Xie *et al.*, 2015). The aggregate stability is defined as the resistance of the aggregate soil breakdown against the external destructive effects of rainfall, runoff and wind. Generally, soils with a higher aggregate stability have higher resistance to erosion and a better water infiltration. Soil aggregate stability depends on multiple soil properties such as soil organic matter content and soil texture (Tisdall & Oades, 1982; Oades, 1984). The SOM have impact on aggregate stability through increasing the mechanical strength, increasing the cohesion within the aggregates and lowering the wettability (Onweremadu *et al.*, 2007; Król *et al.*, 2013). The SOM is influenced by natural factors, such as the changing of rainfall frequency and input of plant residuals to the soil (Chen *et al.*, 2018; Yan *et al.*, 2018). However, humans also influence SOM through the harvesting of the live and dead vegetation, cropping, applying manure or compost, plowing (Mukumbuta *et al.*, 2019), deforestation (Karhu *et al.*, 2011), and afforestation (Feng *et al.*, 2018). Changes in SOM subsequently have substantial impacts on soil aggregate stability (Feng *et al.*, 2018; Miller *et al.*, 2019; Mukumbuta *et al.*, 2019). Different size fractions in soil aggregates even have different percentages of carbon present, which influences their stability and erodibility. The macro-aggregate has less organic matter, less aggregate stability and high erodibility factor thus prone to erosion while the micro-aggregate is less susceptible to erosion (Liu *et al.*, 2017). The amount of carbon present in the size fractions of aggregates enables the determination of the amount of organic matter which will potentially be lost due to the erosion process, which adversely affects the structural condition of the soil (Kadlec *et al.*, 2012).

Poorly-drained soils generate more run-off water than well-drained soils, increasing the wash potential downstream (Morgan, 2005). The impact of raindrops on exposed soil aggregates in semi-arid East Africa, do not only cause detachment but also consolidation within the sort of a surface crust (Mutchler & Young, 1975; Tarchitzky *et al.*, 1984). When the raindrops strike the surface, the impact energy breaks the aggregates their constituent grains. Furthermore, some water is forced into aggregates, compressing air inside them, causing them to explode during process referred to as slaking (Farres, 1987). The loose grains are subsequently washed into pore spaces around intact aggregates, creating an impermeable seal (Tarchitzky *et al.*, 1984). This crust creates an impermeable surface that limits infiltration and increases runoff from subsequent rains (Tarchitzky *et al.*, 1984; Luk & Cai, 1990; Casenave & Valentin, 1992). The actual crusting response of a soil depends on its structural state, moisture content (Bissonnais, 1990; Tarchitzky *et al.*, 1984) and the intensity of the rain (Poesen & Govers, 1986). Moreover, crustability decreases with increasing contents of clay and organic material since these provide greater strength to the soil. In general, loams and sandy loams are the foremost vulnerable to crust formation. However, an exception to the present rule is cracking clay soils, which are widespread in East Africa. When these soils are wetted by the rain, they begin to swell, creating an impermeable soil crust as well (Tarchitzky *et al.*, 1984; Casenave & Valentin, 1992; Le Bissonnais, 1996; Morgan, 2005). Last, the soils within the semi-arid landscapes of East Africa have a natural high erodibility due to a combination of low SOM content, weak structural development and a high vulnerability to soil crusting or cracking (Mati & Veihe, 2001; Veihe, 2002; Morgan, 2005).

2.2.3 Climate

The climatic effect is closely associated with the precipitation amount and intensity. East Africa is characterized by a diverse seasonality in rainfall with a dry season and one or two wet seasons where the rainfall erosivity are often much higher in seasonal high intensity rains than in temperate areas (Moore, 1979; Nicholson, 1996). As a result, the higher precipitation causes a high wash erosion potential through the kinetic energy of flowing water. The entire precipitation amount, however, is not the climatic factor controlling erosion in East Africa, rather the frequency of high intensity rainfall events (Rapp *et al.*, 1972; Hudson, 1981). The study by Hudson (1981) in Zimbabwe revealed this phenomena, where he realized that about 50% of the annual soil loss occurred in just two storms which during one year even 75% of the erosion happened in ten minutes. High inter-annual variability in rainfall (Ngecu & Mathu, 1999) driven by an interplay of multiple global and local factors of which the precise details are still largely unknown (Nicholson, 1996; Souverijns *et al.*, 2016) is another essential factor for climate driven soil erosion in East Africa. El-Nino southern oscillation is behind this driving variations (Ropelewski & Halpert, 1987; Wolff *et al.*, 2011).

2.2.4 Vegetation

Vegetation affects the erosional vulnerability of the soil in various ways (Greenway, 1987; Thornes, 1990). The plant roots act as a natural anchor by increasing the frictional resistance of soil particles, increasing slope stability and eventually keeping the soil from moving (Greenway, 1987; Gyssels *et al.*, 2005; Reubens *et al.*, 2007). Furthermore, vegetation cover directly buffers the impact energy of rain, reducing the rainfall erosivity (Hudson & Jackson, 1959; Morgan, 2005).

Moreover, the vegetation root biomass and residues can potentially obstruct run-off, thus decreasing the flow energy, while contemporarily promoting infiltration. The higher biological activity of fauna and flora also aid infiltration by increasing the input of the organic matter to the soil (Temple, 1982; Greenway, 1987; Thornes, 1990). Furthest, the vegetation cover and therefore the higher content of organic matter strongly reduces crust formation (Thornes, 1990; Rhoton *et al.*, 2002).

2.3 Sediment Connectivity from Hillslope to River

The flux of sediment particles movement within the catchment requires an understanding the notion of sediment connectivity. Sediment connectivity not only details the potential for eroded soil particles movement through the system, but also addresses the spatio-temporal flexibility in sediment delivery and storage (Hooke, 2003; Bracken & Croke, 2007; Croke *et al.*, 2013; Fryirs, 2013). The hydrological connectivity essentially controls the sedimentological connectivity in catchments, and is reliant on six major drivers: (a) Climate, (b) Hillslope runoff potential, (c) Delivery pathway, (d) Lateral buffering, (e) Landscape position, and (f) Sediment propagation (Bracken & Croke, 2007; Bracken *et al.*, 2013; Wynants, 2020). Firstly, climate is the central factor because it impacts rainfall extent, duration and intensity, as well as the antecedent conditions in the catchment. Secondly, hillslopes, as the main landscape unit, are spatially variable in hydrological properties due to complex geological, pedalogical and management histories (Fitzjohn *et al.*, 1998). The factors influence the hillslope runoff, includes the slope gradient, soil characteristics, surface roughness, vegetation type and density, land use etc (Auzet *et al.*, 1993; Lal, 1990a; Puttock *et al.*, 2013; Singer & Le Bissonnais, 1998; Wynants, 2020). In this end, the sedimentological connectivity on hillslopes varies as a linear function of increasing runoff production until sediment exhaustion occurs (Croke *et al.*, 2013). Anthropogenic land use impacts can both increase hillslope connectivity by increasing run-off following vegetation removal (Guzha *et al.*, 2018; Wynants, 2020), or decrease connectivity by installing terraces and planting vegetation strips (Saiz *et al.*, 2016; Wynants, 2020). Delivery pathway of sediment from hillslope to river channel is another major factor influencing sediment connectivity. In the context of run-off flow-paths the channelized flows, has a much higher probability of connectivity to the drainage route in comparison to dispersive flows (Bracken & Croke, 2007) and thus connects sediments further downstream. The delivery pathway, however, is vastly reliant on previously discussed factors such as topography, surface roughness, vegetation and land use (Auzet *et al.*, 1993; Montgomery, 1994; Jacobs *et al.*, 2018). Another important factor is lateral buffering that limits sediment delivery to the channel through physical decoupling of hillslope to channels (Michaelides & Wainwright, 2002). The presence of wetlands, swamps or riparian vegetation in the catchment act like a buffer to the runoff coming from hillslopes thus enable decoupling (Tabacchi *et al.*, 2000; Bracken & Croke, 2007). Study by Harvey (1996) and Michaelides and Wainwright (2002) have presented the important differences in catchment response when hillslopes are directly connected to a drainage network or are decoupled. The distance from sediment source to the outlet referred to landscape position is another factor influences sediment connectivity (Bracken & Croke, 2007). The connectivity will spontaneously be higher as the distance to the stream or outlet is smaller, however, this is complex due to influence of various factors (Lal, 1990a; Bracken & Croke, 2007). Last but not least is the sediment propagation in the river system that is influenced by the dynamics of connectivity such as catchment topography (river gradient, -type and presence of floodplains), transmission losses etc (Hooke, 2003; Fryirs, 2013). Anthropogenic activities can also increase river connectivity by straightening and embanking rivers or decrease connectivity by installing check dams and slowing down river flow (Wynants, 2020). To this end, sediment connectivity is built in the mutual effects of various factors over the entire catchment that leads to vigorous interactions of runoff generations between different zones, their propagation through the catchment and eventually to large-scale response of catchments to produce floods (Bracken & Croke, 2007; Fiedler *et al.*, 2002). For instance, the research by Ambroise (2004) revealed that, although runoff can occur over a large proportion in the catchment, but may not be connected to the downslope outlet. This suggests that connectivity is exclusively depends on the rainfall extent, duration and intensity being high enough to allow transmission of water over hillslopes and into channels, and then to propagate down channels overcoming transmission losses to connect whole catchments (Lavee *et al.*, 1998; Ambroise, 2004; Bracken & Croke, 2007; Wynants, 2020).

2.4 Sediment Budget Approach

Sedimentation is a dynamic process, which involves a number of processes ranging from erosion, detachment, transportation and deposition (Kirkby, 2008). In addition, sedimentation process is controlled by the transport capacity of the flow (flow energy) (Bracken *et al.*, 2013) which when is reduced to a low level short for directing the particles in the channel or streams, leading to their deposition (Meyer & Wischmeier, 1969; Wynants, 2020). Furthermore, sedimentation depends on sediment yield, which is defined as the mass of sediment annually leaving a catchment per unit area or the sediment discharge through a river outlet per unit catchment area per unit time (American Society of Civil Engineers [ASCE], 1982). Elevated rates of erosion areas over sedimentation are named source areas since they have a net movement of soil particles away from the site, likewise, areas with elevated rates of sedimentation over erosion, will have a net deposition (Wynants, 2020). These sink areas will thus increase in soil volume over time, until the dynamics change (Borselli *et al.*, 2008). Sediment control strategies from deposition into the reservoirs after getting eroded from the catchment requires understanding of the relationship between soil erosion, sediment yield and sedimentation in the reservoir, since these parameters deal with the life of a reservoir directly or indirectly (Dutta, 2016). The sediment budget approach provides such a holistic perspective by accounting for the various sediment sources, transport, storage, sinks and redistribution when the sediment is routed through that catchment (Visser, 2003; Smith & Dragovich, 2008; Amasi *et al.*, 2021). However, the preceding studies has showed the dynamicity of storage by realizing that sediment sources, sinks, and fluxes vary widely over time and space (Smith & Dragovich, 2008; Trimble, 1983; 1999). Sediment movement within the catchment should thus be described as a soil-sediment continuum (Croke *et al.*, 2013). Essentially, all of the sediment loss from the suspended main stem river network flows is ascribed to overbank flow and deposition, with only very small amounts of sediment deposited within channel flow (Walling *et al.*, 1986; Lambert & Walling, 1987; Croke *et al.*, 2013). During high flow events, sediment-laden river water spills out over the floodplain, slows and finally becomes nearly stationary over large areas. This reduction in velocity leads to widespread deposition, as the water can no longer maintains the sediment in suspension (Ashbridge, 1995). Sediment deposition thus only occurs when thresholds for bankfull channel capacity have been exceeded and floodplain inundation occurs (Croke *et al.*, 2013). The deposited sediment can however remobilized and transported further downstream following flood surges or changes in river course (Foster, 1995). This bring about vague sediment transfer and gradual (dis) connectivity over various temporal scales (Reid *et al.*, 2007a; Reid *et al.*, 2007b; Wynants, 2020). The duration of intermediate sediment storage in various locations within the catchment will differ from location to location and from catchment to catchment (Walling & Webb, 1983). Larger catchments are thus more spatially variable, have more chances for intermediate floodplain storage and a decrease in slope and channel gradient, resulting in a more complex sediment transport from source to the end of the catchment (Walling, 1983; Wynants, 2020). Moreover, spatial variability in channel-floodplain connectivity disrupts continuity of the downstream sediment transfer between reaches and influence sediment storage alongside floodplains (Croke *et al.*, 2013). Since the catchments ends on ocean or lake and when the river reaches them, the sudden decrease in flow initiates deposition (Meade, 1972; Foster *et al.*, 1986; Foster *et al.*, 2007; Foster, 2010). The continuation series of sediment depositions generates a collective vertical silting-up, which can be presented as a sedimentary record (Trimble, 1983, 1999; Foster *et al.*, 2007; Foster, 2010; D'Haen *et al.*, 2012). Interruption of the catchment soil-sediment continuum by increased erosion, climate change or other anthropogenic factors that disrupt connectivity thus lead to changes in sedimentation (Lavee *et al.*, 1998; Croke & Mockler, 2001; Bracken *et al.*, 2013; Wynants, 2020), which in turn can have detrimental impacts on downstream systems (Pimentel, 2006). Understanding the source-to-sink sediment dynamics in catchments is therefore imperative for scrutinizing the connection between on-site disturbances and off-site response (Bracken & Croke, 2007; Wynants, 2020).

2.5 Impacts of Changing Sediment Flux Dynamics in East African Rivers to Hydropower Production

Sustainable land management and water resource development in many developing countries (Francke, 2009) are susceptible to accelerated erosion and downstream sediment transport (Morris & Fan, 1998; Le Tamene *et al.*, 2006). Siltation of reservoirs is an utmost concern in regions of semi-arid catchments where water is insufficient, and land degradation commonly leads to increased masses of sediments entering rivers and reservoirs (Smetanová *et al.*, 2020). The storage capacity of reservoirs in East Africa is being reduced by accelerated sedimentation, which jeopardizes food, water, and energy security (Oldeman, 1992; Pimentel, 2006; Blaikie & Brookfield, 2015; Wynants *et al.*, 2019). For example, Vanmaercke *et al.* (2014) showed that the sediment yields in East Africa typically range between 100-1000 ($t/km^2/y$). Studies on hydropower reservoirs also indicated similar sediment yields within the hydropower catchments of East Africa as shown in Table 1 (Abernethy, 1987; Food and Agriculture Organization [FAO], 2006; 2008; Milliman & Farnsworth, 2013; Basson, 2008; 2009).

Table 1: Sediment yields in selected catchments in the East African region

Country	Catchment	Area ×10^4 (km^2)	Monitoring dates	SY ($t/km^2/yr$)	References
Ethiopia	Awash	1.01	1959-1973	1468	Ryken *et al.* (2013)
Kenya	Tana	4.2	1968-1983	761.9	FAO (2008)
Tanzania	Rufiji	15.6	1954-1970	106	Basson (2008)
Ethiopia	Koga	0.379	2009-2010	25	Wolancho (2012)
Sudan	Atbara	2.0	1964-1976	3422	Basson (2008)
Sudan	Blue Nile	9.0	1966-1976	957	Basson (2008)

The service lifetime of a number of these reservoirs is thus reduced due to the unexpectedly high siltation rates (Wolancho, 2012). However, sparce information on reservoir sedimentation impedes the spatial analysis of the problem in the region (Basson, 2008) Table 2.

Table 2: Sedimentation rate of East African hydropower reservoir

Country	Number of hydropower reservoirs	Average sedimentation rate (%/year)
Ethiopia	1	0.52
Kenya	4	1.45
Tanzania	1	3.27
Sudan	2	2.66

FAO (2008)

Across Africa, many reservoirs have experienced similar increases in their sedimentation rates through changes in delivery from contributing sources (Shahis, 1993; Teodoru *et al.*, 2006; Hathaway, 2008; Wynants *et al.*, 2020). Sumi *et al.* (2004) noted that by 2100, about half of the global gross reservoir capacity of 6000 km^3 will be lost, ignoring new storage built after that year (Kondolf *et al.*, 2014). Similarly, Annandale *et al.* (2013) revealed that the net world capacity of reservoirs has decreased from its height of 4200 km^3 in 1995 as sedimentation rates outweigh new storage construction rates. Furthermore, Basson (2009) and Dreyer (2018) predicted that an average of 80% of reservoir capacity in several continents of the World will be filled with sediments in the following years; Africa by 2100, Asia by 2035, Europe and Russia by 2080 and Central East and North America by 2060.

The main drivers of increased erosion and accelerated sedimentation in East Africa are increasing land use pressures (Blake *et al.*, 2018; Wynants *et al.*, 2019; Wynants, 2020). The loss of permanent vegetation through the fast expansion of agricultural land (Fleitmann *et al.*, 2007; Kiage, 2013; Maitima *et al.*, 2009; Wynants *et al.*, 2018) has accelerated erosion and downstream sediment transport (Hathaway, 2008; Awulachew *et al.*, 2009). Wood and charcoal also remain the most utilized energy source within the region, which is driving the loss of forests and woodlands (Wynants *et al.*, 2018). Moreover, the increase in the number of livestock and densities on rangelands has led to overgrazing and soil trampling (Little, 1996; Ruttan *et al.*, 1999; Wynants *et al.*, 2018). The extent of the response of a catchment to loss of vegetation depends on the topography, soil, and natural climatic dynamics (Overeem *et al.*, 2013). East Africa is characterized by a steppe climate with a dry season and one or diurnal rainy season (Wynants *et al.*, 2019). These high-intensity runoff events are related to land sliding (Clark *et al.*, 2016), mudflows (Tote *et al.*, 2011), and gully erosion (Molina *et al.*, 2015), and potentially cause catastrophic flooding downstream (Gonzalez & Meneses Claudio, 2019). During such rainfall events, the erosional energy is more significant. It therefore can lead to extreme levels of the sediment transport, which increases the danger of reservoir infilling as well as serious wider ecological consequences downstream (Morera *et al.*, 2013).

While natural climatic variations and global climate change may affect erosion and downstream sediment transport (Zhu *et al.*, 2008), unsustainable land use change is plausible to magnify the impacts of hydro-climatic drivers of erosion by water with unknown outcomes for community resilience and development (Blake *et al.*, 2018). The climate-driven vegetation change that impacts the abrupt change of ecological system and ecosystems has shown to steer to more extreme responses to natural climate fluctuations (Turner *et al.*, 2020). Furthermore, global climate change alters the dynamics of river flow and discharge. The effects of global climate change on hydropower are uncertain due to regional differences, depending on changes within the flow regimes and the variation of the rainfall and temperature (International Energy Agency [IEA], 2012a). The construction of reservoirs also significantly impacts sediment connectivity by halting the downstream sediment flux (Vörösmarty *et al.*, 2003; Syvitski *et al.*, 2005; Li *et al.*, 2018). There is increasing evidence of 'hungry water' effects due to sediment starvation downstream of dams, resulting in increased channel erosion and other ecosystem impacts (Schmidt & Wilcock, 2008; Singer, 2010; Kondolf, 1997; Draut *et al.*, 2011; Ma *et al.*, 2012). Coastal areas and river deltas that depend on the supply of riverine sediment are mostly susceptible to the effects of the supply of reduced sediment (Kondolf *et al.*, 2014; Vörösmarty *et al.*, 2003). This can lead to the disappearance of beaches, increased coastal erosion (Inman, 1985; Gaillot & Piegay, 1999; Kondolf *et al.*, 2014), and the subsidence of deltas (Syvitski *et al.*, 2009). Significant proportions of the sediment transported by many rivers originated from eroded agricultural soil, consequently, the extent of this change quantifies the degraded land and the corresponding soil resource reduction (Walling, 1999). Whilst catchment erosion is known to be responsible for the accelerated sedimentation in the dams/reservoirs (Tamene *et al.*, 2006) little is understood on the spatial and temporal dynamics of erosion–sedimentation processes and sediment connectivity on a catchment scale.

2.6 East Africa's Increasing Demand for Hydropower

East Africa is undergoing rapid economic growth with Gross Domestic Product (GDP) growth rates ranging from 5.7 to 6.1, averaging 5.9% per year between 2016 and 2019 (African Development Bank [AfDB], 2019). Since 2000, the energy consumption in the region has risen by an estimated 45% (AIE, 2014; Ouedraogo, 2017). However, the development of regional energy systems has not met increasing demands (Ouedraogo, 2017). The ineffective and unreliable nature of electricity production in East Africa could limit future economic growth (Foster & Steinbuks, 2009; IEA, 2012a; Khennas, 2012). Over 82 million people in East Africa still have no access to electricity (Ouedraogo, 2017). The distribution is spatially uneven between and within the countries and the areas that do have access are dependent on a high-cost, unreliable supply (AIE, 2014; Ouedraogo, 2017). The combination of the rapidly growing population (United Nations-Economic and Social Affairs [UN-ESA], 2011; Vanmaercke *et al.*, 2014) and projected climate changes (Vanmaercke *et al.*, 2014) creates an urgent need for resilient hydropower management strategies (Vanmaercke *et al.*, 2014). A commitment to the development and sustainable management of hydropower electricity generation plants in East Africa is thus central to achieving sustainable growth (Ouedraogo, 2017; Kichonge, 2018).

Increasing, hydropower capacity, offers the potential to improve the energy security in East Africa, which is critical for the region's socio-economic growth (International Conference on Water Resources and Environment Research [ICWRER], 2013). The Renewable Energy Policy Network for the 21st Century (REN21) estimated that the region has approximately 13.4 GW of hydropower potential (REN21, 2016). However, at the moment, hardly 16% of that potential is being exploited. Currently, hydropower is by far the major source of grid electricity in the region with more than 6000 (MW) followed by geothermal (598 MW), biomass co-generation (110.5 MW), wind (25.5 MW) and solar (9.2 MW) (Otieno & Awange, 2006). In Tanzania, natural gas is also a major source of electricity production, contributing around 892.72 (MW). However, many environmental and organizational challenges impede the region's development of its hydropower potential. These include a shortage of technical know-how in planning (Intergovernmental

Committee of Experts [ICE], 2013), dynamic and unpredictable climatic and environmental conditions, increasing land use pressures, and a lack of legal and institutional frameworks for sediment management (Lumbroso *et al.*, 2015). A better institutional framework is required to effectively integrate climate information into sustainable reservoir management. While the East African countries have drafted renewable energy policies, the approval rate of hydropower technology is unsound because of the lack of financial funds of East African governments and the absence of know-how and cooperation between different stakeholder groups (Sarakikya *et al.*, 2015). Therefore, present renewable energy policies should be coordinated, and the current practice appraised to increase the implementation of these technologies (Sarakikya *et al.*, 2015). In this framework, hydropower can also be regionalized to improve grid stability and to sustain the exploitation of other sporadic renewable energy sources such as wind and solar power (Kichonge, 2018).

In view of this discussion, the mandatory use of climate change information to decid**e** the location of dams is imperative for projecting service life and risk mitigation strategies. Selection of new hydroelectric reservoir sites must consider long-term scientific data on change climate, the dynamics of erosion and sediment transport in the basin, sustainable land management planning, and the benefits of hydropower sustainability and should not be dominated by political and fiscal considerations, petitioning and negotiation.

2.7 Tools for Assessing Soil Erosion, Sediment Yield, and Sedimentation Rates to Support Sediment Management

An evaluation of the scale of sedimentation problems is required before sustainable management plans to prolong reservoir durability. Various approaches can be used to estimate soil erosion and sediment yield including experimental plots (Fullen & Reed, 1986; Nearing *et al.*, 1999), field measurements of erosional forms (Evans & Boardman, 1994; Stocking & Murnaghan, 2001; Prasuhn, 2011, 2012; Evans *et al.*, 2017), sediment tracing (Walling *et al.*, 1990; Quine *et al.*, 1991; Walling & Quine, 1991), historical documents, river sediment yields monitoring (Evans & catchments, 2006), lake and reservoir sedimentation (Walling, 1987; Butcher *et al.*, 1992; Rowan *et al.*, 1995; Foster, 2010; Foster *et al.*, 2011), aerial photography (Evans *et al.*, 2010) and modelling (Evans *et al.*, 2005) as well as remote sensing (Jain *et al.*, 2002; Durbude & Purandara, 2005). The GIS-based model is another method for estimation of soil erosion and site selection. It incorporates the use of satellite images to determine vegetation coverage for the entire basin, which determines the erosion potential of the sub-basins as well as the critical areas

(Bhattarai & Dutta, 2007; Ganasri & Ramesh, 2016). Estimation of erosion rates or modelling, however, has proved problematic for many years in spite of availability of range of approaches. For instance, studies involves modelling have been conducted at a spread of scales ranging from small hillslope plot to large catchment all of which differs in their results of the estimated erosion rates (Evans *et al.*, 2017), the concern being the decrease of sediment delivery ratio with increased catchment basin (Evans *et al.*, 2017). The prospect of various factors affecting the rates of erosion at different scales is another governing concern. Likewise, sediment yield estimates resulting from river monitoring or lake sediment-based reconstruction will underestimate hillslope erosion rates basically since not all eroded material is transported to rivers and streams (Walling *et al.*, 2002; Parsons *et al.*, 2004; Walling, 2006; Walling & Collins, 2008; Parsons, 2012), however, those that reaches the reservoirs not all is buried through the water column as other is lost via outflow due to less retention time (Appleby *et al.*, 2019). In addition, many techniques anticipates on visual approximations (Reid & Dunne, 1996), modelling (Foster, 1988), long-term field records (Gellis *et al.*, 2005), or conventional monitoring techniques. The last technique make use of indirect method and encompasses measurements of erosion activity, including those built on erosion pins to quantity the rates of surface lowering (Slattery *et al.*, 1995; Lawler *et al.*, 1999) erosion plots to document the rates of soil loss from surface sources (Motha *et al.*, 2002b). Similarly, the indirect methods encounter various concerns including: (a) Principal assumptions about the originality of sediment sources, (b) Striving in keeping records of erosion rates due to the spatial inconsistency, and (c) Inability of the methods to estimate sediment delivery to the streams (Walling, 2005).

