SITE AND LAND USE EFFECTS ON SOME PHYSICAL PROPERTIES AND THE DISTRIBUTION OF TOTAL CARBON, ALUMINIUM, AND IRON WITHIN AGGREGATES OF SOME HUMIC SOILS IN KWAZULU-NATAL, SOUTH AFRICA

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DECLARATION 2: PUBLICATIONS

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ABSTRACT

Humic soils occur mainly on old land surfaces, with a warm, misty climate having native forests and grasslands as the predominant natural vegetation, in KwaZulu-Natal (KZN) and along the coast of Pondoland and the eastern escarpment of Mpumalanga. The effects of site characteristics and response to replacement of native vegetation with sugarcane cultivation on these soils, remain unclear. This study investigated the effects of different site conditions, including native forest, grassland, and sugarcane production, on (i) general soil characteristics such as pH, clay mineralogy, and total Al, Fe, and C in bulk soils, (ii) bulk density (BD), total porosity (TP), saturated hydraulic conductivity (Ks), and moisture content at field capacity (FC) and permanent wilting point (PWP), (iii) AS and size distribution, and (iv) the distribution of total Al, Fe, and C within different aggregate size fractions in humic soils.

The humic soils studied were developed on dolerite and were under grassland at Cedara and Karkloof and on sandstone and under native forest and sugarcane at Eshowe in KZN. Disturbed samples were collected from soils under each land use at 5 cm intervals from 0-20 cm, 10 cm intervals from 20-60 cm and 20 cm intervals from 60-100 cm depth to compare (a) forest and grassland sites and (b) forest and sugarcane sites. These samples were air-dried and a portion was used for AS measurements while the rest was ground and sieved. The samples for AS were fractionated into large macro-aggregates (LM; > 2000 µm), small macro-aggregates (SM; 250-2000 µm), micro-aggregates (M; 63-250 µm) and silt + clay (SC; < 63 µm).

Results showed that Karkloof soils had lower pH but a higher TOC content than those at the other sites. Cedara and Karkloof soils had higher Al, Fe, clay, and silt content than soils at Eshowe that had a higher sand fraction. After 30 years of sugarcane cultivation, there was an increase in BD and Ks on the sandy clay loam humic soil at Eshowe. However, TOC could not solely explain the water retention characteristics, suggesting that other factors such as texture, especially silt, played a significant role. Soils at Karkloof had higher TOC and MWD, resulting in lower BD and higher TP, thereby increasing Ks, FC, and PWP compared to soils at Cedara and Eshowe. Moreover, TOC only explained Ks and water retention characteristics in the top 30 cm depth, suggesting that texture played a more significant role in these properties below 30 cm.

Soils at Karkloof and Cedara had higher MWD than those at Eshowe. Exchangeable acidity (and thus Al) and silt played a critical role in aggregation at all depths, while TOC was mainly responsible for the formation of LM and SM in the upper 30 cm. Below 30 cm, clay, total Fe, and Al predominated in all aggregates. The aggregate size distribution showed that LM and SM dominated at all sites at the expense of the SC fraction. The M size fraction in soils at Cedara had higher TOC content, while both LM and SM had a higher concentration in soils at Karkloof and Eshowe. Total Al and Fe concentrations were higher in the LM and SM than in the M fractions at Karkloof and Cedara, with the SC fraction at Eshowe having higher concentrations of these elements. Sugarcane cultivation decreased AS and TOC in both LM and SM, while Al and Fe increased in all aggregate size fractions. It was also found that total Al and Fe did not explain the protection of TOC in the aggregates of the studied soils.

The results of the study imply that total Al and Fe contents increase when these soils are cultivated. Therefore, to ensure sustainable production, agricultural management practices such as lime application are needed to reduce soil acidity, and improve the availability of exchangeable bases. Despite these fertility challenges, the high TOC in these soils promotes the formation of stable aggregates, thereby positively affecting BD, TP, Ks, FC, and PWP, especially in the topsoil. To improve or maintain their TOC status, it is strongly recommended that practices such as adding organic matter, reducing tillage, using cover crops, crop rotation, maintaining soil moisture, and applying lime be used. This is because the stability of aggregates against disaggregation is of paramount importance in preventing soil degradation.

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CHAPTER 1: GENERAL INTRODUCTION

Humic soils are important agricultural soils in South Africa, particularly on old land surfaces in KwaZulu-Natal, and along the Pondoland coast and the eastern escarpment of Mpumalanga, where they commonly occur (Macvicar et al., 1984). These highly leached soils are associated with intense weathering, rapid removal of soluble products and a very low base status (< 4 cmol_c of exchangeable bases per kg clay for every one percent organic carbon present) (Soil Classification Working Group, 2018). The clay mineral fraction consists mainly of kaolinite, aluminous chlorite, gibbsite and iron oxides (goethite and/or hematite) (Macvicar et al., 1984). The iron (Fe) and aluminium (Al) released by weathering are relatively immobile and tend to accumulate residually in these soils (Macvicar et al., 1984). While most humic soils are generally deep, well drained and lack visible macro-structure (i.e. apedal), some soils of this type are shallow (Gubevu, 1997). A high content of organic carbon (>1.8 %) in the topsoil (humic A horizon) is the main characteristic feature of these soils and they are not very susceptible to compaction due to their strong structural aggregation, particularly at the micro-scale (Gubevu, 1997; Fey, 2010).

The favorable climate conditions associated with the formation of humic soils make them suitable for intensive agriculture and they are commonly used for cultivating sugarcane, plantation forestry, and vegetables (Fey, 2010). However, there has been limited focus on understanding the changes from native forest and grassland to arable agriculture in physical properties such as bulk density (BD), total porosity (TP), saturated hydraulic conductivity (Ks), field capacity (FC), permanent wilting point (PWP), and available water content (AWC) in these soils. Because these physical attributes vary with factors such as climate and management practices (Kavvadias et al., 2001; Diego et al., 2006; Lal, 2006), study of humic soils can provide more detailed information about their water dynamics and guide appropriate irrigation and drainage strategies for optimal plant growth and soil management (Rawls et al., 2003; Celik, 2005). Adequate water retention capacity, in turn, helps to create and maintain stable aggregates by supplying the moisture required for soil organic matter (SOM) decomposition and microbial activity (Singer et al., 1992; Strudley et al., 2008).

Aggregate stability (AS) is often used to characterise the resilience of soil structure since it controls the dynamics of SOM and nutrient cycling (Elliot, 1986; Carter, 1992; Six et al., 2000a). Several aspects of a soil's physical behaviour are also influenced by AS. Stable aggregates promote balanced porosity against various stresses such as the impacts of raindrops, erosive forces and shrinking and swelling caused by drying and rewetting. Mechanical breakdown by raindrop impact plays a major role under wet conditions as water weakens the soil aggregates (Le Bissonnais, 1996). The breakdown of aggregates may lead to surface sealing and soil compaction (Carter, 1992; Snyman and du Preez, 2005), thereby decreasing infiltration rate to as little as 1 mm h^{-1} (Le Bissonnais, 1996). The AS is also influenced by the land management practices of a given area. For example, van Antwerpen and Meyer (1996) reported a 28 % decline in sugarcane yields on a Glenrosa soil form (Inceptisol) (IUSS Working Group WRB, 2014) as a result of compaction following structural breakdown of aggregates after cultivation in northern KwaZulu-Natal. However, the precise effects of these practices on humic soils are not yet clearly understood.

On the other hand, other land uses with minimal soil disturbance (such as natural grasslands and forests) have the potential to build-up large amounts of SOM and thus soil organic carbon (SOC) thereby improving the stability of aggregates and soil physical fertility (Bronick and Lal, 2005). However, some research findings have suggested that soil type, rather than management practice or land use, is the key determinant of the extent to which soil AS is affected. For example, Sumner (1957) noted a greater AS in dolerite-derived soils that had a mean weight diameter (MWD) of 2.35 mm than those derived from Beaufort (1.72 mm) and Ecca (1.56 mm) shales in the Tall Grassveld ecosystem of KwaZulu-Natal. This difference was ascribed to the AS conferred by Fe oxides in the Fe-rich doleritic soils. Several other studies (e.g. Haynes and Naidu, 1998; Arshad et al., 2004; Agnese et al., 2011) have also reported higher AS in soils with higher clay content than coarser textured soils. The SOC is, nevertheless, extremely important in the functioning of soil since it improves soil structure thereby increasing infiltration rate and soil water content (Amelung et al., 1998; Haynes and Naidu, 1998).

The role of SOC in AS is, however, still a controversial issue worldwide because other workers (e.g. Shepherd et al., 2001; Carter et al., 2002; Mthimkhulu, 2011; Madikizela, 2014) have found only weak or no correlations between SOC and AS, but rather indicated a stronger correlation between the exchangeable bases and AS. In contrast, in shallow (0-20 cm) non humic soils (i.e. Nitisol, Stagnosol and Fluvisol) under undisturbed grassland, cultivated pasture and arable land uses in northern KwaZulu-Natal, Mbanjwa et al. (2022) found that AS increased with SOC up to values ranging from 3.0 to 3.9 %. In humic soils, bases are in low supply and AS is apparently strong but the role of SOC, Al and Fe in both top-soils and subsoils remains unclear. Consequently, the objective of this study was to investigate the impact of soil organic carbon (SOC), aluminium (Al) and iron (Fe) within aggregates, along with bulk density (BD), total porosity (TP), saturated hydraulic conductivity (Ks), field capacity (FC), permanent wilting point (PWP), and available water content (AWC) in humic soils under various land use types in KwaZulu-Natal, South Africa. The results of this study will aid in establishing the causes of the strong AS in humic soils. This, in turn, will improve understanding of soil physical properties that are linked to AS, thereby improving future management of native forest and grasslands as well as sugarcane cultivation on these soils.

The key specific objectives of the study were thus to:

1. Evaluate how different site conditions affect general soil characteristics such as clay mineralogy, texture, pH and total Al, Fe and C under native forest and grassland.

2. Investigate the effects of (i) sugarcane production compared to native forest and (ii) site conditions under native forest and grassland, on BD and other soil water-related properties of some humic soils.
3. Investigate how different site conditions affect (i) AS and size distribution, and (ii) the distribution of total Al, Fe and C within different aggregate size fractions in some humic soils under native forest and grassland.
4. Determine the effects of land use change from native forest to sugarcane farming on (i) AS and size distribution, and (ii) the distribution of total Al, Fe and C within different aggregate size fractions to a 1m depth in some humic soils.

The thesis is structured as follows:

Chapter 1 gives an introduction and some background to the study.

Chapter 2 presents a literature review of some selected models of aggregation and the factors that affect soil AS.

Chapter 3 describes the study sites and the field and laboratory methods used.

Chapter 4 addresses objective one by presenting and discussing results on the impact of different site conditions on general soil characteristics in some humic soils under native forest and grassland.

Chapter 5 focuses on objective two which addresses (i) the effect of sugarcane production compared to native forest on some physical properties of a sandy clay loam humic soil, and (ii) how different site conditions affect BD and some water-related properties in humic soils under native forest and grassland.

Chapter 6 addresses objective three by providing a detailed understanding on the effects of different site conditions on (i) aggregate stability and size distribution, and (ii) the distribution of total Al, Fe and C within different aggregate size fractions under native forest and grassland.

Chapter 7 focuses on research objective four on the effects of long-term (> 30 years) sugarcane cultivation on AS and size distribution, and the distribution of total Al, Fe and C within different aggregate size fractions compared with humic soils from adjacent native forest.

Chapter 8 presents a General Discussion, and gives conclusions and suggestions for future work.

CHAPTER 2: THE EFFECT OF SELECTED SOIL PROPERTIES AND EXTERNAL FACTORS ON THE STABILITY OF AGGREGATES

2.1 Introduction

The capacity of a soil to store and supply water and air for plant growth is an important indicator of its physical fertility. The amount of plant-available water (PAW) in relation to air-filled porosity at field capacity (FC) is often used to assess soil physical fertility (Blanco-Canqui et al., 2017). The PAW is held primarily by the meso-pores and is defined as the water content between FC (at matric suction of -10 kPa) and the permanent wilting point (PWP) (at matric suction of 1500 kPa) (Celik, 2005). Some studies, however, use FC to be water stored at 10 kPa for coarse and 33 kPa matric suction for fine textured soils (Rawls et al., 2003). The PAW is largely controlled by the volume of pores and pore-size distribution, as well as by the soil's ability to transmit water under saturation, known as saturated hydraulic conductivity (Ks). In general, soils with higher Ks, allow for more efficient infiltration and drainage, reducing the risk of erosion and waterlogging among other factors (Blanco-Canqui et al., 2017). The total PAW, however, also depends on other factors such as soil texture, water retention, organic matter and aggregate stability.

Aggregates are formed by organic and inorganic materials cementing the sand, silt and clay particles together, while soil structure refers to the organization of aggregates and the pore spaces between them (Elliott, 1986; Wuddivira et al., 2010). The stability of any soil structure is defined by its ability to retain the arrangement of solids and void spaces that exists at a given time when exposed to different stresses (Bronick and Lal, 2005). Good soil structure has stable aggregates which sustain agricultural productivity while decreasing erodibility (Blanco-Canqui and Lal, 2004). Aggregates occur as a continuum of sizes but are generally classified into macro-aggregates (> 250 μm) and micro-aggregates (< 250 μm) (Tisdall and Oades, 1982).

Aggregate stability (AS) influences a wide range of physical and biogeochemical processes in the natural and agricultural environments. A number of studies (e.g. Hammad and Dawelbeit, 2001; Lal, 2002) have shown that soil compaction not only reduces porosity, infiltration and water holding capacity, but also effectively increases bulk density (BD) and soil strength, thus lowering crop performance via stunted aboveground and reduced root growth (Haynes and Swift, 1990). Soil compactability depends on texture, organic matter (OM), water content and aggregation. For instance, coarse textured soils are less prone to compaction than fine textured soils (Villarreal et al., 2020) and the susceptibility to compaction increases as soil organic matter (SOM) content decreases (Haynes and Naidu, 1998). On the other hand, the water content of the soil controls the level of compaction that a soil may reach (Kornecki and Fouss, 2011).

