Soil Aggregates Characteristics and Interrill Erosion in Some Weakly Weathered Coarse Textured Ecotopes in Eastern Cape Province, South Africa

By

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DECLARATION

I, Nebo, Godwin Iloabuchi, declare that the dissertation hereby submitted for the degree of Master of Science in Agriculture (Soil Science) at the University of Fort Hare is my work and has not been previously submitted to another University for any other degree.

Signature:…………………………………………………………….

Date:..

Place: University of Fort Hare, Alice Campus

PREFACE

This dissertation consists of six chapters. Chapter one gives the background and justification together with the objectives of the study. Chapter two reviews the literature relevant to the study. Chapter three describes the materials and methods used in the study. Chapter four presents the findings and Chapter five is the discussion. Chapter six is the conclusions and recommendations arising from this study.

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DEDICATION

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ABSTRACT

Aggregate stability and aggregate size distribution on soil surface that is impacted by rain drops affect soil erosion yet little is known about less weathered coarse textured soils. The objectives of the current study were to determine (i) the aggregate stability and associated aggregate fraction size distribution and (ii) the impact of the initial aggregate size on the aggregate stability and the resulting sediment fraction size distribution following rain drop impact in some quartz dominated coarse textured soils in the Eastern Cape Province. Soil samples for this experiment were collected from 14 ecotopes on the surface with a natural slope between 7.5 to 11% and at the depth between 0 to 0.2 m in the Eastern Cape Province. In each ecotope, twenty-five different spots were sampled using a spade at depth 0 to 0.2 m in other to eradicate biasness and ensure homogeneity. Thereafter, the soil samples were mixed to make a composite sample. The composited soil samples were then placed in rigid containers and taken to the soil science laboratory of the University of Fort Hare, Alice Campus where analyses were carried out. The soil properties were determined by passing the < 5 mm soil sample through a 2 mm sieve. The total Na, Ca and Mg contents in the soil samples were also determined using the wet digestion with sulphuric acid method. The total Soil organic matter content (SOM) was determined by the process known as weight loss on ignition. Thereafter, the fraction size distribution and aggregate stability was done by passing < 5 mm soil samples through a 3 mm sieve. The obtained calibrated aggregates between 3 and 5 mm were oven dried at 40^{o} C. Thereafter, five gram (5g) of oven dried calibrated aggregates was immersed in a 50 mL deionized water in a 250 mL beaker for 10 minutes. The soil material left was transferred to a 0.053 mm sieve already immersed in ethanol and moved five times in the ethanol to separate < 0.053 mm from > 0.053 mm fragments. The remaining > 0.053 mm was re-immersed in ethanol and further oven dried at 40^{o} C for 5

minutes. Thereafter, the > 0.053 mm fraction was transferred from 0.053 mm sieve, oven dried at 40^{o} C, dry sieved using Digital Electromagnetic Shaker on a six column of sieves: 2 mm, 1 mm, 0.5 mm, 0.25 mm, 0.106 mm, and 0.053 mm. The aggregate stability was determined using the resulting size distribution in seven classes by calculating the mean weight diameter (MWD, mm). The soils were very stable, moderately stable or unstable. The presence of smectite and cultivation as opposed to pasture lowered aggregate stability. The studied soils showed three different aggregate size distributions. Unstable soils were dominated by 0.106 – 0.25 mm aggregate size and showed a positively skewed aggregate fraction size distribution. Aggregates finer than 0.106 mm were limited because of the coarse nature of the soil texture. Moderately stable soils broke down to both micro aggregates, 0.106 – 0.25 mm and macro aggregates, 2 – 5 mm giving a bimodal distribution. The aggregate size distribution in the very stable soils was dominated by the aggregate fraction size 2 – 5 mm and a negatively skewed aggregate fraction size distribution. The smaller the initial aggregate size the higher was the aggregate stability but the reverse was true for splash erosion. It was thought that the short 5 minutes duration of the rainfall might not have been enough to cause a total breakdown of the aggregates. Alternatively, ecotopes that were dominated by primary soil minerals such as quartz showed different breakdown behaviour compared to those containing secondary minerals such as kaolinite or smectite.

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CHAPTER ONE

1. INTRODUCTION

Soil erosion accounts for almost the entire land degradation (Hudson 1992) and it is the second biggest environmental problem facing the world after population growth (Pimentel 2006, Wuddivira *et al.* 2009). Unfortunately, in most natural ecosystems soil erosion proceeds very slowly, usually unnoticed that only the cumulative impact becomes visible. For example, a soil erosion rate of just 2 t/ha/yr in a flat area under normal vegetation is equivalent to a loss of about 1 ha of land with a soil depth of 0.15 m in 100 years (Pimentel 2006). Moreover, early civilizations thrived on alluvial plains exploiting the fertile sediment deposits for crop and pasture production (Jones 1952). Therefore, soil erosion in the upper ridges of the world's basins was positively viewed for a long time. Today, most of the agricultural production activity is concentrated in formerly uninhabited eroded areas and the cumulative loss of the fertile top soil affects the livelihood of about 2.6 billion people worldwide (UNDP/GEF 2004; Fleitmann *et al.* 2007). Eighty five per cent of South Africa is threatened by soil erosion (van Rensburg 2008) and each year several arable lands are rendered unusable due to erosion (Rienks *et al*. 2000). Besides inherent nature of soils, the semi-arid climate in most of South Africa exacerbates soil erosion due to scarcity of vegetation to protect the soil surface from the impact of the rain (van Rensburg 2008). Meanwhile, soil erosion research in South Africa has concentrated on testing and developing technologies for monitoring at various scales (Le Roux *et al*. 2007) and less on understanding the mechanisms of soil erosion. Nonetheless, the problem continues to grow unabated (Laker 2004, van Rensburg 2008).

Soil erosion is a two stage processes namely: (i) aggregate breakdown due to raindrop impact and (ii) transport by runoff of the resulting soil particles and micro-aggregates on the impacted soil surface (Poesen and Savat 1981). The effects of the impact of rain drop on the aggregates results in the crumbling of the aggregates depending on their resilience. In turn, transportation of the detached soil material ensues (Nimmo 2004). Therefore, aggregate stability affects the susceptibility of a soil to erosion (Zhang *et al.* 2007). Soil aggregates are essentially primary particles held tightly to each other than the surrounding ones (Kemper and Rosenau 1986). Aggregate stability has been variously used to define or explain the ability of a soil to withstand the destructive forces of raindrop impact (Six *et al.* 2000). Therefore, aggregate stability plays a central role in ecosystems' physical condition (Bronick and Lal 2005). Certain soil physical and chemical properties have been used to distinguish aggregate stability in various soils (Kemper and Rosenau 1986, Six *et al.* 2000, Wakindiki and Yegon 2011). For example, Wakindiki and Yegon (2011) reported that there is a relationship between soil texture and aggregate stability.

Soil texture, clay mineralogy, organic matter content and sesquioxides profoundly influence stability of aggregates (Le Bissonnais 1996). Soil organic matter increases the cohesion between aggregate particles (Chenu *et al.* 2000). Le Bissonnais and Arrouays (1997) found a positive correlation (r=0.93) between soil organic matter content and soil aggregate stability. Wakindiki and Ben-Hur (2002) noted that the rate of water flow in the soil was dependent on the quantum of aggregation. A noticeable decrease in soil organic matter content of a soil also results in a decrease in aggregate stability of such a soil, generating crusting and runoff. Organic matter and its humic fractions influence aggregation in two ways: (i) increasing soil hydrophobicity (Zaher *et al.* 2005) and (ii) decreasing aggregate breakdown by slaking (Chenu et al. 2000, Zaher *et al.* 2005). Hydrophobicity in soil is attributed to various humic composition; while aggregate slaking results from the alteration of the pore size distribution with a slight change from the micropores (5 to 30 µm), and mesopores (30 to 75 µm) to ultra micropores (0.1 to 5 µm) (Dal Ferro *et al.* 2012).

Soil minerals form an interface with soil particles to form aggregates because of their cementing patents (Ohashi and Nakazawa 1996, Wuddivira *et al.* 2009, Horpibulsuk *et al.* 2010) or disperse the soil depending on the type when exposed to raindrop impact (Wakindiki and Ben-Hur 2002). The result of the dispersion produces a seal which impedes infiltration rates and consequent runoff (Lado and Ben-Hur 2004). Soil texture and mineralogy are products of weathering (Brady and Weil 2008). Hitherto research on the effects of soil texture and mineralogy on erosion and aggregate stability in weakly weathering soils is scant. Most less weathered soils such as arenosols are shallow (Ande 2011). Rhoton and Lindbo (1997) found that soil erosion is prevalent in shallow soils because of their limited water storage capacity. In spite of the huge investments and knowhow deployed in South Africa especially in the Eastern Cape province, previous erosion control measures in particular terracing has failed to stop soil erosion (Lasanta *et al.* 2001, Van Dijk and Bruijnzeel 2003, Gardner and Gerrard 2003, Machado *et al.* 2010). It appears that little is known about the inherent soil physical behaviour. Weakly weathered soils are made of composite primary minerals with fine sand and coarse silt as the predominant grains (Bryan 1970). Nevertheless soil research involving texture has predominantly focused on clay particles (Wakindiki and Ben-Hur 2002, Blanco-Canqi and Lal 2004). Laffan and McIntosh (2005) reported that less weathered doleritic soils comprise mainly plagioclase, feldspars and pyroxenes with minor minerals such as quartz, orthoclase, iron oxides, pyrites, hornblende and biotite. Nciizah and Wakindiki (2012) determined the soil mineralogy in 14 ecotopes in the Eastern Cape Province and found them to be predominantly dominated by quartz. Dixon *et al.* (2009) found a significant correlation between soil erosion and primary minerals (saprolites) degree of weathering. Weakly weathered saprolites were fast eroding.