Due to the complex source-to-sink dynamics, it is difficult to assess sediment source using traditional monitoring techniques. In addition, the convectional techniques for monitoring sediment source like erosion pins and surveys of erosion features are time consuming and costly (Collins & Walling, 2004; Foster *et al.*, 2007). Following this, the catchment sediment budget have involved a combination of several different techniques/methodologies that mutually offer the required information on sediment mobilization, redistribution, transport, and storage within a catchment (Walling *et al.*, 2001; Van Dijk & Bruijnzeel, 2005; Golosov *et al.*, 2008; Gellis & Walling, 2011; Minella *et al.*, 2014). The potential for integrating contemporary developments in sediment tracing with more conventional monitoring techniques has created new opportunities to collect the required information for sediment budget production (Walling *et al.*, 2001; Walling, 2003; Walling, 2006; Walling *et al.*, 2006; Walling & Collins, 2008). The various approaches for assessing soil erosion, sediment yield, and sedimentation rates to support sediment management are therefore discussed in details in the next subsections.

2.7.1 Quantifying Soil Loss through Experimental Plots and Field Survey

Studies of soil erosion are conducted on various spatial scales ranging from plots to continental catchments (Kirkby *et al.*, 1996). On the most miniature scale, experimental plots (Fullen & Reed, 1986; Nearing *et al.*, 1999; Evans *et al.*, 2017) and field measurements (Evans & Boardman, 1994; Stocking & Murnaghan, 2001; Prasuhn, 2011, 2012; Evans *et al.*, 2017) can be directly used to quantify the rates of erosion rates. However, these small plots (Cerdan *et al.*, 2010) are not necessarily representative of the whole catchment system (Abrahams *et al.*, 1991; Govers, 1991; Mathier & Roy, 1996; Chaplot & Poesen, 2012). Plot studies cannot easily be extrapolated to entire catchment systems and implicate substantial uncertainties when extrapolated to other catchments in different regions (Picouet *et al.*, 2001; Haregeweyn *et al.*, 2008; Meshesha *et al.*, 2011; Schmengler, 2011; Vanmaercke *et al.*, 2014; Evans *et al.*, 2016; Evans & Boardman, 2016). Moreover, plot studies can restrict information on certain types of erosion process, like the periodicity and severity of rill erosion and the components governing the between-field and within-field variations (Govers, 1991; Evans, 2002; Prasuhn, 2011). Hence, erosion rates determined on test plots may not comprehensively reflect the entire erosion in a catchment (Poesen *et al.*, 2003). Furthermore, field studies require measurements over multiple years to capture the variance resulting from natural environmental fluctuations (Pandey *et al.*, 2016).

2.7.2 Remote Sensing and Geographical Information System Models to Evaluate Soil Redistribution

In recent decades, modelling has become an increasingly important method for estimating the dynamics and quantities of eroded sediment (Van Dijk & Sampurno, 2005). Models such as the 'Revised Universal Soil Loss Equation' (RUSLE) (Renard *et al.*, 1997) the 'European Soil Erosion Model' (EUROSEM) (Morgan *et al.*, 1998), and the 'Water Erosion Prediction Project' (WEPP) (Flanagan *et al.*, 2001) have been developed to estimate erosion at different spatial and temporal scales (Karydas *et al.*, 2014). These models differ in terms of origin (e.g. empirical versus process), processes considered, complexity, data requirements and implementation potential (Pandey *et al.*, 2016). While the process-based models require larger quantities of input data and calibration routines (Fenta *et al.*, 2020), empirical models require less input data while maintaining the most factors like the physical characteristics (e.g. topography, geology, land use, climate) that effects the erosion process (Renard *et al.*, 1997; Fenta *et al.*, 2020), as long as the conditions for model development are relevant to the world of application. The process-based models are also limited in the accuracy of the soil loss rate estimation (Tamene *et al.*, 2006; Fenta *et al.*, 2020) but arguably capture process interaction and feedback more realistically. In East Africa, the combination of environmental heterogeneity and poor data availability (Fenta *et al.*, 2020) constrains the use of complex, data-hungry process-based erosion models in larger spatial domains (Fenta *et al.*, 2020). East African erosion modelling applications often must use the models in data-poor catchments (Ndomba *et al.*, 2005; Mulungu & Munishi, 2007; Ndomba *et al.*, 2008). In this context, current empirical methods such as RUSLE are extensively applied in the East African region, principally due to their average demand for data and ability to incorporate with GIS databases, which aids the upscaling process (Tamene & Le, 2015; Borrelli *et al.*, 2017; Haregeweyn *et al.*, 2017; Fenta *et al.*, 2020). With the advantage of GIS, the RUSLE model can foresee the likely erosion on a cell-by-cell basis (Shinde *et al.*, 2010), which is useful when striving to spot the spatial pattern of the soil loss present within an outsized area (Ganasri & Ramesh, 2016). The soil loss computed by RUSLE model for every pixel predicts the erosion related to runoff like the landscape heterogeneity factors (soil type, slope, topography, vegetation, geology, land use, climate) that impact the soil erosion process (Renard *et al.*, 1997; Fenta *et al.*, 2020). However, the model represents only one aspect of the entire erosion spectrum because it was established solely to predict sheet and rill erosion (Fenta *et al.*, 2020) and did not account for other erosion processes. Therefore, in areas where gully erosion, streamline incision processes are dominant (Renard *et al.*, 1997; Blake *et al.*, 2018), this model does not achieve the goal. Additionally, the RUSLE model does not predict on-site changes in susceptibility to erosion in response to process change and is less effective for studying source-to-sink dynamics in large and complex catchments (Wynants *et al.*, 2018). Furthermore, the model does not consider certain important factors for erosion dynamics, such as sediment supply and overland flow initiation dynamics (Wynants *et al.*, 2018).

Applications of the RUSLE model therefore benefit from combination with other sediment evaluation tools like sediment tracing source techniques which will provide complementary evidence to explore the knowledge of source-to-sink dynamics within the catchment. This complementarity also provides a reciprocal validation of the proportional contribution from areas of high erosion risk (Owens *et al.*, 2016; Wynants *et al.*, 2018). Coupling RUSLE models with other models for plotting susceptibility to other erosion processes (e.g. mass movements, gully-, riverine- and wind erosion), would provide an improved representation of the entire erosion susceptibility (Aksoy & Kavvas, 2005; Wynants *et al.*, 2018). Not all approaches to monitor, assess and estimate erosion are suitable at all scales (Evans *et al.*, 2017).

For example, no model matches all hydrologic conditions (Yanda, 1995; Ndomba, 2007) because each model has specific assumptions and limitations. Therefore, different methods to monitor, assess and estimate sedimentation will be appropriate at different spatial and temporal scales.

There are no particular models specifically designed for East African conditions, so critical values of model parameters for current models are likely to be beyond the constraints under which these models have been created (Visser, 2003). Most models assume a steady-state, whereby modifications in catchment environments are directly propagated to the sediment flux at the catchment outlet (Geeraert *et al.*, 2015), but ignore temporal changes in sediment connectivity. The concept of connection–disconnection between the slopes and the channel network (hillslope-sediment delivery ratio) is thus vital since the quantity of the sediment getting into the river network predominantly depends on the catchment connectivity (Brosinsky *et al.*, 2014; Vercruysse, 2017).

2.7.3 Apportionment of Sediment Sources to Identify Soil Erosion Hotspots

Pinpointing and mitigating hotspot soil erosion areas contributing to high sediment yields is a key factor for building sustainable soil-water conservation measures in reservoir catchments (Liu *et al.*, 2018; Shi *et al.*, 2019). Thus, sediment control strategies require confirmation on the relative and absolute contributions of sediment from different sources (Nosrati *et al.*, 2019). As highlighted in previous sections, traditional techniques are commonly constrained by spatial and temporal scale challenges and data availability (Peart & Walling, 1986; Collins & Walling, 2004; Nosrati *et al.*, 2019). Therefore, sediment source fingerprinting techniques have emerged to couple upstream erosion with downstream sedimentation measurements (Walling, 2013; Owens *et al.*, 2016; Walling & Foster, 2016). These techniques can offer comprehensive information of source-to-sink dynamics within the catchment and ensure the proportional source contribution and pinpointing areas of high risk (Walling *et al.*, 2014; Owens *et al.*, 2016). Sediment source fingerprinting techniques were established to underpin the similarities between the physical or chemical traits of downstream sediments with the catchment potential sediment sources (Pulley *et al.*, 2015b; Collins & Walling, 2004; Nosrati *et al.*, 2019). The technique can produce valued evidence on the relative significance of specific possible sources contributing to the downstream sediment flux of a river and reservoir (Chalov *et al.*, 2017). Such information is vital for supporting evidence on the connections between upstream potential sediment sources and downstream sediment yield (Walling & Collins, 2008), essential for targeted sediment control measures. The technique also provides essential information about the transfer of sediment through the landscape at various temporal and spatial scales (Guzmán *et al.*, 2013).

Different properties of soil and sediment can be used as tracers to distinguish between specific land use types, erosion processes and catchment zone. Fallout radionuclides (FRN) activities are usually greater in topsoil materials and less in sub soil materials (Caitcheon *et al.*, 2012; Walling, 2005) making them useful in distinguishing surface from sub surface materials as well as cultivated and uncultivated agricultural surface soils (Smith & Blake, 2014). Subsequently, sediment source apportionment using FRNs (Collins *et al.*, 2001; Collins & Walling, 2007; Smith & Blake, 2014; Pulley *et al.*, 2017) tends to be at a more generic surface-subsurface level. In this context, the use of single component signature to distinguish between the potential sources of the sediment features a high uncertainty and sometimes leads to false associations between source and sediment (Collins & Walling, 2002). Most fingerprinting studies use multivariate and composite fingerprints that encompass various distinctive diagnostic signatures affected by different environmental factors, thus improving the validity of discrimination of sediment sources (Walling *et al.*, 2006). The integration of many parameters forms a multivariate fingerprint (Walling *et al.*, 1993) that permits for an increased number of sources to be modelled and is assumed to be more reflective of the associations between sediments and their sources (Laceby *et al.*, 2017). This reduces risk of unlikely matches that might be theorized to occur with individual tracer properties (Collins *et al.*, 1996; Laceby *et al.*, 2017). Subsequently, the quantitative examination is performed to ascertain the relative contribution of every possible source to the collected target sediment samples and these often depends on Frequentist or Bayesian un-mixing models (Nosrati *et al.*, 2019). These models use multivariate fingerprints for source tracking and ascertain the relative significance of specific sediment source types in various circumstances (Walling & Woodward, 1995; Russell *et al.*, 2001; Motha *et al.*, 2003; Collins & Walling, 2007). Routinely, these models need tracer data that interpret both the sources and mixture; these qualities are anticipated to conservatively transfer from sources to mixtures through a mixing process (Stock *et al.*, 2018).

The ability of any mixing model to accurately represent source contributions to a mixture will ultimately be determined by the error assumptions and model structural choices made by the modeler. The Frequentist models commonly minimize the sum of squared residuals as outlined by Collins *et al.* (1997) with more recent approaches typically coupling parameter optimization with Monte Carlo based stochastic sampling to represent uncertainties associated with source area and target sediment variability (Collins *et al.*, 2013; Wilkinson *et al.*, 2013). However, these models are often inconsistent in their uncertainty representation and they lack the structural flexibility to coherently translate all sources of error into model results. Consequently, Bayesian mixing models have come to increasing prominence over a decade as a more robust alternative for comprehensively incorporating uncertainty into models (Massoudieh *et al.*, 2012; D'Haen *et al.*, 2013; Dutton *et al.*, 2013; Nosrati *et al.*, 2014, Blake *et al.*, 2018a). In environmental and ecological mixing problems, a key advantage of Bayesian over conventional linear mixing models is their flexible likelihood-based structure which permits better representation of inherent variability in source and mixture tracer data due to environmental processes (Cooper *et al.*, 2015; Stock & Semmens, 2016; Cooper & Krueger, 2017). Bayesian models also enable existing knowledge, in the form of 'prior' probability distributions, to be combined with new tracer data to obtain updated 'posterior' probability distributions for parameters of interest (Stock *et al.*, 2018). Fundamentally, the Bayesian approach is advantageous over Frequentist methods as it enables all known and residual uncertainties associated with the mixing model and the data set to be coherently translated into parameter probability distributions in a hierarchical framework. Since the Frequentist optimisation lacks the structural flexibility to coherently translate all sources of uncertainty into mixing model results, means that adopting Bayesian inference in sediment fingerprinting studies is preferable to the more commonly applied Frequentist approach (Cooper & Krueger, 2017).

2.7.4 Reconstructing Reservoir Sedimentation Rates

Reconstructing changes in reservoir sedimentation rates is crucial for evaluating the size of siltation problems and, therefore, the durability of hydropower reservoirs. Both non-radiometric and radiometric dating methods often estimate sedimentation rates. The non-radiometric methods (such as ecological or pollution markers) can provide distinct stratigraphic time-markers, which can be used to estimate the average rate of sedimentation between the dated layers. Radiometric dating, however, can provide a continuous age determination for lake/reservoir sediments (Carroll & Lerche, 2003; Mabit *et al.*, 2014). The FRNs, ^{210}Pb and ^{137}Cs are employed to study erosional records of a catchment and, therefore, the effects of land use and climate by presenting data over the last 100-150 years for different time windows (Mabit *et al.*, 2014). The fundamental ability of $^{210}Pb_{ex}$ to provide evidence on the chronology of a sediment deposit and thus estimate the sedimentation rate depends on its source, its moderately long half-life, its global distribution, and its retrospective assessment that provides a longer-term (ca 100 year) chronology or age-depth relationship (Appleby, 2001). The ^{137}Cs is an anthropogenic radionuclide from weapon testing fallout that peaked in early 1960s. However, its fallout in tropical Africa was low, challenging its application (Walling & He, 2000). The ^{210}Pb is a natural geogenic radionuclide its deposition is continuous and constant from year to year (Du & Walling, 2012). Generally, the rate of decrease of $^{210}Pb_{ex}$ (i.e. the fallout component) activity with depth in a sediment core offers the foundation for developing an age-depth correlation and estimating sediment accumulation rates (SAR) (Mabit *et al.*, 2014). From its activity profile, it is feasible to determine the sedimentation rate and, in some conditions to reconstruct environmental changes (Du & Walling, 2012) through time using numerous models includes the Constant Flux: Constant Sedimentation (CFCS), the Constant Initial Concentration (CIC) and the Constant Rate of Supply (CRS) models accounting for a number of different assumptions (Krishnaswamy *et al.*, 1971; Appleby & Oldfield, 1978; Du & Walling, 2012; Sanchez-Cabeza & Ruiz-Fernández, 2012).

For example the Constant Flux: Constant Sedimentation (CFCS) model developed to interpret the rate of decrease of ^{210}Pb with depth in marine and lake sediment deposits assumes that the flux of $^{210}Pb_{ex}$ from water to sediment is constant through time (i.e. from year to year) and also that the sedimentation rate is constant (Robbins, 1978). The model did not consider the season or climatic variability factors such as precipitation, for instance the annual atmospheric depositional fluxes of ^{210}Pb vary considerably in a large number of places, depending on the amount of precipitation as a result sedimentation rates will be variable as well as the ages of the sedimentary layers (Robbins & Edgington, 1975; Robbins, 1978; Appleby, 1998; Appleby & Oldfield, 1992; Swarzenski *et al.*, 2006; Jweda & Baskaran, 2011;).

The constant initial concentration (CIC) model in the other hand assumes that the initial $^{210}Pbex$ concentration in the deposited sediment is constant (Appleby & Oldfield, 1978) and therefore that an increased flux of sediment particles from the water column will be accompanied by a proportionally increased flux of $^{210}Pb_{ex}$ from the water to the sediment (Appleby & Oldfieldz, 1983) i.e the ^{210}Pb supply varies directly in proportion to the sedimentation rate (Shukla *et al.*, 1989). While bioturbation and mixing rates are the main processes affecting the vertical distribution of radionuclides in the lake or seabed, the vertical profiles of the $^{210}Pb_{ex}$ will therefore be affected since the vertical profiles of $^{210}Pb_{ex}$ are governed by the changes in the fluxes of sedimentary particles and ^{210}Pb. Bioturbation enhances interactions between sediments, interstitial waters and the overlying bottom water, thereby greatly influencing early sediment diagenesis. This process affects the physical structure of the sediments and burial efficiency (Ming-Yi *et al.*, 1993; Green *et al.*, 2002) as well as fluxes of nutrients, oxygen, contaminants and pollutants, and more generally strongly influences the process of organic matter mineralization near the water–sediment interface (Lee *et al.*, 1980; Aller, 1982; Gilbert *et al.*, 1995; Kristensen, 2000; Turnewitsch *et al.*, 2000; Furukawa *et al.*, 2001). As such there are still considerable gaps in knowledge of this process in sedimentation studies especially in sedimentary systems where biorturbation occurs (Wilkinson, 1997; Green *et al.*, 2002).

Most of the East Africa hydropower reservoirs are located on complex catchments which encounter catchment-wide environmental changes (Wynants *et al.*, 2020). In this context, the Constant Rate of Supply (CRS) model developed by (Appleby & Oldfield, 1978) is the most applicable to account for changes in the rates of sedimentation the initial concentration of $^{210}Pb_{ex}$ activity in the sediment (Wynants *et al.*, 2020). The CRS model (Appleby & Oldfield, 1978; Robbins, 1978; Benoit & Rozan, 2001) depends on the assumption that the ^{210}Pb flux to sediment is constant over time, while the sedimentation rate may vary (Sanchez-Cabeza *et al.*, 2000; Persson & Holm, 2011). In the model, the attention is focused to the downcore reduction in $^{210}Pb_{ex}$ activity, which in turn reflects the sedimentation rate and natural radioactive decay, whereby high sedimentation rates will result in slower declines in the vertical $^{210}Pb_{ex}$ activity profiles. On the other hand, lower sedimentation rates will result in steeper decreases of the vertical $^{210}Pb_{ex}$ activity profiles (Du & Walling, 2011, 2012).

Geochronological model assumptions might be challenged, however, when a substantial proportion of $^{210}Pb_{ex}$ supply entering the water column derived from mobilized catchment material (Appleby *et al.*, 2019; Wynants *et al.*, 2020), where differences in the existing $^{210}Pb_{ex}$ activities of the transported and deposited sediment might occur due to the natural differences in the geological prevalence of ^{238}U and or variation in dominant erosion processes (Wynants *et al.*, 2020). Additionally, the changes in dominant abrasion processes within a channel network can alter the fraction of topsoil versus subsurface material within the transported sediment, thereby affecting the $^{210}Pb_{ex}$ activity of input sediment to sediment column (Aalto & Nittrouer, 2012; Du & Walling, 2012; Baskaran *et al.*, 2015; Wynants *et al.*, 2020). This variability in the input of $^{210}Pb_{ex}$ requires independent methods to scrutinize the CRS model (Smith, 2001). Most often, the ^{137}Cs ($t_{1/2}$ = 30.17 years) peak fallout has been used (Appleby, 2008). In the southern hemisphere, however, the activity concentration of ^{137}Cs in soil and sediment is low, and in some cases, the geochemical profiles of sediment cores have been shown to exhibit changes that might have been associated with hydrological or volcanic events (Arnaud *et al.*, 2006; Łokas *et al.*, 2010; Wynants *et al.*, 2020) that pre-concentrate detrital ^{137}Cs input (e.g. through selective erosion of fine sediment from the catchment) instead of direct fallout intrinsically. Other limitations for the determination of SAR using ^{210}Pb occur when the environmental settings posed special interpretive problems like depositional regime dominated by episodic large-scale turbidity currents or debris flow. In this situation it is difficult to estimate SAR quantitatively because the stratigraphic sequences are either reworked or mixed by gravity flows or are interspersed with occasional event layers that compromise $^{210}Pb_{ex}$ profiles (Krishnaswami & Lal, 1978) but in many cases an indication of broad rates of SAR change can still be determined which is of value to managers.

2.7.5 The Sediment Budget as a Foundation for Sustainable Reservoir Sediment Management

Understanding the processes that result in erosion and its connectivity to the river channel, storage in hillslopes, floodplains, and sediment accumulation in the reservoirs is vital for the choice of dam location and for the sustainable management of the reservoirs (Zarfl & Lucía, 2018). Sediment connectivity processes through time integrate sediment transfer processes across and sinks along the soil-sediment continuum of detachment, transport and deposition (Bracken *et al.*, 2015). The process of sediment delivery in the catchment is complex; it involves the interaction of multiple factors and processes on different spatial and temporal scales (Jetten & de Roo, 2001; Porto *et al.*, 2011). These complex systems cannot be understood by examining outcomes alone (e.g. sediment yield or SDR) (Visser, 2003). The complexity of processes, feedbacks and consequences require a system-wide perspective (Visser, 2003). The sediment budget approach provides such a holistic perspective by accounting for the various sediment sources, transport, sinks and redistribution when the sediment is routed through that catchment (Visser, 2003). Policy makers and catchment managers can use the sediment budget approach as a realistic mechanism for targeting mitigation measures/ strategies (Wilkinson *et al.*, 2005; Walling & Collins, 2008).

Development of suitable sediment management strategies entails the quantification of sediment flux and links their transport dynamics to drivers both within the channel and the broader catchment to reliably forecast sediment discharge in rivers over relevant time scales of management (Gao, 2008; Taylor & Owens, 2009; Vanmaercke *et al.*, 2011; García-Ruiz *et al.*, 2015; Vercruysse, 2017). Nonetheless, the spatial and temporal aspects of sediment transport factors and process interactions in rivers have not been fully captured and understood yet (Vercruysse, 2017). The potential of employing sediment budgets to improve understanding on the catchment fluxes has increased following the latest established advanced techniques and further evolved insights (Brown *et al.*, 2009). The quantification of catchment-wide sediment budgets involves a large number of components to be integrated at various spatial scales and for prolonged timescales.

Although the essential requirements of budgeting sediments are steadily developed and extensively used (Brown *et al.*, 2009), there has been a limited application to support mitigation of hydropower sediment problems. Nonetheless, there is much potential here to be exploited. Field assessment measurements can provide an empirical quantification of sediment storage, erosion processes, flux rate or water/particle residence time (Evans *et al.*, 2017). Modelling has the potential to provide the functional relationships between erosional processes and dominant factors influencing rates of erosion and predict sediment yield within catchments both in spatial and temporal scales (Nyssen *et al.*, 2004; Syvitski & Milliman, 2007; Kettner & Syvitski, 2008; Brown *et al.*, 2009; Vanmaercke *et al.*, 2014). Sediment source tracing has the potential to establish hillslope-channel connectivity knowledge that provides new opportunities and skills for establishing sediment sources, obtaining spatially distributed and temporally integrated data on sediment mobilization, delivery and storage (Brown *et al.*, 2009). Sediment core dating techniques provide an opportunity to reconstruct changes in sedimentation rates over time, which ultimately allows the association of sediment flux with forcing factors, including climate and human activity (Brown *et al.*, 2009). The age-depth model is often taken as a proxy for the assembly of a chronostratigraphy for sediment budgets and to estimate catchment erosion (Dearing & Foster, 1986; Dearing & Zolitschka, 1999). However, the notion of sediment budget involving the quantification of sediment storage components remains challenging and time-consuming (Hinderer, 2012). Following this, most studies that have been undertaken to determine a catchment sediment budget have involved a combination of several different techniques/methodologies that mutually offer the required information on sediment mobilization, redistribution, transport, and storage within a catchment (Walling *et al.*, 2001; Van Dijk & Sampurno, 2005; Golosov *et al.*, 2008; Gellis & Walling, 2011; Minella *et al.*, 2014). The potential for integrating contemporary developments in sediment tracing with more conventional monitoring techniques has created new opportunities to collect the required information for sediment budget production (Walling *et al.*, 2001; Walling, 2003; Walling, 2006; Walling *et al.*, 2006; Walling & Collins, 2008). To this end, poor reservoir planning during the design phase remains the main reason for the rapid sedimentation and anticipated sediment yield. The absence of sediment yield data and absence of suitable methodologies to forecast sediment yield is an attribute of poor planning of reservoir during the design phase. In this context, sediment budgeting remain**s** an imperative method for comprehending and forecasting sediment delivery to the reservoir basin as one of the mitigation strategy goals. This method should not be replaced by faster sediment flux quantification approaches, instead, the synergistic application of both approaches improve tackling of hydropower sediment challenges.

CHAPTER THREE

MATERIALS AND METHODS

3.1 Overview

This Chapter provides the detail description of the study area, sampling strategies and a comprehensive narration of the analytical methods. The fundamentals of the Bayesian mixing modelling from which the model is draws to quantitatively compare different sources with the lake mixture and the assumptions from which the model was built in the MixSIAR framework are given in details. Data analysis includes radiometric, geochemical and statistical analysis using SPSS (Statistical Package for Social Science) are also presented in this Chapter.

3.2 Description of the Study Area

Nyumba ya Mungu is a man-made reservoir that largely replaced a natural wetland when the rivers Kikuletwa and Ruvu were impounded. The NYM catchment includes the highlands of Africa's highest peak, Mt. Kilimanjaro (5985 m), and fifth highest peak, Mt. Meru (4566 m). The reservoir is part of the upper Pangani River Basin (PRB) and receives water from two main tributaries, the Kikuletwa and the Ruvu River (Fig. 2). The Kikuletwa sub-catchment covers about 6650 km^2 out of 13 000 km^2 and the Ruvu approximately 5350 km^2 of the total catchment. The dam was erected in 1965 for hydropower generation but in later years, irrigation potential was realized and integrated into strategies (Lein, 2002; Ndomba *et al.*, 2008; Lalika *et al.*, 2015). The NYM reservoir is about 150 km^2, has a live storage capacity of about 875 million m^3 (Mzuza *et al.*, 2017) and has a maximum depth of 40 m (Shaghude, 2006; Hellar-Kihampa *et al.*, 2012). However, due to the highly variable climate and changing sedimentation these factors fluctuate seasonally and between years.