Aggregate stability is affected by a number of soil internal factors such as texture, clay mineralogy, iron (Fe) and aluminium (Al) oxides, pH, cation exchange capacity (CEC), soil moisture content, porosity as well as external factors such as climate and land use (Six et al., 2000b). Due to the complex nature of the processes and mechanisms involved in aggregation, several theories and models have been proposed to relate soil aggregation to soil organic carbon (SOC) dynamics (e.g. Tisdall and Oades, 1982; Oades, 1984; Six et al., 2000a) and these models have influenced the way research and its application have subsequently developed (Blanco and Lal, 2004). For example, the holistic concept of aggregation embodied in the model by Tisdall and Oades (1982) has been used (with minor modifications) in a wide range of soil studies (e.g. Elliott, 1986; Haynes and Swift, 1990; Six et al., 2000b; Shepherd et al., 2001) to study the composition and turnover of SOM within the aggregate matrix. This review outlines the relationship between water-related physical properties and AS under different land uses, some of the conceptual models of soil aggregation, and the effects of both internal and external soil factors on aggregate stability. Although the focus is on South African research, this review refers to work done in other countries as well in order to put the local research into a wider perspective.

2.2 Soil water retention

The soil water retention (SWR) refers to the ability of soil to hold water (Celik, 2005), and is closely related to AS. For instance, the high quantity of Fe and Al oxides, which are commonly present in highly leached soils (e.g. Oxisols), induce the formation of stable aggregates (Totsche et al., 2018). Consequently, these soils have a well-developed structure and contain an abundance of inter- and intra-particle pores, which allow for high porosity and large amounts of water to be retained at various matric potentials (Fey, 2010). Therefore, any changes in pore space distribution are generally accompanied by changes in the soil water status due to the re-arrangement of capillary (between soil particles) and non-capillary (between soil aggregates) pores (Lal, 1978; Strudley et al., 2008). Soil aggregates should generally possess pores that are between 30 and 0.2 µm, to hold capillary water (Blanco-Canqui et al., 2017). Large pores between the soil aggregates allow rapid infiltration of water as well as free drainage so that the soil remains aerobic (Rawls et al., 2003). In general, an increase in aggregation will lead to a decrease of soil BD, which tends to increase the total pore space bringing about variations in the SWR (Rawls et al., 2003; Blanco-Canqui et al., 2017).

Although the relationships outlined above highlight the interconnectivity between soil structure, BD and SWR, results reported on the relationship between SOC and SWR are contradictory. Haynes and Naidu (1998) for instance reported a 1.5 % increase in soil water content at FC with a 1 % increase in SOC. Similarly, Emerson and McGarry (2003) showed that a 50 % increase in water content is achieved per gram of additional SOC at FC. On the other hand, Lal (1978) did not find any effects of SOC on SWR whereas Calhoun et al. (1973) did obtain positive results at FC, but the results for PWP were ambiguous. McBride and MacIntosh (1984) further reported that SOC affected SWR at PWP only if its concentration was > 5 %. The conflicting results between SOC and SWR may, nevertheless, be attributed to the synergistic effects of textural components and SOC. Rawls et al. (2003) conducted a study to assess the effect of soil organic carbon on soil water retention and develop pedotransfer functions to incorporate this effect using 1200 soil samples. Regression tree modelling and group method of data handling were applied to relate water retention at -33 kPa to organic carbon content and texture. The results showed that the sensitivity of water retention to changes in organic carbon content varied at different organic carbon contents, with the highest sensitivity observed at a low carbon content of 1 % (Figure 2.1). The study also found that there were differences in the relative importance of organic carbon content in coarse and fine-grained soils. The study demonstrated the complex relationship between soil texture and organic carbon content and how they jointly affect water retention. The results of this study have important implications for soil management and protection, as they provide a better understanding of how soil organic carbon content affects water retention in soil and can be used to develop more accurate pedotransfer functions for estimating water retention.

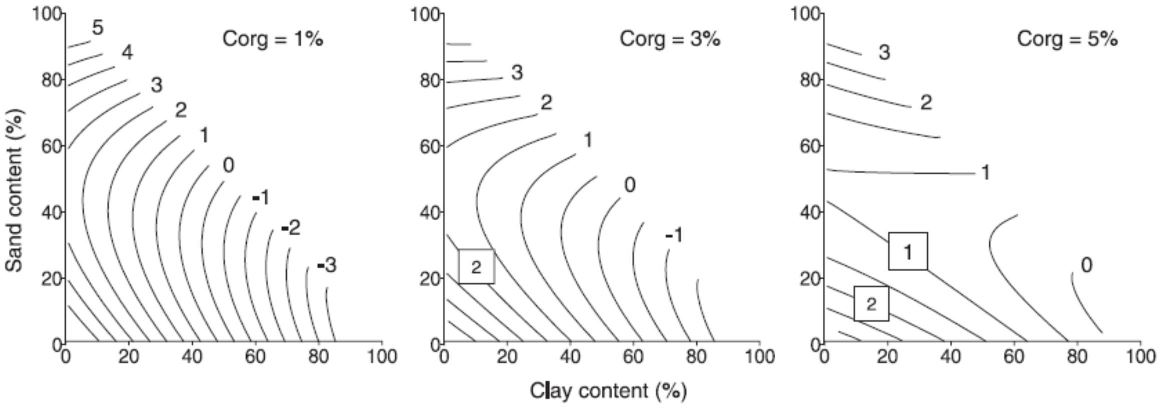

Figure 2. 1: Changes in soil water content at -33 kPa (Vol. %) per 1 % change in organic carbon (C_{org}) content (redrawn from Rawls et al., 2003).

2.3 Selected models of soil aggregation

The models of Tisdall and Oades (1982), Oades (1984), and Six et al. (2000a) are examples of those that have been proposed to explain the processes of soil aggregation that are important to plant growth and biological activity (Figure 2.2). Tisdall and Oades (1982) proposed a pioneering conceptual model for aggregate hierarchy in which SOM is the main binding agent of aggregate formation. The model proposed four stages of aggregation such that micro-aggregates are linked together to form macro-aggregates. In this model, the small micro-aggregates (< 2 µm) are associated with polyvalent metal cations and strongly sorbed polymers, whereas medium micro-aggregates (2-20 µm) are an organic- rich fraction that is held together by glutinous material. The aggregates between 20 and 250 µm are linked by a combination of alluminosilicates, hydrous oxides of Al, transient (e.g. microbial- and plant-derived polysaccharides) and persistent agents while the largest aggregates (> 250 µm) are enmeshed by temporary agents such as roots and fungal hyphae.

Oades (1984), however, revised the model and proposed only three stages of aggregation, and that (i) micro-aggregates do form within macro-aggregates, and so micro-aggregates (and not macro-aggregates) mainly protect the SOC, (ii) effect of root exudates, which act as glues or alter the wetting and drying characteristics of the rhizosphere, and many polysaccharides (that have a sticking effect on soil particles, improving aggregate stability and preventing their breakdown), is primarily mediated through associated mycorrhizal fungi and (iii) a hierarchy of pores exists along with the hierarchy of aggregates (Figure 2.2). Similarly, a model by Six et al. (2000a) identified four dynamic stages of macro-aggregate turnover, micro-aggregate formation and SOC stabilization in micro-aggregates (Figure 2.2). According to the model, SOC-enriched residues initially form larger aggregates, known as macro-aggregates. These macro-aggregates subsequently transform into inter-aggregate particulate organic matter (POM), contributing to the formation of smaller aggregates called micro-aggregates. Eventually, the micro-aggregates are disintegrated or dispersed, allowing them to participate in the next cycle of macro-aggregation. Consequently, micro-aggregates are denser and have a higher internal strength than macro-aggregates (Bronick and Lal, 2005).

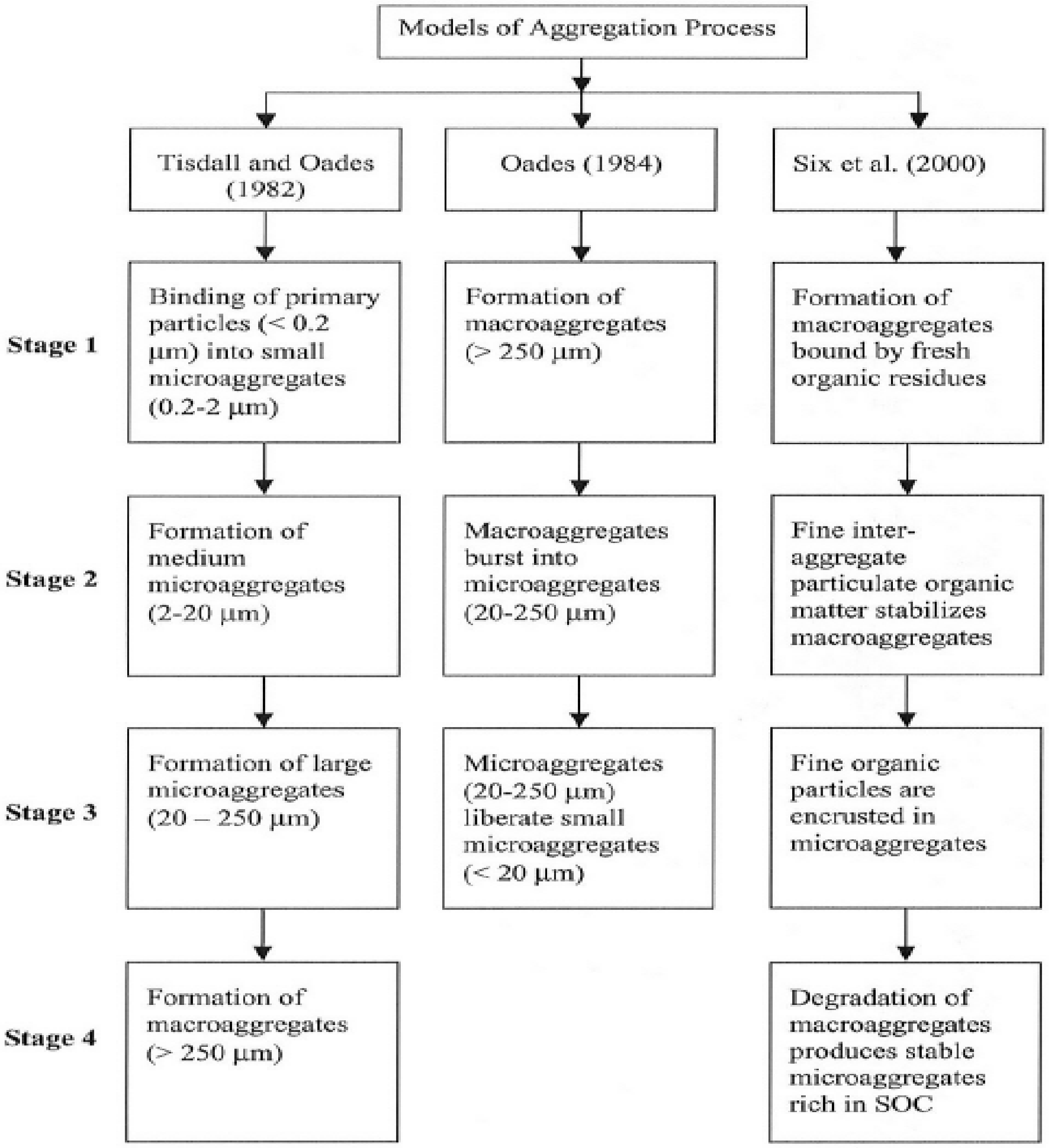

Figure 2. 2: Selected models of the formation of macro-and micro-aggregation of soils (Blanco-Canqui and Lal, 2004).

2.4 Internal factors of soil aggregation

The types of internal binding agents that are involved in soil aggregation affect the nature, size, strength and configuration of aggregates (Six et al., 1998; Blanco-Canqui and Lal, 2004).

2.4.1 Soil texture

Soil texture is one of the most important factors that plays a crucial role in the development and maintenance of soil structure leading to the physical protection of SOC. Bronick and Lal (2005) indicated that SOC is not easily accessible to soil microbes in clay and silt textured soils due to chemical adsorption onto the surfaces of clay minerals and also physical occlusion within micro-aggregates. As a result, clay and silt associated SOC has a longer turnover time than that within the sand fraction, regardless of soil type and depth. Sandy soils generally suffer greater losses of OC due to low negativity on the edges of the sand particles (Blanco-Canqui and Lal, 2004; Wuddivira et al., 2010), which reduces their ability to retain and protect the SOC, making it more susceptible to leaching, dissolution, and microbial decomposition (Singer et al., 1992). When SOM is very low or absent, promotion of soil particle aggregation by high clay content has been reported (Shepherd et al., 2001; Lado et al., 2004). In an attempt to confirm these findings, Tayel et al. (2010) analysed three different Egyptian topsoils to determine the soil aggregation percentage and particle size distribution (Table 2.1). They concluded that soils with higher clay content were more strongly aggregated than soils with lower clay content. On the contrary, Lado et al. (2004) studied some soils from Spain and observed increased aggregate breakdown with an increase in clay content. It was concluded that a minimum of 18 % clay is required before clay could contribute significantly to soil aggregation.

Table 2. 1: The influence of different texture classes on soil aggregation in some topsoils in Egypt (modified from Tayel et al., 2010).

Profile number	Depth (cm)	Aggregation (%)	Coarse sand	Fine sand	Silt	Clay
			……………….......... (%)	....		
1	0-25	63.86	7.35	35.67	23.02	33.96
2	0-30	81.45	1.55	16.37	31.36	50.71
3	0-20	80.50	1.23	3.85	24.65	70.28

2.4.2 Clay mineralogy

The weathering of primary minerals leads to the formation of highly reactive secondary clay-sized minerals in soils that consist of a range of phyllosilicates with varying degree of aggregation (Totsche et al., 2018). Lado and Ben-Hur (2004) reported higher AS in a soil dominated by kaolinite and vermiculite compared to soils dominated by chlorite and illite. This was due to the stabilizing effect caused by interaction between kaolinite, and Fe and Al oxides that increased the AS (Six et al., 2000b). The adsorption of these oxides onto the planar surfaces of kaolinite decreased the CEC and increased the number of positive charges hence encouraging aggregation (Lado and Ben-Hur, 2004). In similar vein, Oades and Waters (1991) established that under unstable conditions, such as high sodium adsorption ratio values and low electrolyte concentration, soils high in kaolinite and sesquioxides are relatively stable. On the other hand, soils with a high content of montmorillonite are relatively unstable while those with low or no montmorillonite have an intermediate behaviour. All these studies showed that the frequent association of kaolinite with Fe oxides might be responsible for the very strong resistance of aggregates to slaking (Six et al., 2000b).