Since climate drives erosion and weathering processes, it is important to incorporate the ecotope concept in soil erosion studies. Van Averbeke and Marais (1991) evaluated crop production under rainfed agriculture in Eastern Cape Province and found it to be ecotope specific. An ecotope was defined by Mac Vicar et al. (1974) as a unit of agricultural land that is relatively homogeneous in terms of climate, soils and topography. Bastian *et al*. (2003) submitted that ecotopes were landscapes uniformly alike, unambiguously distinct units that are vital for classifying, mapping, delineating and measuring landscapes into economically discrete purpose and transformation. In other words, ecotopes differ from one another and are suitable entities for agro-technology transfer (van Averbeke and Marais 1991). Ecotopes were initially delineated on the basis of soil fertility (P-fixation and N-leaching) (MacVicar et al. 1974) and later Marais (1978) added water sufficiency as a necessary criterion. Soil erosion was not included in the ecotope concept although it affects both nutrient capacity and water regime in soils.

1.2 Objectives

The objectives of the current study were to;

(1) Determine the aggregate stability and associated aggregate fraction size distribution in some quartz dominated coarse textured soils in the Eastern Cape Province.

(2) Determine the impact of the initial aggregate size on the aggregate stability and the resulting sediment fraction size distribution following rain drop impact in some quartz dominated coarse textured soils in the Eastern Cape Province.

CHAPTER TWO

2. LITERATURE REVIEW

2.1 Aggregate Stability and Fraction Size Distribution

Soil aggregate stability studies were first done in 1930s by Yoder and Henin (Le Bissonnais 1996) but a lot remains unclear in terms of understanding the mechanisms involved. Aggregate stability is the structural resistance of a soil to the destructive powers of raindrop impact, disintegration, dispersion and slaking (Le Bissonnais 1996, Six *et al.* 2000, Nimmo and Perkins 2002). Aggregate breakdown has been found to be the dominant factor that controls soil crusting and erosion rate. Therefore, unstable soil evolves due to the breakdown of aggregates (Sophie and Le Bissonnais, 2004). Aggregate stability influences many physical properties of the soil (Cho *et al.* 2006) including its ability to detach and be transported. Detached soil particles are not uniform but vary in size. In other words, soils differ in their aggregate stability. In a given soil, aggregate stability and the fractional size distribution are related (Reichert *et al.* 2009, Teh 2012). Fraction size distribution differs according to the soil type, clay content and organic matter (Teh 2012). Six *et al.* (2000) credited the strong aggregation observed in some soils to the prevalence of aluminium and/or iron compounds (Al or Fe). The determination, evaluation and prediction of the impact of erosion can be ascertained by studying aggregate stability and fraction size distribution (Nimmo and Perkins 2002) because aggregate stability has wide effects on sundry physical behaviour of soils such as infiltration rate and runoff (Nimmo and Perkins 2002, Sophie and Le Bissonnais 2004).

Aggregate stability is commonly estimated using the mean weight diameter (MWD), which involves segregating the products of aggregate breakdown through sieves. According to Le Bissonnais (1996), MWD is calculated by multiplying the weight of aggregate fractions in each sieve by the mean aperture of that sieve. A soil with a MWD < 0.4 mm is termed very unstable, 0.4 to 0.8 mm is termed unstable, while the range 0.8 to 1.3 mm is regarded as medium stability. If MWD is 1.3 to 2.0 mm the soil is stable and soil with MWD > 2 mm is declared very stable. The first four aggregate stability classes have a higher capacity to form crusts, and hence are more erodible.

2.2 Interrill Erosion

Soil erosion is the disintegration and detachment of soil aggregates and primary particles, entrainment down the slope or the direction of the force of detachment and eventual deposition (Stern *et al.* 1991, Le Roux *et al.* 2008). Although wind is an agent of erosion (Le Roux *et al.* 2008), its impact is negligible (Garland *et al.* 2000) compared to water erosion in cultivated areas in South Africa. Soil erosion has been variously and inexplicably viewed as one of the significant land degradation despoliation of South African milieu. Hence, the focus of this research study (interrill erosion) was to uncover the incipient nature and the detrimental effect of water erosion in our environment and the ecosystem. The two major forms of soil erosion in cultivated areas include rill and interrill (Salles and Poesen 2000). The extreme form of rill erosion is called gully or locally christened "donga" erosion in South Africa (Hoffman and Ashwell 2001). Gully erosion incises the soil while on course. Gully erosion studies have dominated international space and recognition over interrill erosion because of its obvious physical manifestation (Machado *et al.* 2010). Like other forms of erosion, it begins with the detachment of soil aggregates and particles as runoff flows on the soil surface. Eventually, runoff gathers momentum and concentrates its flow causing a massive flood with an attendant force of flow which abrades the soil leading to gullies.

Interrill erosion otherwise referred to as sheet or splash erosion is that form of erosion that involves detachment and transportation of soil particles through the impact of raindrops without the formation of permanent rills (Salles and Poesen 2000). The crushing impact of raindrops and the consequent detachment results in surface sealing, which is the prelude to soil erosion and runoff due to a decreased infiltration rate. The detached particles often in suspension are entrained, homogeneously resulting in soil surface loss or removal. Although interrill erosion is not dramatic like rill erosion, its impacts is formidable (Gumiere *et al.* 2009). They found out that interrill erosion is the major cause of sediment deposition in the ecosystem especially in the lower slope of cultivated fields. The insidious nature of interrill erosion, because of its seemly unseen threat, has undermined its potential as a veritable harbinger for land degradation (Salles and Poesen 2000). Splash erosion on sloping soils has a tremendous net transfer of sediment particles (van Dijk *et al.* 2002). Previous studies have shown that interrill erosion contributes significantly to the land degradation as well as sediments deposition which are a major source of pollution and threats to human and aquatic life (Sharifah-Mastura *et al.* 2006, Parlak and Özaslan-Parlak 2010). Many studies have been done to describe erosion process (Stroosnijder 2005, Kinnell 2005) but few deal with the mechanisms of soil erosion. A few workers have shown the importance of detachment stage in the erosion process and identified soil organic matter content, texture and soil type to have a profound influence (Parlak and ÖzaslanParlak 2010). Furthermore, the proneness of soils to surface sealing and crusting is dependent on a combination of soil physical and chemical properties, mainly soil mineralogy (Stern *et al.* 1991), clay content (Le Bissonnais 1996, Wakindiki and Ben-Hur 2002) and organic matter (Le Bissonnais and Arrouays 1997).

2.2.1 Organic Matter Content

Soil organic matter is an active property of the soil (Blanco-Canqui *et al*. 2004). Incorporation and presence of soil organic matter or place in a soil cannot be overstated because of its enduring positive impact not only on the crops and tree plants but on the quality of the soil (Carter 2002). Soil organic matter is one of the four major components of the soil (other components are mineral matter, air and water) (Brady and Weil 2008). Tisdall and Oades (1982) classified organic binding agents into three categories: (i) transient, which is chiefly polysaccharides, (ii) temporary, comprising of roots and fungal hyphae and (iii) persistence, consisting of polyvalent metal cations, resistant aromatic components and highly sorbed polymers. Organic matter is reputed to play a vital role in combating the excessive force of raindrops that leads to detachability of particles through its binding activity with the soil particles forming stable aggregates (Chenu *et al.* 2000). Organic matter inhibits compaction by raindrop impact thereby enhancing the infiltration rate which reduces the rate of runoff and crusting (Le Bissonnais 1996).

Le Bissonnais and Arrouays (1997) commented on the invincibility of organic matter to influence aggregate breakdown under rainfall conditions. Danga *et al.* (2010), Sultani *et al.* (2007), Wakindiki and Yegon (2011) acknowledged that organic matter plays an essential role in soil aggregation and surface flow. Furthermore, Fuentes *et al.* (2009) found that soil organic matter had a major influence on both macro and micro aggregates in the soil. Chenu *et al.* (2000) reported that soil organic matter decreases the wettability of aggregates thereby increasing their cohesion through the cementing of mineral particles via organic polymers and/or by enmeshment of physical particles (De Gryze *et al.* 2006). Lado and Ben-Hur (2004) studied aggregate sizes of <2 mm and 2 to 4 mm without rainfall impact and found out that the saturated hydraulic conductivity of soil with higher organic matter content (3.5%) was greater than that of lower organic matter content (2.5%) soil.

Le Bissonnais and Arrouays (1997) found that measurement of soil organic matter level of soils offers an insight in the determination of the structural degradation risk and threat by degradation forces as erosion. Water movement, therefore, is resisted by stable soil aggregates greater than the primary particles of silt. Furthermore, soil pores originating from aggregation of soil particles promotes infiltration rate thereby decreasing runoff and transport of soil particles by erosion.

2.2.2 Soil Texture

Soil texture is an indispensable factor that immensely contributes to erodibility of soils (Bryan 2000) because of its great influence on several other properties (Brady and Weil 2008). This distinguishing characteristic of soil texture has placed a demand to understand its critical role on soil erosion (Brady and Weil 2008). Soils which are high in silt and very fine sand are highly erodible. Sandy soils are highly susceptible to detachment but are unfavourable to entrainment by runoff due to their lack of cohesion. Soils with high clay content are difficult to detach but easy to transport (Neyshabouri *et al.* 2011).