The catchment of the NYM reservoir is located between Latitudes 3°00'00" and 4°3'50" South, and Longitudes 36°20'00" and 38°00'00" East, and its altitude ranges between 700 and 5825 m.a.s.l. The ice cap at the peak of Mount Kilimanjaro forms the highest ground in the catchment. The catchment occupies a total land and water area of about 13 000 km^2 (Ndomba *et al.*, 2007) and experiences a tropical climate that provides high levels of precipitation with average annual rainfall (AAR) of 900-2200 mm/year at 800 m.a.s.l and 2200 m.a.s.l, respectively (Rohr & Killingtveit, 2003; Shaghude, 2006; IUCN/PBWO, 2008; Hellar-Kihampa *et al.*, 2012) (Fig. 3). The catchment experiences bimodal rainfall; occurring mainly in March to May with short rains in November and December (Kijazi & Reason, 2009; Mahongo & Shaghude, 2014). The temperature gradient of the catchment is closely related to altitude ranges from 15^{o}C to 33^{o}C where the maximum and minimum temperatures occur between February and July, respectively (IUCN/PBWO, 2008) (Fig. 4). Catchment's geology is volcanic comprising olivine and alkaline basalts, phonolites, trachytes, nephelinites and pyroclastics (IUCN/PBWO, 2008; Schlüter, 2008; Mzuza *et al.*, 2017). The major soil types in the watershed comprise Nitisols, Luvisols, Solonchaks, Chernozems, Leptosols and Histosols (Dewitte *et al.*, 2013) (Fig. 5). The Nitisols cover the highlands to the lowlands and are predominantly developed on volcanic material. They are usually deeply and well-drained, have a stable structure, and a high clay and nutrient content. With proper management, they have medium to high potential for rain-fed agriculture. The Luvisols are mostly constrained to the lowlands of the catchment. They are highly weathered with a subsurface accumulation of clay and are characterized by low nutrient retention, and a high susceptibility to surface crusting and erosion. However, with proper management, they have a medium agricultural potential. The Solonchaks are located in lowland depressions or salt pans are characterized by high rates of evaporation of runoff water leaving a high concentration of soluble salts. They have a limited potential for cultivation, only with salt tolerant crops. Most solonchaks are therefore used for extensive grazing or as natural reserves. Histosols are acidic, organic soils that form when fallen plant material decomposes more slowly than it accumulates (McClaugherty, 2001). They are constrained to wetlands on the upper parts of Mount Kilimanjaro and Mount Meru, where they have formed under almost permanently saturated conditions. Chernozems are the dominant soil type in the Kikuletwa sub-catchment. They are fertile soils that are currently mostly used for agricultural production. These soils are characterized by a high degree of biological soil mixing and soil organic carbon, leading to the formation of biologically stabilized soil aggregates on the soil surface (Schaetzl & Thompson, 2015). Leptosols are generally weak aggregated coarse or medium-textured soils with limited profile development, mostly located in the highlands and in the Kikuletwa sub-catchment. Soil erodibility factor of the dominant soil types in the catchment ranges from 0.012 to 0.026 t ha h ha^{-1}MJ^{-1}mm^{-1}, according to Fenta *et al.* (2020) which suggests that the catchment has significant soil aggregate stability. The catchment land cover types changes in response to the changing elevation ranging from montane forests on the higher altitudes to semi-arid in the lower slopes. The major land cover types include the natural forests, woodlands, grassland thickets with emergent trees, bushland and plantation forests (Turpie *et al.*, 2005). The majority of the population settlements is located on the lower slopes between 900 and 1800 m.a.s.l where most of agricultural activities are concentrated. The ever increasing demand for food with an increasing population in the NYM catchment in the Pangani River Basin in the

Northern Tanzania has led to rapid expansion of agricultural land, thus accelerating soil loss rates and downstream siltation of the reservoir (Hathaway, 2008; Awulachew *et al.*, 2009; Amundson *et al.*, 2015; FAO, 2015; Borrelli *et al.*, 2017; Wynants *et al.*, 2020).

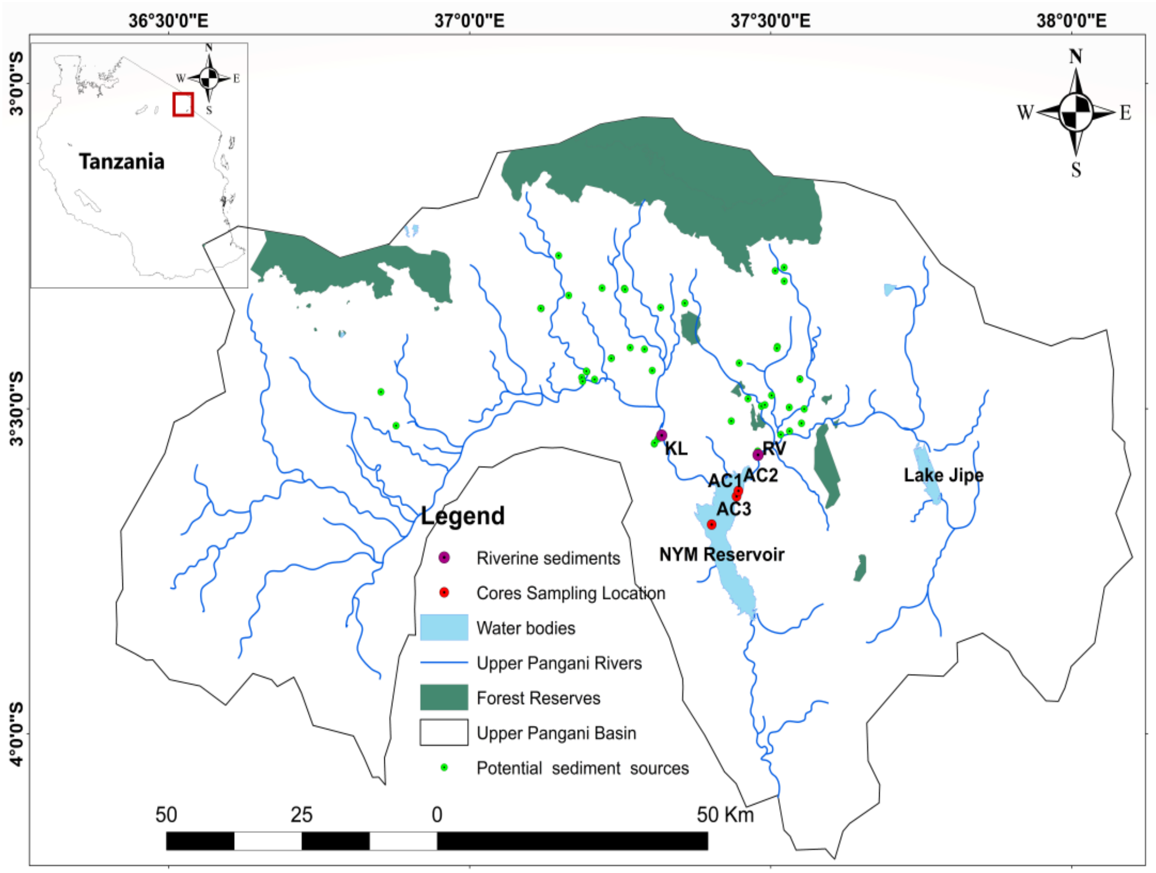


Figure 2: Location of the Nyumba ya Mungu catchment detailing the catchment major tributary inlets, the Kikuletwa (KL) and Ruvu (RV), the riverine sampling locations (purple) and sediment cores sampling locations (red marks) and potential sediment sources (green)

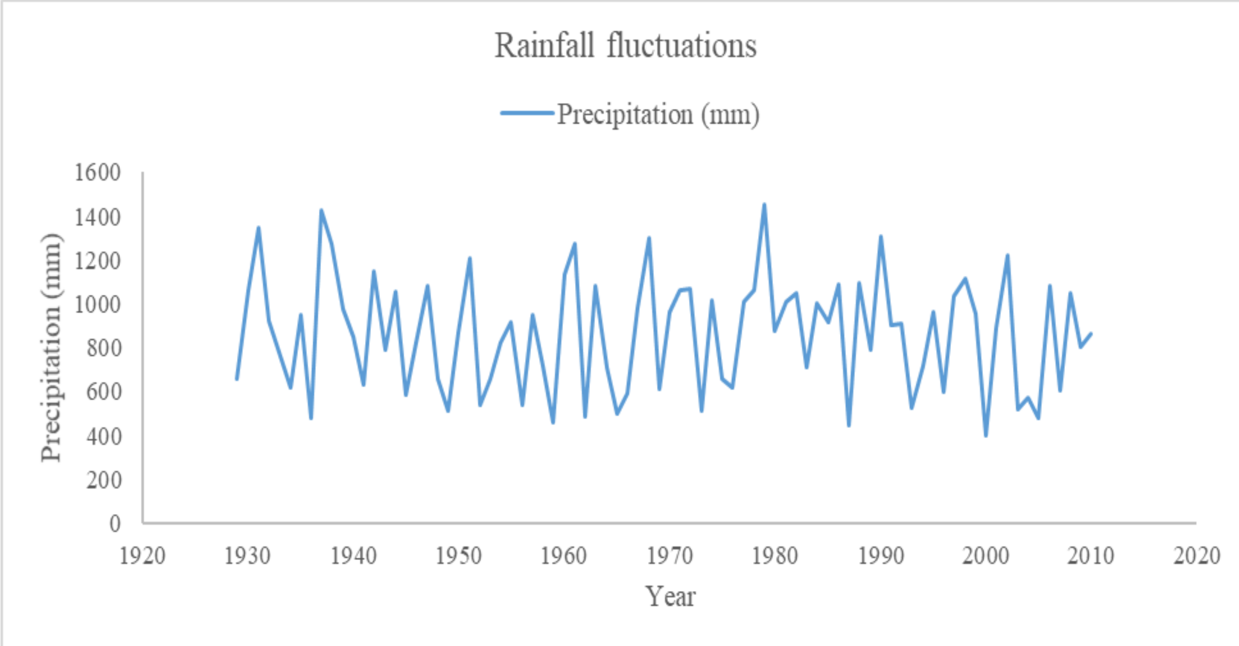


Figure 3: Changes in annual precipitation (mm) measured at Moshi Airport in Moshi Municipal, Kilimanjaro station No: 9337004 coordinate 3° 21' 0'' S and 37° 19' 48'' E

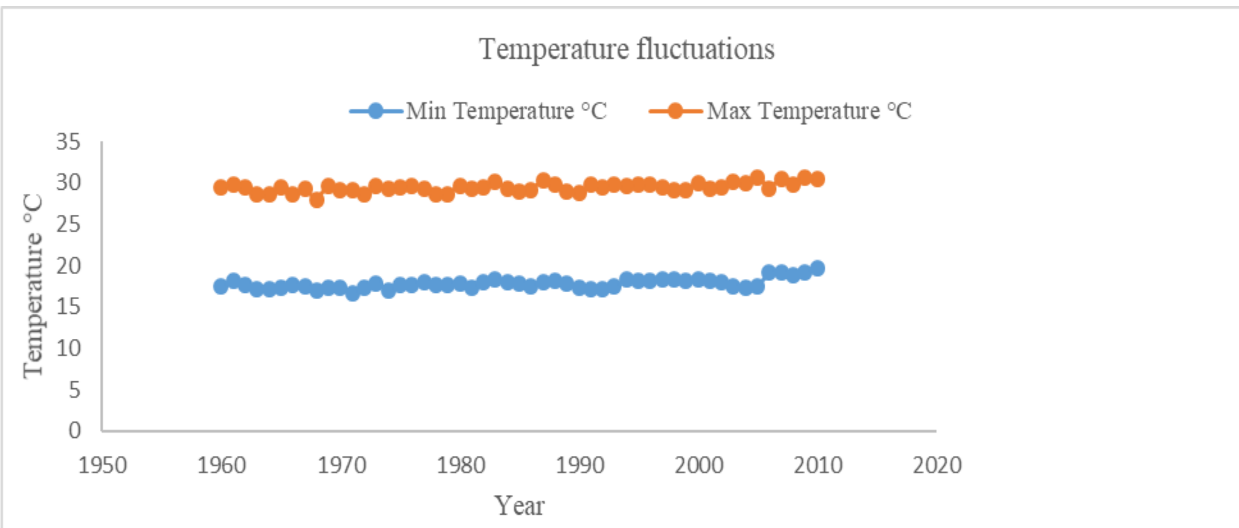


Figure 4: Changes in annual temperature (°C) measured at Moshi Airport weather station No: 9337004 Moshi Municipal, Kilimanjaro coordinates 3° 21' 0'' S and 37° 19' 48'' E

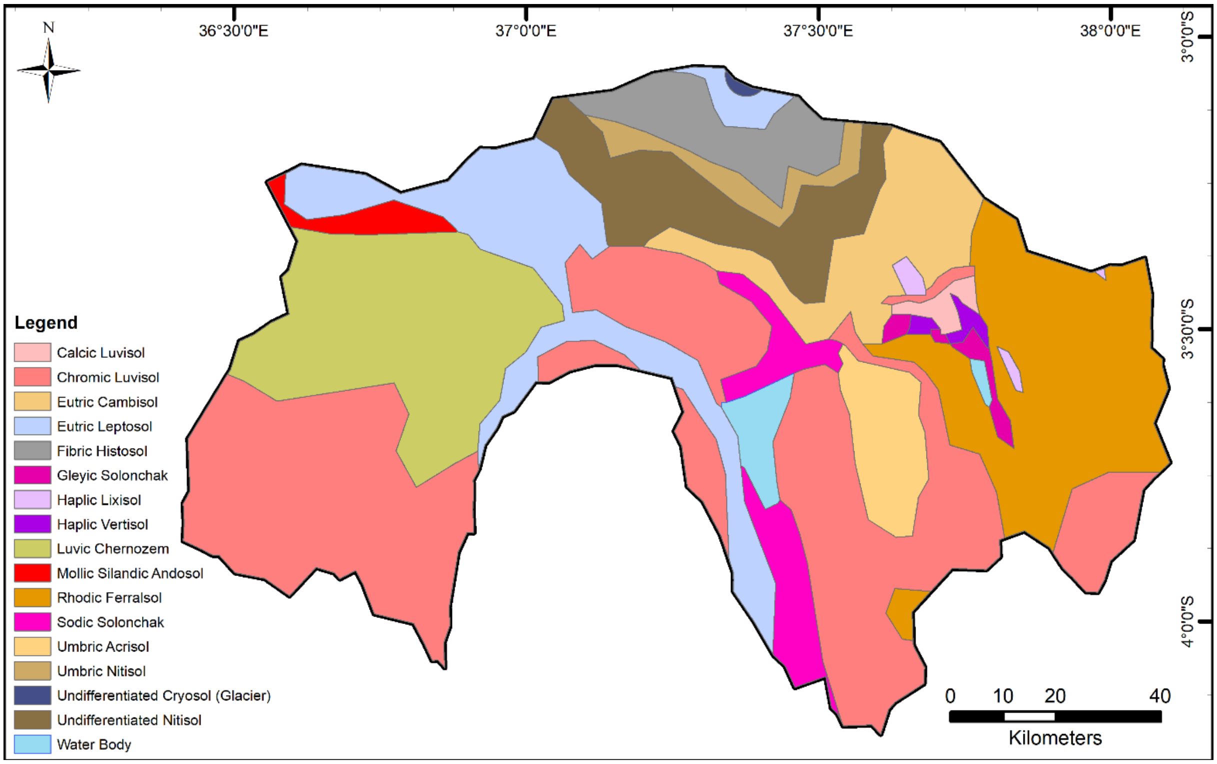


Figure 5: The Map of the Upper Pangani Basin details the major soil types in the catchment

3.3 To Reconstruct the Sedimentation Rates Overtime in the Nyumba ya Mungu Hydropower Reservoir

3.3.1 Sampling Strategy

Three cores, with respective depths of 28 to 32 cm were taken from the reservoir, two close to the mouth of the adjacent river input and one more distant (Fig. 6). The core locations were subject to logistics and chosen to include the variety of deposited sediment in the reservoir. Frequent flooding in the villages nearest the headwaters of the reservoir was an indication of the high siltation in the inlets. Shallow depth in the middle of the lake and presence of mostly coarse (sand to gravel) sediments in some locations in the middle was the reason for the abandonment in sampling in the middle of the reservoir. In addition, due to the presence of hippopotamus certain areas in the lake were also avoided. In this end, the efforts were mostly concentrated close to the inlets to account for spatially specific localized sedimentation effects. The water depths of the cores spanned between 1.6 and 3.0 m (water depth of the cores AC1=1.8 m; AC2 =1.6 m; AC3 = 3.0 m). A corer, inside fixed with PVC tubes (60 cm in height and 8 cm in diameter) was manually pushed in the sediment through diving. The cores were kept in upright positions to the boat for some hours where their overlying waters were gently decanted. The PVC tubes were consequently opened from one end, after which each core was sub-sectioned at 2 cm intervals, to obtain a sufficient temporal resolution, while keeping a high analytical accuracy (Kirchner, 2011). The core section were stored in polyethylene bags and transported to the laboratory, where they were weighed wet, then oven dried at 50°C to constant weight, in order to determine dry bulk density. The dried sliced cores were subsequently disaggregated using a mortar and pestle and sieved < 2 mm following standard procedures.

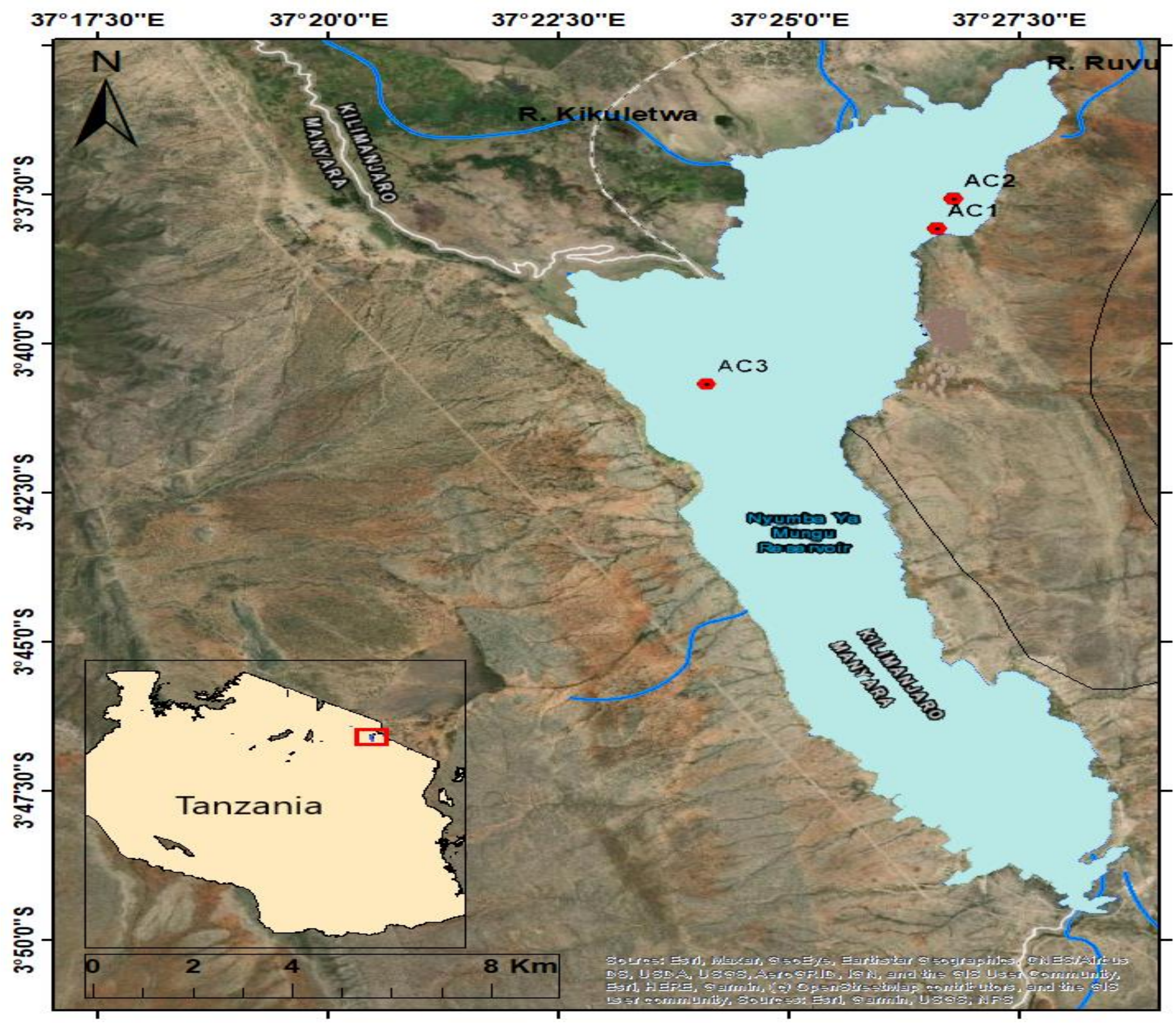


Figure 6: Sediment cores sampling locations (red marks) at the *Nyumba ya Mungu* reservoir

3.3.2 Radiometric Analysis

The sieved sliced core sections were filled into aluminium canisters and sealed for a minimum of 21 days to account for the secular equilibrium between ^{226}Ra and its progenies. Activity concentrations of the radionuclides in the subsequent section cores were analyzed at the Consolidated Radioisotope Facility (CoRiF) Laboratory of the University of Plymouth under a quality management system (QMS) certified to the ISO9001: 2015 using low background EG&G Ortec planar (GMX50–83-LB-C-SMN-S) and well (GWL-170-15-S) HPGe gamma spectrometers. Sediment cores were counted for 24 hours, analyzing the gamma spectra of the natural radionuclides ^{210}Pb (46.5 keV), ^{226}Ra (via ^{214}Pb and ^{214}Bi peaks at 352 and 609 keV, respectively) and for the anthropogenic radionuclide ^{137}Cs (662 keV). The samples with low masses were counted for 48 hours, and their results were cited with a 2σ counting error. Excess ^{210}Pb ($^{210}Pb_{ex}$) activity was determined as the difference between total ^{210}Pb and supported ^{210}Pb (in equilibrium with the parent radionuclide ^{226}Ra). A natural homogenized soil, with low background activity, spiked with a radioactive traceable standard solution (80717–669 supplied by Eckert & Ziegler Analytics, Atlanta, USA), was used to perform the calibration of the gamma spectrometer. The GammaVision software was used to establish the geometry-specific calibration relationships. Analytical performance was assessed by participation in International Atomic Energy Agency (IAEA) worldwide proficiency using example soils (IAEA-CU-2009-03 and soil IAEA-TEL-2012-03). The results are discussed in section 4.

3.3.3 Sediment Chronology and Mass Accumulation Rates

The rate of change in ^{210}Pb activity with mass depth in a sediment core provides the basis for an age-depth relationship and for estimating sediment mass accumulation rates (MARs) (Goldberg, 1963; Krishnaswamy *et al.*, 1971; Robbins, 1978; Wynants *et al.*, 2020). The comparatively long half-life of ^{210}Pb ($t_{1/2}$ = 22.23 years) provides the basis for the age determination processes of up to 5–6 half-lives, i.e. ~100 years. For the purpose of reconstructing variations in sedimentation rates over time, the Constant Rate of Supply (CRS) model was used because it assumes a constant ^{210}Pb flux but allows the sediment supply to vary (Appleby & Oldfield, 1978) Equations 1 to 4 (CRS-standard approach).

With s_n as the experimentally derived dry bulk density in section n, the cumulative dry mass m_n above sediments at depth x_n can be calculated as:

$$m_n = m_{n-1} + s_n(x_n - x_{n-1}) \tag{1}$$

Where C_n is the experimentally derived $^{210}Pb_{ex}$ activity at layer n, the cumulative $^{210}Pb_{ex}$ inventory can be calculated using the trapezium rule:

$$\hat{A}_n = \hat{A}_{n-1} + \frac{C_{n-1} - C_n}{\ln\left(\frac{C_{n-1}}{C_n}\right)}(m_n - m_{n-1}) \tag{2}$$

The total $^{210}Pb_{ex}$ (in Bq m^{-2}) inventory $A(0)$ of the sediment core is then equal to the $\hat{A}_n$ value in the deepest layer. The residual $^{210}Pb_{ex}$ (in Bq m^{-2}) inventory in the sediment core below depth n can subsequently be easily calculated by subtracting $\hat{A}_n$ from $A(0)$. Following the CRS model and with λ_{Pb} as the ^{210}Pb radioactive decay constant of 0.03114 y^{-1}, the age t of the sediment layer at depth n can be estimated by:

$$t = \frac{1}{\lambda_{Pb}} \ln\left(\frac{A(0)}{A(n)}\right) \tag{3}$$

The sedimentation rate r at depth z can subsequently be calculated as follows:

$$r = \frac{\lambda_{Pb} A(n)}{C(n)} \tag{4}$$

This model has been successfully applied in a nearby complex catchment that has experienced catchment-wide environmental changes (Wynants *et al.*, 2020). However, a major limitation of this technique in complex East African catchments is that a significant fraction of ^{210}Pbex supply to the sediment deposits may originate from older catchment material (Appleby *et al.*, 2019). In large catchments there might be natural variability in the terrestrial geological prevalence of ^{238}U that can influence the ^{210}Pbex activity from the secondary ^{210}Pbex activity. In addition, the natural variability might be caused by the dissimilarity in prevalent erosion process (He & Walling, 1997) that can change the fraction of topsoil vs subsoil in the transported sediment material, impacting the ^{210}Pbex activity (Aalto & Nittrouer, 2012; Du & Walling, 2012; Baskaran *et al.*, 2015; Wynants *et al.*, 2020). Consequently, whether the atmospheric ^{210}Pbex flux to the reservoir environment acts steadily stable over time, the arriving secondary ^{210}Pbex fingerprints from deposited sediment might differ significantly (Appleby *et al.*, 2019). Owing to potential differences in ^{210}Pbex fluxes, the CRS model outcomes were scrutinized through comparison with another independent marker, ^{137}Cs ($t_{1/2}$ = 30.17 years) from its 1965 peak fallout (Appleby, 2008) in the southern hemisphere using the fitting approach (CRS-fitted) as described by Appleby (2002) and in Equations 5 to 6 (CRS-fitted approach).

If $\hat{A}_{ref}$ denotes the entire $^{210}Pb_{ex}$ inventory above the reference level t_{ref}, the inventory below that level can be obtained by the following formula:

$$A_{ref} = \frac{\hat{A}_{ref}}{e^{\lambda t_{ref}} - 1} \tag{5}$$

The total inventory is then:

$$A(0) = \hat{A}_{ref} + A_{ref} \tag{6}$$

Sediment dates and accumulation rates can subsequently be calculated using Equations 3 and 4 respectively.

However, due to the low levels of ^{137}Cs fallout in tropical Africa and the known date of built of the dam, the deepest sediment layers were also fitted to the reference date of 1969 using the fitting approach (CRS-fitted) as described by Appleby (2002) and from the above Equations. Finally, the geochemical profiles of the cores were also scanned for distinct changes or peaks that could be linked to hydrological or sedimentological changes (Łokas *et al.*, 2010; Wynants *et al.*, 2020).