In a review of South African literature on soil mineralogy studies, Bühmann et al. (2004) noted a two-fold role of smectite in AS, firstly as a strong stabilizing agent, and secondly as a dispersive agent in cases where the cohesion is overcome by swelling. Differences in the dispersivity of clays are generally associated with their morphology and surface electric charge but the effect is difficult to assess because soils usually contain a mixture of clay minerals, so the association with other minerals alters their behaviour (Oades and Waters, 1991). Smectitic soils are, however, the most dispersive compared to kaolinitic soils (Bühmann et al., 2004). Lado and Ben-Hur (2004) showed that kaolinitic soils have a higher aggregate mean weight diameter (MWD) compared to montmorillonite dominated soils (Table 2.2). Wakindiki and Ben-Hur (2002) also observed that the MWD in a kaolinitic soil was more than ten times higher than in smectitic soils from Israel (profile 2), although both soils had similar clay content. While this difference in the case of profile 2 was attributed to the strong interactions between the positive charge at the edge and the negative charge at the flat surface of the kaolinite soil, the much higher clay content of kaolinite compared to montmorillonite clays in profiles 3 and 4 was another reason attributed to the observed differences.

Table 2. 2: The relationship between the mean weight diameter (MWD), clay mineralogy and clay content in soils from different locations in Israel and Kenya (modified from Lado and Ben -Hur, 2004).

Country	Profile	Clay mineralogy	MWD (mm)	Clay (%)
Kenya	1	Kaolinite	2.80	64.0
Israel	2	Montmorillonite	0.25	63.0
Israel	3	Montmorillonite	0.84	30.4
Kenya	4	Montmorillonite	0.31	10.0

2.4.3 Aluminium and iron

There is some contradiction among available studies in explaining AS in relation to Fe and Al oxide dynamics. Some researchers (e.g. Oades and Waters, 1991; Le Bissonnais, 1996; Lado and Ben-Hur, 2004; Igwe et al., 2009; Totsche et al., 2018) state that higher structural stability in highly leached, acidic soils is a result of Fe and Al polyvalent cations. In contrast, other studies (e.g. Bartoli et al., 1992; Shainberg and Levy, 1994), showed that the effect of Al and Fe on AS is negligible, which could be due to differences in the mineralogy of the soils used in these studies (Ben-Hur, 2004). Although the mechanisms are not clear as yet, there is evidence that Fe (hematite) and Al (gibbsite) oxides play an important role in soil aggregation. Churchman et al. (1993) for instance observed that Fe and Al are adsorbed on kaolinite surfaces, inducing a cementation effect, leading to the development of strong aggregates in highly leached Oxisols. Barthès et al. (2008) made similar observations and concluded that such an effect was a result of the association of positively charged polyvalent cations with the negatively charged clay minerals that stabilized soil particles thereby improving aggregation. In general, the effect of Al and Fe varies with particle size and aggregate size distribution (Oades and Waters, 1991; Golchin et al., 1994; Totsche et al., 2018).

It is, however, still debatable whether Fe or Al is more effective at aggregating soil particles. According to Keren and Singer (1990), Al is more effective in aggregation than Fe due to (i) the minerals' higher point of zero charge leading to a higher charge density, and consequently to stronger clay-polymer attraction forces; and (ii) the planar shape of the Al precipitates, compared to the spherical shape of the Fe polymers, enabling a closer binding surface to the clay particles. On the other hand, Bullinger-Weber et al. (2007) found that Fe was a more effective stabilizing agent compared to Al. They concluded that the stabilizing effect was associated with the charge on the polymers rather than the total amount extracted. This suggests that the charge on the polymers is determined by the dissociation of the monomer side groups. As the number of dissociated monomer side groups increases, so does the charge on the polymer, consequently affecting the stabilizing effect of Fe. The effect of Al and Fe is nonetheless to encourage aggregation and especially the AS of the micro-aggregates (Churchman et al., 1993; Igwe, 2005; Totsche et al., 2018). Peng et al. (2015) assessed the contributions from sesquioxides and SOM to aggregation in Ultisols. They found that removal of Fe/Al oxides broke down the 2–53 µm aggregates most intensively whereas removal of SOC disrupted only the 250–2000 µm aggregate size fraction, indicating that SOC was only responsible for aggregation in this fraction. It was concluded that the aggregating role of SOC could be less effective in soils that contain high concentrations of Fe/Al oxides.

2.4.4 Organic matter

Although the influence of SOC on AS has been amply addressed, inconsistencies exist because some researchers have found weak (Zhang et al., 1996; Mthimkhulu, 2011; Madikizela, 2014) or no correlation (Perfect and Kay, 1990; Carter et al., 2002) between SOC and AS while others (e.g. Haynes and Beare, 1997; Chenu et al. (Figure 2.3), 2000; Haynes, 2000; Zhao et al., 2017) have showed strong positive correlations between the two variables. Even the correlations reported between aggregate sizes and SOC are somewhat contrasting. Jastrow et al. (1998) for example found low correlations (r = 0.28 for > 2000 µm, 0.43 for 1000-2000 µm and 0.02 for 212 -1000 µm). On the other hand, Golchin et al. (1994) observed a strong positive correlation (r = 0.86) between SOC and 1000-2000 µm aggregates. In central Spain, Hernánz et al. (2002) reaffirmed the results of Golchin and co-workers by showing a positive correlation (r = 0.62) between SOC and 1000-2000 µm aggregates. A positive correlation between SOC and AS is likely as polysaccharides have a transient aggregating effect on micro-aggregates, while roots and hyphae have a temporary stabilizing effects on macro-aggregates, and polymers and aromatic compounds have a persistent aggregating effect on micro-aggregates (Le Bissonnais 1996). Conversely, Zhang et al. (1996) and Carter et al. (2002) indicated that the addition of fulvates, citrates or oxalates could increase clay dispersion and prevent aggregate formation.

According to Amézketa (1999), there are probably two hypotheses to explain this apparent contradiction. Firstly, the effect of SOM depends on the type of union between the humic substances and the clay and, in particular, on the size of the organic anions. Only if the organic anion is longer than the clay edge, will it attach to the edges of several clay particles and bind them together. Secondly, SOM acts differently at the two levels of macro- and micro-aggregation. Thus, organic bonds stabilize aggregates against slaking and disaggregation, but once these bonds are broken and disaggregation has occurred, the SOM acts as a deflocculant. In this sense, Pulleman and Marinissen (2004) suggested that the effect of SOM on soil structure is a function of the size of the soil particles analysed. As a result, in clay-sized aggregates, OM forms a hydrophobic coating around the aggregates, reducing soil wettability, and consequently reducing the sensitivity to slaking. By contrast, in coarse sand-sized aggregates, OM acts as a binding agent, through roots and hyphae. As a result, SOM would have different effects on macro-aggregates compared to micro-aggregates.

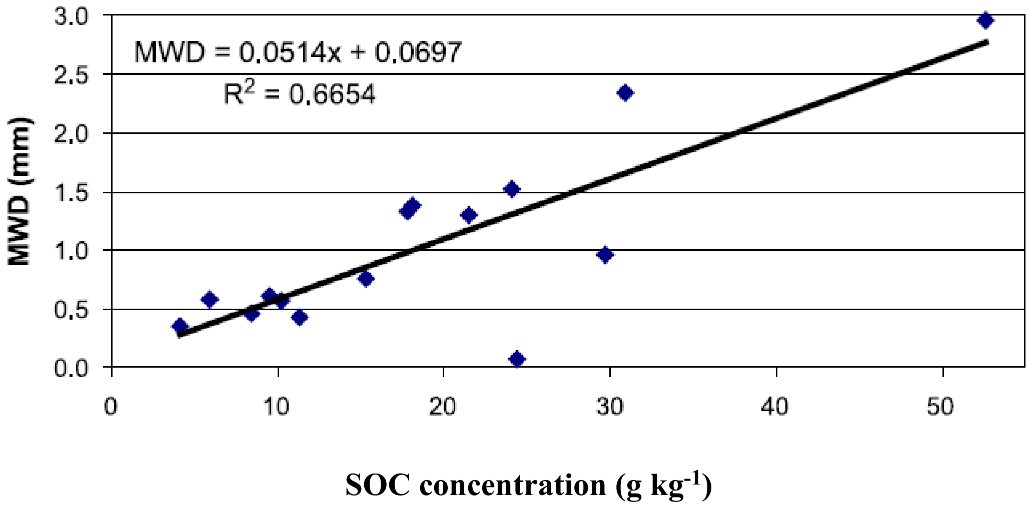

Figure 2. 3: Correlation between soil organic carbon (SOC) and mean weight diameter (MWD) in soils from southwest France (Chenu et al., 2000).

However, the total SOC or SOM may not be a good indicator of AS, mainly because the different pools that make up the bulk SOC differ in their physical and chemical properties and hence have specific functions (Baldock and Skjemstad, 2000). The humic fraction for example primarily contributes to a soil's CEC, while soil structure relies on both humic and particulate organic carbon (POC) fractions. Soil thermal properties, on the other hand, are influenced by humus and the inert carbon pool (Figure 2.4). The POC is mostly important in providing energy for biological processes while humus is an essential source of nutrients (Krull et al., 2013). In the Eastern Cape Province of South Africa, Nciizah and Wakindiki (2014) found a stronger correlation between the AS and total SOM (r = 0.78) than POC (r = 0.53). It was concluded that some other SOM fraction and not POC influences this relationship in weakly weathered soils. The contradictory results found in the correlations between SOC and AS are therefore due to one or more of the following reasons: (i) only part of the OM is responsible for aggregation (Haynes and Beare, 1997), (ii) there is a content of SOC above which there is no further increase in aggregation (Haynes, 2000; Mbanjwa et al., 2022), (iii) organic materials may not be the major binding agents in some soils (Carter et al., 2002), and (iv) aggregation is more related to free organic materials than total SOM content (Amézketa, 1999).

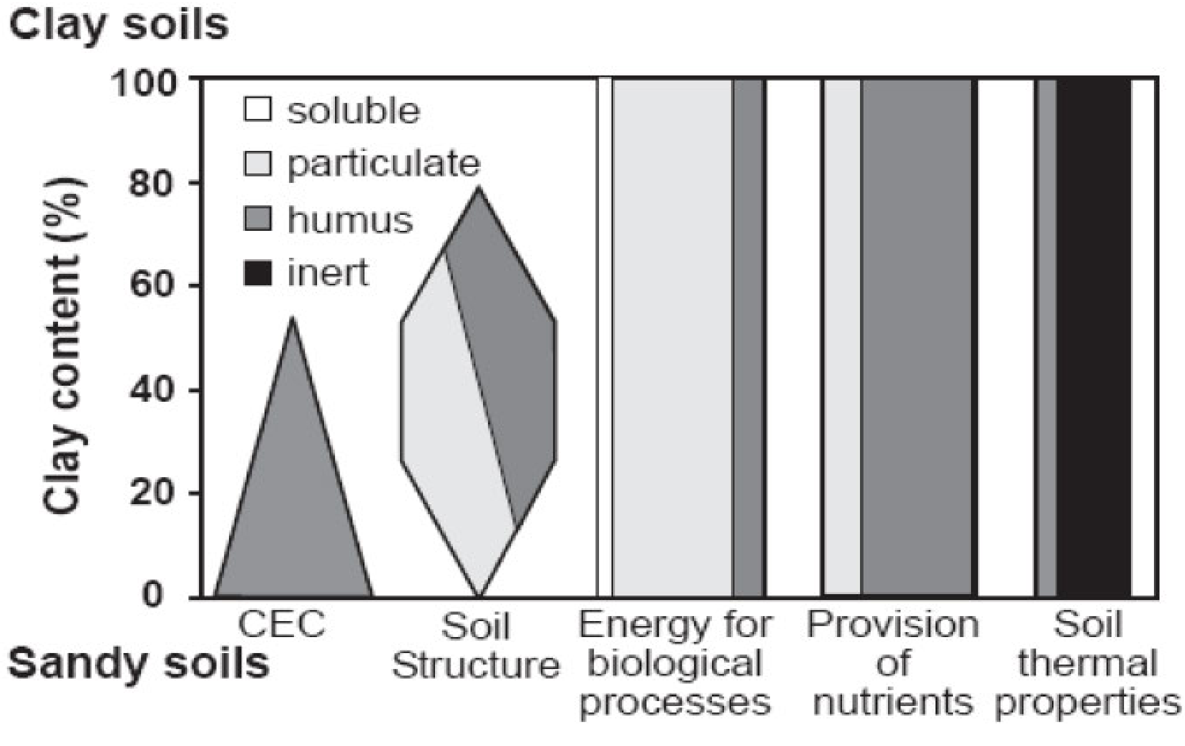

Figure 2. 4: The optimal expression of soluble, particulate, humus and inert organic pools (Krull et al., 2013).

2.4.5 Natural wetting-drying cycles

Periodic changes in moisture and temperature (wetting-drying cycles) are considered to be one of the processes responsible for the formation of aggregates (Six et al., 2000a; Bronick and Lal, 2005). Wetting-drying cycles also repair soil structure due to rearrangement of the soil particles previously disturbed by compaction (Calhoun et al., 1973; Singer et al., 1992; Rabbi et al., 2014). According to Six et al. (2000a), the effects of wetting-drying cycles on aggregate formation and stabilization depend on the rate of wetting and the chemical composition of wetting solutions generated by root exudates. Soil aggregation adjacent to plant roots is increased by intense and periodic drying of soil because of plant transpiration. Rapid wetting by rainfall can cause unstable soil aggregates to collapse and result in a hard and structureless mass of soil (massive structure) when the soil becomes dry (Bronick and Lal, 2005). Singer et al. (1992) observed a progressive development of soil structure from massive to complex crumb, blocky and platy after sequential wetting-drying cycles. The effect of wetting-drying cycles on soil structure can, however, not be generalized since both increases and decreases in water-stable aggregation have been observed following drying due to mineralogy or texture (McBride and MacIntosh, 1984; Bronick and Lal, 2005).