Duplex soils with weak structural and textural development are easily eroded (Igwe and Ejiofor 2005, Pimentel 2006). Therefore, soil texture influences aggregate stability and consequently infiltration rate and soil loss during rainstorms (Wakindiki and Ben-Hur 2002, Ben-Hur and Wakindiki 2004). Sandy loam, sandy clay and loamy clay soils exhibit and behave differently upon wetting. Coarse textured soils have higher infiltration rates than their medium to fine texture counterpart (Gregory et al. 2005). The infiltration rate characteristics exhibited by sandy loam soils is courtesy of their coarse texture and large pores unlike the medium to fine textured sandy clay and loamy clay soils with restricted pore spaces. Soil Survey Staff (1996) reported that the amount of available water in sand is usually low, more in clay and maximum in loamy and silty soils. Conversely, clay soils have poor drainability compared to sand. Soil texture which is the building block for aggregate formation, therefore, plays a vital role in the resistance to erosion and stability of a soil (Bryan 2000). Although the resistance of soils to interrill erosion are still sketchy a compelling need to understanding soil parameters responsible for resistance to interrill erosion (Bryan 2000). While Gumiere *et al.* (2009) reported that soil texture is critical in interrill erosion and crusting (Kosmas *et al.* 1999). Soils formed on marl deposits with high silt contents are found to be susceptible to crust formation with very high surface runoff and sediment loss (Kosmas *et al.* 1999). Krull *et al.* (2001) found that the spatial arrangement of soil texture influenced the biological stability of organic matter. Therefore, soil texture is the physical protective armour of soil organic matter (Krull *et al.* 2001).

2.2.3 Soil Mineralogy

Soil mineralogy is an essential parameter with profound effect in unmasking the dynamics of the soil when exposed to the forces of rainfall (Le Bissonnais 1996, Wakindiki and Ben-Hur 2002, Ben-Hur and Wakindiki 2004). Clay mineralogy is reputed for its ability to influence binding in aggregates and their resistant to drop impact (Ben-Hur and Wakindiki 2004) and hence disintegration, breakdown and dispersion of aggregates (Stern *et al.* 1991). Soil aggregate breakdown leads to surface sealing and crusting of soils (Wakindiki and Ben-Hur 2002, Ben-Hur and Wakindiki 2004).

Clay minerals play a unique role in the understanding of soil forming processes and weathering (Johnson 1970). This unique role and interplay between clay mineral and soil forming processes explains the reasons why various occurrences and behaviour in the soil complex like clay dispersion, seal formation, runoff and soil loss are influenced by soil mineralogy (Lado and Ben-Hur 2004). Usually, clay minerals are viewed as that part of soil that evolved through the processes of weathering on the prevailing minerals and amorphous material (Johnson 1970) thus, making clay mineralogy that principal factor which influences aggregate stability and interrill erosion (Ben-Hur and Wakindiki 2004). Six *et al.* (2002) found that soils rich in 1:1 clay minerals and oxides are very high in aggregate stability with a little correlation between aggregate stability and organic matter especially in the tropics. Commenting on the effect of clay minerals on aggregate stability, Denef *et al.* (2002) observed that soils with varied mineralogy exert macro aggregate stabilization on interaction with soil organic matter. Denef *et al.* (2002) also further observed that at low organic matter concentration, soil mineralogy of variable charge 1:1 clay minerals and oxides will form stable aggregates. On the other hand, Trakoonyingcharoen *et al.* (2012) found that there is neither relationship nor association between the content of clay, type of clay mineral with aggregate stability.

The surface area of clay minerals affects dispersibility, charge density, and cationic exchange capacity (Trakoonyingcharoen *et al.* 2012), thus enhancing the inter and intra aggregates pore spaces (Morgan 1979). Smectitic soils are very dispersive and unstable, which means that they are seal forming as well as seal prone. Wuddivira et al. (2006) found out that there is increased slaking and seal formation of smectitic mineral soils with <3% low organic matter content in the humid tropics. Wuddivira *et al.* (2006) also reported that smectitic mineralogy with high organic matter content can greatly enhance the soil stability as well as lower slaking and seal formation. However, kaolinitic soils are very stable, less dispersive and not susceptible to sealing. In their studies, Wakindiki and Ben-Hur (2002), Ben-Hur and Wakindiki (2004) found that unstable soils accounts for a higher soil loss under interrill condition compared to the stable soils. They submitted that smectitic rich mineral soils have greater soil loss than the kaolinite and/or illite unstable dominated clay soils. Norton *et al.* (2006) noted that aggregate stability increases in uncultivated soils with 2:1 clay, suggesting that soil use moderates the effect on soil minerals on erosion. Amezketa (1999) found that there are conflicting differences based on climatic zones and soils in the quantification of role of soil properties and breakdown mechanisms to soil aggregate stability.

2.3 Weakly Weathered Soils

Weakly weathered soils generally are composed of and/or formed from ancient Pleistocene, recent volcanoes and ash-fall layers. The ash-fall layers on further weathering forms plagioclase and green hornblende (Ruxton 1970). Therefore, the intensity of weathering determines how deep, well weathered or developed it becomes and vice versa. Most weakly weathered soils are shallow and are dominated by primary minerals especially quartz and feldspars (Laffan and McIntosh 2005). Although they are chemically poor, these soils are known to exhibit high capacity and potentials of boosting farming and agricultural activity when fertilized (Laffan and McIntosh 2005). Weakly weathered soils like Aridisols (e.g. Augrabies form) and the Inceptisols (e.g. Glenrosa form) are found in many regions and environments. Weakly weathered soils are prone to erosive forces. For example, the dry nature of Aridisols is an indication of the absence of moisture in the profile. Soil development is limited to the top horizon and the soil is generally low in organic matter (FAO 1983). On the other hand, inceptisols are mostly derived from alluvial and colluvial deposits. Though the organic matter content varies from low to high, inceptisols are good for agricultural purposes and development and have a fine texture (FAO 1983). The soils are generally thin and moderately fertile. However, their limited depth implies that the profile water storage capacity is low and there is high risk of surface runoff, degradation and soil erosion. It must be pointed out here that these soils are less investigated probably because they often occur in agriculturally marginal areas such as arid and semi-arid areas or the alluvial zones.

Erosion rates are naturally very high on steep slopes at elevated topography with shallow to moderately shallow soils. This however, has given credence to the mineralogical investigation of weakly weathered soils in order to ascertain their weathering pattern and processes and the underlying parent rock and/or material from which they are formed. Mineralogically, soil weathering is classified into three categories namely: (i) the young (ii) the intermediate and, (iii) the strong weathering (Fendorf 2004). The young weathered soils are the weakly weathered soils which may be composed of the fine grained mica or quartz, chlorite, and vermiculite minerals. These minerals are dominant in Entisols and the Inceptisols. The intermediate weathered soils are dominated by vermiculite, smectite, and kaolinite e.g. in Mollisols, Alfisols and the Ultisols. While the strong weathered soils are dominated by kaolinitic and hydrous oxides minerals and these are mostly Ultisols and Oxisols. Figure 1 is an illustration of the hierarchy of soil mineralogy reflecting its weathering processes.

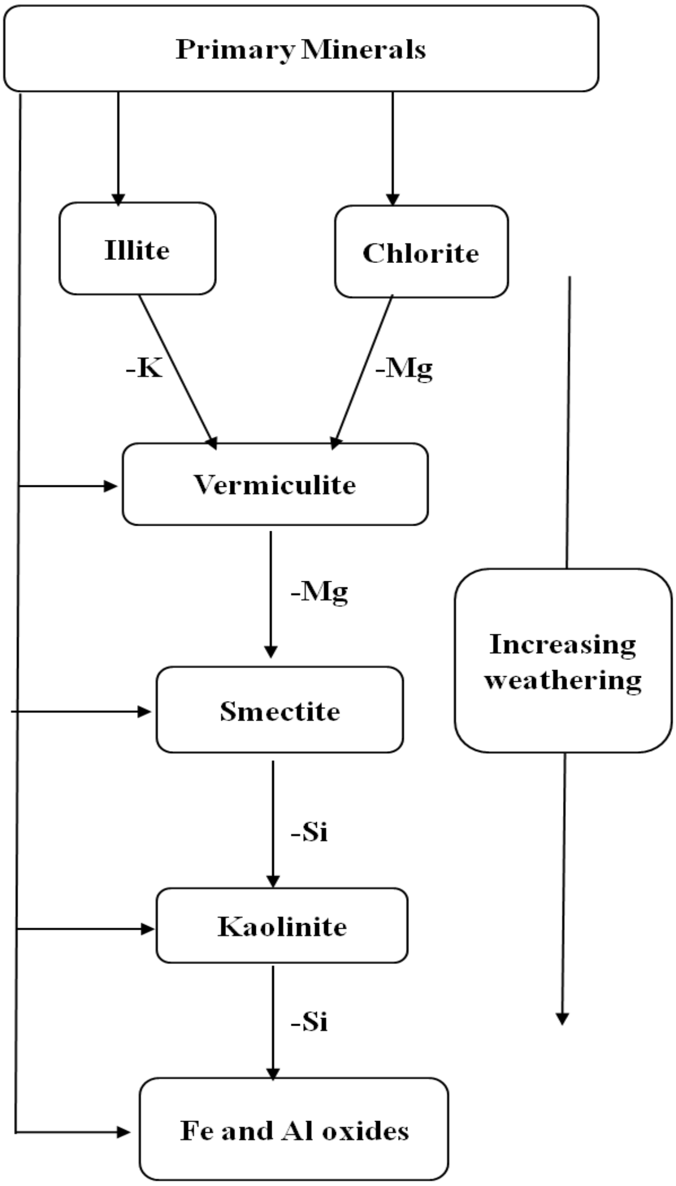

Figure 1. An illustration of the hierarchy of soil mineralogy reflecting its weathering processes (Fendorf, 2004).

Eighty per cent of South Africa's landmass is in the semi-arid and arid dry lands and only 18% is said to be dry sub humid to humid (van Rensburg 2008). Average rainfall decreases rapidly from the east (> 800 mm) to west (< 200 mm). Investigations also revealed that over 60% of South Africa receives less than 600 mm of rainfall per annum, while 20% of her land receives less than 200 mm per annum (ARC-ISCW 2004). Consequently, the ARC-ISCW

(2004) researchers found that over 30% of South Africa’s soils are young, weakly developed and shallow. These weakly weathered soils are overall very poor and low in organic matter content and also highly susceptible to erosion.