3.4 To Establish the Relative Proportions of Catchment Sediment Sources

The tributary riverbed sediment samples (DS) were collected from lower reaches of the two major tributaries, Kikuletwa (KL) and Ruvu (RV), assuming that the transported and deposited sediments offer a representative sample of the composite mixture from their respective sources in the entire catchment. Fourteen and eighteen samples of the respective tributaries were collected over a range length of about 200 m to include potential spatial differences in riverine sediment deposition (Gellis & Noe, 2013; Wilkinson *et al.*, 2013; Wynants *et al.*, 2020). The DS sample collection depended on the environmental and logistical constraints in the system. Due to the ephemeral nature of the rivers, DS samples were generally collected in the dry season from the exposed beds. Land use soil samples were recovered from agricultural topsoil (CU), bushland topsoil (BS), channel banks (CB) and mainstem river banks (RB). The agricultural and bushland top soils involved surface soil (0–5 cm) sampled from areas presumed vulnerable to water erosion and their connectivity to river network. Sampling of subsurface/channel bank material was done in upstream areas characterized by exposed banks devoid of vegetation with actively eroding bank sections due to flow incision by high water energy released during heavy rainfall. Eroding mainstem river banks were also sampled. Sampling locations for the land use samples depended on the accessibility, necessary permits and safety. At each site, samples comprised a composite of 10 to 15 random scoops pooled into a single composite sample to ensure the representativeness of the corresponding fingerprint property datasets. A total of 57 samples were collected to characterize four main potential sediment sources: (a) bushland (BS, n = 15), (b) channel banks (CB, n = 15), (c) cultivated agricultural land (CU, n = 14), and (d) mainstem river banks (RB, n = 15), all collected in one-year season (Appendix 1). Soil and tributary sediments were also oven-dried at 55–60°C to constant weight and consequently disintegrated using a mortar and pestle and then sieved.

3.4.1 Geochemical Analysis

Before analysis, all dried samples were homogenized and sieved to <63 μm fraction to minimize particle size effects on tracer signals that can bias fingerprint property (Owens *et al.*, 2016; Collins *et al.*, 2017; Laceby *et al.*, 2017; Kroese, 2020). The elemental concentrations are generally enriched in the fine, < 63 μm, particle size fraction in comparison to < 2 mm bulk fraction of the soil (Rawlins *et al.*, 2010; Laceby *et al.*, 2015; Kroese, 2020). For comprehensive reviews on issues of particle size effects on sediment fingerprinting, readers are referred to Laceby *et al.* (2017). Subsequently, about 4 g of dried and sieved sample material was mixed with about 0.9 g of cellulose binder (FLUXANA®), homogenized in a pulverizer and pressed into a pellet of approximately 32 mm diameter. The method was validated by using the IAEA Soil 7 certified reference materials (CRM) described by IAEA (2000). The core samples were analyzed for minor and major elemental geochemistry by wave length dispersive X-ray fluorescence (WD-XRF; PANalytical Axios Max; OMNIAN application) as loose powder at CORiF Laboratory of the University of Plymouth. The sediment certified reference material was used to validate the analyses (GBW07318, LGC, Middlesex, UK). The dried soil and tributary potential sediment sources were analyzed by an energy dispersive X-ray fluorescence (EDXRF) spectrometer coupled with Xlab ProTM software as pressed pellets at Tanzania Atomic Energy Commission (TAEC). For assessment of the analytical variability and sample homogeneity, triplicates were made from arbitrarily selected samples about once every 3 samples. Only those elements returning measurements above the limit of detection (DL) were employed in the analysis (DL varies with the element and depends upon several factors including the sample matrix). The difference in the analytical methods for the major and minor elements may have influenced the accuracy of the model. However, the inter-laboratory comparison was performed and the results were shown to be directly comparable.

3.4.2 Bayesian Mixing Model for Source Apportionment

After the broad spectrum geochemical analysis, each sample can be represented as a multi-elemental concentration data point. The fingerprints of the potential sediment sources and the lake mixture form multivariate concentration matrices on which the model is draws to quantitatively compare different sources with the lake mixture. A Bayesian mixing model (BMM) was built in the open-source MixSIAR framework (Stock & Semmens, 2016; Stock & Semmens, 2017; Stock *et al.*, 2018) and utilized as first demonstrated by Blake *et al.* (2018a) for river basin sediment source apportionment at tributary and land use levels. For comprehensive details of the mathematical formulation of MixSIAR readers are referred to Stock *et al.* (2018). The "deconvolutional MixSIAR" (D-MixSIAR) methodology was used to hierarchically unmix the tributary sediment against the core sediments and subsequently unmix the land use source fingerprints against the tributary sediment. For the BMM to accurately represent the system, it depends on the following four assumptions: (a) the model includes all dominant sources contributing to the sediment, (b) the value of the tracers are known in both sources and mixture, (c) tracers are behaving conservatively throughout the mixing processes, and (d) fingerprint variability between sources is larger than within sources.

3.4.3 Tracer Conservation Test

In order to make a direct comparison of the properties of the sediment samples with those of the potential source materials, tracers need to behave independently and conservatively (assumption c) in the environment (Motha *et al.*, 2002a; Blake *et al.*, 2018a) This implies that the chemical composition of the tracers does not alter during detachment, transport or after deposition (Koiter *et al.*, 2013; Belmont *et al.*, 2014; Laceby *et al.*, 2017).

The source apportionment results were achieved using a tracer selection procedure that only excluded tracers on the basis of non-conservative behaviour from literature evidence of potential mobility in aquatic systems (Smith *et al.*, 2018). Initially, all elemental concentrations from the depositional samples that fell outside the minimum detection limits were removed. Thereafter, the basic tracer screening approach of Blake *et al.* (2018a) and Sherriff *et al.* (2015) was adopted with additional evaluation of geochemical behaviour. For each set of sources and associated mixtures for all tracers, boxplots were produced and the means of the mixture data assessed to see if they largely fell within or outside of the mean concentrations of the different sources (Blake *et al.*, 2018a) (Fig. 7). Tracers wherein the mixture fell outside the source range were removed. In addition, the tracers that were found to be higher in intra-source variance than the inter-source variance were also removed. Finally, the normality assessment using the 'Shapiro-Wilk test' for the individual tracer mixtures was done because the model assumes normal distribution of the mixture tracer data (Stock *et al.* 2018). Seven tracers passed the range test (P, Ti, Mn, Fe, Zn, Sr, Nb) and seven borderline tracers (S, Co, Ni, Cu, Ga, Ba, Hf) were also retained and engaged (Fig. 5). Model efficacy was evaluated using the gelman-rubin diagnostics (Semmens *et al.*, 2009; Gelman *et al.*, 2013). Seventeen (17) tracers (Na, Mg, Al, Si, Cl, K, Ca, Cr, Br, Rb, Zr, Ce, Pb, Th, La, Y, Sn) were eliminated from the analysis based on the evidence of non-conservative behaviour or a high intra-source variability. These exclusions warrant some geochemical clarification. The Sr and Rb may have been caused by their known variability in the soil depth as a function of weathering processes and mixing of soil horizons by cultivation (Tyler, 2004). Non conservative behaviour of Mg, Na, Ca, Cl, F, K, and Br can be explained by their tendency to form highly soluble salts (Blake *et al.*, 2018a) driven by evaporation in the lake (Horowitz, 1991). The La, Ce and Th were removed because of the observed inter-source variability, which are potentially an artefact of analytical challenges due to a low abundance or a high variability in the terrestrial source concentrations (Wynants *et al.*, 2020). Various elements, such as Al, are known to exhibit non-conservative behaviour during fluvial transport and short-term storage in river channels (Withers & Jarvie, 2008), while other tracers (Si Cr, Y, Pb and Zr) are known to undergo transformations in medium- to long term storage elements such as floodplains, lakes and wetlands due to changes in redox, pH, salinity and other environmental conditions (Hudson-Edwards *et al.*, 1998; Owens *et al.*, 1999; Pulley *et al.*, 2015a). The Al, Si and Zr may be due to wider fluvial sorting i.e. textural, controls on mineral composition i.e. changing proportions of silt *versus* clay minerals in mixtures which has been shown to exert a strong influence on sediment concentrations (Cuven *et al.*, 2010). The older sediment deposits however, may have undergone diagenetic processes within the sediment column (D'Haen *et al.*, 2012).

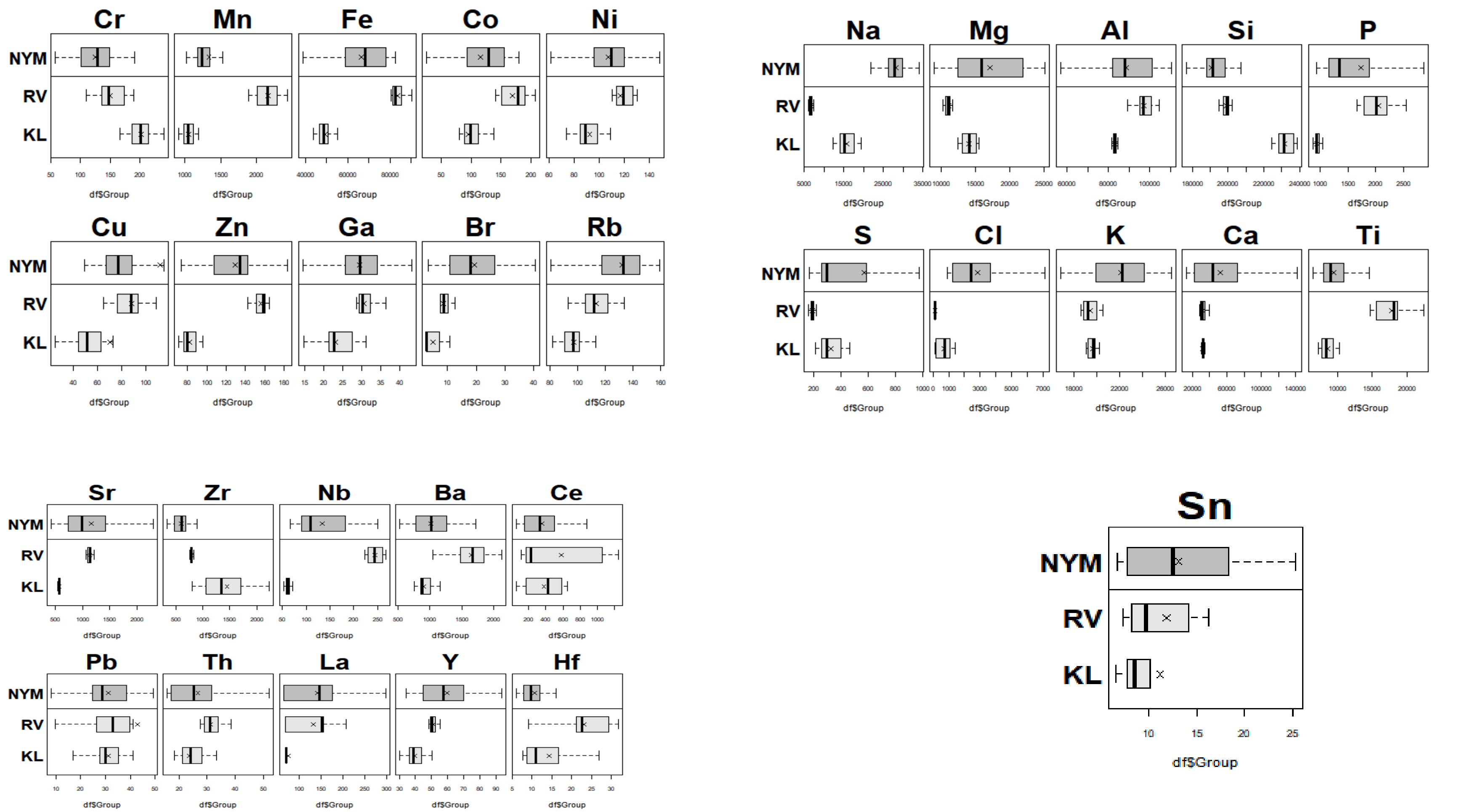


Figure 7: Boxplots for tracer selection of the tributary sources KL (Kikuletwa), RV (Ruvu) and NYM (Nyumba ya Mungu) as a mixture. In box plots, median is shown by central line, interquartile range by box, range by whiskers with circles indicating outliers

3.4.4 Principal Component Analysis

The PCA was performed to accommodate assumption 4 which requires the inter differences of various source fingerprints to be larger than the intra differences of each source fingerprint. It is in this context that the PCA was performed to analyze variance in multi-tracer datasets and reduce dimensionality (D'Haen *et al.*, 2012) from the potential sediment sources data into land use specific and tributary categories according to their geochemical composition.

3.4.5 Model Build and Running Protocols

(i) Error Formulation

Eroded soils with distinct geochemical properties from the wider catchment mix from the source to a mixture through mixing process. However, in these complex systems it is impossible to capture the total variability in sediment source sample by sampling. Therefore, a 'residual error' formulation was integrated in the model. Since the transport of sediment from channel networks to the reservoir is random and constant, a 'process error' was not included (Stock & Semmens, 2016; Stock *et al.*, 2018).

(ii) Uninformative Prior

Since there are no other sources of empirical information about the sediment source contributions to the reservoir, an uninformative prior was used; (1, 1, 1, 1) and (1, 1) for land use and tributary sources, respectively.

(iii) Fixed Categorical Effect

A mixture of sediment samples was analyzed without fixed or random effects to infer the proportions of the tributaries and land uses to the 'total' reservoir sediment. Afterwards, a fixed categorical effect of the sediment mixture from reservoir sampling locations was established to infer the proportions of different land uses and tributaries to the specific sampling location in the reservoir. The ascription of sediment delivery over time was inspected by using individual sediment cores whereby the "age" was introduced as a fixed continuous effect. Subsequently, the sediment core sliced samples were grouped into distinct classes whereby the depth was established as fixed categorical effect.

(iv) Model outputs

The model outputs were assessed under different modes of covariate structure. For all model runs, the following stipulations were used: A residual error term only and an uninformative Dirichlet prior ($\alpha = 1$) (Stock *et al., 2009).* Model convergence was assessed by the Gelman-Rubin diagnostic (variables < 1.05), rejecting model output if >5% of total variables was above 1.05. Model convergence indicates that the model has found a singular solution to the problem. The MIXSIAR model using the selected 14 tracer fingerprints on the sediment sources passed the Gelman–Rubin convergence diagnostic with the parameters of the Markov Chain Monte Carlo (MCMC) chain run length set as follows: chain length = 1 000 000, burn = 700 000, thin = 300, chains = 3.

3.5 To Evaluate the Soil Carbon as a Proxy for Soil Erosion Risk in the Catchment

3.5.1 Data Acquisition for Land use Classification

Ortho-rectified and geometrically corrected Landsat images (Landsat 4-5 and Landsat 8) with a resolution of 30 m were obtained from the USGS Earth Explorer website (https: glovis.usgs.gov/). For this specific study, Landsat images captured on February 1987 for Landsat 4-5 and on September/October 2017/18 for Landsat 8, were selected based on the lack of interfering cloud cover and ability to reconstruct land cover change over time. Before the analysis, the images were projected to UTM zone 37S allowing spatial assessment in combination with other spatial data of the study area (Liang *et al.*, 2018).

3.5.2 Land use/cover Image Classification

Geo-tagged photos and field notes were gathered during multiple ground-truthing campaigns to offer a comprehensive documentation of the land cover spectrum. By using these ground observations complemented with Google Earth images, the major land cover types in the area were delineated into spectral signature files. The supervised classification by maximum likelihood algorithm method in ArcMap uses these signature files to extrapolate across the full Landsat image database into the pre-defined cover classes. A visual examination and comparison with high resolution aerial imagery from Google Earth was used to remove potential incorrectly classified features. A raster calculator function was used to direct the correct elevation of a particular landuse class based on the expert knowledge of the study area as explained by Taweesuk and Thammapala (2005). The Expert classification is aiming at improving classification accuracy thus used to integrate remote-sensed data with other sources of georeferenced information such as digital elevation model (DEM), land-use data, and spatial texture. A total of nine major classes of land cover were classified representing both changes due to natural drivers and human influence (Table 3).

Table 3: Land use/cover classes and their description

S/N	Land use/cover class	Description
1	Agricultural land (AGL)	All cultivated land with crops and harvested crops
2	Water (WT)	Including water in wetlands, rivers, irrigated areas and fish ponds
3	Grassland (GRL)	Areas dominated by short and tall grasses and bare soils in the dry season.
4	Bush land (BSL)	Areas dominated with shrubs and less closed canopy
5	Bare land (BAL)	Areas includes gullies, bare soils, rocky, sand and quarry
6	Built-up area (BLA)	Man-made infrastructure (urban and rural settlements) and roads (tarmac or paved)
7	Forest (FRL)	Includes natural and planted forests with closed trees and closed canopy
8	Wetland (WTL)	Areas moderately saturated with water seasonally or permanently
9	Glacier ice (GLA)	Includes areas enclosed with glacier and ice

3.5.3 Soil Scanning to Estimate Soil Organic Matter

An AgroCares scanner, a portable handheld Near Infrared (NIR) sensor for soil scanning (Saskia *et al.*, 2020), was used to scan the soil samples to evaluate SOC content. The scanner is connected to an app ("soil cares app" downloaded from Google play/Apple store) using a smartphone via Bluetooth. A spectral analysis of the scanned soil is sent to the application on the smartphone via Bluetooth. Subsequently, the smartphone application connects to 'AgroCares' global calibration database to convert the spectral image into the required soil data.

A hand trowel and scoop were used to collect a soil sample between 0-5 cm depth which was put in a bucket and well mixed. The samples were clearly labelled, and the coordinates of the sampling location were recorded using an Infinix Hot 9 android 10 XOS 6.0. The scanner was calibrated in situ following manufacturer instructions. The scanner was placed on the sub-samples (drawn from the bucket) on the sample tray and per soil sample the scanner performed 5 scans. Using the reflectance signature and global calibrated database, the application estimated the following soil parameters: Soil organic carbon (g/kg), pH, soil texture class, total Phosphorus (g/kg), Potassium (mmol+/kg), soil temperature (◦C) and Cation exchange capacity (mmol+/kg). The SOM was derived from the estimated SOC using the conversional van Bemmelen factor of 1.72 (Nelson & Sommers, 1996; Pribyl, 2010). The conversion factor is based on the assumption that the organic matter is 58% carbon (Sprengel, 1827).

3.5.4 Loss on Ignition to Estimate Soil Organic Matter

Loss-on-ignition was determined on the oven-dried subsamples of soil fractions (Jensen *et al.*, 2018). Approximately 20 grams of air-dry soil was added to previously ignited and weighed porcelain crucibles, dried at 105ºC for 12 hours in a ventilated oven, cooled in a desiccator and weighed again. Finally, the crucibles were ignited at 550°C for 4 hours in a muffle furnace (Cole-Parmer® StableTemp). After ignition, the crucibles were cooled in a desiccator and weighed. The LOI was calculated as the difference between the oven-dry weight (DW) before and after ignition and related to oven-dry soil, Equation 7.

$$LOI_{550} = \frac{DW_{105} - DW_{550}}{DW_{105}} \times 100 \ldots\ldots (7)$$

The values of the SOM obtained in percentages were compared with the SOM derived from the loss on ignition using the same samples.

3.5.5 Soil Aggregate Stability (Slake Test)

Soil aggregate stability in water was assessed using a semi-quantitative method adapted from the USDA-ARS Soil Slake test method (Herrick *et al.*, 2001), wherein a value was assigned to the assessed soil samples based on the stability of soil aggregates in water. Soil sample aggregates with a diameter of approximately 10 mm were collected using a trowel from different land use types and subsequently air-dried at room temperature. The air-dried aggregates were placed on a 6 mm mesh that was fixed on a basket cup. The basket cup with soil aggregates was subsequently immersed with water on top of the mesh. Following the behavioural criteria of the aggregates in water (Table 2), the slaking away of the soil fragments was recorded for five minutes. For each soil sample, a soil stability score was rated according to the time required for 50% of the soil aggregates and the proportion of the soil fragments remaining on the mesh after the five minutes of immersion.

3.5.6 Statistical Analysis

Initially, a regression analysis was conducted to determine the correlation between SOM content % derived from the scanner and SOM content % derived from the LOI experiments. Data were subsequently tested for normality, where for the WSA only values for forest land and cultivated land were normally distributed $p < 0.0005$ and $p < 0.018$ respectively. For LOI only forestland and bushland were also normally distributed $p < 0.036$ and $p < 0.008$ respectively and normally undistributed to the rest of land uses while for SOM only bushland data were normally distributed. Following these results, the non-parametric Krukal-Wallis test was carried out in SPSS (Statistical Package for Social Science) to test if there were significant differences in the measured LOI, scanned SOM and estimated WSA between the different land use types. The differences were subsequently visualized using boxplots allowing a comparison of the mean values and variability of LOI, SOM and WSA within and between land use sites.

Table 4: Criteria for scoring soil stability in water

Stability class	Criteria for assignment to stability class (for Standard Characterization)
0	Soil too unstable to sample (falls through sieve)
1	50 % of structural integrity lost within 5 seconds of insertion AND or < 10% remains after agitation
2	50 % of structural integrity lost 5 - 30 seconds after insertion AND or < 10% remains after agitation
3	50 % of structural integrity lost 30 - 300 seconds after insertion AND or < 10% remains after agitation
4	10 - 25% of soil remains after 5 minutes agitation
5	25 - 50% of soil remains after 5 minutes agitation
6	50 - 75% of soil remains after 5 minutes agitation
7	75% - 90% of soil remains after 5 minutes agitation
8	> 90% of soil remains after 5 minutes agitation

Herrick *et al.* (2001)

CHAPTER FOUR

RESULTS AND DISCUSSION

4.1 Overview

This Chapter depicts the main findings of the research and a comprehensive analysis of the results. The Chapter further illustrates the necessity of the integration of the methods (geochronology and geochemical fingerprints) that explicitly maintain sediment control strategies for sustainable management of food, water and energy security in Eastern Africa.

4.2 Reconstructing Sedimentation Rates Over Time in the Nyumba ya Mungu Hydropower Reservoir

4.2.1 Changing Sedimentation Rates Over Time

The $^{210}Pb_{ex}$ profiles (Fig. 8a-c) of all three cores did not follow an exponential decline with depth and were instead characterized by multiple peaks and troughs indicating episodic increase in sedimentation rate which would have diluted the $^{210}Pb_{ex}$ activities of the more recently deposited sediments (Appleby, 2002; Wynants *et al.*, 2020). However, a number of processes could have also resulted in the vertical mixing in the sediment cores. First of all, high energy water flows and wave action at the river inlets could have led to vertical mixing of the sediment deposits and a flattening of the $^{210}Pb_{ex}$. Second, human and biological activities might also have led to mixing of the upper sediment layers in the shallow parts of the reservoir. Third, is the variation in the remineralization of sediment particles downcore. Furthermore, previously discussed variations in the direct atmospheric depositional flux of ^{210}Pb due to fluctuations in the annual amount of rainfall and variation in the secondary input of the ^{210}Pb from the catchment could have influenced the $^{210}Pb_{ex}$ profile (Appleby *et al.*, 2019).

The ^{137}Cs activity concentration, where detected, in all of the cores was low, with most of the sections below the limit of detection. These results are similar to the global ^{137}Cs fallout estimates in tropical regions (Walling & He, 2000). The low ^{137}Cs activities obstructed the assessment of ^{137}Cs peak integrity and the comparison of ^{210}Pb radiometric dates with independent ^{137}Cs peaks (Appleby, 2002). Contrary to core AC2 and AC3, core AC1 had a significant single measure of ^{137}Cs activity above the detection limit (Table 5). The detectable layer was assumed to be the 1965 southern hemisphere peak deposition (Cambray, 1989;

Walling & He, 2000), however, the 'peak' might also be related to secondary transport of catchment surface material to the reservoir, meaning that it could be younger than 1965 (Mabit *et al.*, 2008; Appleby *et al.*, 2019). For these reasons, the ^{137}Cs dating and fitting the CRS models of the NYM reservoir cores using the ^{137}Cs dates would result in a high level of uncertainty. Therefore, two versions of the CRS model were run. The first 'standard' approach estimated dates based solely on fallout $^{210}Pb_{ex}$. In a second 'fitted' model, it was opted to fit the deepest sediment layers to the construction date of the dam. In the former case, the assumption in the core penetrates the post impoundment soft sediment into the original substrate. In the latter, there is an assumption that the coring captured the full soft sediment profile.

The MAR output of the CRS-standard approach in cores AC1, AC2 and AC3 shows a general trend of increase in sedimentation rate before commissioning of the dam, assuming in this case the core penetrates the soft sediment into former marsh sediment and sharp increase after commissioning of the dam (Table 5) and (Fig. 8 d, e, f). Reservoir sedimentation generally increased from 0.1 g cm^{-2} yr^{-1} in the lower sediment column to 1.7 g cm^{-2} yr^{-1} in the most recent deposits. Under CRS standard approach model version, the observed changes could be associated with the changing sedimentation dynamics following the construction of the dam, but also by higher levels of soil erosion following deforestation and agricultural degradation. The changes in 1970's correspond with the adoption of the villagization policy in Tanzania which led to practicing of a cooperative economy (communal production) through collective farming (Bryceson, 2002; Fouéré, 2014). The sharp MAR peaks 0.86 g cm^{-2} yr^{-1} and 1.01 g cm^{-2} yr^{-1} in the mid of 1990s in cores AC1 and AC3 respectively might probably be linked to ENSO, 1997/1998 El Niño events (Kane, 1999) (Fig. 8 d & f). These general trend of increase of sedimentation rate in cores AC1, AC2 and AC3 in 2000's to 2010's is related to growing urbanization, extension of agricultural activities and the loss of permanent vegetation through the fast expansion of agricultural land (Mbonile *et al.*, 2003; Tadross & Wolski, 2010; NBS, 2012; NBS, 2018; Said *et al.*, 2019). The latest peak can be spotted at 2019 in both cores being 0.88 g cm^{-2} yr^{-1}, 1.42 g cm^{-2} yr^{-1} and 1.69 g cm^{-2} yr^{-1} in cores AC1, AC2 and AC3 respectively, whereas the 2010 MAR peak (1.4 g cm^{-2} yr^{-1}) in core AC2 is substantially higher than the peak in core AC1 (0.879 g cm^{-2} yr^{-1}).