2.4.6 Soil type

The mechanisms that control aggregation differ in different soil types (Six et al., 2000a; Bronick and Lal 2005). The Al-humus complexes and noncrystalline Al hydroxides for example are dominant aggregants in both Oxisols and Ultisols while carbonates are more common aggregants for Aridisols. Allophane, OM, and Fe and Al-oxyhydroxides on the other hand dominate the aggregation of Andisols in volcanic regions (Velde and Barré, 2010). Ćirić et al. (2012) studied the effects of soil type on the AS of five different soil types (Arenosols, Fluvisols, Chernozems, Gleysols and Solonetz) in Vojvodia, Serbia. They found better aggregation in Fluvisols, Gleysols, Solonetz and Chernozems than Arenosols due to the presence of numerous cementing agents and high clay content. The relationship between soil aggregation and soil type is, however, complex because it is the basic combinations of soil properties that strongly influence aggregation (Schmidt et al., 2011). Soil type can thus impact aggregation through its mineral composition, OM content and erosion susceptibility. Aggregation is often stronger in clay-rich soils, and may be weaker in sandy soils. However, regardless of soil type, OM content and land use management practises have a significant impact on aggregation (Pulleman and Marinissen, 2004; Ćirić et al., 2012).

2.5 External factors of soil aggregation

The external factors influence aggregates through altering the internal factors in a direct or indirect manner. Research (e.g. Sumner, 1957; Calhoun et al*.,* 1973; Oades, 1984; Haynes and Swift, 1990; Six et al., 2000a; Bronick and Lal, 2005) has identified climate and land use (i.e. forest, native grassland and cropland) as key external factors affecting the stability and size distribution of aggregates.

2.5.1 Climate

Temperature and rainfall are important factors that affect soil AS because both play important roles in biomass production thereby affecting the net balance of SOC. Soil structure and SOC are interrelated in that SOC holds minerals together to form aggregates in soils and these in turn, offer protection to SOC against decomposition (Tisdall and Oades, 1982). This effect, however, becomes less evident with an increasing depth of soil as texture instead of OM plays a more important role in the stability of aggregates (Diego et al., 2006). In addition, warmer climates cause soils to weather more quickly, resulting in higher levels of Fe and Al and more stable aggregates (Lal, 2002). Soils of humid regions are often found to have aggregates of high stability compared to those of dry regions due to a greater production of OM in humid regions

(Lal, 2002). Dry soils have limited biomass addition, resulting in less microbial activity and SOC addition, but wetter soils have higher SOC decomposition if there is significant air circulation (Lal et al., 1979).

In general, the effect of rainfall on micro-aggregate formation is largely through slaking due to raindrop impact and rapid wetting. The mechanical breakdown of aggregates by raindrop impact usually occurs in combination with the other mechanisms if the kinetic energy of the raindrops is great enough (Le Bissonnais, 1996). Raindrop impact plays a dominant role when the soil aggregates are wetter and hence weaker (Diego et al., 2006, Liu et al., 2011). The breakdown of macro-aggregates by raindrops often results in the development of micro-aggregates < 100 µm (Le Bissonnais, 1996; Podwojewski et al., 2014), depending on the rainfall intensity, pattern and duration (Bronick and Lal, 2005; Diego et al., 2006). Ramos et al. (2003) showed that low intensity rainfall causes less soil structural degradation (i.e. MWD = 1.26 mm) than high intensity rainfall (i.e. MWD = 0.82 mm). It was concluded that the differences were as a result of the disaggregation of the soil aggregates due to fast wetting under the high intensity rainfall. The extent of the soil aggregate degradation largely depends on the intensity of the rainfall and state of the soil before rainfall begins (Liu et al., 2011; Podwojewski et al., 2014).

2.5.2 Land use

2.5.2.1 Forest

Native forest ecosystems cover approximately 30 % of the world's land area and are classified into four categories, each with its own set of characteristics (Park, 1971; Evrendilek et al., 2004; Mucina and Rutherford, 2006). Temperate deciduous and coniferous forests are found in areas with moderate climate and have a dense canopy of trees that provides great soil erosion protection (Evrendilek et al., 2004). Tropical rainforests are distinguished by extensive biodiversity and dense vegetation, which enables effective nutrient cycling by allowing different species to contribute to nutrient release, recycling and availability (Lawes et al., 2004). Boreal forests of high-latitude locations with cold climates produce OM that decomposes slowly due to low temperatures and acidic conditions. Finally, mangrove forests thrive in coastal settings, where brackish water and salt-tolerant flora operate as natural filters, trapping silt and pollutants and so enhance water quality (Mucina and Rutherford, 2006).

Several studies (e.g. Hudson, 1994. Leifeld et al., 2005; Wang et al., 2009; Jarvis et al., 2013; Kar et al., 2017) have found a positive relationship between native forests and soil water content under a wide range of soil and climate conditions around the world. In general, the dense canopy cover provided by trees helps to intercept rainfall while reducing direct raindrop impact on the soil surface. This promotes water infiltration and reduces runoff from the soil (Blanco-Canqui and Lal, 2004). Moreover, the litter layer serves as a natural mulch, reducing evaporation and maintaining stable soil moisture levels (Leifeld et al., 2005). It has also been noted that the extensive root systems of trees in natural forests provide pore spaces that promote water-holding capacity and soil moisture retention (Kaiser et al., 2002; Celik, 2005; Jarvis et al., 2013). In addition, several studies (e.g. Park, 1971; Hudson, 1994; Blanco-Canqui and Lal, 2004; Evrendilek et al., 2004; Kar et al., 2017) have found lower BD under native forests than other land uses or disturbed areas, which has been related to the presence of OM, and the activities of soil organisms within the forest ecosystem that reduce soil compaction. All these studies concluded that lower BD in natural forests is beneficial as it allows for improved root penetration, water infiltration and nutrient exchange and so promote plant growth and soil health.

Native forests are important not only for maintaining low BD and increasing soil water content, but also for aggregating soil particles through abundant litter and extensive root systems (Blanco-Canqui and Lal, 2004). Several studies (e.g. Kaiser et al., 2002; Blanco-Canqui and Lal, 2004) have indicated that forest trees improve soil aggregation while stabilizing old SOC pools (physically protected SOM), associated with the formation of micro-aggregates. The extent of stabilization is determined by organo-mineral interactions, type and nature of clay surfaces and SOC location within micro-aggregates (Blanco-Canqui and Lal, 2004). Kaiser et al. (2002) found that 45 % of the SOC in forest subsoil was bound within micro-aggregates of < 20 µm in diameter. Del Galdo et al. (2003) also found an improvement in the stabilization of SOC within micro-aggregates (53-250 µm) and silt + clay fraction (< 53 µm) following afforestation. However, Jacinthe et al. (2001) observed higher SOC concentration in macro-aggregates than micro-aggregates in both forest and cropland soils.

In general, macro-aggregates (250–2000 µm) may play an important role in aggregate dynamics in mature forests due to differences in annual OM inputs and litter quality (Oades and Waters, 1991; Evrendilek et al., 2004), as well as the enmeshing effect of roots and associated mycorrhizal hyphae (Blanco-Canqui and Lal, 2004). All of these effects may contribute to stabilization of topsoil macro-aggregates (Six et al., 2000b) because the SOM makes conditions favourable for proliferation of soil organisms, although this depends on the tree species as some trees create an environment which is unfavourable for earthworms e.g. pine trees (Kavvadias et al., 2001).

2.5.2.2 Grassland

Native grassland ecosystems cover about 26 % of the world's land area and are divided into many categories based on climate, geographical location, plant communities and ecological dynamics, among other characteristics (Snyman and du Preez, 2005; Aucamp, 2008). They are distinguished by tall grasses and rich, fertile soils that can support intensive agriculture. Tropical grasslands for instance occur in regions with distinct wet and dry seasons, although the region maintains relatively high temperatures. They are characterised by a combination of grasses and scattered trees or shrubs that are resistant to drought. Mediterranean grasslands are found in areas with moderate, wet winters and hot, dry summers and are dominated by drought-resistant plants, grasses, and herbs. Desert grasslands occur in arid regions with low rainfall and high temperatures. They have scant vegetation made up of drought-tolerant grasses, succulents, and scattered bushes (Arshad et al., 2004; Lawes et al., 2004).

Similar to native forests, a number of studies (e.g. Oades and Waters, 1991; Evrendilek et al., 2004; Snyman and du Preez, 2005; Wang et al., 2009) have found that native grassland has a positive influence on SWR, which has been attributed to the fibrous root systems of grasses that penetrate deep into the soil, forming channels and pathways that improve water infiltration. The extensive network of grass roots not only improves SWR but also reduces the risk of erosion (Mills and Fey, 2003; Podrázský et al., 2015). Grass also serves to intercept rainfall, reducing the impact of raindrops on the soil surface and facilitating gradual infiltration of water (Aucamp, 2008). In addition, the fibrous root systems of grasses, as well as the activities of soil organisms such as earthworms help to produce bigger pore spaces within the soil. This reduces soil compaction and BD (Nemes et al., 2005; Li et al., 2007; Wang et al., 2016).

On the other hand, some studies have revealed that the AS of grassland soils may be equal (Cerdá, 1998; Podrázský et al., 2015), higher (Gijsman and Thomas, 1995), or lower (Evrendilek et al., 2004) than that of forest soils. Evrendilek et al. (2004) found that grassland produces more SOM and SOC that leads to higher aggregate MWD and lower BD than natural forest (Table 2.3). Mills and Fey (2003) reported similar findings in South African grassland soils and concluded that the increased root biomass under grassland results in higher biological activity that contributes to higher SOC, thereby improving aggregation. Further comparisons of different types of grasses (e.g. Cambardella and Elliot, 1992; Mills and Fey, 2003) showed that perennial grass systems provide greater SOM accumulation than annually cropped grasses. The positive effect of perennial grass on total carbon within macro-aggregates (> 1 mm) decreases in micro-aggregates (0.25 – 1 mm) compared with annually cropped grasses (Cambardella and Elliot, 1994). Soil AS was higher under the perennial grasses bromegrass (*Bromus inermis* Leyss.) and red fescue (*Festuca rubra* L.) than under annually cropped grasses (MWD of 2.1 and 1.6 mm under perennial and annual systems, respectively) (Arshad et al., 2004). According to Cerdá (1998), differences in root mass and litter composition between grasses explain the differences in AS among grass species.

Table 2. 3: Comparison of soil properties (mean ±S.D.) of forest and grassland soils (0-20 cm depth) (modified from Evrendilek et al., 2004).

Soil property	Forest	Grassland
Bulk density (Mg m^{-3})	1.27±0.03b	1.19±0.050c
Soil organic matter (g kg^{-1})	38.83±3.80b	41.27±4.50b
Soil organic carbon (kg ha^{-1})	56480±132b	57317±261c
Mean weight diameter (mm)	2.89±0.60b	3.50±0.70c

Different letters in each row indicate significant differences ($p < 0.05$).

2.5.2.3 Tillage

In agricultural practices, various types of tillage systems are employed, each with its unique characteristics and objectives. Conventional tillage, for instance, involves significant soil disturbance, deep tilling, and the incorporation of organic matter. It is typically utilised to address compacted soil layers and integrate crop residues (Lal and Kimble, 1997; Six et al., 1998). On the other hand, conservation tillage focuses on minimising soil erosion and enhancing soil health by reducing soil disturbance and leaving crop residues on the soil surface. On the other hand, conservation tillage of various types focuses on minimising soil erosion and enhancing soil health by reducing soil disturbance and leaving crop residues on the soil surface. No-till, in particular, eliminates the use of traditional tillage implements, promoting planting directly into untilled soil with minimal soil structure disruption. This method is effective in reducing soil erosion, improving water infiltration, and increasing organic matter content. Ridge tillage creates raised beds or ridges with furrows in between, and is mainly used in row crop production to enhance water drainage, decrease soil erosion, and create an optimal seedbed for planting. Strip tillage involves tilling narrow strips for planting while leaving the remaining field untilled. Mulch tillage, which retains crop residue on the soil surface, helps conserve moisture, reduce erosion, and enhance soil structure, often complementing other tillage systems like strip-till or zone tillage. Zone tillage, a modified deep tillage method, exclusively agitates the soil in narrow strips between the rows, addressing soil compaction issues and enhancing internal soil drainage without disturbing the soil (Puget et al., 1995; Franzluebbers and Arshad, 1996).

Although tillage is widely reported as a major driver of destabilization of soil aggregates, particularly when native ecosystems are converted to agricultural land (Arshad et al., 2004: Bronick and Lal, 2005), different tillage systems have different impacts on AS (Six et al., 2000a; Chivenge et al., 2007). Puget et al. (1995) showed that plant residues left under no-till resulted in greater macro-and micro-aggregation that physically sheltered SOC within aggregates and this was attributed to the increased biological activity. In another study, Arshad et al. (2004) observed a drastic decrease in AS following conventional tillage. It was concluded that repeated tillage brings about loosening of soil particles especially when it involves the use of a mouldboard plough, followed by disking and harrowing before planting. According to Chivenge et al. (2007), reduced tillage increases below-ground OM, and improves soil aggregation. Minimally disturbed soils may generally have micro-aggregates with twice the stability of conventionally tilled soils, promoting SOC sequestration within the micro-aggregates due to slower macro-aggregate turnover (Franzluebbers et al., 1999; Wuddivira et al., 2010). The SOC rich residues within micro-aggregates remain undecomposed for a long time because they are protected from enzymatic and microbial actions.