CHAPTER THREE

3. MATERIALS AND METHODS

3.1 Soil Sampling and Preparation

Soil samples were collected from 14 ecotopes (Figure 2) with natural slopes between 7.5 to 11% in Eastern Cape Province (van Averbeke and Marais 1991; Nciizah and Wakindiki, 2012). Twenty-five spots were sampled in each ecotope using a spade between 0 to 0.2 m depth, and mixed to make one composite soil sample. The soil samples were placed in rigid containers and transported to the soil science laboratory at the University of Fort Hare, Alice campus for analysis. Soil sampling collection was done within two weeks to minimize seasonal variation in soil properties (Mulla *et al.* 1992). Thereafter, the soil samples were spread on benches for 72 h to air-dry. Afterwards, the air-dried samples were gently crushed by hand to pass through a 5 mm sieve after removing stones and roots. Consequently the samples were stored in dry containers at room temperature for subsequent analysis. Three subsamples were prepared from each composite sample for investigation.

3.2 Determination of Soil Properties

A portion of the < 5 mm soil samples was passed through a 2 mm sieve, and the fraction < 2 mm was used to determine the soil properties as described by Rowell (1994). The total Na, Ca and Mg in the soil samples were determined using wet digestion with sulphuric acid method. Soil organic matter content was determined by weight loss on ignition procedure adapted by (Cambardella et al. 2001). The bulk density was determined using the clod method described by Blake (1965). The hydrometer method patterned after (Gee and Or 2002) was used to determine the soil texture after treating the soil samples with calgon solution to disperse the mineral particles and destroying organic matter with hydrogen peroxide. Soil mineralogy was determined by the Rietveld method for XRD quantitative analysis (Zabala *et al.* 2007) as described by Nciizah and Wakindiki (2012). Briefly, after milling, the samples were then prepared for XRD analysis using the back-loading preparation method. They were analysed with a P Analytical X'Pert Pro powder diffractometer with X'celerator detector and variable divergence and fixed receiving slits with Fe filtered Co-Kα radiation. The phases were identified using X'Pert Highscore Plus software.

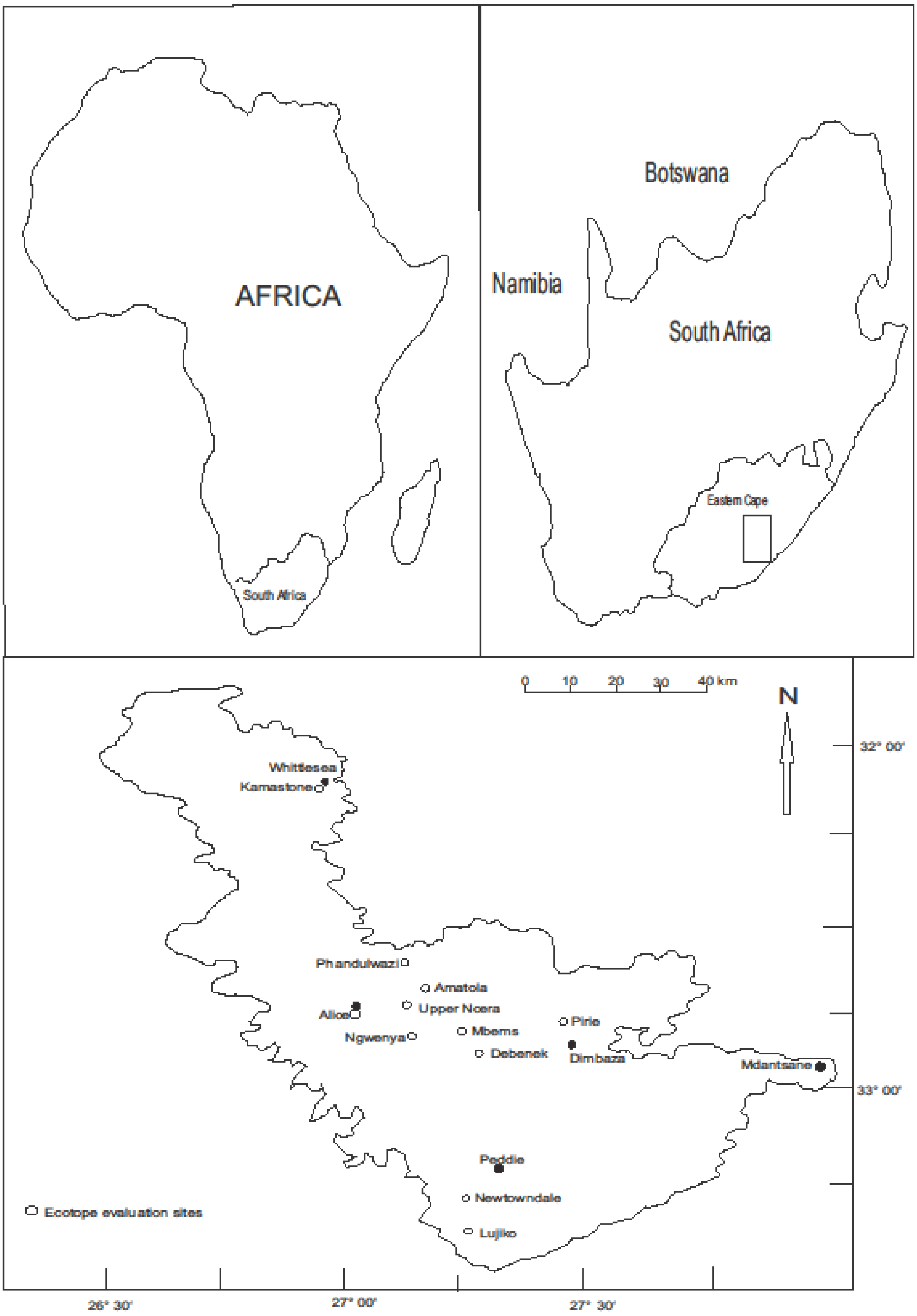

Figure 2. Soil sampling sites.

3.3 Fraction Size Distribution and Aggregate Stability

A portion of the < 5 mm soil samples was passed through a 3 mm sieve to obtain calibrated aggregates between 3 and 5 mm size. The calibrated aggregates were placed in an oven at 40°C for 24 h to equilibrate their matric potential and then subjected to the fast wetting method described by Le Bissonnais (1996). Five grams (5g) of the calibrated aggregates was carefully immersed in a 250 mL beaker filled with 50 mL of deionized water for 10 minutes. A pipette was then used to suck off the water while the slaked aggregates remained. The soil material that was left was then transferred to a 0.053 mm sieve already immersed in ethanol for the measurement of fragment size distribution. The 0.053 mm sieve initially immersed in ethanol, which also contained the soil material after the fast wetting treatment was meticulously and gently moved up and down five times in ethanol to separate fragments < 0.053 mm from those > 0.053 mm. The remaining > 0.053 mm fraction was thereafter re-immersed in a little quantity of ethanol before it was further oven-dried at 40° C for 5 minutes and dry-sieved in order to avoid re-cementing of fragments and particles during drying.

Afterwards, the > 0.053 mm fraction was transferred from the 0.053 mm sieve, oven-dried at 40° C, dry-sieved using a Digital Electromagnetic Sieve Shaker Model FTLVH-0200 on a column of six sieves: 2 mm, 1 mm, 0.5 mm, 0.25 mm, 0.106 mm, and 0.053 mm. This was done in one revolution per minute to avoid further aggregate breakdown. The aggregate stability was expressed using the resulting fragment size distribution in seven classes by calculating the mean weight diameter (MWD, mm) being the sum of the mass fraction of soil remaining on each sieve after sieving multiplied by the mean aperture of the adjacent mesh as shown in equation 1.

$$MWD = \sum_{i=1}^{7} \bar{x}_i w_i \quad (1)$$

Where: w_i = the weight fraction of aggregates in the size class i with a mean diameter $\bar{x}_i$ (Le Bissonnais 1996).

3.4 Rainfall Simulation

A splash plate that was used by Cheng *et al.* (2008) and adapted by Nciizah and Wakindiki (2012) was used in this study (Figure 3). The plate was made from a thin piece of iron. It had an outside diameter of 0.3 m with a height of 0.1 m, and an inside diameter of 0.1 m with a height of 0.03 m. In the middle of the plate, there was a perforated soil tray with 0.1 m diameter and 0.03 m height. In order to prevent soil loss from the holes, the trays were covered with a piece of gauze. Bulk soil samples with aggregates < 5 mm were packed in a perforated soil tray to a bulk density of 1.35 Mg/m^3 and kept at 9% slope. The soil samples were then subjected to simulated rainfall at 360 mm/h for 15 min. The high intensity rainfall was used to compensate for the short falling distance of 0.4 m of each simulated rain drop and the resulting low volume-specific kinetic energy of the applied shower (Martin *et al.* 2010). A mini rainfall simulator, LUW type (Eijkelkamp Agrisearch Equipment) drop forming was used to simulate rainfall. The rainfall simulator consists of an air-tight water reservoir fitted with a 49 variable intensity rainfall simulating nozzles perforated to the bottom with an inbuilt intensity and pressure regulator.