The alternative CRS-fitted geochronology model approach of Appleby (2002), using 1969 as a fixed date, showed a similar trend of the increase in sedimentation as the CRS-standard approach. While the MAR of cores AC1, AC2 and AC3 and age-depth relationship using the standard CRS approach, were found to be relatively similar (Fig. 8 d-f), the MAR-fitted trends were also found to be similar between the different cores. There were some interesting differences observed between the CRS-standard and CRS-fitted approaches in each individual core. For instance, the MAR-fitted in core AC3, were higher than those in core AC1 and core AC2. Due to the smaller time-range, the MAR-fitted rates and peaks were much higher, giving rise to a more punctuated profile. However, the general observed trend of increasing sedimentation over time was observed in both approaches. While, extreme rainfall during the ENSO, 1997/1998 El Niño events (Kane, 1999) seems to have contributed to higher rates of sedimentation in the NYM reservoir, this was not observed for the higher rainfall in 1978/79 (Fig. 2). Under this model scenario, this seems to confirm that the increase in sedimentation in the NYM reservoir is driven by a complex interaction between natural rainfall variability and increasing vulnerability to soil erosion through land use change. The observed differences in the timing and heights of the MAR peaks between the cores might be due to spatial differences in sedimentation due to the dynamics of sediment transport from the dominant riverine sources, as will be further explored in section 4.3. Difference in particle settling velocity due to different in the location of the cores might be another reason for the variations in sedimentation rates. However, the main limitation of the CRS-fitted approach fitting the curve to 1969 is that, it might have included the activities from the floodplain of the natural wetland before the date the reservoir was commissioned in 1969.

The above alternative geochronology model outputs clearly indicate that ascribing exact dates to this sediment column is a challenge, especially considering the potential impact of changing sediment source on $^{210}Pb_{ex}$ supply, but overall changes in sedimentation rate can offer meaningful interpretation of catchment impacts on sedimentation dynamics in the context of geochemical data. The geochemical profiles of the sediment cores offer a useful indication of changes in tributary sediment delivery to NYM reservoir over time to help constrain and interpret the different geochronological models. Comparing the broad trends of the CRS output of all with their geochemical profiles allowed for a catchment-based contextualization of the modelled MARs (Wynants *et al.*, 2020). The recent MAR peaks in all cores closely matched with peaks in many elements connected to allogenic sediment origins and minima in elements linked to evaporative autogenic tracers as shown in Appendix 2 (Horowitz, 1991). The allogenic tracers: Fe, Ti, Zr and Nb, Rb, Ba, Ga, and the evaporative autogenic tracers: Mg, K, Na, Ca, and Sr have corresponding maxima and minima at the same depth (4 cm) in 2016 and 2017 in core AC1 and AC3 respectively as the MAR peak. However, in core AC2 the allogenic tracers: Al, Ti, Ni, Cr, and Fe and the autogenic evaporative tracers: Na, S, Cl, Sr, and Ca have distinct maxima and minima at 2 cm in 2019 and 5 cm in 2016, respectively while the evaporative autogenic tracers Mg and Na have the minima corresponds to maxima of the allogenic tracers. This high correlation between geochemical tracers of allogenic sediment delivery and reconstructed sedimentation peaks in both cores reciprocally validates both evidence bases, making it highly likely that NYM reservoir recently experienced extreme sedimentation rates driven by increased erosion and sediment transport from the catchment. The difference in concentrations was observed between the cores; the variation in geochemical concentration is not only caused by changes in absolute sediment delivery, it may be attributed by the difference in source contribution (e.g. from different tributaries) and erosion process (e.g. subsurface vs. surface) which also cause changes in profile. The difference in concentration can also be due to location-specific sedimentation effects (e.g. more or less sedimentation and /or higher contribution from one of the tributaries). Interestingly, the difference in concentrations was also observed between the deepest older sections and upper recent sections of the autogenic tracers (Na, S, Cl, Sr, Ca) and the allogenic (Al, Ti, Ni, Cr, Fe). The concentrations of the autogenic tracers were much higher in the deepest sections and lower in the upper recent sections and vice versa for the allogenic tracers. The possible reason is that the deepest core sections are from the before the reservoir construction i.e 'standard' geochronology model, and or the older sediment deposits, may have undergone diagenetic processes within the sediment column (D'Haen *et al.*, 2012). An additional consideration to the limitations of ascribing exact dates to the sediment column is that the $^{210}Pb_{ex}$ profiles are vulnerable to sediment reworking through biological and physical activities such as bioturbation and fishing, respectively (Hu *et al.*, 2019). Following these occasional events the $^{210}Pb_{ex}$ profiles are interspersed and might be difficult to determine (Krishnaswami & Lal, 1978), however, in many cases an indication of broad rates of SAR change can still be determined with notable evidence for changes in sediment provenance being a key observation (Mabit *et al.*, 2014).

Table 5: The Constant Rate of Supply dating results for cores AC1, AC2 and AC3 respectively

Core AC1							
Depth (cm)	**Mass depth (g cm^{-2})**	**Date (yr)**	**MAR (g $cm^{-2}yr^{-1}$)**	**Date fit (yr)**	**MAR fit (g cm^{-2} yr^{-1})**	**$^{210}Pb_{ex}$ (Bq/kg)**	**^{137}Cs (Bq/kg)**
2	2.26	2019	0.879	2018	1.03	60.81	2.67
4	5.13	2016	1.12	2015	1.32	44.01	2.30
6	8.26	2014	1.52	2013	1.83	29.75	1.82
8	11.33	2012	1.2	2012	1.45	35.29	2.14
10	14.62	2009	0.91	2009	1.1	42.98	1.85
12	17.29	2005	0.79	2006	0.99	43.82	1.83
14	20.17	2001	0.65	2003	0.83	47.5	2.83
16	22.57	1997	0.86	2001	1.18	31.07	2.38
18	24.83	1994	0.42	1998	0.58	57.42	2.02
20	27.42	1988	0.44	1994	0.63	46.47	2.75
22	30.18	1981	0.404	1989	0.63	41.05	2.99
24	33.07	1974	0.29	1984	0.48	45.08	2.80
26	36.01	1962	0.17	1977	0.33	53.41	4.90
28	38.78	1937	0.09	1968	0.27	48.14	2.40

Core AC2							
Depth (cm)	**Mass depth (g cm^{-2})**	**Date (yr)**	**MAR (g $cm^{-2}yr^{-1}$)**	**Date fit (yr)**	**MAR fit (g cm^{-2} yr^{-1})**	**$^{210}Pb_{ex}$ (Bq/kg)**	**^{137}Cs (Bq/kg)**
2	1.87	2019	1.4	2018	1.25	59.66	2.32
4	3.99	2017	0.865	2016	1.05	67.26	1.99
6	6.25	2015	1.2	2014	1.48	44.92	2.41
8	8.84	2013	0.863	2012	1.06	58.66	2.45
10	11.22	2010	1.05	2010	1.34	43.7	2.83
12	13.82	2007	0.771	2008	0.99	55.31	4.26
14	16.47	2004	0.701	2005	0.92	54.44	3.06
16	18.96	2000	0.649	2003	0.89	51.9	2.45
18	21.40	1996	0.519	2000	0.73	57.18	2.53
20	23.85	1990	0.404	1996	0.59	62.74	3.00

Core AC2

Depth (cm)	Mass depth (g cm^{-2})	Date (yr)	MAR (g $cm^{-2}yr^{-1}$)	Date fit (yr)	MAR fit (g cm^{-2} yr^{-1})	$^{210}Pb_{ex}$ (Bq/kg)	^{137}Cs (Bq/kg)
22	26.45	1984	0.479	1992	0.77	42.86	3.12
24	29.04	1978	0.61	1989	1.08	28.12	2.13
26	31.86	1973	0.35	1986	0.65	42.11	2.37
28	34.81	1964	0.27	1981	0.57	40.71	2.18
30	37.67	1951	0.383	1977	1.08	19.31	2.47
32	41.0702	1942	0.27	1973	0.56	32.53	2.64

Core AC3

Depth (cm)	Mass depth (g cm^{-2})	Date (yr)	MAR (g $cm^{-2}yr^{-1}$)	Date fit (yr)	MAR fit (g cm^{-2} yr^{-1})	$^{210}Pb_{ex}$ (Bq/kg)	^{137}Cs (Bq/kg)
2	3.37	2019	1.69	2018	2.23	42.77	1.36
4	6.88	2017	1.86	2016	2.26	36.4	2.00
6	11.37	2015	2.12	2015	2.57	30.15	1.61
8	15.18	2013	0.92	2012	1.12	64.73	2.32
10	19.69	2008	1.12	2009	1.38	46.53	1.48
12	23.05	2004	1.26	2006	1.67	36.08	2.00
14	26.5	2001	0.85	2004	1.13	49.38	1.57
16	30.37	1997	1.01	2001	1.39	36.11	1.72
18	35.04	1993	1.21	1998	1.73	26.54	1.47
20	38.68	1989	0.60	1995	0.89	46.78	1.64
22	41.86	1982	0.47	1991	0.76	49.13	2.03
24	45.35	1974	0.61	1987	1.13	29.3	1.49
26	49.63	1968	0.46	1983	0.89	32.5	1.66
28	52.29	1957	0.24	1979	0.58	44.43	1.82
30	55.49	1943	0.19	1974	0.62	35.01	1.82
32	59.62	1920	0.13	1968	0.70	25.46	1.55

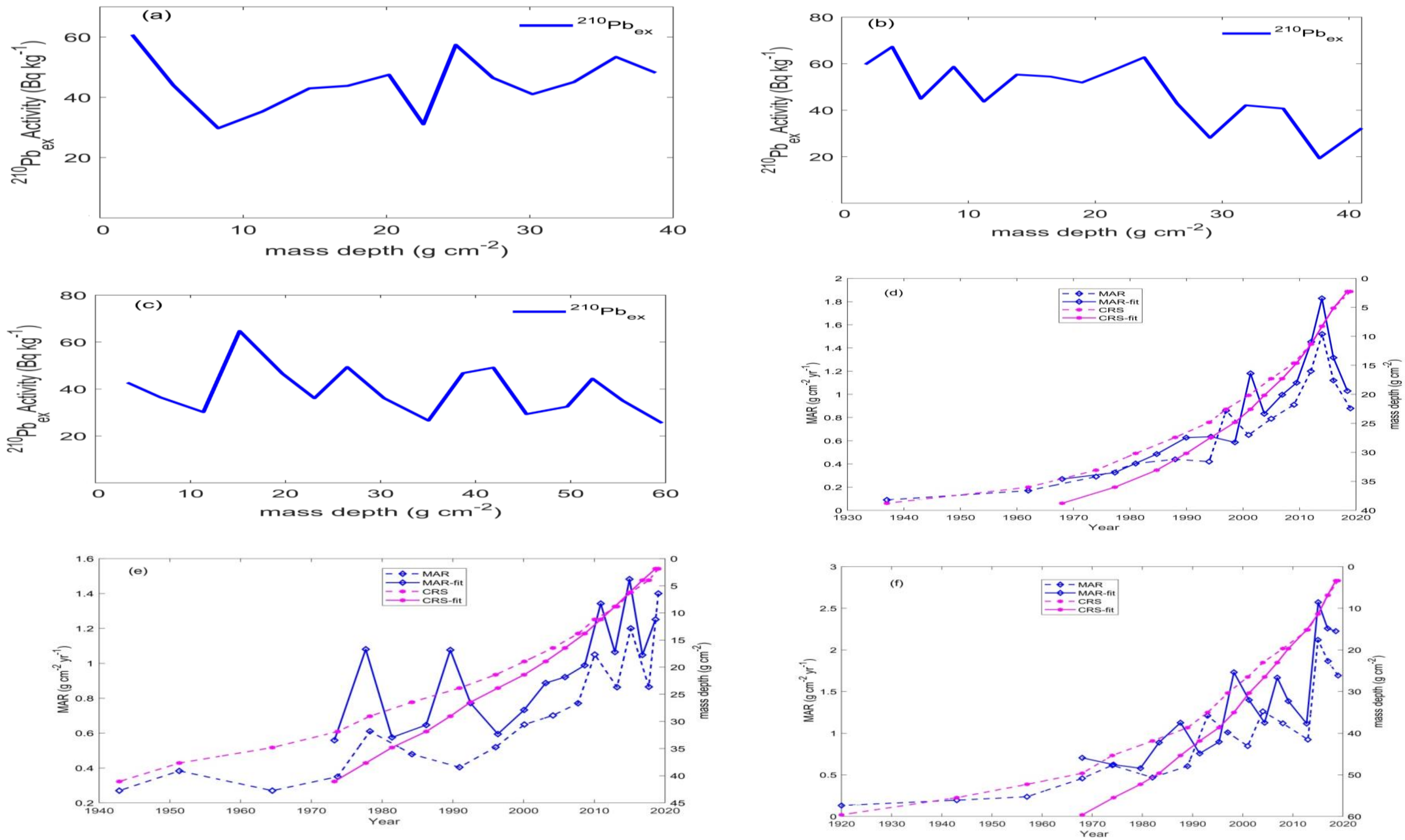


Figure 8: ^{210}**Pbex mass depth profiles for cores (a) AC1, (b) AC2 and (c) AC3, and their respective age–depth relationships in (d–f). Lines are defined in the legend**

4.3 Establishing the Relative Proportions of Catchment Sediment Sources

4.3.1 Principal Component Analysis for Statistical Analysis of Data

The temporal and spatial distinctiveness of the source fingerprints identified show distinct fingerprint clusters between the two tributaries (Fig. 9) and a low level of overlap between the land use sources, the RB and CB, while CU and BS show distinct fingerprints (Fig. 10). The reduced discrimination between the RB and CB may be due to source signatures to resemble a mix of RB and CB, given both are subsurface materials and presumably less weathered. Overall, however, the signatures provide a clear basis for sediment attribution (Smith & Blake, 2014; Smith *et al.*, 2015).

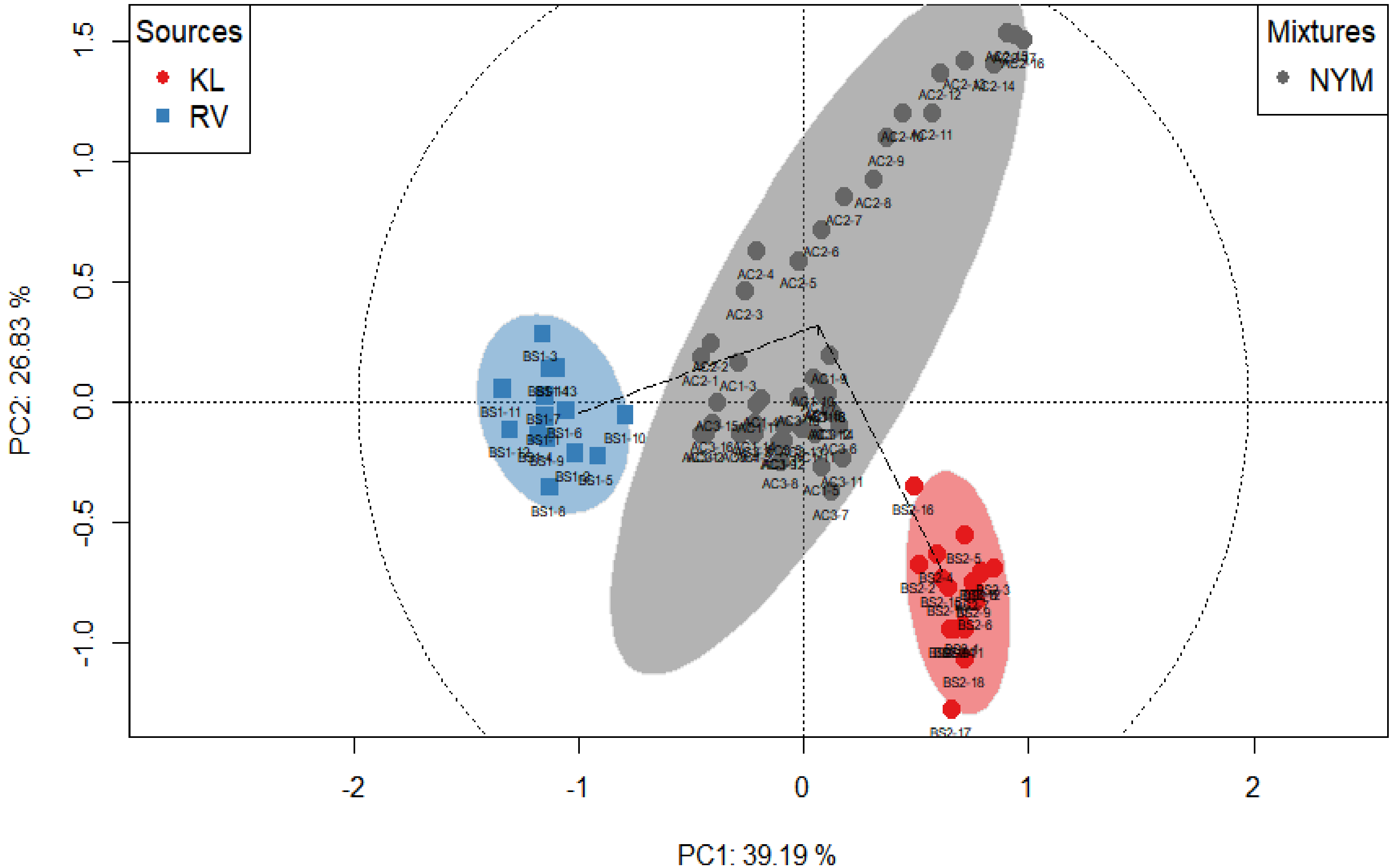


Figure 9: Ordinance biplots visualization of the geochemical drivers of variance in the fingerprint of the tributary sources detailing the intra- and intervariance from a mixture pool

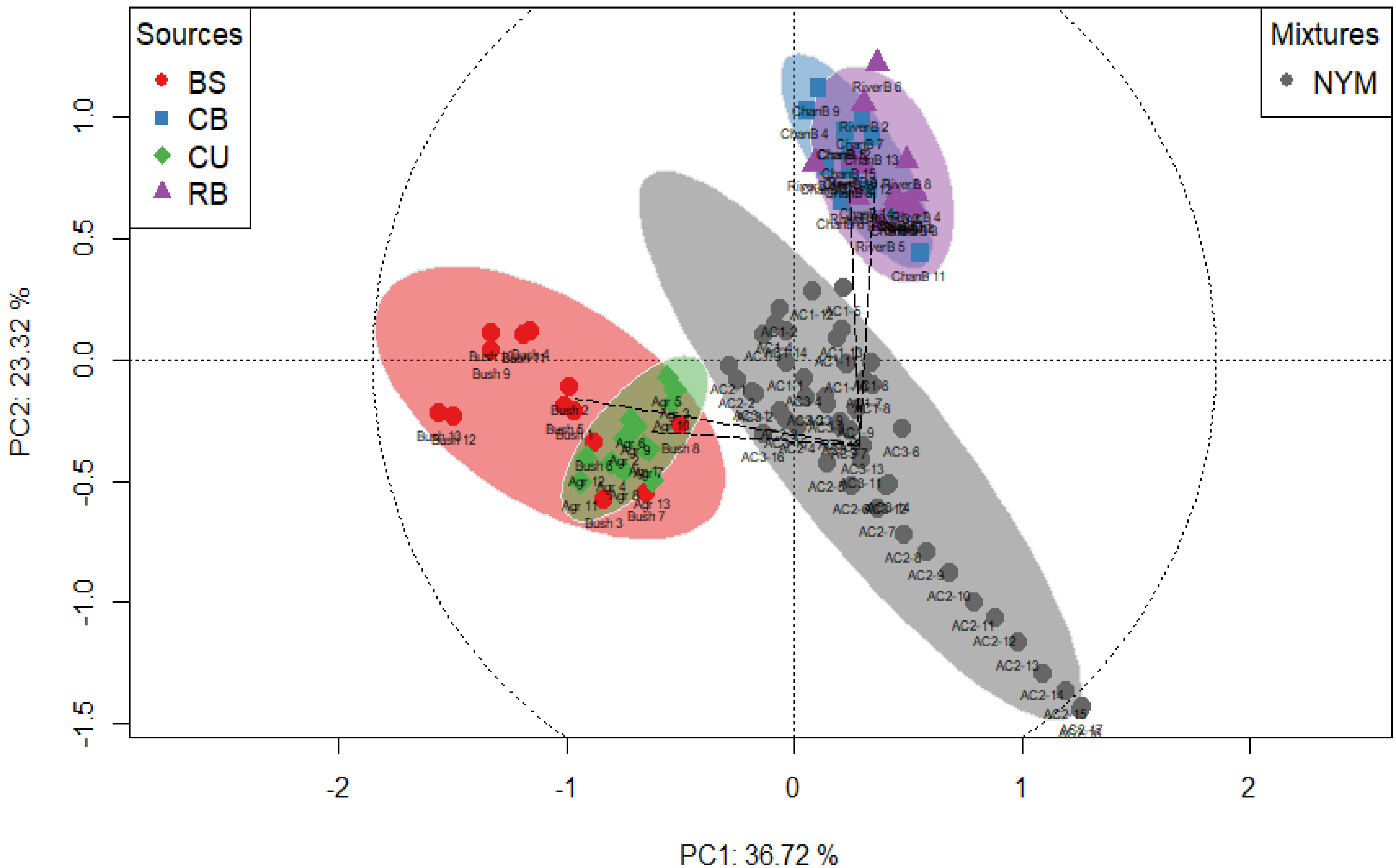


Figure 10: **Ordinance biplots visualization of the geochemical drivers of variance in the fingerprint of land use sources detailing the intra- and intervariance from a mixture pool**

4.3.2 Proportional Tributary and Land Use Contribution

The BMM outputs revealed that the Kikuletwa River contributed 60.3% of “total” reservoir sediment and the Ruvu River 39.7% (Fig. 11). While the Kikuletwa River is the dominant contributing tributary to the total reservoir sediment, cores AC2 and AC3 have a higher proportional contribution of Ruvu sediment with 55.4% and 51.8%, respectively (Table 6). The dominance of Ruvu in AC3 is counterintuitive since the location of AC3 is farther from the Ruvu inlet. However, sedimentation dynamics are also regulated by other factors besides distance to the inlet, such as the dominant flow direction and velocity at the river inlets. Overall, the contribution seems to be well balanced between both rivers.

Table 6: The mean values and Gelman diagnostics (Diag.) of the Bayesian Mixing model runs for both ‘total’ and ‘spatial’ model builds

Tributary	Total		AC1		AC2		AC3	
	Mean	Diag	Mean	Diag	Mean	Diag	Mean	Diag
Kikuletwa	0.603	1.002	0.572	1.002	0.446	1.001	0.482	1.003
Ruvu	0.397	1.001	0.428	1.001	0.554	1.000	0.518	1.002

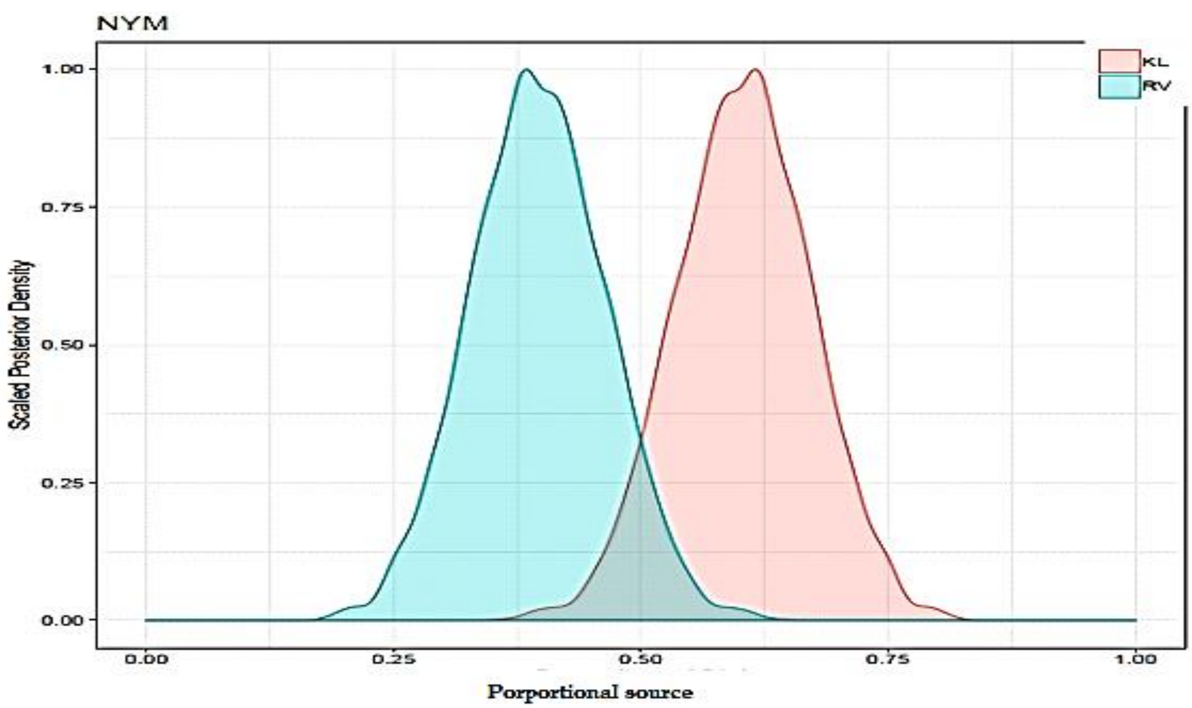


Figure 11: Sediment source apportionment using MixSIAR, where KL accounted with 60.3% and RV with 39.7%

The sign of changes in tributary sediment delivery to the NYM reservoir over time is evidenced by the geochemical stratigraphic record of the sediment cores that experienced distinct periods of high and low sedimentation associated with changes in tributary sediment delivery (Fig. 12). Unmixing the core sections over time yielded changing proportional contributions of the river systems. The proportional contribution of KL in core AC1 seemed to have increased in the lower core before reverting back to almost 50%. The RV decreased from the lower to mid core and then remained stable over time before a slight increase in the upper core. In core AC2, the proportion of KL showed a continuous decrease from 83.1% in the lower core to approximately 50% in the upper, and vice versa for RV. Core AC3 also shows an increase of KL over time, after experiencing a distinct drop in contribution in the upper core (Table 7). Integration of the changing proportional contribution and reconstructed MAR seems to indicate that the most recent increase in sedimentation is mostly driven by increased sediment delivery from the RV system. This finding seems to indicate that increased erosion and sediment transport from the RV system are driven by increased land use change in the catchment. This result corresponds with previous research by Mzuza *et al.* (2017) using magnetic properties showing that RV is the most recent contributing tributary. This association of land use change was also observed using the older and the recent core sections of AC1 and AC3, respectively (Table 8). The comparison was made following the trend of increasing sedimentation rate in core AC1 and AC3 (Fig. 8e, f) with the relative source contribution (Table 8) and (Fig. 13).