Conventional tillage, on the other hand, disrupts aggregation by exposing existing aggregates to microbial processes, thereby accelerating SOC turnover and reducing AS (Chivenge et al., 2007). The processes that result in an interaction between SOC and soil structure cease in drastically disturbed soils leading to the formation of SOC depleted micro-aggregates and unstable macro-aggregates (Figure 2.5; Lal and Kimble, 1997). In addition, in response to fertilization, AS may decrease (Franzluebbers et al., 1999), not change (Celik, 2005) or vary (Milne and Haynes, 2002) depending on the amount and type of fertilizer added. These varied responses indicate the complex relationships between AS factors and tillage and fertilizer changes (Franzluebbers and Arshad, 1996). According to Blanco-Canqui and Lal (2004), the relationship is even more complex in high-C soils due to the interaction between texture, AS and SOC dynamics. For instance, soils with higher clay and silt are generally better aggregated. High-C soils benefit from increased AS due to the presence of OM acting as a cementing agent. Forest and grassland soils often exhibit higher AS and SOC content, with a larger proportion of SOC associated with macro-aggregates. Cropped soils may have lower AS and SOC content, particularly in the macro-aggregate fraction due to intensive agricultural activities.

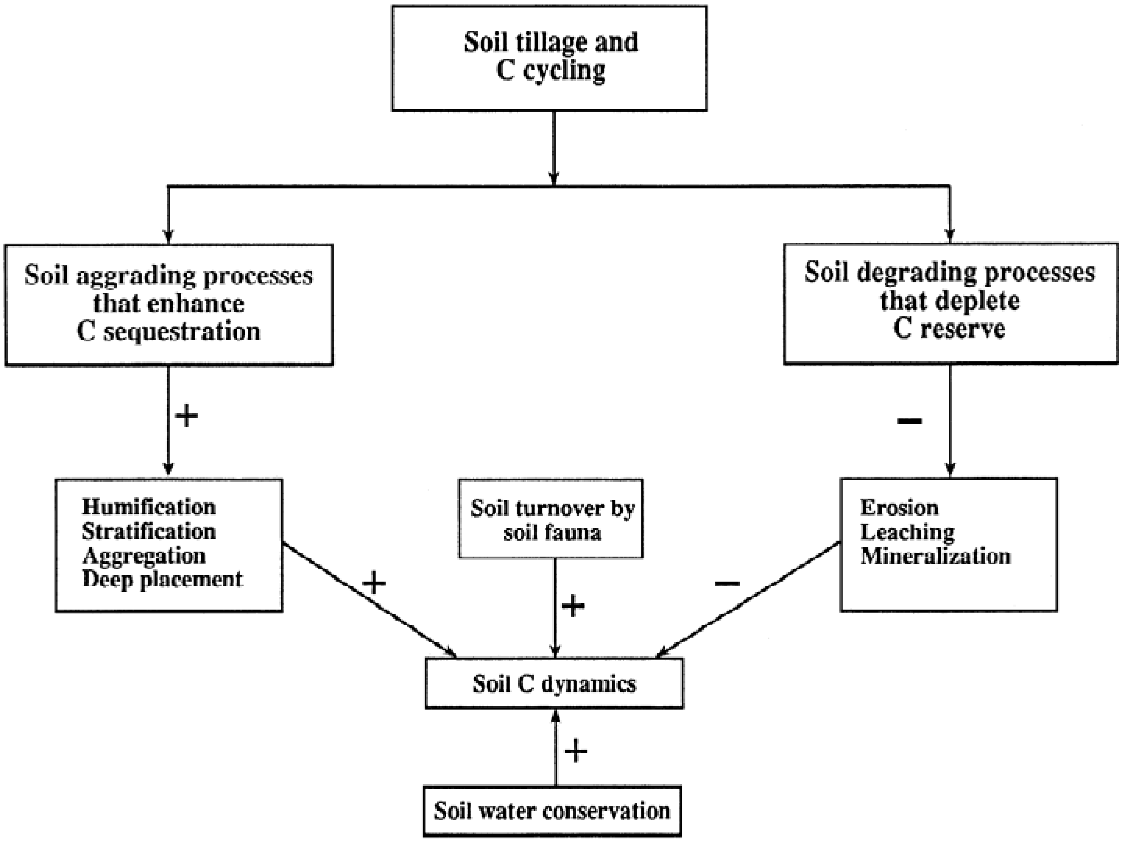

Figure 2. 5: Tillage effects on processes that affect carbon (C) dynamics and reserves in soil (Lal and Kimble, 1997)

2.6 Sampling depth for aggregate stability and soil organic carbon determination

Despite the fact that no studies have been conducted on the factors affecting the structural stability of humic soils in South Africa, some have evaluated the effects of land use on SOM and its relationship to AS and related soil physical properties in other soil types. Dominy and Haynes (2002) studied the effects of land use (grassland, pasture and sugarcane) on SOC content and AS in two profiles of Hutton form (Ferralic Nitisol) (IUSS Working Group WRB, 2014) in the Midlands region of KwaZulu-Natal. The results showed a lower AS under sugarcane than under grassland which was ascribed to the (i) breaking up of soil aggregates by tillage thereby favouring rapid decomposition of SOM and (ii) low above-ground dry matter input in sugarcane soils following pre-harvest burning of crop residues.

It was further indicated that sugarcane soils in the inter-row spaces were predisposed to structural breakdown and compaction as was evidenced by the high bulk density (1.40 Mg m^{-3}) in the 0 – 10 cm layer compared to 1.34 Mg m^{-3} in the 20 – 40 cm layer. Although this study documented important information on the long-term effects of sugarcane mono-cropping on soil physical properties, the sampling depth was limited to 40 cm, which may have underestimated the effect of the SOC and related properties, especially in the subsoil (Jobbágy and Jackson, 2000; Olson and Al-Kaisi, 2015). Selim et al. (2016) found 50 % of the SOC at the 50-100 cm depth under both burnt and unburnt sugarcane trials at the St. Gabriel Sugar Research Station, Louisiana (Figure 2.6). In another study, Skjemstad et al. (1996) reported higher SOC and AS at the 50-100 cm soil depth in some Australian soils.

There is, however, no agreed standardised sampling depth protocols across disciplines as some studies refer to surface samples as the uppermost 10 cm (Six et al., 1998; Celik, 2005) while others use the uppermost 20 cm (Milne and Haynes, 2002; Peng et al., 2015) or 30 cm (Franzluebbers et al., 1999; Chivenge et al., 2007). In general, the upper 10 cm of soil is frequently utilised for forest and grassland soils as most OM is concentrated close to the surface due to the lack of disturbance under these land uses. On the other hand, when investigating cultivated soils, deeper sampling depths such as 0-20 or 0-30 cm are often employed because tillage practices tend to mix the topsoil and subsoil layers together.

Unger (1995) suggested that sampling of the surface soil should be confined to the uppermost 4 cm, as most significant changes in SOC are apparent at that depth and deeper sampling will dilute these effects.

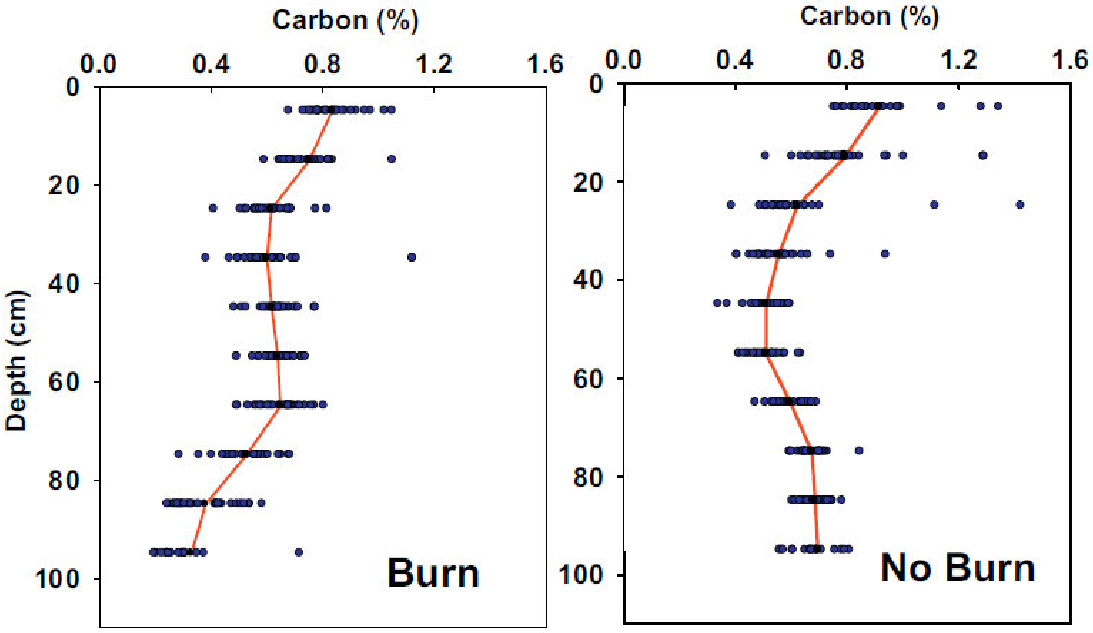

Figure 2. 5: Soil organic carbon variation with depth for burnt and unburnt plots under sugarcane at the St. Gabriel Sugar Research Station, Louisiana. Continuous lines represent averages for each depth (Selim et al., 2016).

When the depth of tillage is, however, sufficient to mix the surface layer with part of the subsoil layer and to a depth below the sampling zone, the SOC-rich surface layer can be incorporated below the shallow sampling zone (Perfect and Kay, 1990; Skjemstad et al., 1996; Jobbágy and Jackson, 2000; Olson and Al-Kaisi, 2015). Generally, the interaction between atmosphere, biosphere, and lithosphere affects the vertical distribution of nutrients in soil resulting in large chemical and physical gradients from surface to bedrock. Therefore, excluding the contribution of lower soil horizons can result in substantial underestimates of total SOC and inaccurate conclusions about the effects of management practices on many soil physicochemical properties (Jobbágy and Jackson, 2000).

2.7 Conclusions

Soil structure is composed of primary particles (sand, silt and clay) and secondary particles which result from the arrangement and binding of primary particles into aggregates by the effect of both organic and inorganic cementing agents. Although several theories of soil aggregate formation exist, the aggregate hierarchy concept is most commonly accepted. The model is based upon the hypothesis that macro-aggregates (> 250 μm) are collections of smaller micro-aggregates (< 250 μm) held together by temporary organic binding agents. Micro-aggregates bind SOC while macro-aggregates contain fresh SOC material.

There is still much uncertainty in the type of relationship that exists between SOC and aggregate stability as there are many other internal factors that play a role. These factors include clay mineralogy, soil texture, oxides and hydroxides of Fe and Al, natural wetting/drying cycles and soil moisture. Many of these factors are affected by different environmental changes. There is good evidence for the positive effect of the Al and Fe oxides on micro-aggregates, although the debate over whether Fe or Al is more effective in aggregation of soil particles is not resolved. Some researchers also believe that the total SOC or SOM may generally not be a good indicator for assessing AS, mainly because the different pools that make up the bulk SOC differ in their physical and chemical properties and hence have specific functions. Besides these internal factors, AS is also influenced by land use systems such as forest, grassland and cropping in a given area. Forest and grassland have high inherent SOM content that supplies plant nutrients and increases soil aggregation. While some forms of cultivation such as no-tillage have been shown to improve the existing stability of aggregates, conventional tillage practices disrupt aggregation thereby exposing aggregates to microbial processes resulting in the disaggregation following the decomposition of SOC.

Although the influence of many factors affecting soil structural stability has been addressed worldwide, no work has been done on the factors affecting the structural stability of humic soils in South Africa, a scenario that warrants further studies. There is a need for comparative studies that directly investigate and compare SOM, soil water content and structural stability between native forest, grassland, and cultivated humic soils in South Africa. Such studies would not only help in identifying the specific differences and similarities in their soil properties, but would also provide useful insights into the long-term effects of land use practices on these soil parameters, as well as help to identify any trends. Understanding these patterns would allow for the development of specific management practises and soil conservation practices. Also with climate change influencing rainfall and temperature regimes, research in this area would enhance understanding of how these soils may respond to future climate scenarios. In light of the lack of research into the structural determinants in humic soils, the current study was designed to establish the relationships between some factors of aggregation in selected humic soils of South Africa. Literature has shown that the stability of aggregates and consequently OC storage in highly weathered bulk soils is determined by several factors including SOM, and Al and Fe. These factors act independently or interact spatially or temporally with each other in their effects on aggregation. This study thus determines the role of SOC, Al and Fe within the aggregates of some humic soils to a depth of one metre under different land use types in KwaZulu-Natal.

CHAPTER 3: DESCRIPTION OF SITES AND LABORATORY PROCEDURES

3.1 Site descriptions

3.1.1 Cedara

The Cedara Agricultural Research Station site (29° 33.399' S; 30° 15.118' E) is located approximately 16 km from Pietermaritzburg and 12 km from Howick (Figure 3.1). The site receives an average annual rainfall of 885 mm and has average temperatures ranging from 19.9 °C in January to 11.3 °C in June (Schulze, 1997). The grassland is located on an undulating landscape position, with slopes up to 16 % and at an altitude of 1018 to 1120 m a.s.l. The soils were classified as Inanda 1200 (Soil Classification Working Group, 2018); Rhodic Ferralsols (IUSS Working Group WRB, 2014) derived from dolerite parent material. The area was planted to Kikuyu grass (*Pennisetum clandestinum*) and had never been fertilized, irrigated or cropped but was burnt regularly.

3.1.2 Karkloof

This site was on a dairy farm (29° 22.722' S; 30° 17.922' E), located approximately 36 km from Pietermaritzburg (Figure 3.1). The area receives an average annual rainfall of 1150 mm and has average temperatures ranging from 19.5 °C in January to 10.9 °C in June (Schulze, 1997). The site is located on a relatively steep landscape position (16-25 % slope), at an altitude that ranges from 1098 to 1165 m a.s.l. The soil type at the site was the same as that at Cedara also with dolerite as the parent material. The area was under Kikuyu grass that was managed as at Cedara.

3.1.3 Eshowe

The third study site was near Eshowe (28° 52.763' S; 31°25.180' E) in northern KwaZulu-Natal (Figure 3.1). The mean annual rainfall is 1109 mm and mean monthly temperatures range from 26.7 °C in January to 15.4 °C in June (Schulze, 1997).