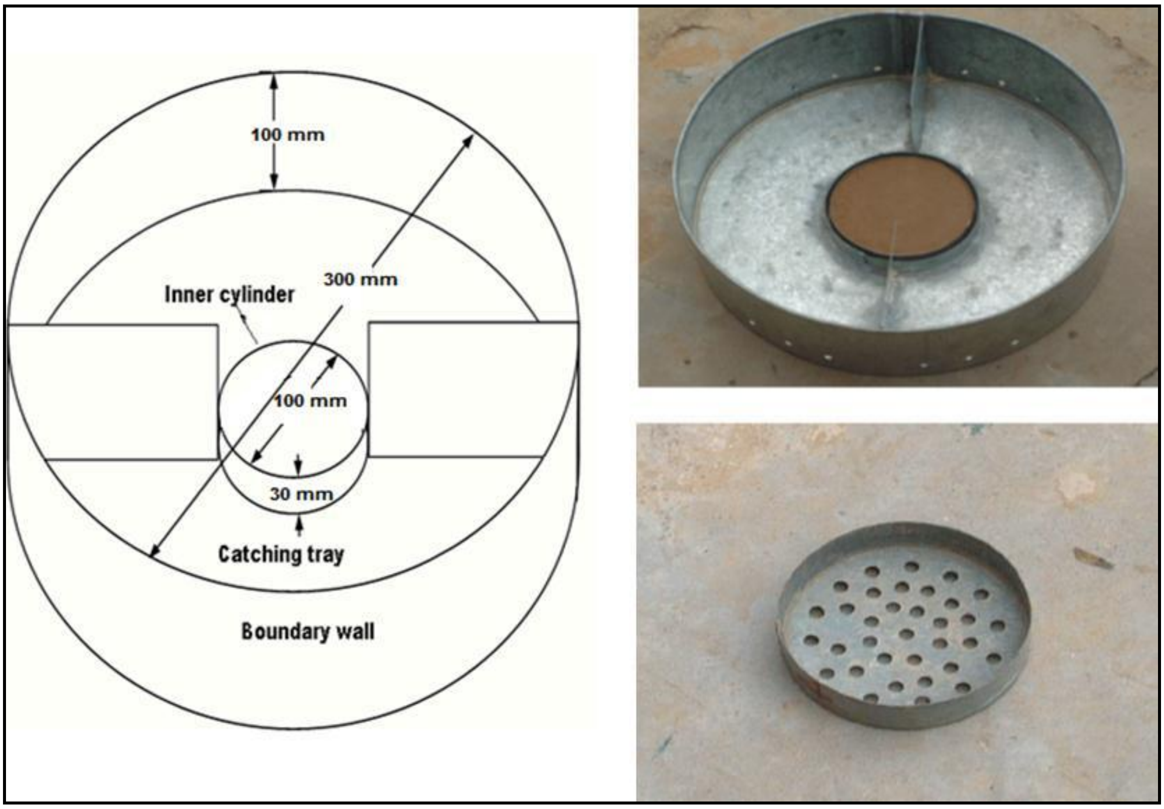

Figure 3. Splash plate apparatus.

3.5. Determination of Interrill Erosion

After the rain application, soil particles and aggregates, which were detached were brushed off and weighed. The average splashed amount (S, g/min.m^2), was then calculated using equation 2.

$$S = \frac{Dt_2 - Dt_1}{(t_2 - t_1)A} \quad (2)$$

Where Dt_1, Dt_2 denotes the total detachment after rainfall time of t_1, t_2, respectively (g); t_1, t_2 are the rainfall duration (min); A represents the area of splash plate (m^2).

3.6 Statistical Analysis

Data were subjected to analysis of variance for a completely randomized design with 14 ecotope treatments using JMP® Release 10 statistical package (JMP® 2012). Mean separations was achieved by using Fishers protected Least Significant Difference (LSD). A probability level of less than 0.05 ($P \leq 0.05$) was designated as significant (Steel and Torrie 1981).

CHAPTER FOUR

4. RESULTS

4.1 Soil Properties

Table 1 shows the soil physical properties while Table 2 shows the chemical properties.

4.1.1 Texture

In general the soils were either sandy loams or sandy clay loams (Table 1). Seven of the 14 ecotopes had a sandy clay loam while six were sandy loams. However, silt was least in Kamastone site with a value of 90 g kg^{-1} of soil. Sand was the dominant particle in all the soils.

Table 1. Selected soil physical properties in the 14 ecotopes in Eastern Cape.

Ecotope	Management	Texture (g kg^{-1})			Textural class	Organic matter (g kg^{-1})	Density (Mg m^{-3})	
		Sand	Silt	Clay			Bulk	Particle
Alice Jozini	Cultivation	600	280	120	SL*	35.7	1.6	2.6
Amatola Jozini	Cultivation	470	150	370	SCL	66.1	1.4	2.7
Debenek	Cultivation	560	260	180	SL	24.0	1.8	2.1
Kamastone	Cultivation	720	90	190	SL	31.8	1.5	2.6
Lujiko Leeufontein	Cultivation	680	110	190	SL	38.2	1.4	2.5
Mamatha	Cultivation	610	210	180	SL	29.9	1.3	2.5
Mbems Koedosvlei	Pasture	560	230	210	SCL	34.3	1.4	2.5
Mbems Koedosvlei	Cultivation	560	220	220	SCL	42.7	1.3	2.5
Ncera Kinross	Cultivation	480	260	260	SCL	41.9	1.5	2.6
Newtondale	Cultivation	650	140	210	SCL	51.4	1.3	2.5
Ngwenya Jozini	Cultivation	720	100	180	SL	36.4	1.8	2.6
Ngwenya Swartland	Pasture	670	120	210	SCL	28.4	1.6	2.6
Phandulwazi Jozini	Pasture	580	210	210	SCL	24.7	1.6	2.7
Pirie Shorrocks	Cultivation	500	280	220	L	44.1	1.7	2.5

*SL= Sandy Loam, SCL = Sandy Clay Loam, L = Loam

4.1.2 Soil Organic Matter

Soil organic matter was highest (6.61%) in Amatola Jozini ecotope followed by Newtondale with 5.14%, Pirie Shorrocks and Ncera Kinross at 4.41% and 4.19% respectively. (Table 1) Ecotopes with least organic matter content were Debenek and Phandulwazi Jozini at 2.40 % and 2.47% respectively.

4.1.3 Soil Density

The highest bulk density was approximately 1.8 Mg m^{-3} in Debenek and Ngwenya Jozini ecotopes (Table 1). Newtondale and Mamatha ecotopes had a bulk density of 1.3 Mg m^{-3}, which was the lowest. Meanwhile, particle density was highest in Amatola Jozini at 2.74 Mg m^{-3}. The least particle density was in Debenek (2.13 Mg m^{-3}) and Alice Jozini (2.14 Mg m^{-3}) (Table 1).

4.1.4 Soil Mineralogy

Apart from Debenek and Kamastone ecotopes, primary minerals dominated the soils and quartz contributed over 61% of the soil minerals (Table 2). The soil solution was slightly acidic.

Table 2. Selected chemical and mineralogical properties in the 14 ecotopes in Eastern Cape.

	Na	Ca	Mg	EC		Soil Mineralogy (%)∗						
Ecotope		(mg kg^{-1})		(µSm^{-1})	pH	H#	K	Mi	Mu	P	Q	S
Alice Jozini	5.29	198.39	57.36	47.90	5.78	0.29	-	4.40	6.10	12.2	77.01	-
Amatola Jozini	2.30	414.92	108.77	28.47	5.80	1.91	32.4	4.36	2.74	9.29	28.88	14.7
Debenek	3.35	75.10	34.23	29.23	5.79	0.30	2.1	4.59	8.50	84.5	-	-
Kamastone	1.29	226.69	67.29	66.47	6.27	0.67	8.56	10.0	18.8	5.9	5.96	-
Lujiko Leeufontein	3.67	108.54	32.90	52.23	5.45	0.63	-	8.61	5.14	10.4	75.14	-
Mamatha	2.40	82.24	26.53	34.50	5.50	0.43	-	5.52	6.46	12.2	75.32	-
Mbems Koedosvlei	3.30	208.75	53.59	55.17	5.65	1.10	-	4.99	6.58	9.97	77.35	-
Mbems Koedosvlei	4.23	158.07	33.49	80.97	5.76	0.65	-	4.69	7.76	10.5.8	76.37	-
Ncera Kinross	2.66	101.75	31.82	61.50	5.08	1.12	9.3	4.48	3.12	8.23	61.90	9.9
Newtondale	8.45	256.76	69.75	40.34	6.25	0.76	-	10.5	7.83	8.11	72.74	-
Ngwenya Jozini	3.00	144.23	29.95	41.27	6.49	0.56	-	8.83	5.78	16.6	68.22	-
Ngwenya Swartland	3.66	125.82	32.02	53.57	5.53	0.66	-	7.50	6.51	17.2	68.11	-
Phandulwazi Jozini	2.86	55.50	20.59	37.80	5.49	0.58	-	0.98	3.95	7.64	86.85	-
Pirie Shorrocks	3.94	126.26	56.85	45.89	5.33	0.55	-	6.98	6.19	17.7	68.55	-

H# = Hematite, K = Kaolinite, Mi = Microline, Mu = Muscovite, P = Plagioclase, Q = Quartz, S = Smectite.

∗ Soil mineralogy was done by Nciizah and Wakindiki (2012).

4.2 Aggregate Stability

There were significant differences ($p < 0.05$) in aggregate stability among the fourteen ecotopes. The MWD seemed to be dependent on the ecotope. In general, the soils could be grouped into ten according to their MWD as shown in Table 3. Lujiko had the highest MWD (3.01 mm) while Alice Jozini had the lowest (0.49 mm).

Table 3.The mean weight diameter (MWD) values in the various ecotopes.

Ecotope	MWD, mm
Lujiko	3.01^{a}
Mbems Koedosvlei-Pasture	2.51^{b}
Phandulwazi Jozini	1.85^{c}
Debenek	1.74^{c}
Ngwenya Swartland	1.51^{d}
Mamatha	1.39^{de}
Newtondale	1.26^{e}
Kamastone	1.20^{e}
Mbems Koedosvlei	0.95^{f}
Pirie Shorrocks	0.70^{g}
Ncera Kinross	0.69^{gh}
Ngwenya Jozini	0.61^{gh}
Amatola Jozini	0.61^{gh}
Alice Jozini	0.49^{h}
p-value	<0.0001

Note: Different superscript letters following each value indicate significance difference between the ecotopes, $P \leq 0.05$.