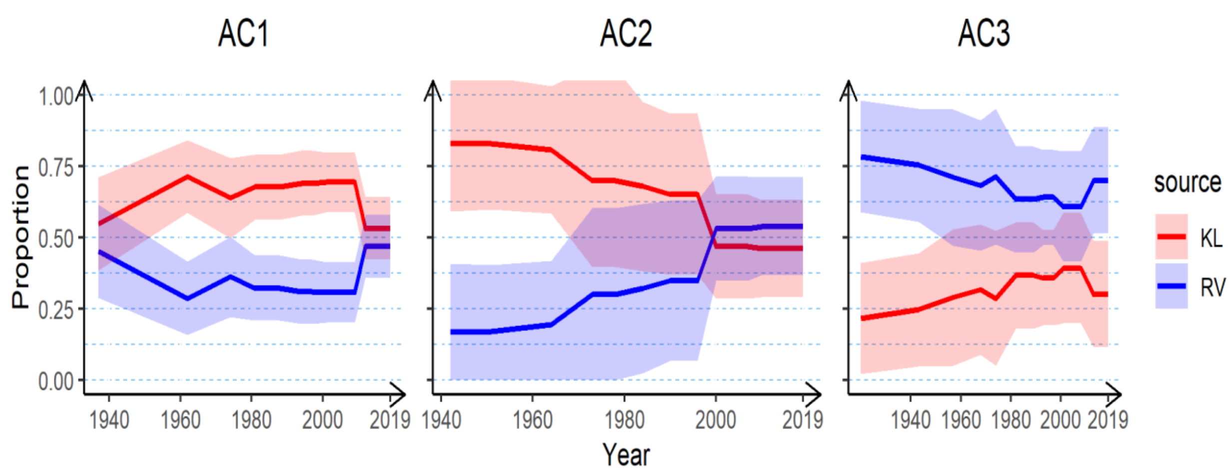


Figure 12: Changing proportional sediment contributions of tributary rivers over time to cores AC1, AC2 and AC3, based on age estimates from the standard CRS model used here as relative age markers

Table 7: The mean values and Gelman diagnostics (Diag.) of the Bayesian Mixing model output of cores AC1, AC2 and AC3 specified for grouped core sections

Core groups	AC1				AC2				AC3			
	Kikuletwa		Ruvu		Kikuletwa		Ruvu		Kikuletwa		Ruvu	
	Mean	Diag	Mean	Diag	Mean	Diag	Mean	Diag	Mean	Diag	Mean	Diag
1	0.532	1.002	0.468	1.001	0.461	1.001	0.539	1.003	0.301	1.003	0.699	1.003
2	0.693	1.002	0.307	1.001	0.469	1.001	0.531	1.002	0.392	1.003	0.608	1.003
3	0.69	1.001	0.31	1.001	0.652	1.005	0.348	1.005	0.359	1.002	0.641	1.002
4	0.677	1.002	0.323	1.001	0.679	1.004	0.321	1.004	0.367	1.001	0.633	1.002
5	0.638	1.002	0.362	1.001	0.7	1.001	0.3	1.001	0.287	1.001	0.713	1.001
6	0.713	1.000	0.287	1.002	0.806	1.005	0.194	1.005	0.317	1.000	0.683	1.001
7	0.548	1.001	0.452	1.001	0.829	1.004	0.171	1.004	0.29	1.001	0.71	1.001
8					0.831	1.000	0.169	1.001	0.247	1.002	0.753	1.002
9									0.216	1.002	0.784	1.002

Unmixing of the core sections directly against the land use pattern demonstrated a significant increase of sediment contribution from the agricultural land (CU) and decrease of the BS, CB and RB in older and younger sections of cores AC1 and AC3. The CU increased from 47.6% (lower core) to 59.6% (upper core) while BS, CB and RB decreased from 11.6%, 20.5%, and 20.3% to 10.1%, 15.4% and 14.9%, respectively, in core AC1 (Table 8) and (Fig. 13). A similar trend was observed in core AC3 where CU increased from 53.5% to 71.0% while BS, CB and RB decreased from 16.1%, 15.1% and 15.3% to 8.7%, 10.3% and 10.0%, respectively (Table 8) and (Fig. 13). The increase in contribution from CU corresponds to the previous studies that observed high levels of deforestation and the loss of permanent vegetation through the rapid expansion of agricultural land and growing urbanization (Mbonile *et al.*, 2003; Tadross & Wolski, 2010; NBS, 2012; NBS, 2018; Said *et al.*, 2019). The decrease in the contribution of the BS, CB and RB is proportional to the increase in CU and the increase in sedimentation in younger core sections, increasing the evidence for changing dynamics of soil erosion and sediment deposition in the reservoir during recent years.

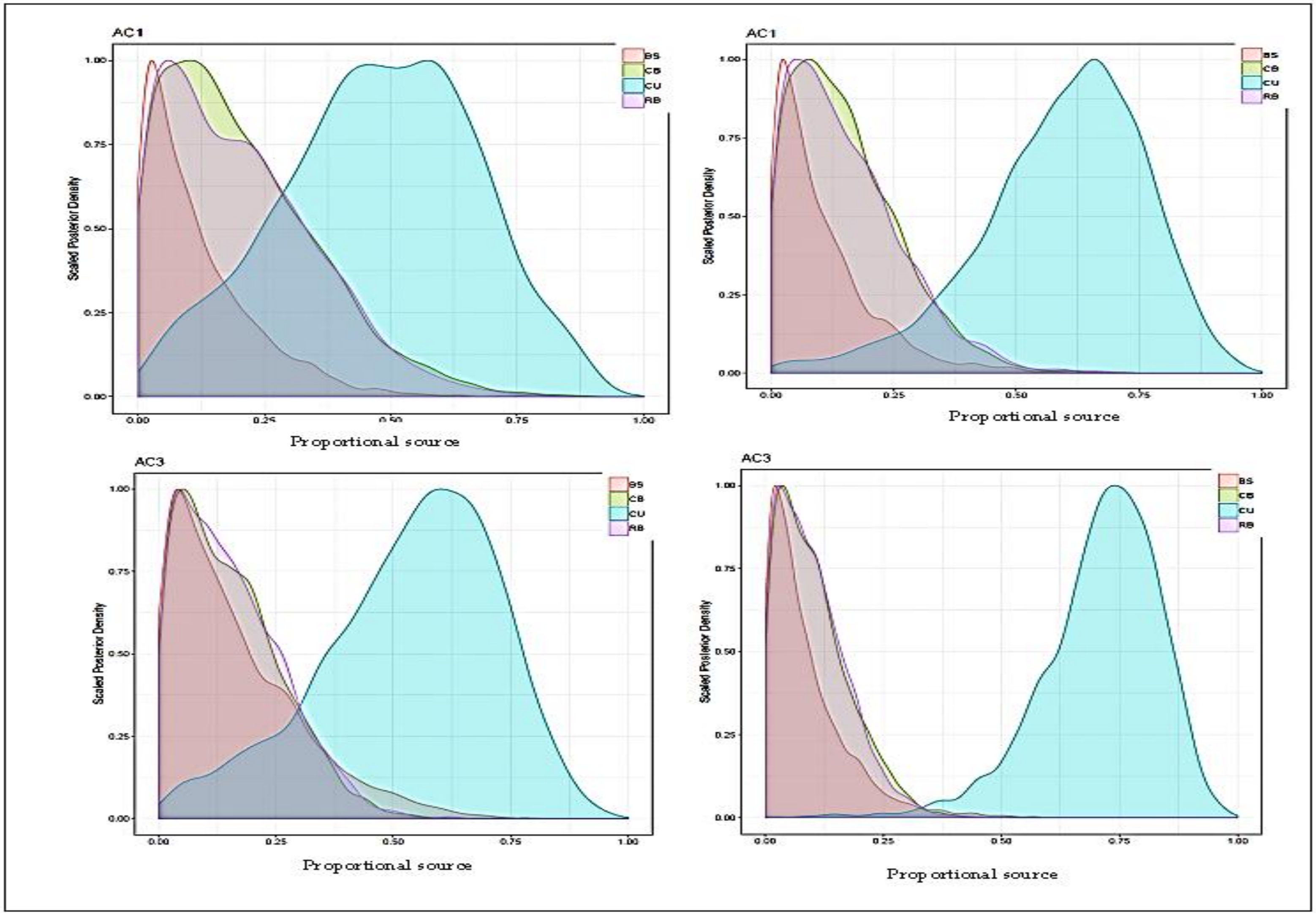


Figure 13: Changing proportional contribution of land use from older to the recent core sections of sediment cores in AC1 and AC3

Table 8: The mean values and Gelman diagnostics (Diag.) of the Bayesian Mixing model output of core AC1and AC3, specified for recent and older core sections

AC1	Bush (BS)		Channel Bank (CB)		Cultivated (CU)		River Bank (RB)	
	Mean	Diag	Mean	Diag	Mean	Diag	Mean	Diag
Recent sections	0.101	1.002	0.154	1.002	0.596	1.001	0.149	1.001
Older sections	0.116	1.001	0.205	1.001	0.476	1.000	0.203	1.000

AC3	Bush (BS)		Channel Bank (CB)		Cultivated (CU)		River Bank (RB)	
	Mean	Diag	Mean	Diag	Mean	Diag	Mean	Diag
Recent sections	0.087	1.001	0.103	1.002	0.710	1.002	0.100	1.002
Older sections	0.161	1.001	0.151	1.002	0.535	1.005	0.153	1.002

Unmixing of the land use types against the riverine sediment revealed that the Kikuletwa tributary had a greatest contribution of sediments from CU (38.4%), and the other types also had significant contribution, RB with 25.6%, CB with 25.5% and BS with 10.5% (Fig. 14, left, and Table 9). The CU on Ruvu River accounted for 44.6% of the sediment contribution, CB with 31.0%, RB with 18.0% and BS with 6.4% (Fig. 14, right, and Table 9). These findings corroborate with previous research showing that both sub-catchment have high soil erosion risk due to unsustainable farming practices (Mbonile *et al.*, 2003; IUCN/PBWO, 2008; Tadross & Wolski, 2010; NBS, 2012; NBS, 2018; Said *et al.*, 2019). The farming practices in both sub-catchments may have increased the sediment connectivity to the channel networks due to less buffering of the soil (Mbonile *et al.*, 2003; IUCN/PBWO, 2008; Tadross & Wolski, 2010; NBS, 2012; NBS, 2018; Said *et al.*, 2019). Although most of the irrigated large-scale plantations are located on the lowlands, sediment sources might have originated from the sloped small-scale agricultural plots and not from the lowland irrigation agriculture. In addition, the use of floodplains for agriculture might have increased the sediment routing to the lake. The evidence of eroding river and channel banks as important sediment sources in fluvial systems (Lawler, 1986; Collins *et al.*, 1997; Lawler *et al.*, 1999; De Rose *et al.*, 2005) is underscored by the concentration of the geochemical tracers in the Ruvu River (Appendix 3). The steady increase in the concentration of the Mn, Sr and S (McLaughlin, 1954; Horowitz, 1991; Marques *et al.*, 2004) with depth supports the hypothesis of the channel banks and gully erosion. The anthropogenic land use changes or natural alteration of geomorphology dynamics of channel and river bank dimension in the watershed may have influenced the subsurface erosion of the allogenic sediment delivery to the reservoir. In addition, the changes in the catchment land cover types from montane forests on mountain slopes to semiarid grasslands (Said *et al.*, 2019) in response to climate change impacts (Blake *et al.*, 2018) and land use pressures (Neff *et al.*, 2000; Hemp, 2009) may have increased the hillslope erosion thus increasing the structural sediment connectivity to the channel networks.

Overall, the fingerprinting analysis shows that cultivated land was the dominant source of the riverine and total reservoir sediment. However, these results should be inferred as estimates and not complete observations because of the temporally and spatially constrained representations of geochemical fingerprints, the geochronological model assumptions and structure and tracer selection protocols. Nevertheless, the compatibility of model output proves the general strength of the methods and the significance of sediment input from this system with important messages for agricultural land management.

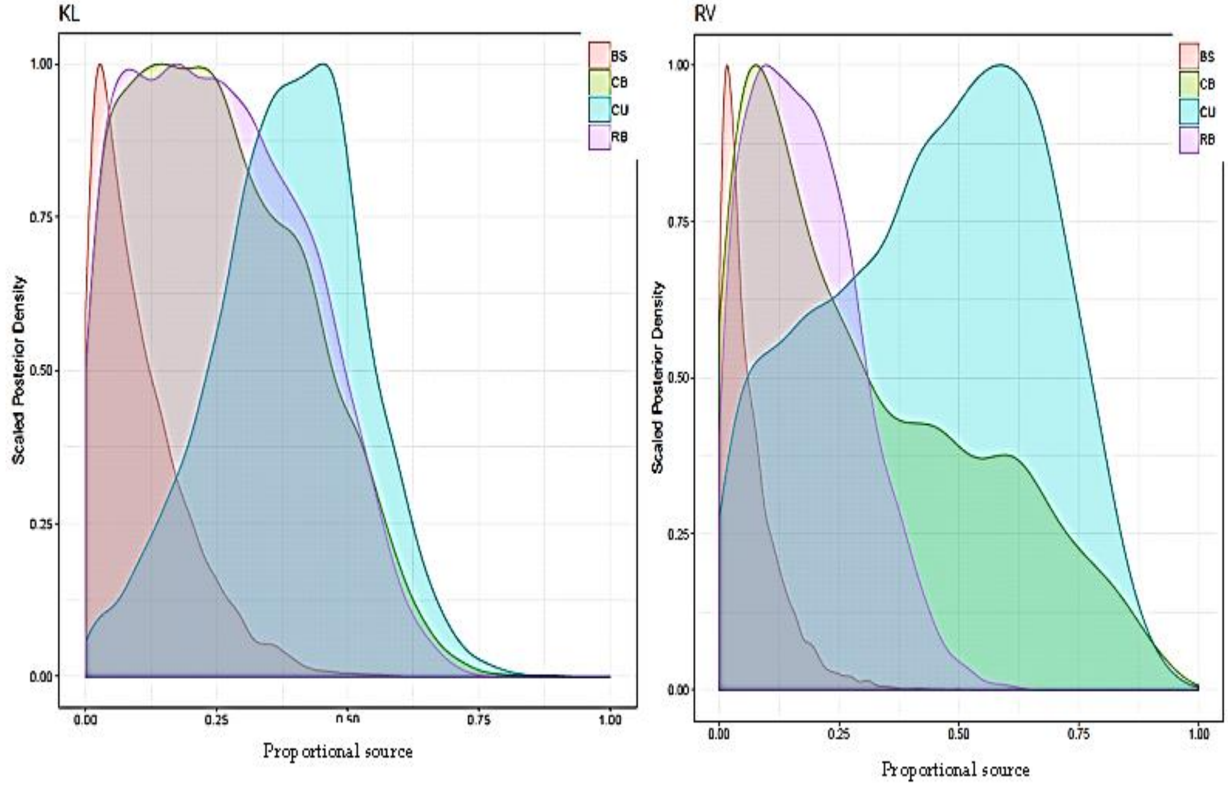


Figure 14: D-MixSIAR of the land use from KL and RV tributaries where CU accounted for 38.4%, CB with 25.5%, RB with 25.6% and bush land with 10.5% for KL (left) and where CU accounted for 44.6%, CB with 31.0%, RB with 18.0% and BS with 6.4% of the total RV sediment (right)

Table 9: The mean values and Gelman diagnostics (Diag.) of the Bayesian Mixing model runs for "land use" in "tributaries" model builds

Land uses	Kikuletwa		Ruvu	
	Mean	**Diag**	**Mean**	**Diag**
Bush (BS)	0.105	1.006	0.064	1.005
Channel Bank (CB)	0.255	1.006	0.310	1.003
Agricultural land (CU)	0.384	1.002	0.446	1.002
River bank (RB)	0.256	1.002	0.180	1.001

4.3.3 Limitation of the Methods and Robustness of the Model

A major challenge of the approach used here is that the riverine sediment fingerprint originates from samples spanning one year, while the cores integrate >60 years. The riverine sediment thus does not include potential variations in sediment fingerprint over time. However, given the strong difference in signature between the Kikuletwa and Ruvu river sediment, it is assumed that intrasource variations will remain smaller than intersource difference. In addition, the model may have difficulty in distinguishing between clusters with similar geochemical fingerprints. For example, RB samples are often a natural integration due to deposition from upstream sources, often leading to large overlaps with other sources. Differences in tracer concentrations between land uses replicated the processes distinguishing physical and chemical properties of subsurface soils. The prevalence of deeply weathered soils at the study site could have led to the overlap between CB and RB (Dawson *et al.*, 1991; Cornu *et al.*, 1999). The CB comprised mostly subsurface soil; however, due to sampling of the entire bank profile it might include a mixture of some surface and subsurface soil.

4.4 Evaluating Soil Carbon as a Proxy for Erosion Risk in the Catchment

4.4.1 Land Cover Changes

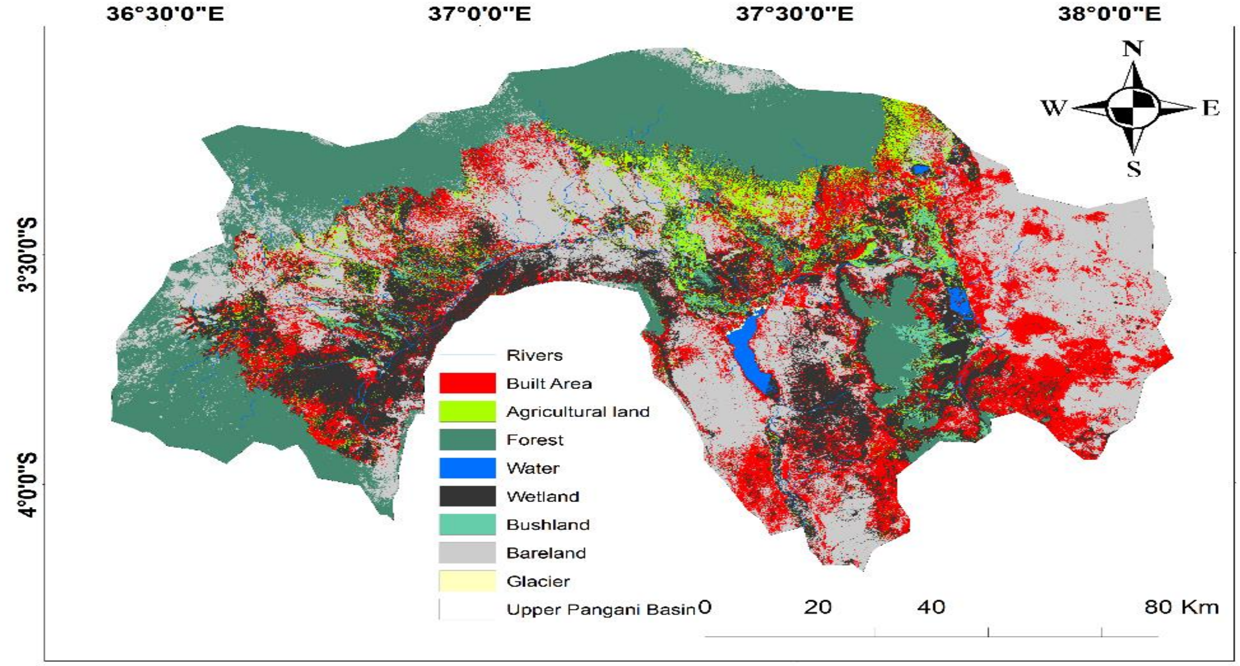
A visual and numerical representation of land cover changes are summarized in Fig. 15 a and b and in Table 10. The summary provides the information on the specific land cover types that has been converted to others and those that has been persistence to change. The net decrease in forest, bare land, grassland and bush land are clear distinguishing trends that evidence conversion to mostly agriculture land by 34.6% (4542 km^2). Built up areas were observed to have significant increased by 6.17% (809.9 km^2). This corresponds to previous studies that observed high levels of deforestation and the loss of permanent vegetation through the fast expansion of agricultural land and growing urbanization (Mbonile *et al.*, 2003; Tadross & Wolski, 2010; NBS, 2012; NBS, 2018; Said *et al.*, 2019). Local manifestation of urbanization includes the establishment of the Siha district; and emergence of many villages and urban suburbs along roads across the catchment (Soini, 2005; Hyandye & Martz, 2017). The expansion of agricultural land and settlements has also led to disappearance of riparian forests and the degradation of riverbanks in the lowlands (Mbonile *et al.*, 2003). Another pronounced change is the considerable net decrease in wetland by -3.96% (519.8 km^2) that would have been caused by drainage and potentially climatic change and variability. The conversion of the catchment land cover types for instance the montane forests on mountain slopes, grassland and bush land to small and large scale plantations in the lowlands is an evidence of increased land use pressures (Neff *et al.*, 2000; Hemp, 2009) and response to climate change impacts (Blake *et al.*, 2018). Another notable change evidenced by the literature is the decrease of glaciers in volcanic peaks of Mt. Kilimanjaro which is an important indicator of environmental changes in the region (Thompson *et al.*, 2002; Mckenzie *et al.*, 2010). Although the decrease in glacier corresponds with the previous studies, however, the cloud cover in the top of the mountain may have influenced the classified image interpretation from which the spectral signatures from the Landsat images obscure the parts of the glacier thus affects the training samples as a result impacts the absolute classification accuracy (Rastner *et al.*, 2019).

Table 10: Losses/gains in land use/cover areas

LU-Classes	1987		2018		1987-2018
	Area (km^2)	% of total	Area (km^2)	% of total	% change
Built up	2203.35	16.8	3015.1413	22.9	6.17
Agricultural land	644.587	4.9	5191.2486	39.5	34.6
Forest	3302.27	25.16	1576.431	12.01	-13.15
Water	79.7715	0.61	106.7616	0.81	0.2
Wetland	562.757	4.29	42.7257	0.33	-3.96
Bush land	2147.58	16.36	1009.1025	7.69	-8.36
Grassland	224.251	1.71	9.4374	0.07	-1.64
Bare land	3950.95	30.1	2167.5091	15.66	-29.94
Glacier	10.5264	0.08	9.6413	0.92	-0.012

The historical land use land cover change in the catchment is confirmed by other previous studies, that include the conversion of shrub and grassland and light vegetation to cultivated land from 1987 to 2005 in Kahe plains (GITEC, 2011), increased forest degradation on the lowlands from 1606 ha to 5170 ha between 1973 and 2000 (Mbonile, 2005), conversion of about 39.5% of bush land to agriculture between 1973 and 2000 years (Mbonile, 2005) degradation of more than 41 km^2 of the forest between 1952 and 1982 (Yanda & Shishira, 2001), conversion of about 49.97 km^2 of shrubs and bush land to agriculture and other uses from 1961 to 2000 years in Kirua Vunjo division (Mbonile, 2005), and increased cultivated land from 54% (in 1973) to 63% in 2000 on the southern and eastern slopes (Misana *et al.*, 2012).

a



b

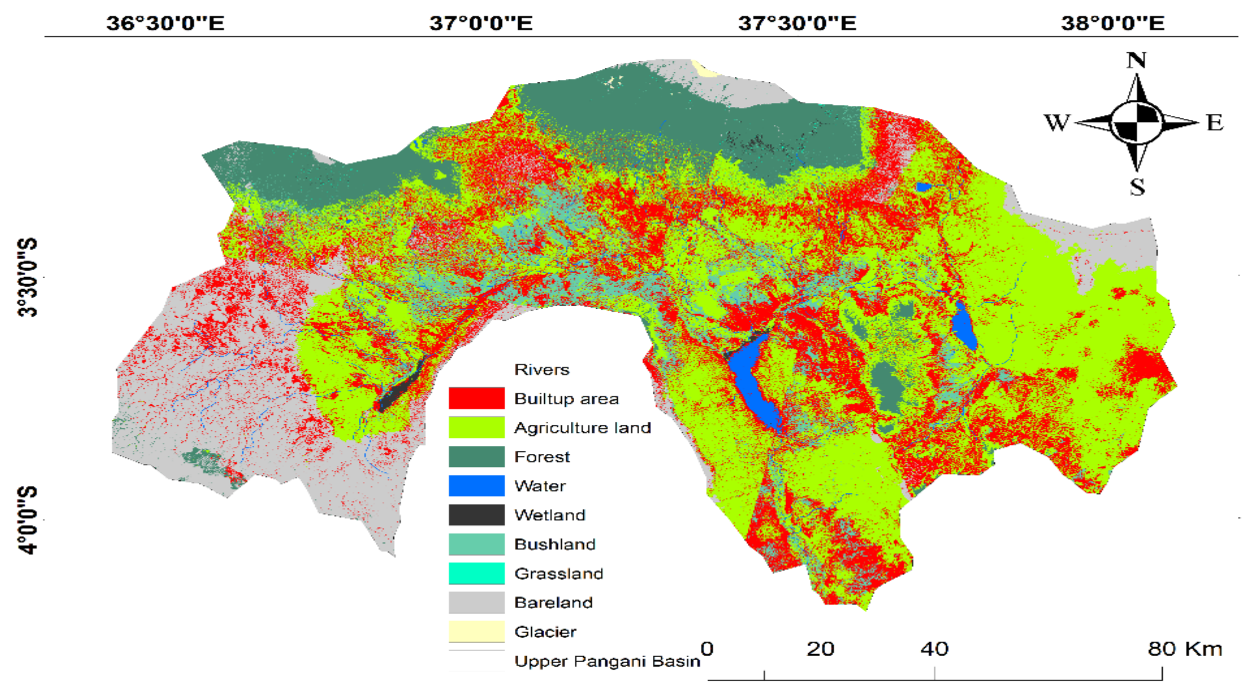


Figure 15: Land cover maps derived from Landsat imagery of 1987 (a) and 2018 (b), respectively

4.4.2 Comparison of Scanner and Laboratory Estimates of Soil Organic Matter

While SOM derived from the AgroCares scanner and converted using the van Bemmelen factor of 1.72 had higher values in all land use types than those derived from LOI, a simple regression analysis between LOI and estimated % SOM from the AgroScanner showed a strong positive correlation (LOI_{FL} R^2 = 0.85, r = 0.93, p = 0.0001; LOI_{CU} R^2 = 0.86, r = 0.93, p = 0.0001; LOI_{GL} R^2 = 0.68, r = 0.83, p = 0.003; LOI_{BS} R^2 = 0.88, r = 0.94, p = 0.0001; LOI_{BL} R^2 = 0.83, r = 0.91, p = 0.0002, Fig. 16). From an environmental perspective, LOI and SOM were significantly different between land uses decreasing in the following order: Bare land > forest land > bushland > cultivated land > grassland and bare land > forest land > grassland > bushland > cultivated land respectively (Appendix 4). Based on the null hypothesis (H_o) that "there is no correlation between the estimated SOM in the laboratory (LOI) and from Agroscanner" the (H_o) was therefore rejected since there is strong correlation between the two variables ($p < 0.05$). However, the SOM for forest land and bare land were close at approximately 5% each, while the LOI for cultivated land and grassland were also similar at 2%. Most soils in the catchment are characterized by a high clay content (Ndomba, 2015) that has the ability to contain more carbon (Arnalds, 2004; Walker & Desanker, 2004). Although the SOM stocks were significantly different among the land use systems, a similar pattern was observed between the forest land and bare land (Appendix 4). This similar pattern might be explained by their dominating soil textures that were primarily loam to clay loam because these textures supports the function of the soil biological community by providing a large and moist surface area in water films around loam and clay loam particles that are often protected within aggregates (Collier *et al.*, 2021; Dungait *et al.*, 2019). The strong relationship between the laboratory measured SOM (LOI) and SOM from the AgroCares scanner showed that LOI is a good method for the determinations of SOM where formal measurements are limited.