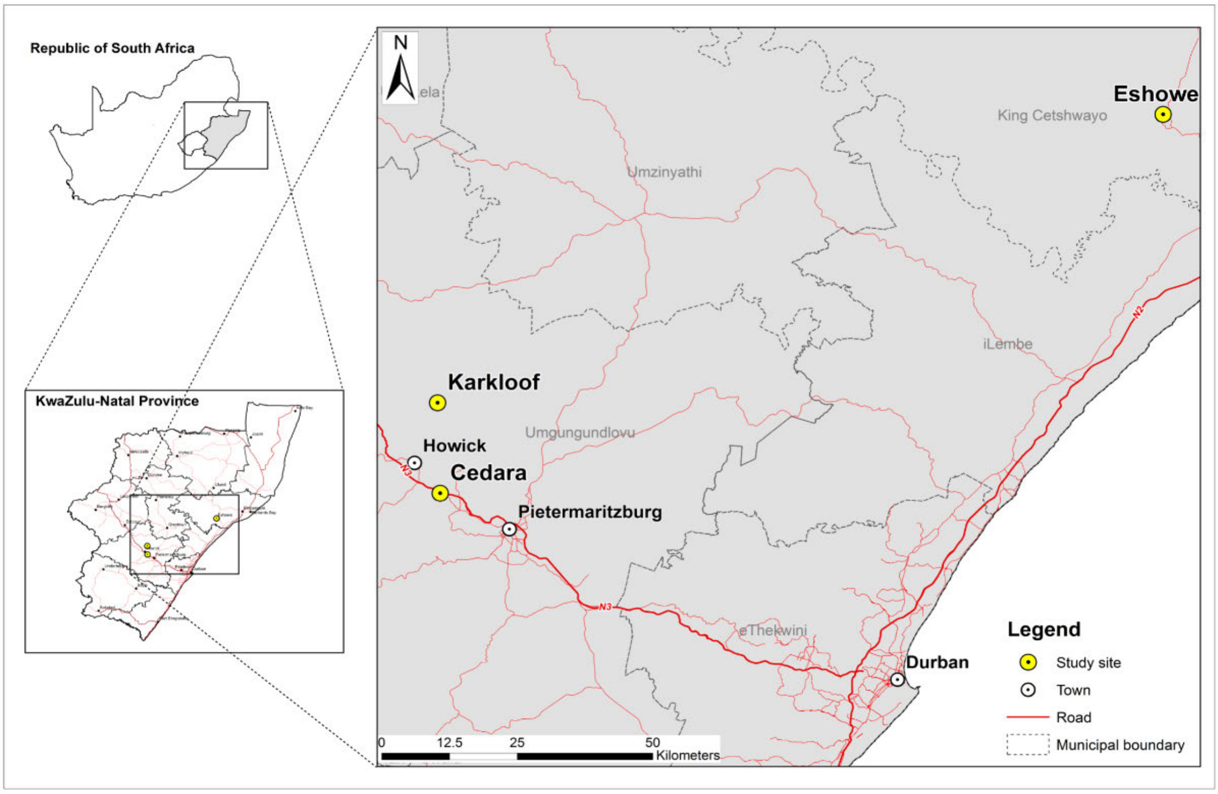

Figure 3. 1: Map showing the location of the study sites within the Province of KwaZulu-Natal, South Africa.

The site is located on a nearly flat landscape (0-2 % slope) at an average altitude of 550 m a.s.l. The soil type was Magwa 2200 (Soil Classification Working Group, 2018); Xanthic Ferralsol (IUSS Working Group WRB, 2014) formed on Natal Group sandstone (Eshowe member) parent material (McBride, 1963). The Eshowe member of this sandstone generally consists of 85-95 % coarse to very coarse-grained, immature, poorly-sorted sandstone with subordinate interbedded reddish micaceous shales, siltstones and unweathered feldspar (Marshall and von Brunn, 1999). The soil supports indigenous coastal scarp forest, and therefore had not received fertilizer, lime or irrigation. The forest was approximately 50 m from the sugarcane field site. The area under commercial sugarcane (*Saccharum officinarum*) had undergone pre-harvest burning for more than 30 years (van Antwerpen and Meyer, 1996). Although, the sugarcane is not irrigated, it is commonly fertilized annually to facilitate growth of the ratoon crop with 5:1:5 (46) at 650 kg ha^{-}

^{1}, approximately 45 days after harvesting (Nxumalo, 2015). Dolomitic lime is also applied (1 to 10 t ha^{-1}) to reduce acid saturation levels to 20 % at least once every 10 years. Sunn-hemp (*Crotalaria juncea*) or oats (*Avena sativa*) are usually planted as rotation crops before replanting sugarcane (van Antwerpen and Meyer, 1996).

3.2 Field sampling protocol

The approach taken in collecting soil samples for the determination of aggregate stability was intentionally designed to minimise disturbance and preserve the integrity of the aggregates, including the macroaggregates. To avoid major disturbance in the field, soil samples were meticulously collected using a spade at each depth rather than an auger. This decision was made to mitigate the shearing effects associated with auger-based sampling method. The use of a spade allowed for more precise control during sampling and ensured that the macroaggregates remained as intact as possible.

Each site was demarcated into three subplots of approximately 0.1 to 0.3 ha that were at least 20 m apart. In each subplot, a 1 x 1 x 1.2 m pit was dug for the collection of bulk soil samples. Three replicate bulk soil samples were collected from the face of each profile pit at depth intervals of 0-5, 5-10, 10-15, 15-20, 20-30, 30-40, 40-50, 50-60, 60-80 and 80-100 cm, giving a total of 90 samples from each site. All the samples were air-dried and about one third of each sample was used for aggregate stability (AS) measurements, while the remainder was ground with a pestle and mortar and passed through a 2 mm sieve. Following Malepfane et al. (2022) the cut-off depth limit of the topsoil was taken to be at 30 cm with material below this point (30-100 cm) termed subsoil.

3.3 Laboratory analysis

3.3.1 pH, particle size distribution and clay mineralogy

Soil pH was determined with a Metrohm E396B meter in 1M KCl at 1:2.5 soil: solution ratio using a glass electrode (Yeomans and Bremner, 1988). The pH of the supernatant was measured after shaking the suspensions for 30 min and equilibrating for a further 30 min (Thomas, 1996). Particle size distribution was determined using the hydrometer method (Bouyoucos, 1962). The pre-treatment of the samples before sedimentation involved the removal of organic matter with hydrogen peroxide. The samples were further dispersed with 10 % Calgon and stirred on a high speed electric stirrer for 3 minutes.

Clay mineralogy was determined by X-ray diffraction (XRD) on Ca-saturated, oriented samples using a Bruker D8 ADVANCE diffractometer equipped with a Lynx Eye detector with Ni-filtered Cu-Kα radiation (λ =0.154 nm) at 40 kV and 40 mA. The air-dried, glycerolated and heated (500 ºC for 3 hours) clay samples were scanned from 2° to 15° 2θ with a scanning step size of 0.01313° at 0.779 s per step (Hillier, 2003).

3.3.2 Aggregate stability

The samples used for AS were sieved to collect sufficient aggregates between 2.8 and 5.0 mm. Aggregate size separation was carried out using the method adapted from Elliott (1986). The method involved separation of aggregates by wet sieving the air-dried soil through a nest of three sieves to isolate four aggregate size classes, which are referred to in this study as i) large macro-aggregates: LM (> 2000 µm), ii) small macro-aggregates: SM (250–2000 µm), iii) micro-aggregates: M (63–250 µm), and iv) silt + clay: SC (< 63 µm). A 100 g sub-sample was evenly spread on top of the 2000 µm sieve, submerged in deionized water at room temperature for 5 min, resulting in slaking of the soil, which was subsequently sieved to separate water-stable aggregates by moving the sieve up and down 50 times over a period of 2 min. The material remaining on the 2000 μm sieve, i.e. the LM, was back-washed into a beaker for drying. Soil plus water that passed through the sieve was poured onto the 250 μm sieve and the sieving procedure repeated. This was repeated for the 63 μm sieve. All aggregate classes were oven dried at 40 °C (48 hours), weighed and stored at room temperature for the analysis of total Al, Fe and C. Mean weight diameter (MWD) was calculated using Equation 1 (Kemper and Rosenau, 1986):

$$MWD = \sum_{i=1}^{n} x_i w_i \qquad \text{……………………………………………………………Equation 1}$$

where: x_i is the mean diameter (mm) of any particular size range of aggregates separated by sieving, w_i is the weight fraction of aggregates remaining on the sieve (%), and n is the number of aggregate classes separated.

3.3.3 Total organic carbon

Total C in the bulk soil samples and the aggregate size fractions was determined on finely ground samples (< 0.5 mm) using the automated Dumas dry combustion method on a LECO CNS 2000 analyser (Matejovic, 1996). Since these soils were acidic and no carbonates were present, the total C measured was considered to represent organic C and is henceforth referred to as total organic C (TOC).

3.3.4 Total aluminium and iron

Total aluminium (Al) and iron (Fe) were determined in both the bulk samples and the soil aggregates. Samples of aggregates from each of the three largest size fractions were manually ground with a pestle and mortar to pass a 63 μm sieve. Each soil sample (2 g) was mixed with 1 g of cellulose flakes as a binder, compressed into a pellet (19 mm diameter) under a pressure of 2 t cm^{-2}, and analysed for total Al and Fe using a polarized energy dispersive X-ray fluorescence spectrometer (X-LAB 2000 PED-XRF) with rhodium as the excitation source.

3.3.5 Exchangeable cations

Exchangeable potassium (K), calcium (Ca) and magnesium (Mg) were extracted with 0.0356 *M* $SrCl_2$ (Manson and Roberts, 2000) and measured in the filtrate by atomic absorption spectrophotometry (AAS, Varian SpectraAA-2600). Exchangeable acidity was measured after extraction with 1 M KCl solution by titration with 0.005 M sodium hydroxide (Hunter, 1974).

3.3.6 Water retention characteristics

From the same pits, located as described in Section 3.2, undisturbed soil cores were taken at the same 10 depth intervals using steel cylinders (5 cm height and 7.8 cm in diameter) with an internal volume of 238.9 cm^3.

The soil cores were prepared and analysed for water content at field capacity (FC) and permanent wilting point (PWP), saturated hydraulic conductivity (Ks), bulk density (BD), and total porosity (TP) using methods described by Klute (1986).

3.3.6.1 Water retention properties

The method for determining water retention properties involved tying a pre-weighed piece of nylon cloth with an elastic band onto the lower end of the soil core that had been trimmed level with the upper and lower surface of the core ring. The samples were saturated by immersing them in water for a sufficient period of time to ensure complete saturation by capillary wetting. When the samples were fully saturated, excess water was removed from the surface of the soil cores with a blotting paper before the wet weight was determined (0 kPa). Dry weight was determined after the soil samples were placed in an oven at a temperature of 105°C to 110°C and dried until a constant weight was reached (24 to 48 hours). The gravimetric water content was calculated by subtracting the dry weight of the soil sample from the wet weight and dividing the result by the dry weight. The soil cores were then placed in different pressure pots and pressures of 10 and 1500 kPa were consecutively applied to mimic equivalent soil potentials at field capacity (FC) and permanent wilting point (PWP), respectively. Water retention at 10 kPa and 1500 kPa was calculated by subtracting the weight of the soil sample after pressure was applied from the dry weight and dividing the result by the dry weight. The available water content (AWC) was calculated as the difference between FC and PWP (Klute, 1986).

3.3.6.2 Saturated hydraulic conductivity

The Ks was determined based on Darcy's law directly after the measurements of water retention using a brass permeameter. The undisturbed soil core sample was supported vertically on the outflow funnel and then water was applied to the top of the permeameter (US Salinity Laboratory Staff, 1954). A fixed head of water was maintained (30 mm) at the top of the permeameter using a Marriott bottle system. The time water was first applied and time taken to percolate through the base was recorded. At regular intervals (± 2 or 3 times a day) the amount of water percolating per unit time was measured. This process was continued until the volume percolating in a fixed time remained constant. The Ks was calculated using Equation 2.

$\text{Ks (cm hr}^{-1}) = [(V/ (A*t)) * (L / \Delta H)]$……………………………………......……….Equation 2

where V is the volume of water (mm) collected for time period of t (minutes), A is the cross sectional area of the core (mm^2), L is the length of the soil core (mm) and ΔH is the hydraulic head drop (mm).

3.3.6.3 Bulk density and total porosity

The soil core samples were oven-dried at 105 ºC for 48 hours and BD was calculated using Equation 3. The TP was calculated using Equation 4.

$\text{BD (g cm}^{-3}) = [\text{mass of oven-dry soil / volume of soil}]$…………..…………………...Equation 3

$\text{TP} = [1 - \text{BD/PD}] * 100\ \%$ ……………………………………………………….Equation 4

where BD is bulk density and PD refers to the particle density assumed to be 2.65 g cm^{-3} (Li et al., 2007).

CHAPTER 4: SOME CHARACTERISTICS OF HUMIC SOILS UNDER NATIVE FOREST AND GRASSLAND

4.1 Introduction

The complex interactions between pH, mineralogy, texture, organic carbon (OC), aluminium (Al) and iron (Fe) in highly leached soils greatly influence the quality, capacity and functioning of these soils (Motavalli et al., 1995). For this reason, many studies on soil quality worldwide including South Africa (e.g. Hartemink, 1998; Dominy and Haynes, 2002; Barthès et al., 2008; Inagaki et al., 2019; Yost and Hartemink, 2019) have often considered these basic soil parameters to measure the influence of pedogenic processes on the profile and because of their sensitivity to variations in management and climatic conditions. There is little information on these qualities in humic soils that are covered with natural vegetation, particularly in South Africa, despite the fact that all these studies have documented very valuable information about soil properties after cultivation of heavily leached soils.