4.3 Fraction Size Distribution

Figures 4a – 4c show the aggregate fraction size distribution in each ecotope. The soils were grouped into three according to the pattern of the aggregate fraction size distribution. In group one (Figure 4a) the aggregate size 0.106 – 0.25 mm was dominant, giving the distribution a sharp peak that was tapering at both ends. There were six ecotopes in this group; Alice Jozini, Ngwenya Jozini, Amatola Jozini, Pirie Shorrocks, Mbems Koedosvlei and Ncera Kinross.

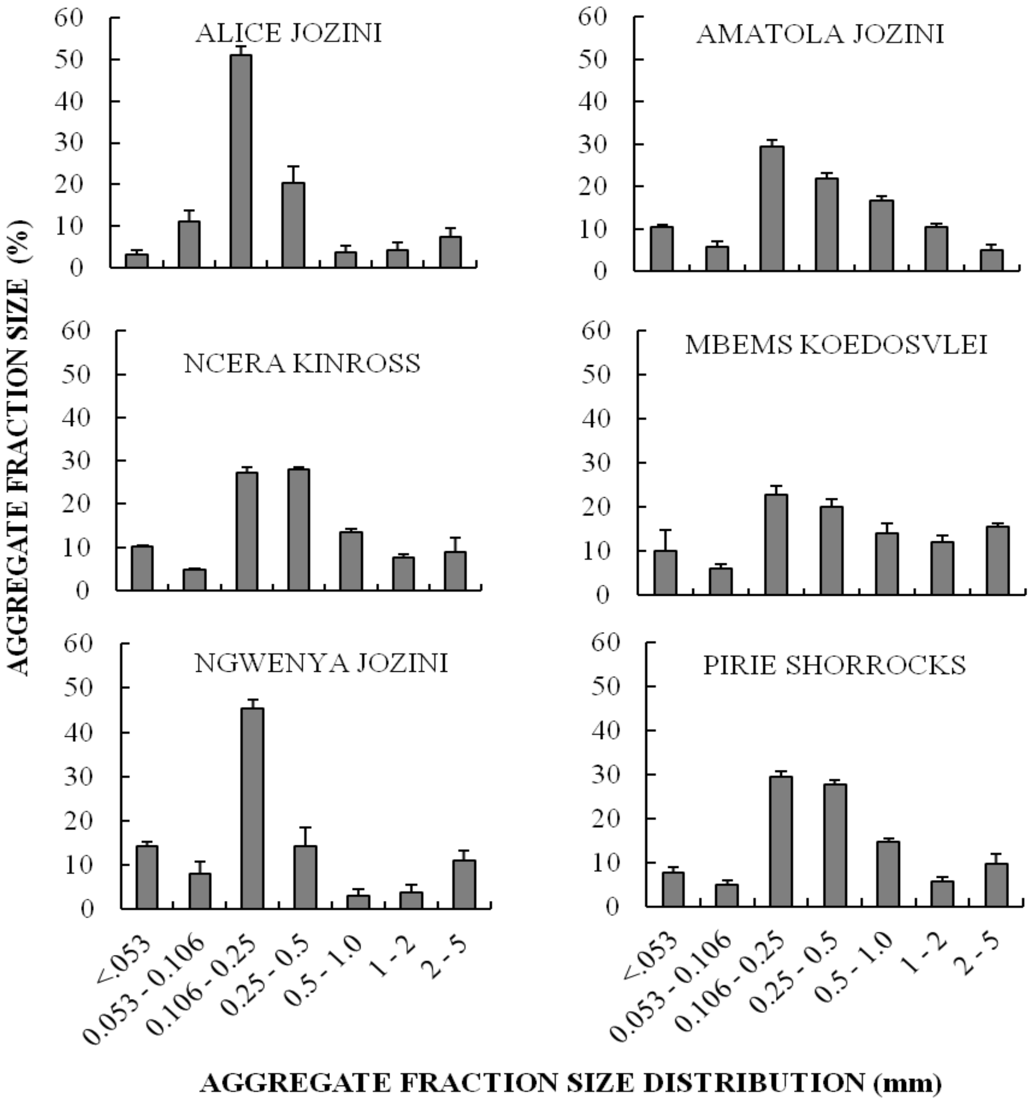

Figure 4a. Aggregate fraction size distribution in six ecotopes dominated by 0.106 – 0.25 mm aggregates.

Group two was made up of five ecotopes namely; Debenek, Lujiko, Mamatha, Mbems Koedosvlei (pasture) and Phandulwazi Jozini (Figure 4b). In group two, the dominant aggregate fraction size was 2 – 5 mm, which are the coarsest fraction and the distribution was a negative skew.

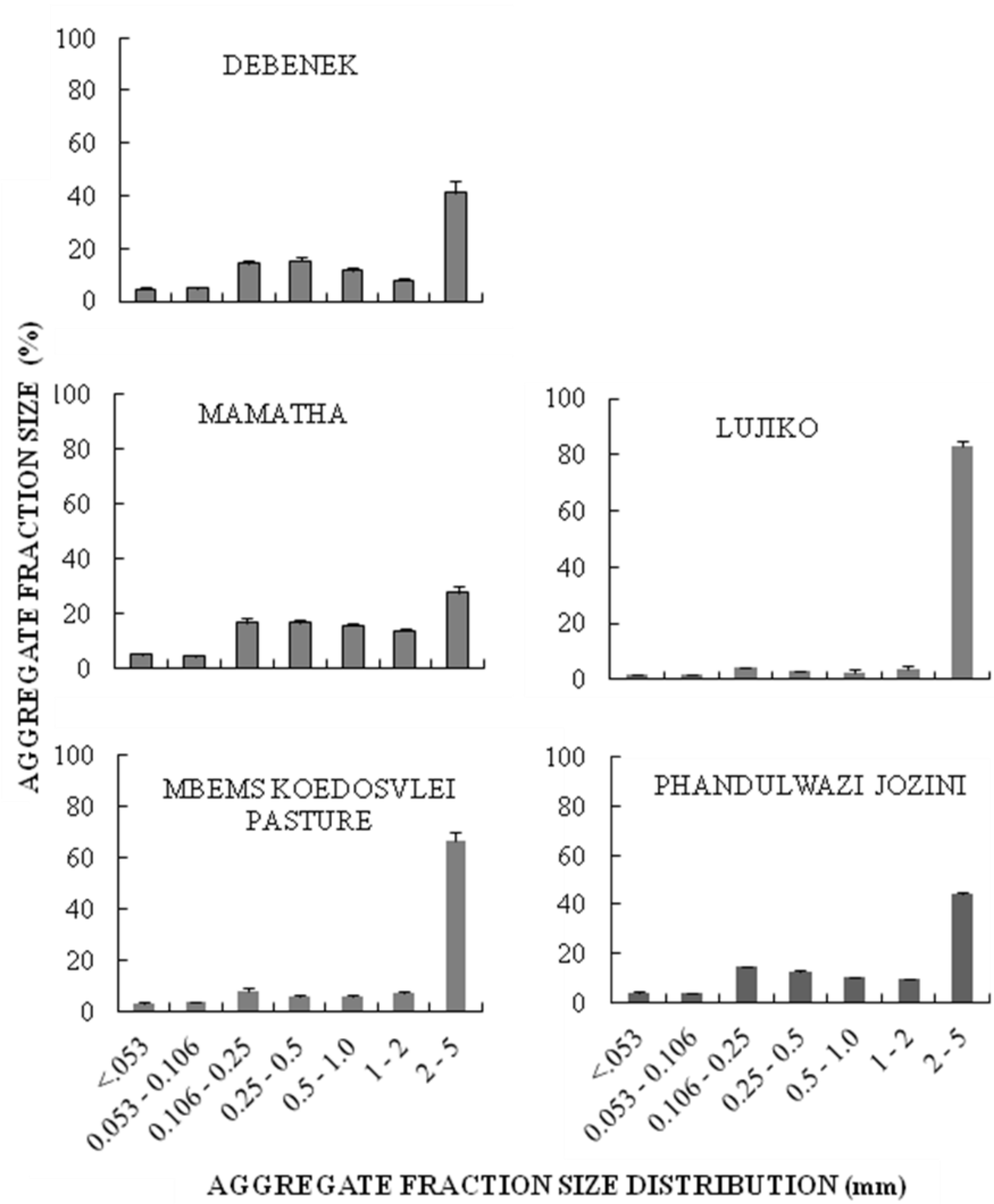

Figure 4b. Aggregate fraction size distribution in five ecotopes dominated by 2 – 5 mm aggregates.

Three ecotopes; Kamastone, Newtondale and Ngwenya Swartland had a bimodal distribution (Figure 4c). The aggregate sizes 0.106 – 0.25 mm and 2 – 5 mm were most abundant and both of them were the major peak.