4.4.3 Soil Organic Matter on Soil Aggregate Stability in Different Land Uses

This study showed that there are significant differences in the soil aggregate stability and SOM stock between the different land use types in the upper Pangani basin. The soil aggregate stability decreases approximately in the following order: Forest land > grassland > bare land > cultivated land > bush land (Appendix 4). The results indicated that SOM and WSA were influenced by land use management type (Fig. 17). Similar to SOM stocks, the WSA in arable soils was typically less than in forest land, grassland and bush land. However, the significant difference between WSA in cultivated land in comparison to other land use types was observed with a wide range in WSA here indicating high variability in soil behaviour under low SOM conditions.

While the means of SOM in forest land and bare land were almost similar, there was a substantial difference in the median values of WSA which might be attributed by difference in soil texture. Soil textures supports the function of the soil biological community by providing a large and moist surface area in water films around loam and clay loam particles that are often protected within aggregates (Dungait *et al.*, 2019; Collier *et al.*, 2021). This similar pattern observed within and between land use types imply that the SOM and WSA not only related to land use management types (Fig. 17 & Appendix 4) but also influenced by other factors such as soil textural properties, geology, clay content, exchangeable cations and other human activities. A multiple linear regression was run to evaluate the influence of SOM and LOI in soil aggregate stability in water (WSA) where all variables were statistically insignificantly to the prediction $p < 0.05$ as follows: Forestland (WSA, $R^2 = 0.235$, $p > 0.05$ (0.299)), cultivated land (WSA, $R^2 = 0.425$, $p > 0.05$ (0.083)), grassland (WSA, $R^2 = 0.119$, $p > 0.05$ (0.642)), bare land (WSA, $R^2 = 0.161$, $p > 0.05$ (0.540)), bushland (WSA, $R^2 = 0.072$, $p > 0.05$ (0.769)). Since there was no significant differences that existed between the variables the null hypothesis (H_o) was retained implies that the SOM has influence on aggregate stability. The Kruskal Wallis test also showed that the difference between the medians of LOI and SOM were not significantly different in all land uses across categories of WSA, $p > 0.05$ (0.164 and 0.195) respectively for bare land, $p > 0.05$ (0.277 and 0.267) respectively for bush land, $p > 0.05$ (0.692 and 0.441) respectively for grassland, $p > 0.05$ (0.386 and 0.262) respectively for cultivated land and $p > 0.05$ (0.127 and 0.197) respectively for forestland (Table 11). Since the null hypothesis (H_o) states that the "populations medians are equal" and these results revealed that the medians of LOI, SOM and WSA were not significantly different in all land uses ($p > 0.05$), the null hypothesis was therefore retained implying that the soil aggregate stability is influenced by the SOM. The strong relationship between LOI, SOM and WSA indicated that LOI approximation for WSA and augmentation of organic matter in soil is a good strategy for farmers to reduce risk of erosion.

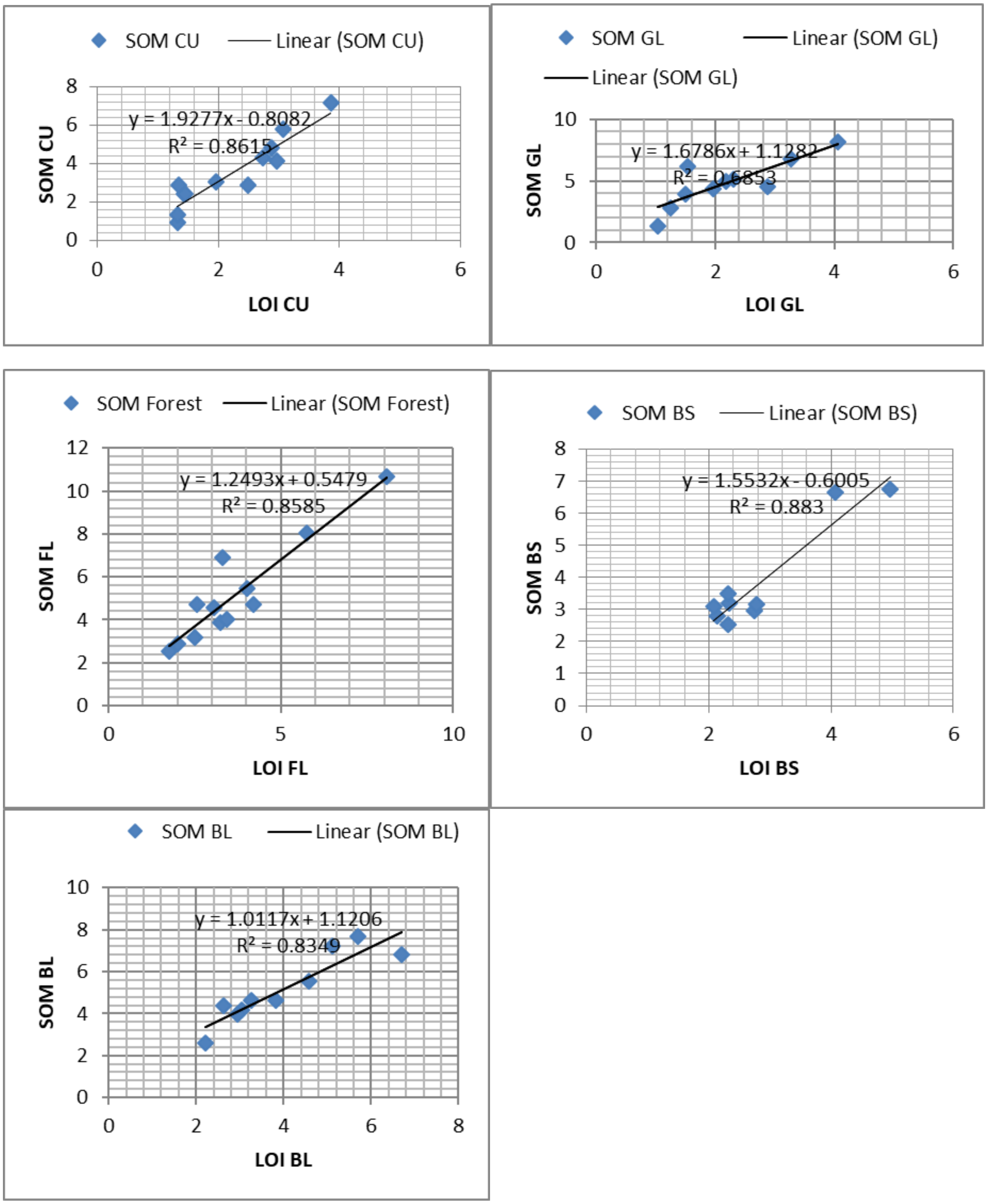


Figure 16: Relationship between loss on ignition and soil organic matter

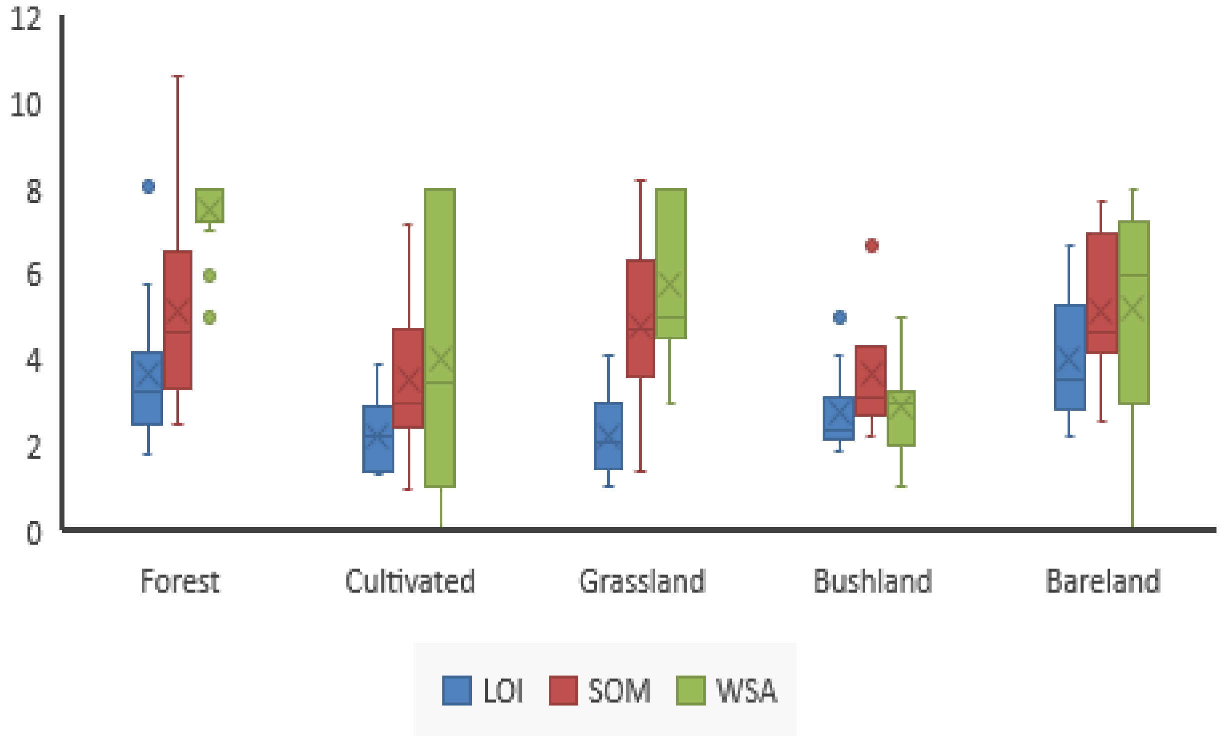


Figure 17: Boxplot comparing the ranges of LOI, SOM and WSA in each land use classification

Hypothesis Test Summary

	Null Hypothesis	Test	Sig.	Decision
1	The distribution of LOI is the same across categories of WSA.	Independent-Samples Kruskal-Wallis Test	.163	Retain the null hypothesis.
2	The distribution of SOM is the same across categories of WSA.	Independent-Samples Kruskal-Wallis Test	.067	Retain the null hypothesis.

Asymptotic significances are displayed. The significance level is .05.

Figure 18: Kruskal Wallis test results from SPSS

The higher values of WSA in forest land and grassland indicate the stability of their soil structures in relation to longer-term vegetation cover. The forest soil exhibited the highest degree of the aggregate stability, which may be due to the higher SOM content input to soil vegetation cover, higher biological activity, and the protection of soils against degradative processes. The WSA was also high in the grasslands, which might be due to higher root biomass and the return of residuals to the soil. High root biomass and residues increase the organic matter content as the carbon source that support the water holding capacity in soil in turn becomes a conducive environment for the decomposition of organic matter (Sartori *et al.*, 2007). The higher WSA and SOM in bare land was unexpected and might be an influence of the high clay content i.e independent of land cover. Soil aggregate stability is affected by various parameters that acts as binding agents, includes organic matter, soil texture, iron and aluminium oxides, carbonates, and metal cations (Six *et al.,* 2004). Increasing organic matter content enhances the stability of soil aggregates, and this is more distinct in soils with higher clay-fraction content. The soil texture in the bare land was mostly clay loam and sandy clay loam which have characteristics to immobilize different macro- and micronutrients and accelerate changes in the microbiological activity of the soil (Xie *et al.,* 2015) (Appendix 4). The low WSA and SOM content in cropland is typically an indication of tillage practices that reduce the soil aggregation process, aggravates soil loss through erosion. The weakening in the structural stability of cultivated soils may apparently be attributed to aggregate disruption and SOC distribution in various physical fractions, including tillage operations and other erosion-facilitating practices that lead to rapid breakdown (Six *et al.,* 2000; Emadodin *et al.*, 2009). The results of this study indicated that the cultivated lands have a lower SOM content compared to the natural land cover types. These results are in line with other studies elsewhere, indicating that cultivation usually markedly decreases soil carbon (Sartori *et al.,* 2007; Deng *et al.,* 2014; Cerdà *et al.,* 2018; Dungait *et al.,* 2019; Seeger *et al.,* 2019; Collier *et al.,* 2021). The forest conversion to cultivation land and settlement (Table 3) likely influences the level of organic carbon due to soil loss, and the more rapid oxidation process of SOC leads to rapid reductions in the SOM values of surface soils (Abrishamkesh *et al.,* 2011). During cultivation, SOM can be lost through multiple processes such as tillage, increased erosion, reduction of vegetative input and biological activity. Therefore, the lower SOM in the catchment is evidence of the increasing conversion of land cover types to large scale plantations in the lowlands that may have influenced the hillslope erosion. In particular, the removal of plant residue from the soil surface layer through different land management practices including cultivation destroys soil macro-aggregate formation which significantly alters soil texture and SOM increase soil erodibility. However, some crop land soils in the Msitu wa Tembo and Soko scheme had higher SOM and aggregate stability (Appendix 4). This might be due to higher fertilizer input in cropland which was revealed during the sampling campaign. The application of NPK fertilizer in the fields could have increased biological activity that promotes the formation of water-stable aggregates, which in turn improved the mechanical stability of soil aggregates by binding soil mineral particles (Šimanský *et al.,* 2019). The labile organic carbon (LOC) of SOM is responsible for organic amendments that enhance the water-stable aggregation process. In turn, the water- stable aggregation increases the availability of organic compounds that promotes soil microorganism growth, which produces more extracellular polysaccharides and promotes aggregate formation (Dai *et al.*, 2019). The low WSA and SOM in bush land might be due to low root biomass production influenced by regular animal grazing that lead to soil degradation. The present study and other studies by Tang *et al.* (2016), Delelegn *et al.* (2017) and Nath and Lal. (2017) show that land use change has a major and important effect on soil aggregate stability, structure and, consequently, on water erosion.

Development of strategic land management plans based on the observed relationship between the slake test/ aggregate stability and SOM is highly appropriate in soils with high clay contents because the distribution of clays in soil is associated with reduced infiltration and run-off, sediment load and crust formation (Watts & Dexter, 1997). Although the catchment soil sample were composed of large clay contents, substantial differences between the stability of aggregates in water was observed. From a sustainable land management perspective soil organic carbon increases soil porosity and improves the mechanical flexibility to compression stress (Zhang *et al.*, 2005). The cohesive effect of organic matter and its behaviour to sustain soil microbial activities makes soil organic carbon content a good proxy for soil degradation (Dungait *et al.*, 2019). The relationship between SOC content and improved physical quality of soil, and the subsequent benefits for the quality of farmed soils are widely acknowledged (Dungait *et al.*, 2012; Paustian *et al.*, 2019).

CHAPTER FIVE

CONCLUSION AND RECOMMENDATIONS

5.1 Conclusion

This study has demonstrated the potential use of geochemical profiles of reservoir sediment columns as a valuable alternative tool for independent confirmation of changing sedimentation rates in the context of disturbance in East African hydropower catchments. The allogenic maxima and autogenic minima of Nyumba ya Mungu reservoir freshwater have reciprocally aligned with $^{210}Pb_{ex}$ reconstruction of sedimentation dynamics over time. This study successfully deployed a quantitative sediment fingerprinting technique to apportion recent potential sediment sources in relation to land use change. The integration of geochemical fingerprinting within a Bayesian Mixing Model framework pointed toward one of two tributaries, the Kikuletwa River, as the dominant contributing tributary to the total reservoir sediment. In addition, the fingerprinting analysis shows that cultivated land was the dominant source of the riverine and total reservoir sediment. Moreover, the integration of the changing proportional contribution and reconstructed accumulation rates seems to indicate that the most recent increase in sedimentation is mostly driven by increased sediment delivery from the other tributary, the Ruvu system. This result corresponds with previous research by Mzuza *et al.*, 2017 using magnetic properties that Ruvu is the most recent contributing tributary, which indicates the pace of land use change in this system is having a profound effect on sediment supply. The assertion is also backed by sedimentary evidence from cores AC2 and AC3, which had a higher proportional contribution of Ruvu-derived sediment. This finding seems to indicate that increased erosion and sediment transport from the RV system are driven by increased land use change in the catchment. Overall, the study revealed major changes in the sedimentation dynamics over time, which is probably driven by a complex interaction between land use changes, climate changes and natural rainfall variability. Since sedimentation is always highly localized in large reservoirs, evaluation of localized sedimentation effects would provide a full representation of spatially specific sedimentation issues and a deeper understanding of the driving processes within the catchment. To this end, future studies using Bayesian Mixing Model in large reservoirs should aim to include spatial factors in the model setup concerning receptor sediment mixtures; thus, a better representation of complex sedimentation dynamics in space could be obtained. The results underscore the necessity for targeted erosion mitigation strategies on the potential sources to limit soil erosion and reduce further impact on the reservoir water quality. In addition to the sedimentation dynamics over time, this study also revealed that, land use/cover changes through anthropogenic activities and climate changes have direct impact on soil organic matter and on water stability aggregate. Land use change can thus potentially increase the susceptibility of soils to erosion by water when soil organic matter is reduced. The potential to use the slake test is due to its wide applicability for many years to specific conditions of the soils that were tested and adapted. The slake test scoring protocols seemed to reasonably increase the sensitivity of the test without compromising the feasibility of its application by land managers than the existing USDA version. The robust associations between water stability aggregate, soil organic matter and land management practices in this study suggest that, where soil and climate conditions are similar within a defined region, the rapid assessment of water stability aggregate using this approach offers an inexpensive means of assessing and providing a numerical score of ‘soil health’, and potential proxy for direct measurement of soil organic carbon which in turn used to detect changes imposed by management. The understanding of the relationship between soil erodibility and the top soil organic carbon, thus, help to support sustainable land management for soil conservation. The prospect of using water stability aggregate as a rapid proxy for soil organic carbon change by agronomists where advanced measurements are limited would offer agronomists with a new tool for monitoring soil health. More research is essential to initiate its potential in different soil types in a range of management scenarios.

5.2 Recommendations

This study aimed to reconstruct the sedimentation rates over time and identify the changing sources of sediment in a major hydropower reservoir in Tanzania, the Nyumba ya Mungu. The study also aimed to evaluate the soil carbon as a proxy for soil erosion risk in the catchment. Based on the findings of this study, the following are the recommendations:

(i) Future studies that will lead to the understanding of the contemporary land-river-reservoir connections through sediment budget approaches. The understanding of the processes that result in erosion and its connectivity to the river channel, storage in hillslopes, floodplains and sediment accumulation in the reservoirs is vital for the sustainable management of the reservoir. Sediment budget approaches that involves the integration of several different techniques such as remote sensing GIS models, sediment source fingerprinting, FRNs for dating and soil redistribution, etc. provides such a holistic perspective by accounting for the various sediment sources, transport, sinks and redistribution when the sediment is routed through that catchment.

(ii) Future investigation of the sediment dispersion pattern through lithological mapping of the lake bottom sediments using gamma natural radiation. This could provide useful information about the origin and transport of sediments in the reservoir.

(iii) Establishing the "Structural hierarchy'' of high erosion risk areas in a river basin in terms of sub-watershed distribution through integration of Deconvolutional-MixSIAR (D-MixSIAR) with GIS models.

(iv) Future studies using BMM in large reservoirs should aim to include spatial factors in the model setup concerning receptor sediment mixtures; as a fixed categorical effect to account for localized sedimentation effects thus, a better representation of complex sedimentation dynamics in space could be obtained. Furthermore, future sediment tracing studies that want to constrain tributary sources of increased sedimentation should, to the extent possible, account for the overlap between source fingerprints and increased sedimentation signals, and include sediment cores from different reservoir areas in their study design.

(v) Future studies are required for mapping of the hotspots areas of increased erosion which provides a spatially qualitative proxy of erosion risk changes following land conversion. Pinpointing hotspots areas of increased risk will therefore support targeting these areas for more detailed investigation of controls on erosion processes and guide stakeholders and policy makers in land management decisions of soil conservation measures and possible action

(vi) The implementation of sustainable soil management practices that both permits production of food and soil conservation to protect hillslope and downstream systems: these includes afforestation, revegetation, and sustainable grazing management programs and, similarly, agricultural practices that emphasize soil conservation, such as tillage and crop management; terrace construction.

(vii) Future application of Compound Specific Stable Isotope (CSSI) to study the agricultural river basin sediment delivery under specific crop cover for effective mitigation of sediment sources. The informed and efficient catchment management decisions in agricultural land require crop-specific information on sediment source. The current geochemical and radiometric approaches does not provide such information because they depend much on physical characteristics of the soil (e.g. the conservative behaviour) which might be changed by the external influences in the catchment.

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APPENDICES

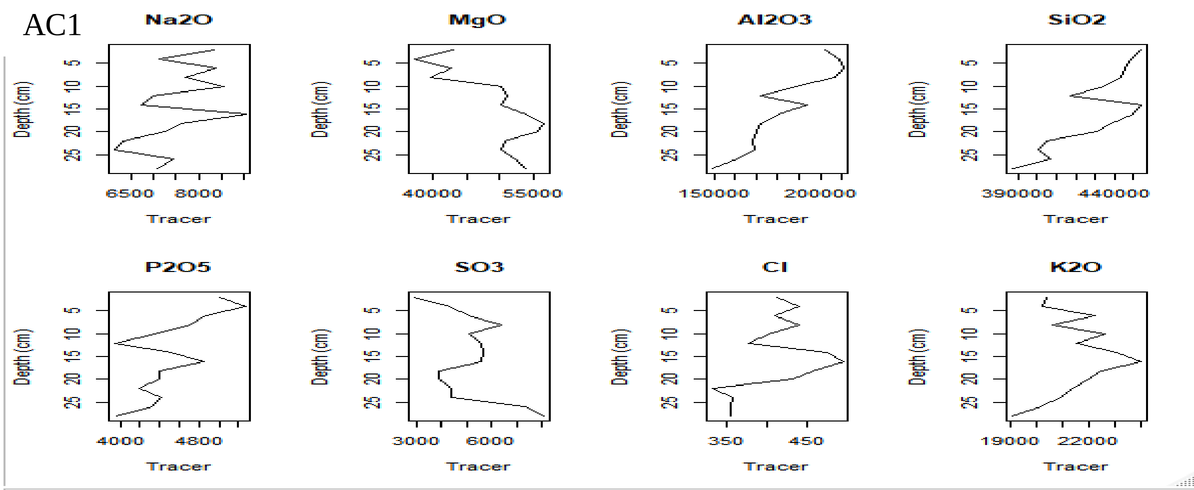
Appendix 1: Sampling locations of potential sediment sources

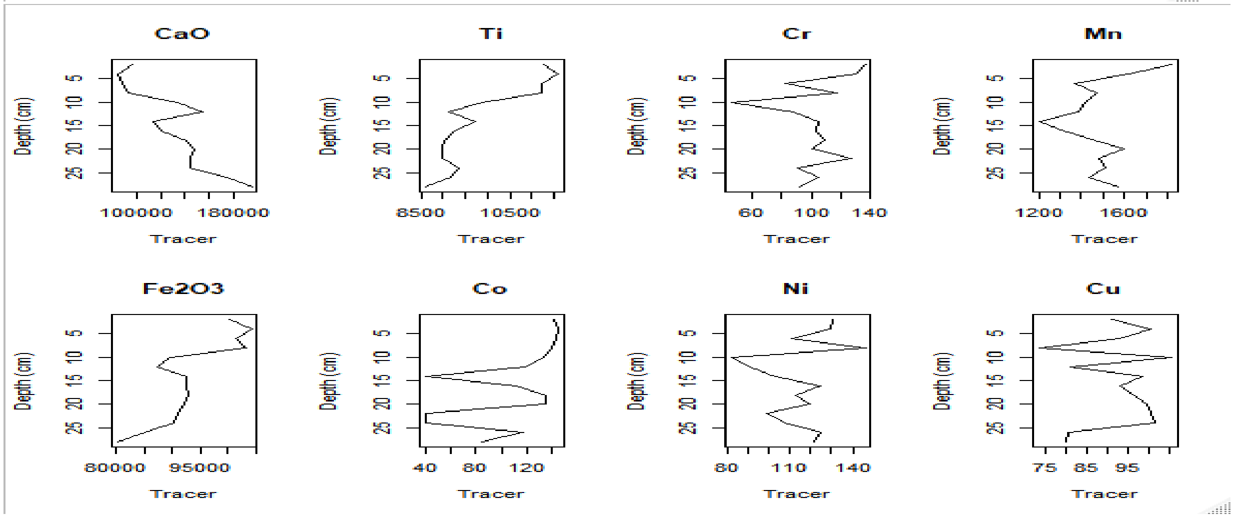
S/N	Land use type	Field Name	Coordinates		Sampling date
			Latitude	Longitude	
	Bushland (BS)				
		Handeni	-3.56527	37.47837	16/09/2019
1		Kituri Old	-3.53449	37.53131	17/09/2019
2		Maweni	-3.47413	36.85271	17/09/2019
3		Kifaru High School	-3.52236	37.55109	17/09/2019
4		Soko village	-3.49409	37.49027	17/09/2019
5		Masama Rundugai	-3.42217	37.23557	17/09/2019
6		KDC	-3.33745	37.35742	19/09/2019
7		Karanga	-3.34442	37.31731	18/09/2019
8		Chemka hot spring	-3.44293	37.19461	18/09/2019
9		Chemka	-3.45201	37.18612	19/09/2019
10		TPC	-3.54638	37.31287	19/09/2019
11		Kiruani	-3.62363	37.32218	18/09/2019
12		Msitu wa Tembo	-3.57876	37.30535	19/09/2019
14		Msitu wa Mbogo	-3.52608	36.87787	20/09/2019
	Channel Bank (CB)				
15		Handeni	-3.56527	37.47837	16/09/2019
16		Kikafu juu	-3.31439	37.21994	19/09/2019
17		Kikafu downstream	-3.408449	37.29042	19/09/2019
18		Msitu wa Tembo	-3.58782	37.30905	18/09/2019
19		Kiruani	-3.62363	37.32218	19/09/2019
20		Mabungo Mue	-3.40466	37.51104	17/09/2019
21		Karanga	-3.34442	37.31731	18/09/2019
22		Longai Hai	-3.40601	37.26669	19/09/2019
23		Chemka Spring	-3.44228	37.19378	19/09/2019
24		Sanya station R	-3.34572	37.11877	19/09/2019
25		KDC	-3.33778	37.35733	18/09/2019
26		Kware Hai	-3.32579	37.16451	19/09/2019
27		Dehu_Soko village	-3.49602	37.48468	17/09/2019
28		Soko village	-3.49409	37.49027	17/09/2019
39		Ngasini Rau	-3.48454	37.46228	17/09/2019
30		Kituri Old	-3.53449	37.53131	17/09/2019
	Agricultural land (CU)				
31		Chekereni	-3.45422	37.54852	17/09/2019
32		Mawala	-3.51898	37.43456	18/09/2019
33		New Kituri	-3.50029	37.55564	17/09/2019
34		Ruvu Old Kituri	-3.53967	37.51661	17/09/2019
35		Kiomu	-3.47948	37.50149	17/09/2019
36		Miwaleni	-3.42979	37.44771	17/09/2019
37		Weruweru	-3.31525	37.25692	19/09/2019