Humic soils play a vital role in soil fertility, water management, carbon sequestration, biodiversity preservation, preventing soil erosion, and sustainable land management in South Africa, according to Macvicar et al. (1984) and Fey (2010). Understanding how ecosystems function, guiding land management and conservation efforts, adapting to climate change, preserving biodiversity, and promoting sustainable agriculture all depend on the study of humic soils in their natural state under various climatic conditions in South Africa (Everson et al., 1998; Everson, 2000). This study was thus designed to evaluate how different site conditions affect general soil characteristics such as clay mineralogy, texture, pH and total Al, Fe and C contents in some humic soils under native forest and grassland.

4.2 Materials and Methods

4.2.1 Description of soils and sites

The bulk soils sampled (Section 3.2) under grassland at Karkloof and Cedara (Sections 3.1.1 and 3.1.2, respectively) and under native forest at Eshowe (Section 3.1.3) were used for this study.

4.2.2 Laboratory analysis

The pH, clay mineralogy, particle size distribution (Section 3.3.1), total organic carbon (TOC; Section 3.3.3), total Al and Fe (Section 3.3.4) and exchangeable bases (Section 3.3.5) in the bulk soils were analysed as described earlier.

4.3 Statistical analysis

Data were analysed using the Genstat 18th edition (VSN International, 2015) by a two-way analysis of variance ($p \leq 0.05$) to test the effects of site, depth, and their interaction. Differences between the means of the significant factor were assessed with Duncan's multiple range test ($p \leq 0.05$). Least significant differences (LSD) at $p = 0.05$ were computed to separate treatment means for all properties. All results were based on three replications in the field.

4.4 Results

4.4.1 Properties of the bulk soil samples

The characteristics of the bulk soils are given in Tables 4.1 and 4.2. The topsoil (0-30 cm) of both grasslands was a thin, dark brown (10YR 2/1 to 3/1), powdery humic A horizon. The subsoil was dark red (5YR 4/4 to 4/6) at Cedara, and red (7.5R 4/6 to 5/6) at Karkloof, with porous, friable structure. At Eshowe, the topsoil was a very dark brown (10YR 2/1 to 2/2), fine, subangular blocky, thick humic A horizon and the subsoil was dark yellowish brown (5YR to 7.5YR 4/6) with apedal to weak structure. The clay mineralogy of the soils at all sites was dominated by kaolinite and quartz. Eshowe samples also contained goethite with subsidiary interlayered chlorite while gibbsite and hematite were only found at Cedara and Karkloof.

The clay content in the average (0-100 cm) was significantly higher at Cedara (62 %) than Karkloof (48 %) and Eshowe (13 %). The clay content increased with depth at Cedara and Karkloof but decreased at Eshowe.

The mean silt content was higher at Karkloof (39 %) than Cedara (21 %) and Eshowe (18 %) ($p \leq 0.05$; Table 4.1). The silt decreased with depth at Cedara and Karkloof while no significant differences were observed with depth at Eshowe. The sand content decreased with depth at Cedara but increased at Karkloof and Eshowe. The average sand content was much higher at Eshowe (70 %) than Cedara (16 %) and Karkloof (13 %).

The TOC significantly decreased with depth at all sites (Table 4.1). The average (0-100 cm) TOC was greater ($p \leq 0.05$) at Karkloof (42 g kg^{-1}) than Eshowe (35 g kg^{-1}) and Cedara (25 g kg^{-1}). Total Al significantly increased with depth at all sites. The average Al was greater ($p \leq 0.05$) at Karkloof (75 g kg^{-1}) than Cedara (67 g kg^{-1}) and Eshowe (25 g kg^{-1}). Similar to Al, total Fe significantly increased with depth at all sites. The average Fe was lower ($p \leq 0.05$) at Eshowe (20 g kg^{-1}) than Karkloof (50 g kg^{-1}) and Cedara (49 g kg^{-1}).

The mean (0-100 cm) pH (KCl) values were acidic at all sites and significantly lower at Karkloof (3.96) than Eshowe (4.20) and Cedara (4.32) (Table 4.2). The concentrations of Ca and Mg were higher at Cedara than the other sites at all depths with the exception of the 0-10 cm depth at Eshowe where it was higher ($p \leq 0.05$; Table 4.2). The average (0-100 cm) Ca was higher ($p \leq 0.05$) at Cedara (2.40 $cmol_c$ kg^{-1}) than Eshowe (1.42 $cmol_c$ kg^{-1}) and Karkloof (0.55 $cmol_c$ kg^{-1}) while Mg was not significantly different between sites. The concentration of K was higher ($p \leq 0.05$) at Cedara than other sites at all depths. The average K followed the order: Cedara (0.95 $cmol_c$ kg^{-1}) > Eshowe (0.52 $cmol_c$ kg^{-1}) = Karkloof (0.01 $cmol_c$ kg^{-1}).

Table 4. 1: Total organic carbon (TOC), aluminium (Al), iron (Fe), particle size distribution and mean weight diameter (MWD) of the humic soil profiles under Eshowe forest, and Cedara and Karkloof grasslands (n = 3).

Site	Depth (cm)	TOC	Al	Fe	Clay (< 0.002 mm)	Silt (0.002-0.05 mm)	Sand (0.05-2 mm)	Texture class	MWD (mm)
		……. ($g\ kg^{-1}$)………							
					……………….(%)……….....				
Cedara	0-5	63.2^{mn}	58.8^{gh}	45.4^{hij}	57.1^{hij}	26.2^{cdef}	16.7^{bc}	C	1.4^{efg}
Karkloof		96.5^{p}	55.2^{f}	35.0^{f}	43.9^{def}	40.1^{hi}	16.0^{defgh}	SC	1.8^{kl}
Eshowe		71.7^{no}	20.5^{a}	17.6^{a}	16.5^{abc}	19.2^{bcde}	64.3^{j}	SCL	1.5^{ghij}
Cedara	5-10	57.2^{lm}	59.6^{gh}	44.7^{h}	52.3^{ghi}	29.3^{fg}	12.3^{cd}	C	1.3^{ef}
Karkloof		77.7^{o}	57.5^{fg}	38.0^{g}	47.0^{defgh}	44.7^{i}	8.5^{ab}	SC	1.7^{l}
Eshowe		49.1^{kl}	19.6^{a}	17.4^{a}	12.5^{ab}	24.6^{ef}	62.9^{j}	SCL	1.5^{ghij}
Cedara	10-15	30.4^{fghi}	60.7^{h}	44.5^{h}	59.3^{jkl}	23.7^{def}	17.0^{efgh}	C	1.4^{efg}
Karkloof		70.9^{no}	67.8^{j}	44.5^{h}	46.3^{defgh}	45.0^{i}	6.3^{a}	SC	1.8^{kl}
Eshowe		40.8^{ijk}	22.1^{a}	19.4^{abc}	16.5^{bc}	20.5^{bcde}	63.0^{j}	SCL	1.4^{fgh}
Cedara	15-20	23.2^{defgh}	63.7^{i}	45.1^{hi}	57.3^{ijk}	21.4^{bcde}	21.3^{i}	C	1.3^{efg}
Karkloof		51.9^{kl}	73.1^{k}	54.3^{m}	48.7^{efgh}	45.1^{i}	6.2^{a}	SC	1.7^{jkl}
Eshowe		35.3^{hij}	19.9^{a}	18.5^{ab}	11.2^{ab}	16.5^{abcd}	72.3^{k}	SCL	1.1^{abc}
Cedara	20-30	19.4^{cdef}	63.6^{i}	47.0^{ij}	65.0^{l}	16.4^{abcd}	18.7^{fghi}	C	1.3^{a}
Karkloof		45.0^{jk}	87.3^{n}	62.3^{p}	50.3^{fghi}	43.3^{i}	6.3^{a}	SC	1.7^{ijkl}
Eshowe		30.1^{fghi}	25.2^{b}	17.7^{a}	14.4^{abc}	11.8^{a}	73.8^{k}	SCL	1.1^{abc}
Cedara	30-40	17.5^{cde}	69.7^{j}	47.4^{j}	64.3^{l}	22.4^{cde}	14.5^{cde}	C	1.4^{efg}
Karkloof		32.1^{ghi}	83.1^{lm}	60.2^{o}	51.0^{fghi}	34.0^{gh}	15.0^{cdefg}	SC	1.7^{kl}
Eshowe		28.3^{efgh}	26.5^{bc}	20.1^{bcd}	11.9^{ab}	17.2^{abcde}	70.9^{k}	SCL	1.0^{ab}
Cedara	40-50	14.4^{bcd}	75.5^{k}	50.1^{k}	56.0^{ijk}	20.4^{bcde}	14.5^{cde}	C	1.4^{efg}
Karkloof		23.6^{defgh}	82.0^{l}	49.9^{k}	40.4^{d}	42.6^{i}	17.2^{efgh}	SC	1.8^{kl}
Eshowe		26.6^{efgh}	30.7^{de}	21.3^{cd}	13.8^{abc}	14.5^{ab}	71.7^{k}	SCL	0.9^{a}
Cedara	50-60	12.5^{bcd}	68.6^{j}	45.9^{hij}	61.0^{kl}	19.8^{bcde}	18.5^{efgh}	C	1.4^{efg}
Karkloof		12.0^{abcd}	83.7^{lm}	51.3^{kl}	45.8^{defg}	39.7^{hi}	14.7^{cdef}	SC	1.8^{kl}
Eshowe		26.9^{efgh}	28.9^{cd}	17.7^{a}	9.9^{ab}	16.5^{abcd}	73.6^{k}	SCL	1.0^{ab}
Cedara	60-80	10.5^{abc}	74.5^{k}	57.8^{n}	72.9^{m}	14.6^{ab}	12.5^{cd}	C	1.5^{fgh}
Karkloof		4.7^{ab}	85.4^{mn}	52.4^{lm}	41.8^{de}	39.0^{hi}	19.3^{hi}	SC	1.8^{kl}
Eshowe		21.1^{cdefg}	29.7^{de}	21.9^{de}	11.9^{ab}	15.8^{abc}	72.3^{k}	SCL	1.1^{bcd}
Cedara	80-100	5.7^{ab}	75.4^{k}	57.9^{n}	74.1^{m}	14.6^{ab}	12.6^{cd}	C	1.5^{fgh}
Karkloof		0.62^{a}	84.6^{lm}	53.2^{lm}	60.9^{j}	18.5^{abcde}	20.6^{i}	C	1.6^{hijk}
Eshowe		18.4^{cdef}	31.6^{e}	23.7^{e}	9.2^{a}	18.5^{abcde}	72.3^{k}	SCL	1.3^{cde}
Cedara	0-100	24.8^{a}	66.8^{b}	48.9^{b}	62.0^{c}	21.0^{a}	16.0^{a}	C	1.38^{a}
Karkloof		41.8^{b}	75.1^{b}	50.0^{b}	47.8^{b}	39.1^{b}	13.3^{a}	SC	1.75^{b}
Eshowe		34.8^{c}	25.0^{a}	19.3^{a}	12.9^{a}	17.7^{a}	69.7^{b}	SCL	1.20^{c}

C: clay; SC: silty clay; SCL: sandy clay loam. The MWD is discussed in detail in Chapter 6, however it is included here for completeness.

Means within the same column followed by different letters are significantly different ($p \leq 0.05$).

Table 4. 2: Exchangeable calcium (Ca), magnesium (Mg), potassium (K), acidity and pH of the humic soil profiles under Eshowe forest, and Cedara and Karkloof grasslands (n = 3).

Site	Depth (cm)	Ca	Mg	K	Exch. Acidity	pH (KCl)
		……………………	($\mathrm{cmol_c\ kg^{-1}}$)…	……	…………………	
Cedara		2.61^{r}	1.81^{t}	1.37^{u}	0.97^{de}	4.05^{abcdef}
Karkloof	0-5	0.88^{i}	0.58^{ij}	0.15^{i}	3.35^{l}	3.94^{abcde}
Eshowe		5.43^{u}	4.57^{w}	0.51^{l}	0.27^{ab}	4.75^{h}
Cedara		2.53^{q}	1.74^{s}	1.13^{s}	1.36^{f}	4.03^{abcdef}
Karkloof	5-10	0.56^{e}	0.38^{b}	0.15^{i}	3.46^{l}	3.79^{a}
Eshowe		2.66^{s}	2.46^{v}	0.29^{k}	0.94^{de}	4.18^{bcdef}
Cedara		2.49^{p}	1.70^{r}	1.19^{t}	1.09^{e}	4.14^{abcdef}
Karkloof	10-15	0.88^{i}	0.60^{j}	0.10^{fg}	3.52^{l}	3.94^{abcde}
Eshowe		1.16^{k}	1.44^{q}	0.17^{j}	1.80^{ghi}	4.05^{abcdef}
Cedara		2.88^{t}	1.79^{t}	1.01^{r}	0.83^{cd}	4.29^{defg}
Karkloof	15-20	0.78^{h}	0.45^{d}	0.07^{bc}	2.92^{k}	3.92^{abcd}
Eshowe		0.93^{j}	0.97^{m}	0.14^{h}	1.94^{i}	4.03^{abcdef}
Cedara		2.48^{p}	1.42^{q}	0.85^{p}	0.88^{d}	4.16^{abcdef}
Karkloof	20-30	0.77^{h}	0.50^{ef}	0.09^{e}	2.82^{jk}	4.06^{abcdef}
Eshowe		0.73^{g}	0.76^{l}	0.13^{h}	1.89^{hi}	4.14^{abcdef}
Cedara		2.48^{p}	1.16^{n}	0.74^{m}	0.91^{de}	4.19^{bcdef}
Karkloof	30-40	0.77^{h}	0.52^{fg}	0.09^{ef}	2.82^{jk}	3.98^{abcde}
Eshowe		0.67^{f}	0.68^{k}	0.10^{g}	1.93^{i}	4.13^{abcdef}
Cedara		2.27^{o}	1.18^{n}	0.84^{p}	0.86^{d}	4.22^{cdef}
Karkloof	40-50	0.17^{a}	0.22^{a}	0.08^{cd}	1.95^{i}	3.84^{ab}
Eshowe		0.54^{e}	0.54^{gh}	0.09^{ef}	1.70^{gh}	4.13^{ab}
Cedara		2.24^{n}	1.24^{o}	0.81^{o}	0.65^{c}	4.34^{fg}
Karkloof	50-60	0.26^{b}	0.49^{ef}	0.09^{ef}	2.63^{j}	4.02^{abcdef}
Eshowe		0.37^{c}	0.42^{cd}	0.06^{ab}	1.68^{g}	4.14^{abcdef}
Cedara		2.20^{m}	1.30^{p}	0.92^{q}	0.35^{ab}	4.57^{gh}
Karkloof	60-80	0.25^{b}	0.48^{e}	0.08^{de}	2.91^{k}	3.89^{abc}
Eshowe		0.39^{d}	0.58^{ij}	0.07^{bc}	1.39^{f}	4.14^{abcdef}
Cedara		2.13^{l}	1.99^{u}	0.75^{n}	0.11^{a}	5.15^{i}
Karkloof	80-100	0.19^{a}	0.40^{bc}	0.07^{bc}	2.84^{k}	4.19^{bcdef}
Eshowe		0.37^{c}	0.50^{ef}	0.06^{a}	1.47^{f}	4.31^{efg}
Cedara		2.40^{b}	1.53^{a}	0.95^{b}	0.80^{a}	4.32^{a}
Karkloof	0-100	0.55^{a}	0.46^{a}	0.01^{a}	2.92^{b}	3.96^{b}
Eshowe		1.42^{ab}	1.29^{a}	0.52^{a}	1.50^{a}	4.20^{a}

Means within the same column followed by different letters are significantly different ($p \leq 0.05$)

The exchangeable acidity was significantly higher at Karkloof than the other sites at all depths while at Eshowe the concentration was higher than at Cedara in the 15-100 cm depth. Similar to the average Al concentrations, the average exchangeable acidity was greater ($p \leq 0.05$) at Karkloof (2.92 $\text{cmol}_c\ \text{kg}^{-1}$) than Eshowe (1.50 $\text{cmol}_c\ \text{kg}^{-1}$) and Cedara (0.80 $\text{cmol}_c\ \text{kg}^{-1}$).