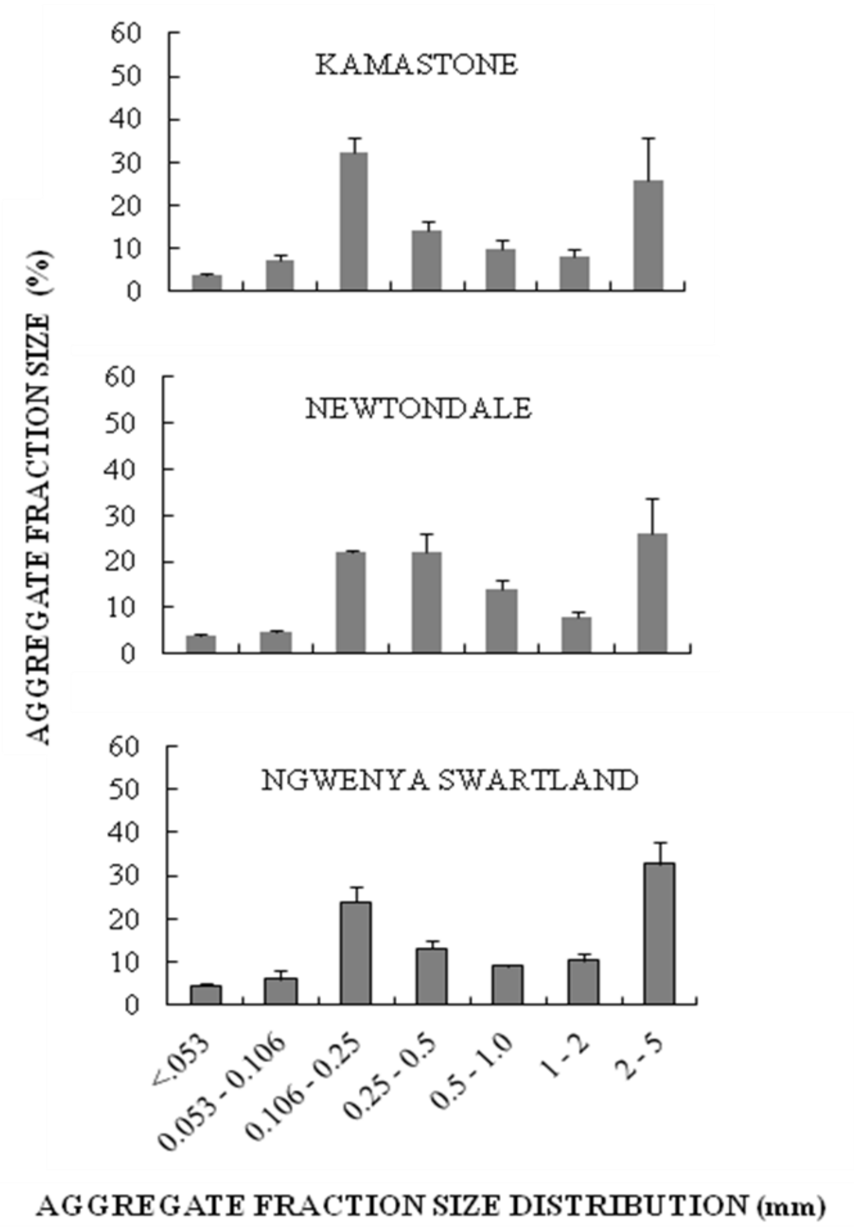

Figure 4c. Bimodal aggregate fraction size distribution in three ecotopes with dominant 0.106 – 0.25 mm and 2 – 5 mm aggregate sizes.

4.4 Interrill Erosion

4.4.1 Effect of Initial Aggregate Size on Splash Erosion in the Various Ecotopes.

Table 4 shows the results of the effect of the initial aggregate size on the amount of splash erosion that took place in the various ecotopes. There was significantly higher splash erosion in the less than 2 mm aggregate size compared to 2 - 3 mm and 3 – 5 mm. Splash erosion was similar in both 2 - 3 and 3 – 5 mm aggregate sizes.

Table 4. Effect of initial aggregate size on splash erosion in the various ecotopes.

Ecotope	Splash erosion (g m^{-2})
Amatola Jozini	93.88^{a}
Alice Jozini	55.69^{b}
Pirie Shorrocks	51.46^{bc}
Ngwenya Jozini	43.69^{bc}
Ngwenya Swartland	43.17^{bc}
Ncera Kinross	37.71^{bc}
Mbems Koedosvlei	35.26^{bc}
Debenek	31.36^{bc}
Kamastone	30.32^{bc}
Mamatha	25.38^{c}
Phandulwazi Jozini	24.76^{c}
Mbems Koedosvlei-Pasture	24.25^{c}
Lujiko	23.86^{c}
Newtondale	23.82^{c}
Aggregate size	
<2mm	93.60^{a}
2-3 mm	9.64^{b}
3-5 mm	13.47^{b}
ANOVA	
Ecotope	0.0006
Aggregate size	<.0001
Aggregate size × Ecotope	0.0027

Figure 5 shows the interaction effects between the initial aggregate size and ecotopes on splash erosion. There was a significant (P ≤ 0.05) interaction between the initial aggregate size and ecotope on splash erosion. The < 2 mm aggregates influenced splash erosion most irrespective of the ecotope. The interaction effect was most evident in Amatola Jozini probably because it had the highest smectite content (14.7%).

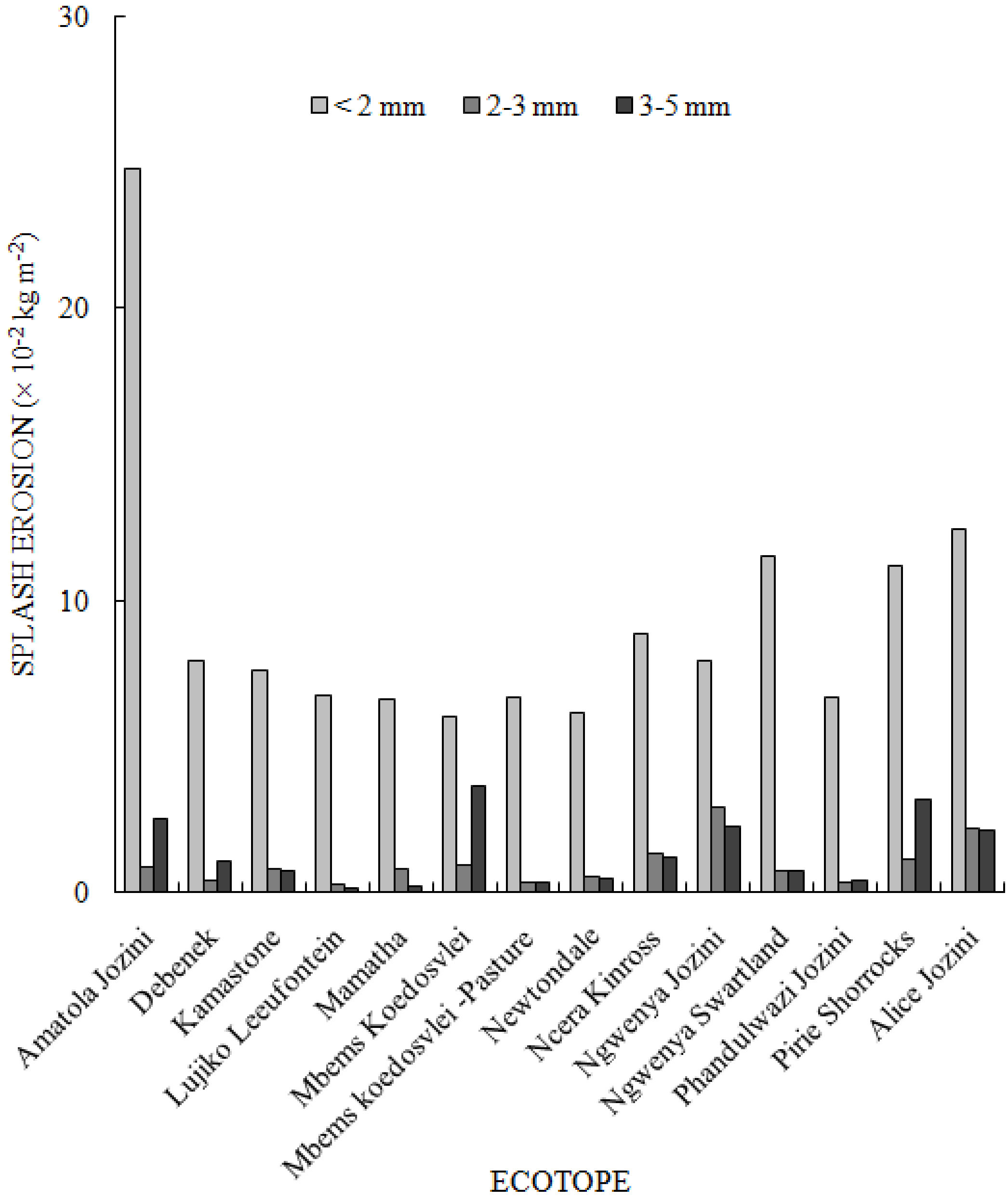

Figure 5. Interaction effect of ecotope and the initial aggregate size on splash erosion.

4.4.2 Effect of Initial Aggregate Size on Aggregate Stability in the Various Ecotopes.

The effect of the initial aggregate size on the aggregate stability in the various ecotopes is shown in Table 5. The smaller the initial aggregate size, the higher was the aggregate stability. Aggregate stability was significantly higher in the less than 2 mm and 2 – 3 mm aggregate size compared to 3 – 5 mm. Aggregate stability was similar in both less than 2 mm and 2 – 3 mm aggregate sizes.

Table 5. Effect of initial aggregate size on aggregate stability in the various ecotopes

Ecotope	MWD, mm
Debenek	0.66[a]
Amatola Jozini	0.49[ab]
Lujiko	0.49[ab]
Mbems Koedosvlei–Pasture	0.34[bc]
Pirie Shorrocks	0.30[bc]
Ncera Kinross	0.30[bc]
Mbems Koedosvlei	0.29[bc]
Kamastone	0.27[bc]
Mamatha	0.27[bc]
Alice Jozini	0.27[bc]
Ngwenya Jozini	0.20[c]
Ngwenya Swartland	0.20[c]
Newtondale	0.19[c]
Phandulwazi Jozini	0.12[c]
Aggregate size	
<2mm	0.38[a]
2-3 mm	0.35[a]
3-5 mm	0.21[b]
ANOVA	
Ecotope	0.0011
Aggregate size	0.0046
Aggregate size × Ecotope	0.0002

Figure 6 shows the interaction effects between the initial aggregate size and ecotopes on aggregate stability. There were significant ($P \leq 0.05$) interaction effect between the initial aggregate size and ecotope on the aggregate stability. Aggregate stability in ecotopes; Amatola Jozini, Debenek and Kamastone was mostly sensitive when using the < 2 mm size aggregates. These ecotopes had either relatively very low quartz or none, they contained kaolinite and Amatola Jozini had smectites (Table 2). Aggregate stability in Lujiko Leeufontein, Mamatha, Mbems Koedosvlei, Mbems Koedosvlei-pasture, Newtondale, Ncera Kinross,Ngwenya Jozini and Alice Jozini were next in sensitive when using the 2-3 mm size aggregates. These ecotopes are high in quartz and smectite free except for Ncera Kinross with a very low smectite (Table 2). For the 3-5 mm aggregates size, we found them to be least sensitive in Lujiko Leeufontein, Mamatha,and Mbems Koedosvlei-pasture.