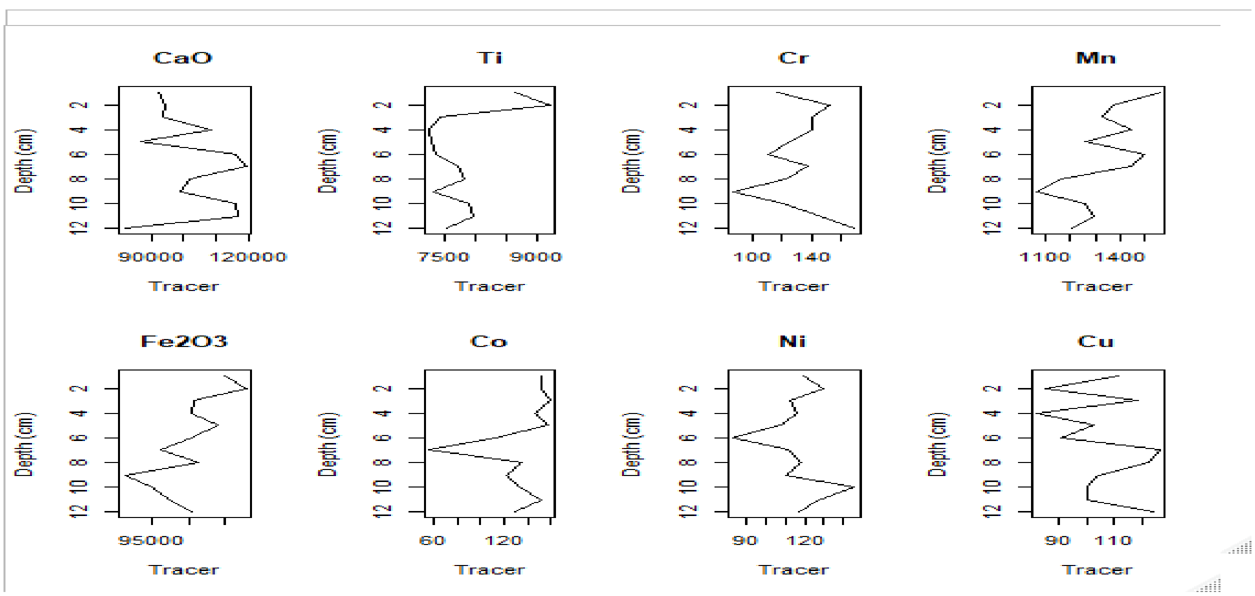
S/N	Land use type Bushland (BS)	Field Name	Coordinates		Sampling date
			Latitude	Longitude	
38		Msitu wa tembo TPC	-3.55385	37.30664	18/09/2019
39		Machame tool	-3.32695	37. 23079	18/09/2019
40		Kuanisira	-3.26456	37.14784	19/09/2019
41		Sanya	-3.34568	37.11808	19/09/2019
42		Kochakindo	-3.4979	37.53082	17/09/2019
43		Rundugai Masama	-3.42258	37.23517	19/09/2019
	River bank (RB)				
44		Marangu Mtoni	-3.28282	37.52229	16/09/2019
45		Whona, kilema	-3.28794	37.50742	16/09/2019
46		chekereni	-3.45464	37.54852	17/09/2019
47		Ruvu bridge	-3.53967	37.51661	17/09/2019
48		Lukaranga	-3.44128	37.30331	19/09/2019
49		Weruweru	-3.31651	37.25808	19/09/2019
50		Samanga	-3.30392	37.52247	16/09/2019
51		Kikafu upstream	-3.31439	37.21994	19/09/2019
52		Upper Kikuletwa	-3.45493	37.20751	19/09/2019
53		Mbuguni	-3.45786	37.18764	19/09/2019
54		Kuanisira	-3.26451	3714718	19/09/2019
55		Kware river Hai	-3.32579	37.16451	19/09/2019
56		Kikuletwa	-3.55248	37.30713	17/09/2019
57		Mue River Mbungo	-3.40711	37.51034	17/09/2019
	Riverine sediments				
		Ruvu riverine	3°34'21.29"S	37°28'02".39E	17/09/2019
		Kikuletwa riverine	3°33'8.93"S	37°18'25".67E	18/09/2019

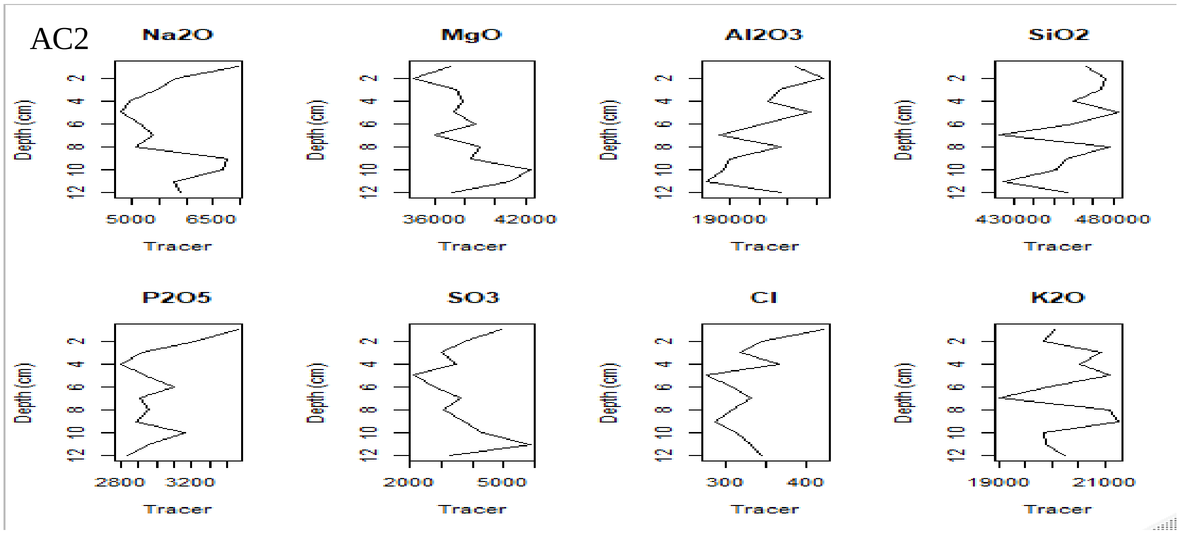
Appendix 2: Geochemical depth profiles of core AC1, AC2 and AC3 respectively Concentration in ppm

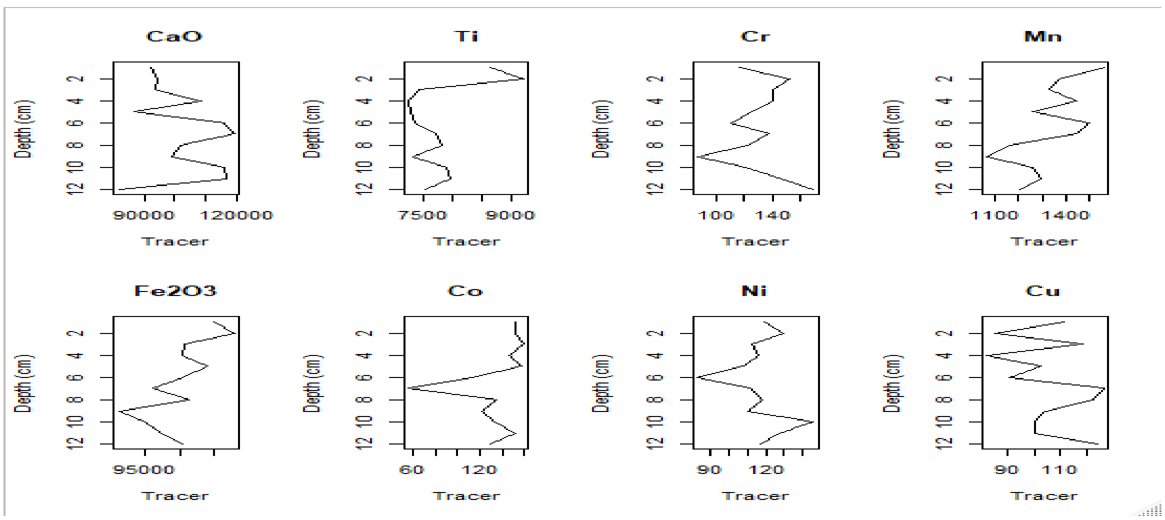
AC1

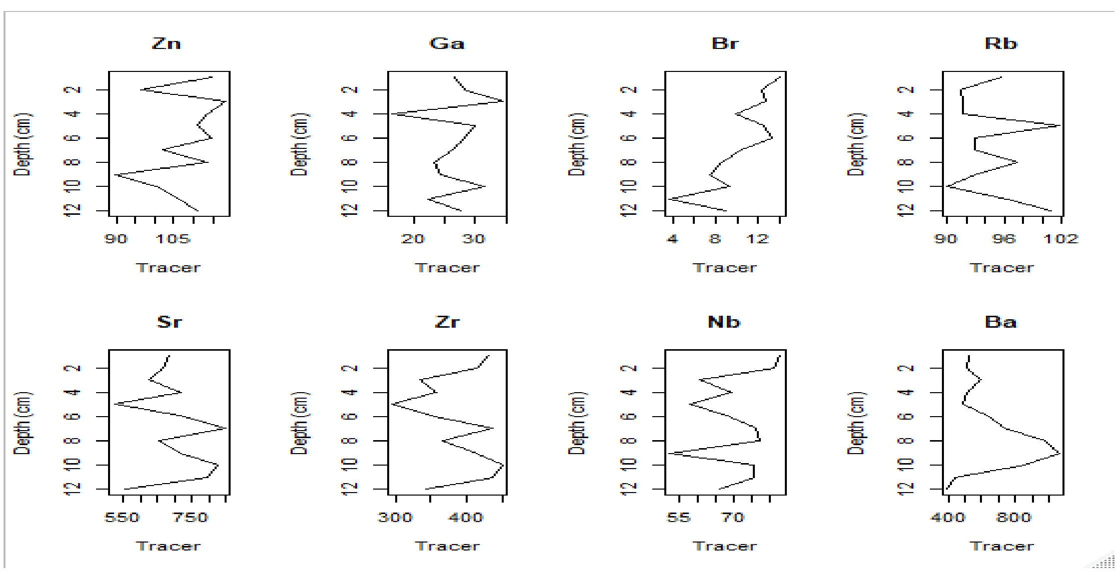


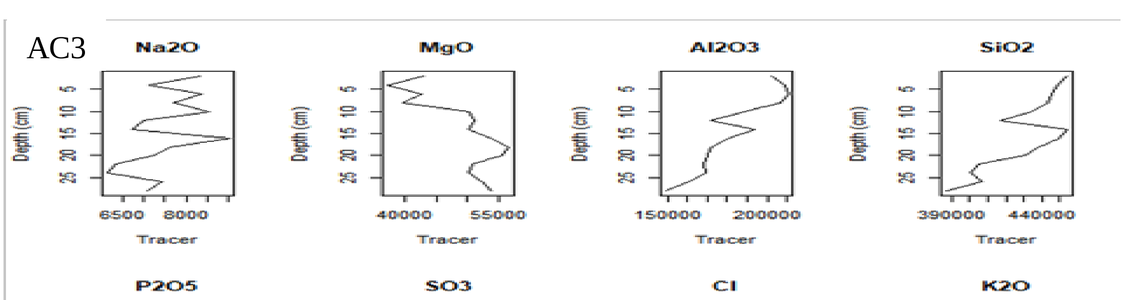


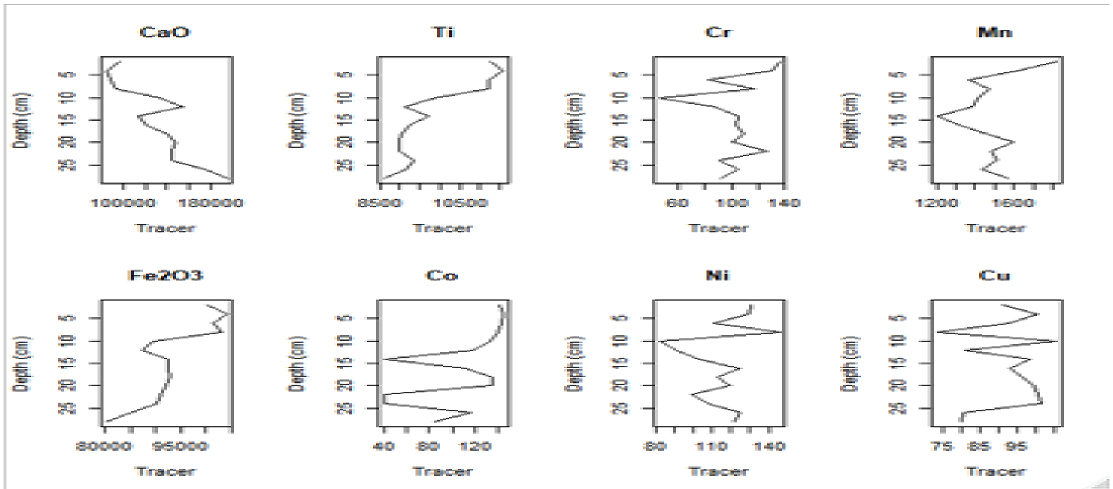



AC2
Na2O
MgO
Al2O3
SiO2
P2O5
SO3
Cl
K2O
Depth (cm)
Tracer
5000 6500
36000 42000
190000
430000 480000
2800 3200
2000 5000
300 400
19000 21000

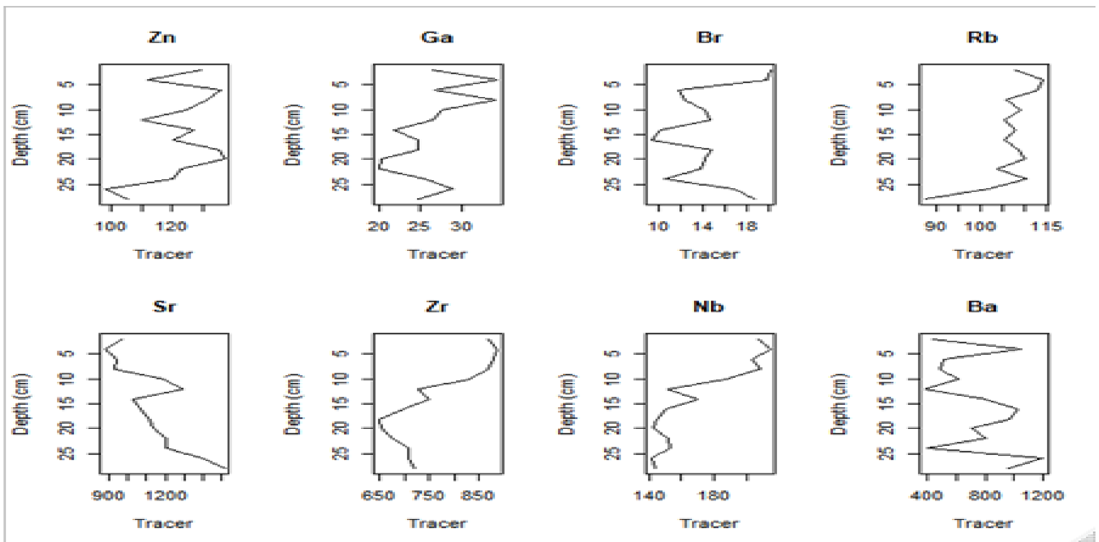

CaO
Ti
Cr
Mn
Fe2O3
Co
Ni
Cu
Depth (cm)
Tracer
90000 120000
7500 9000
100 140
1100 1400
95000
60 120
90 120
90 110


Zn
Ga
Br
Rb
Sr
Zr
Nb
Ba
Depth (cm)
Tracer
90 105
20 30
4 8 12
90 96 102
550 750
300 400
55 70
400 800


AC3
Na2O
MgO
Al2O3
SiO2
Depth (cm)
Tracer
6500 8000
40000 55000
150000 200000
390000 440000
P2O5
SO3
Cl
K2O



CaO
Ti
Cr
Mn
Fe2O3
Co
Ni
Cu
Depth (cm)
5 10 15 20 25
100000 180000
8500 10500
60 100 140
1200 1600
80000 95000
40 80 120
80 110 140
75 85 95
Tracer



Zn
Ga
Br
Rb
Sr
Zr
Nb
Ba
Depth (cm)
5 10 15 20 25
100 120
20 25 30
10 14 18
90 100 115
900 1200
650 750 850
140 180
400 800 1200
Tracer

Appendix 3: Tributary sources tracer concentrations in ppm in means and standard deviations and the sample size n

Sources	MeanP	SDP	MeanS	SDS	MeanTi	SDTi	MeanMn	SDMn	MeanFe	SDFe	MeanCo	SDCo
KL	950.2616	48.20195	326.3333	84.79141	8580.739	1220.911	1051.489	82.5998	49217.61	3679.799	94.91481	29.39974
RV	2053.058	258.8152	187.1743	16.75044	17823.24	2159.02	2159.326	170.9282	83718.41	2917.351	168.9583	35.11014

Sources	MeanNi	SDNi	MeanCu	SDCu	MeanZn	SDZn	MeanGa	SDGa	MeanSr	SDSr	MeanNb	SDNb
KL	92.2	13.10954	70.27222	83.57394	82.18333	7.642855	23.11667	5.156064	566.9944	19.58032	61.56481	5.145065
RV	116.7857	18.39945	87.95238	17.01872	155.919	6.807609	30.65952	3.143632	1131.869	52.35281	244.7286	15.60219

Sources	MeanBa	SDBa	MeanHf	SDHf	n
KL	900.1019	218.458	14.36111	8.813769	18
RV	1647.236	327.3033	23.2	7.059854	14

Appendix 4: Results of different land use classes for LOI, SOM and WSA

S/N	Place	Coordinates (Latitude, Longitude)		Soil texture	LOI	SOM	WSA
Forest land							
1	Msitu wa Mbogo	-3.52608	36.87787	Clay loam	2.01	2.9	5
2	Mbuguni B/Kubwa	-3.56427	36.9433	Loam	3.073	4.59	8
3	Mawalla TPC	-3.50929	37.43439	Clay loam	1.77	2.5	7
4	Kifaru H School	-3.52821	37.55433	Clay loam	8.077	10.66	8
5	Kochakindo Kahe M	-3.50574	37.52873	Clay loam	5.74	8.03	8
6	Kikuletwa Bridge	-3.54595	37.31344	Silty loam	3.26	3.85	8
7	Sakilla Meru 1	-3.33825	36.96451	Loam	4.2	4.7	8
8	Sakilla Meru 2	-3.33333	36.96379	Clay loam	3.43	4.01	8
9	TPC Msarakia	-3.50892	37.34286	Clay loam	3.31	6.88	8
10	Sakilla Meru 3	-3.33056	36.95861	Loam	4.02	5.45	8
11	Sakilla Meru 4	-3.33958	36.96833	Clay loam	2.57	4.72	6
12	Bwawani	-3.54067	36.85773	Clay loam	2.496	3.16	8
Average					**3.66**	**5.12**	**7.5**
Cultivated							
1	Soko Scheme	-3.47948	37.50149	Clay loam	1.44	2.39	8
2	Kikuletwa bridge	-3.55414	37.30683	Silty loam	1.46	2.43	7
3	Kituri Mwanga	-3.50029	37.55564	Sandy Clay loam	1.97	3.04	4
4	Longoi kwa sadala	-3.40601	37.26669	Clay	1.35	2.89	8
5	Machame Gabriella	-3.33576	37.23301	Clay	2.73	4.32	5
6	Kochakindo Kahe/Msh	-3.4979	37.53082	Loam	3.86	7.14	3
7	Mnadani Machame	-3.32695	37. 23079	Clay	2.49	2.86	1
8	Msitu wa Tembo	-3.57212	37.30344	Clay loam	2.88	4.82	8
9	Kituri Proper	-3.53708	37.53488	Sandy Clay loam	3.07	5.78	0
10	Hai Town	-3.3258	37.16489	Clay	2.96	4.15	1
11	Chekereni	-3.458273	37.541108	Silty loam	1.32	0.93	1
12	Kiomo Kahe/Msh	-3.48152	37.52188	Loam	1.33	1.32	1
Average					**2.24**	**3.51**	**3.92**
Grassland							
1	Kochakindo Kahe M	-3.50184	37.53162	Clay	2.88	4.51	8
2	Kituri Proper	-3.53377	37.53155	Loam	3.27	6.76	3
3	Tindigani Masaini	-3.43491	37.12334	Clay loam	4.07	8.20	5
4	Kiomu Majengo	-3.49196	37.53977	Loam	1.03	1.36	5
5	Chemchem	-3.58688	37.33495	Loam	1.25	2.79	8
6	Arusha Airport	-3.36487	36.61352	Loam	2.18	4.98	8
7	USA Leganga	-3.37275	36.84336	Clay loam	1.5	3.90	8
8	Mikocheni Kirungu	-3.59006	37.40077	Silty loam	1.54	6.17	5
9	Kahe Mashariki	-3.51569	37.51675	Loam	1.97	4.40	3
10	Ngaramtoni juu	-3.33712	36.62212	Loam	2.3	5.11	5
Average					**2.19**	**4.82**	**5.8**
Bush land							
1	Maweni Kikwe	-3.45577	36.83135	Clay	2.74	2.98	3
2	Kituri Proper	-3.5346	37.53488	Loamy	2.13	2.79	3

S/N	Place	Coordinates (Latitude, Longitude)		Soil texture	LOI	SOM	WSA
Forest land							
3	Karangai USA	-3.4814	36.86903	Sand Loam	2.32	2.51	5
4	Bwawani	-3.5486	36.85565	Clay loam	2.34	3.18	2
5	Soko village	-3.49737	37.48351	Loam	4.97	6.76	3
6	Mikocheni B	-3.59063	37.42077	Sandy Loam	2.32	3.49	3
7	Masama Rundugai	-3.42217	37.23557	Clay loam	2.79	3.15	3
8	Mawalla	-3.55303	37.42863	Clay loam	4.07	6.66	2
9	Msitu wa Tembo	-3.57212	37.30344	Clay loam	2.08	3.09	4
10	Chekereni Majengo	-3.47629	37.53977	Silty loam	1.89	2.22	1
Average					**2.86**	**3.7**	**2.9**
Bareland							
1	Kia Kaloleni 1	-3.4401	37.04131	Clay loam	3.27	4.63	3
2	Lengijave	-3.20137	36.62547	Loam	3.05	4.18	8
3	Kia njiapanda	-3.37544	37.0482	Clay loam	2.62	4.39	0
4	Leisinyai	-3.46666	37.05372	Loam	2.95	3.99	8
5	Mikocheni A	-3.59006	37.40077	Silty loam	2.22	2.59	3
6	Mererani	-3.45943	37.03773	Clay loam	3.82	4.61	5
7	Sanya Palestina	-3.339693	37.08968	Clay loam	5.11	7.24	6
8	Sanya roadtall	-3.35777	37.09375	Sandy Clay loam	6.7	6.83	6
9	Sanya Power station	-3.37396	37.06587	Sandy Clay loam	5.69	7.69	6
10	Kia kaloleni 2	-3.43999	37.04177	Clay loam	4.58	5.54	7
Average					**4.00**	**5.2**	**5.2**

REASEARCH OUTPUTS

(i) Publications

Aloyce I. M. A., Maarten, W., William, B. & Kelvin, M. (2021). Drivers, Impacts and Mitigation of Increased Sedimentation in the Hydropower Reservoirs of East Africa. *Land*, *10*(6), 638. https://doi.org/10.3390/land10060638

Aloyce, I. M. A., Maarten, W., Remigius, A. K., Shovi, S., Linus, M., William, H. B. & Kelvin M. M. (2021). Reconstructing the Changes in Sedimentation and Source Provenance in East African Hydropower Reservoirs: A Case Study of Nyumba ya Mungu in Tanzania. *Earth*, 2(3), 485-514. https://doi.org/10.3390/earth2030029

Aloyce I. M. A., Maarten, W., Remigius, A. K., Shomvi, F. S., William, H. B., Kelvin, M. M. (2021). Evaluating Soil Carbon as a Proxy for Erosion Risk in the Spatio-temporal Complex Hydropower Catchment in Upper Pangani, Northern Tanzania. *Earth*, 2(4) 764-780

(ii) Poster Presentation