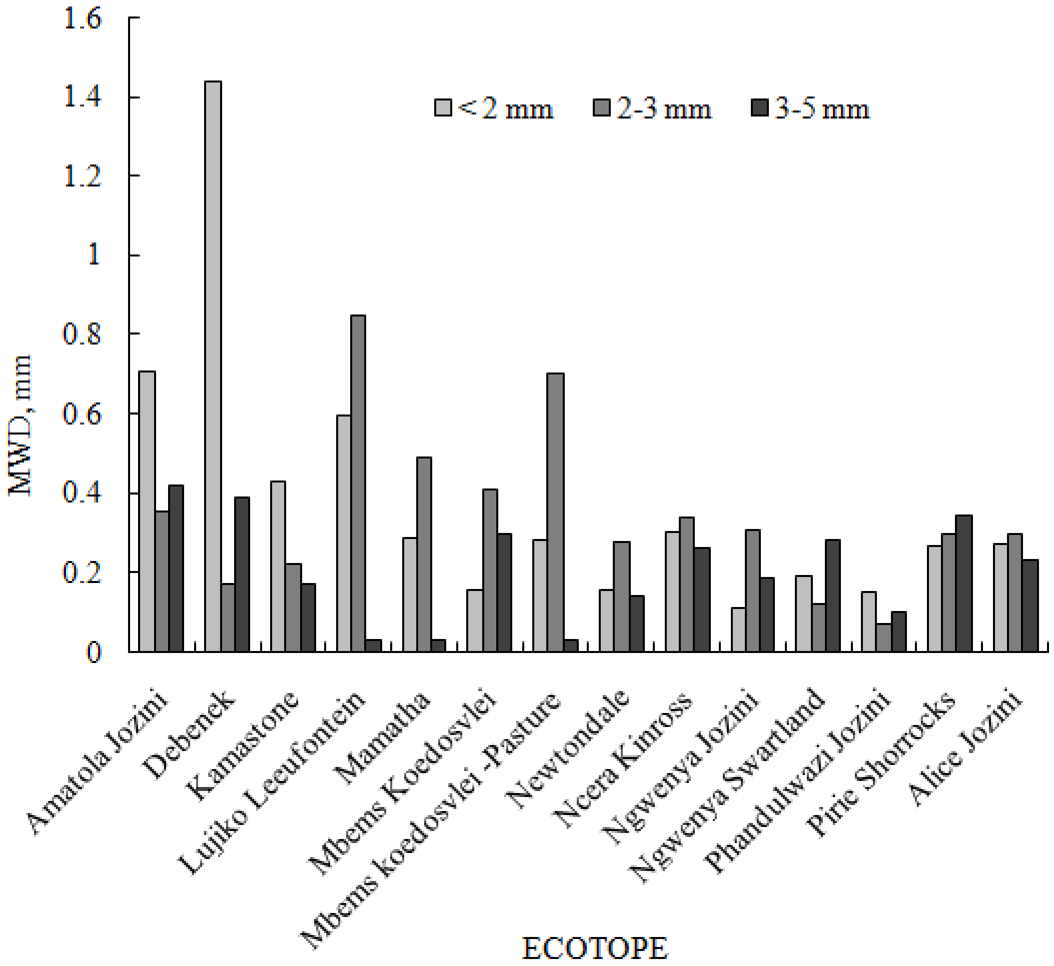

Figure 6. Interaction effects between the initial aggregate size and ecotopes on aggregate stability.

CHAPTER FIVE

5 DISCUSSION

5.1 Soil Properties

The studied ecotopes were coarse textured and organic matter content varied from low to high (Table 1). The difference between highest and the lowest organic matter content was approximately three fold. The bulk density was slightly above the 1.35 Mg m^{-3}average for most mineral soils. In general, the soils were less weathered because they were dominated by primary, resistant minerals especially quartz (Table 2). The high incidence of soil organic matter content in both Amatola Jozini and Newtondale ecotopes was due to a long period of fallow. These ecotopes were pasture lands left for a period of five years which gave rise to accumulated litters hence the difference.

5.2 Aggregate Stability

Aggregate stability increased with increase in the MWD (Le Bissonnais 1996). Therefore, the soils with higher MWD were more stable compared to those with lower MWD (Table 3). Lujiko and Mbems Koedosvlei-Pasture were very stable, Phandulwazi Jozini, Debenek, Ngwenya Swartland and Mamatha were stable while Newtondale, Kamastone and Mbems Koedosvlei were moderately stable. Pirie Shorrocks, Ncera Kinross, NgwenyaJozini, Amatola Jozini, and Alice Jozini were unstable. Unstable soils have a higher capacity to form crusts, and hence are more erodible compared to the stable soils (Le Bissonnais 1996). In general, it is accepted that organic matter increases aggregate stability in soils (Carter 2002, Chenu *et al*. 2000, Le Bissonnais and Arrouays 1997). However, in this experiment some ecotopes that had higher organic matter such as Amatola Jozini (6.61%) were less stable than those with three-times less organic matter such as Phandulwazi Jozini (2.47%). The presence of smectites in Amatola Jozini (Table 2) could have lowered its aggregate stability.

5.3 Fraction Size Distribution

Most of the ecotopes in group one (Figure 4a) that were dominated by 0.106 – 0.25 mm aggregate size were also found to be unstable (Table 3). The unstable ecotopes included Pirie Shorrocks, Ncera Kinross, Ngwenya Jozini, Amatola Jozini, and Alice Jozini. Therefore, mechanical aggregate breakdown of unstable soils resulted in micro aggregates, 0.106 – 0.25 mm. Aggregates finer than 0.106 mm were limited because of the coarse nature of the soil texture (Table 1). Lujiko and Mbems Koedosvlei–Pasture were very stable, Phandulwazi Jozini, Debenek, Ngwenya Swartland and Mamatha were stable while Mbems Koedosvlei was moderately stable. These very stable to moderately stable ecotopes were in group two (Figure 4b) and were dominated by aggregate fraction size 2 – 5 mm, which give them a distinctive negative skew. Ecotopes; Kamastone and Newtondale had a bimodal distribution (Figure 4c) where both 0.106 – 0.25 mm and 2 – 5 mm aggregate sizes were dominant. These ecotopes in group three were moderately stable according to Le Bissonnais (1996). Therefore, moderately stable soils breakdown to give rise to both macro aggregates and micro aggregates.

5.4 Interrill Erosion

There was significantly higher splash erosion in the less than 2 mm aggregate size compared to 2 - 3 mm and 3 – 5 mm. Splash erosion was similar in both 2 - 3 mm and 3 – 5 mm aggregate sizes. Tensile strength of soil aggregates is known to decrease with increase in the initial aggregate size (Abu-Hamdeh et al. 2006). Therefore, splash erosion was expected to increase with increase in the initial aggregate size. The results from this study show that there was more splash erosion in the < 2 mm aggregates as opposed to the 2 - 3 mm and 3 – 5 mm aggregates (Table 4). Furthermore, the results of this study indicated that the < 2 mm aggregates were more stable compared to the 2 - 3 mm and 3 – 5 mm aggregates (Table 5). The observed result could be explained as follows. The soil samples were subjected to simulated rainfall at 360 mm/h for 15 min. The duration of the rainfall might not have been enough to cause a total breakdown of the aggregates. Therefore, the lighter aggregates < 2 mm were splashed without much breakdown while the larger aggregates 2 – 3 mm and 3 – 5 mm remained intact. A second reason could have been due to soil mineralogy. Ecotopes such as Amatola Jozini, Debenek and Kamastone were most responsive to raindrop impact when using the < 2 mm size aggregates (Figure 6). These ecotopes had either relatively very low quartz or none, they contained kaolinite and Amatola Jozini had smectites (Table 2).

CHAPTER SIX

6 CONCLUSIONS AND RECOMMENDATIONS

1. It was observed that the quartz dominated coarse textured soils in most instances and at various ecotopes appeared very unstable, and at other ecotopes it was stable. The instability was most likely promoted by the presence of smectites and cultivation even when soil organic matter content was high.
2. Mechanical breakdown of unstable soils resulted in the formation of micro aggregates, 0.106 – 0.25 mm and a positively skewed distribution. The moderately stable soils broke down to micro aggregates, 0.106 – 0.25 mm, and macro aggregates, 2 – 5 mm giving a distinct bimodal distribution. The aggregate fraction size distribution in the very stable soils was dominated by macro aggregates, 2 – 5 mm and the distribution was negatively skewed.
3. Splash erosion was more in the less than 2 mm aggregate size compared to 2 - 3 mm and 3 – 5 mm. However, aggregate stability was significantly higher in the less than 2 mm and 2 – 3 mm aggregate size compared to 3 – 5 mm, probably due to the effects of the primary soil minerals.
4. The effect of the initial aggregate size on the tensile strength and splash erosion in such soils that are dominated with sand particles and by primary minerals is not completely clear and required further investigation.

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