4.4.2 Correlation between pH, particle size distribution, organic carbon, aluminium and iron

Table 4.3 gives the correlation coefficients between some of the bulk soil properties at 0-30 cm (topsoil), 30-100 cm (subsoil) and 0-100 cm (whole soil profile) depths across the studied sites. The TOC correlated positively with silt ($r = 0.63$) in the top 30 cm but negatively with Fe ($r = -0.53$), Al ($r = -0.55$) and clay ($r = -0.52$) in the 30-100 cm depth. The Al correlated positively with clay and silt and negatively with sand at all depths. The Fe correlated positively with clay and Al at all depths and silt ($r = 0.50$) at 30-100 cm depth but negatively with sand at all depths. The sand correlated negatively with clay and silt at all depths.

Table 4. 3: Pearson's correlation coefficient between some bulk soil properties at 0-30 cm (topsoil), 30-100 cm (subsoil) and 0-100 cm (whole soil profile) depths across the three sites (Eshowe forest, Cedara and Karkloof grasslands) (n =3).

Parameter	pH	Clay	Silt	Sand	TOC	Al	Fe
				Topsoil (0-30 cm)			
pH	1.00						
Clay	-0.21	1.00					
Silt	-0.42	0.39	1.00				
Sand	0.35	**-0.91**	**-0.70**	1.00			
TOC	-0.20	-0.01	**0.63**	-0.29	1.00		
Al	-0.39	**0.76**	**0.50**	**-0.78**	0.05	1.00	
Fe	-0.35	**0.74**	0.39	**-0.73**	-0.02	**0.98**	1.00
				Subsoil (30-100 cm)			
pH	1.00						
Clay	0.38	1.00					
Silt	-0.44	0.15	1.00				
Sand	-0.12	**-0.91**	**-0.50**	1.00			
TOC	-0.31	**-0.52**	-0.23	0.45	1.00		
Al	0.01	**0.81**	**0.65**	**-0.97**	**-0.55**	1.00	
Fe	0.17	**0.89**	**0.50**	**-0.97**	**-0.53**	**0.97**	1.00
				Whole soil profile (0-100 cm)			
pH	1.00						
Clay	0.18	1.00					
Silt	-0.43	0.25	1.00				
Sand	0.06	**-0.91**	**-0.61**	1.00			
TOC	-0.28	-0.16	0.32	-0.04	1.00		
Al	-0.08	**0.78**	**0.53**	**-0.85**	-0.26	1.00	
Fe	0.01	**0.82**	0.42	**-0.84**	-0.24	**0.97**	1.00

TOC: total organic carbon; Al: total aluminium; Fe: total iron. Correlation coefficients in bold are statistically significant at $p \leq 0.05$.

4.5 Discussion

At all study sites (Table 4.2), the pH was found to be acidic, primarily due to the decomposition of organic matter and respiration. In many terrestrial ecosystems, the surface litter and upper soil layers contain a significant amount of organic matter and roots. During the process of decomposition and respiration, high concentrations of carbon dioxide (CO_2) are released, leading to the formation of carbonic acid (H_2CO_3) in the soil water. This carbonic acid contributes to the overall soil acidity, as noted by Wilson (1999). The process of nutrient uptake by trees and other plants may have also contributed to the observed acidity at the studied sites. In general, as plants absorb essential nutrients like Ca, Mg, and K, they excrete hydrogen ions, which can reduce the soil's buffering capacity and result in increased acidity (Vazquez, 1981; Zhang and Horn, 2001). Additionally, the acidity status at all sites could be due to leaching of basic cations. However, there was an increase of basic cations with depth at all sites (Table 4.2), which could be a result of the weathering of minerals from the parent rock containing Ca, Mg, and K, leading to the release of these elements into the soil solution and their subsequent accumulation at depth (Jobbágy and Jackson, 2000; Olson and Al-Kaisi, 2015).

Although the clay mineralogy of the soils reflected their highly weathered nature (Watanabe et al., 2006), the absence of gibbsite at Eshowe was not expected considering the high rainfall at this site. The formation of gibbsite is, nevertheless, difficult to establish as it may form in a soil environment but then be transformed back into kaolinite when in contact with silica-rich solutions, or may be subject to dissolution with solubility increasing with decreasing pH (Vazquez, 1981; Wilson, 1999). As expected, the dolerite-derived soils at Cedara and Karkloof had greater clay and silt content than the sandstone-derived soils at Eshowe. The higher clay content at Cedara than Karkloof probably reflects the topographical features of the two sites (Wakindiki and Ben-Hur, 2002). The low inclination and concave slope favours water accumulation and clay deposition at Cedara, whereas steeper slopes and convex curvature at

Karkloof could have encouraged clay loss through erosion (Mills and Fey, 2004; Wiesmeier et al., 2019).

The decrease in rainfall from Karkloof (1150 mm) to Eshowe (1109 mm) and Cedara (885 mm) was accompanied by a decrease in the average (0-100 cm) TOC content from 42 to 35 and 25 g kg^{-1}, respectively. The higher rainfall at Karkloof and Eshowe could have caused greater soil acidification and reduced SOM decomposition (Chaplot et al., 2010; Chaplot et al., 2011; Wiesmeier et al., 2019) compared to Cedara. This is supported by the slightly lower mean pH, exchangeable Ca and K and higher exchangeable acidity at Karkloof and Eshowe than Cedara (Table 4.2). Similar results to these findings have been reported by several studies all around the world. For instance, Li et al. (2016) performed a meta-analysis of 142 studies on grassland soils (0-50 cm depth) with a wide range of textures and average annual air temperatures between 6.9 and 28.4 °C and found greater TOC in soils in areas with rainfall and temperature greater than 500 mm and 15 °C, respectively. A recent study by Mureva et al. (2018) compared TOC (0-100 cm) in grassland soils along precipitation (300 to 1500 mm) and temperature (-7.2 to 37.5 °C and 3.5 to 35 °C) gradients in South Africa. They found higher TOC at the high rainfall (1500 mm) than at the driest sites (300–350 mm). In well drained, Nordic forest soils (0-100 cm), Callesen et al. (2003) reported a positive linear relationship between TOC and both mean annual precipitation (200-1100 mm) and temperature (-2 to 8.5 °C) with a greater increase in coarse than medium textured soils. With the exception of the study by Mureva et al. (2018) where no effect of temperature was mentioned, all these studies concluded that higher annual precipitation and temperature under native ecosystems enhance net primary productivity and input of litter to the TOC pool, provided that nutrients are not limiting growth. In general, the greater TOC content near the surface in bulk soils at all sites stems from the surface layer being fed with fresh litter from aboveground biomass (Feller and Beare, 1997; Mills and Fey, 2004; Dlamini et al., 2016). In addition, the greater TOC at 40-60 cm depth at Eshowe could be due to the translocation of C to greater depths in the coarser textured soils compared to the clayey grassland soils that may have more limited C movement down the profile (Carvalho et al., 2017; Malepfane et al., 2022). Another possible reason relates to the allocation of the assimilated C below ground by tree root biomass and turnover (Callesen et al., 2003; Blanco-Canqui and Lal, 2004).

The higher Al and Fe concentrations in soils at Karkloof and Cedara is consistent with a number of studies done on dolerite-derived soils from different regions (Vazquez, 1981; Wilson, 1999; Castro et al., 2002; Watanabe et al., 2006; Igwe et al., 2009). All these studies have attributed the higher concentration of these elements to the presence of gibbsite, goethite and hematite as a result of weathering of the dolerite parent material. The positive correlations between clay, and Al and Fe at all depths, silt and Al at all depths, and between silt and Fe in the 60-100 cm depth is a function of the higher concentration of the Al and Fe in the finer fractions of the soils (Ko¨gel-Knabner et al., 2008; Kleber et al., 2015; Rasmussen et al., 2018).

The positive correlation between clay and Al at all depths is also a function of the clay minerals containing Al (Vazquez, 1981; Wilson, 1999). The lack of correlation between silt and Fe in the top 30 cm, while the correlation was significantly positive (r=0.50) in the 30-100 cm depth is, however, contrary to the findings of Igwe et al. (2009) who showed a positive relationship between these variables in the topsoil (r=0.58) with no significant differences in the subsoil. The difference in the results may perhaps be reflecting differences in the parent material of the two studies. For instance, the chemical weathering of the shale-derived soils of Nigeria would result in lower Fe in the subsoil compared to the dolerite-derived soils in the present study (Wakindiki and Ben-Hur, 2002; Watanabe et al., 2006). In the 30-100 cm depth, lower TOC correlated negatively with increasing Al (r= -0.55) and Fe (r= -0.53) contents as a function of depth (Bronick and Lal, 2005) in these highly leached soils.

4.6 Conclusions

The main objective of this study was to assess the effects of different site conditions on general soil properties, including clay mineralogy, texture, pH, and total Al, Fe, and C in some humic soils under native forest and grassland. The humic soils studied across different sites exhibited an acidic nature due to extensive leaching of basic cations and their replacement by acidic cations on the exchange complex. Kaolinite and quartz were found to be the dominant minerals in the clay fraction across all sites, indicating their prevalence in the soil composition. Cedara and Karkloof sites exhibited higher mean clay and silt contents (0-100 cm) compared to Eshowe, while the opposite trend was observed for the sand fraction. This difference can be attributed to variations in parent materials. In the soils at Karkloof and Eshowe, the average TOC was greater compared to Cedara. This indicates that SOM accumulation is related to climate, particularly the higher precipitation in Karkloof and Eshowe.

The higher average Al and Fe contents were accompanied by an increase in clay, silt, and a decrease in sand content under grasslands compared to the forest soil, largely as a result of the different parent materials. When creating land management strategies, it is critical to take site-specific soil qualities into account. This is illustrated by the differences in soil parameters analysed. While addressing soil acidity, appropriate soil amendments techniques can contribute to sustainable land management at all the studied sites. For example, at Cedara land management practices that promote the addition of OM, such as organic amendments, may contribute to sustainable soil management. Studying the effects of site and land use characteristics on the bulk density and water-related properties of these humic soils is, however, necessary for long-term agricultural production and environmental stewardship.

CHAPTER 5: BULK DENSITY AND SOME WATER-RELATED PROPERTIES OF HUMIC SOILS AS AFFECTED BY SITE AND LAND USE FACTORS

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5.1 Introduction

Soil physical properties are important in determining the availability of oxygen, and the ease of (i) movement of water and (ii) root penetration in soils (Shepherd et al., 2001; Abu, 2013). The texture and structure of the soil control water dynamics through their effects on pore size distribution, continuity and tortuosity (Bronick and Lal, 2005; Villarreal et al., 2020). Soil bulk density (BD) and porosity, which are both directly or indirectly influenced by soil organic matter (SOM) and other physico-chemical properties, also affect the dynamics of water and aeration (Bronick and Lal, 2005). Increasing soil organic carbon (SOC) generally improves aggregation, decreases BD, and increases total porosity (TP), which affect saturated hydraulic conductivity (Ks), field capacity (FC), permanent wilting point (PWP) and available water capacity (AWC) in the soil (Kay and VandenBygaart, 2002; Blanco-Canqui and Lal, 2004). In general, coarse textured soils with low OM have low TP and a large proportion of macro-pores that result in high Ks. In contrast, fine textured soils with high OM often have high TP, with a smaller proportion of macro-pores, and the Ks is often lower (Cousin et al., 2003; Celik, 2005; Villarreal et al., 2020).

The studies of BD and water-related properties and their dependence on cultivation have often produced contrasting results. For instance, Hammad and Dawelbeit (2001) reported that deforestation and subsequent sugarcane cultivation practices in a semi-arid area of Sudan resulted in no significant differences in BD in the surface layer (0-30 cm) of a clayey soil. At the same site, Mubarak et al. (2005) found lower BD in soil under long-term (> 40 years; 1.5 Mg m^{-}

^{3}) sugarcane cultivation subjected to pre-harvest burning, with synthetic fertilizer and organic compost, than under both short-term cultivation (< 10 years; 1.8 Mg m^{-3}) and native forest (1.7 Mg m^{-3}). A recent study in northeast China by Li et al. (2021), found that conversion of Mollisols from natural forest to soybean (*Glycine max*) cultivation decreased Ks from 2.2 to 0.8 cm hr^{-1} in the 0-15 cm depth but increased it from 0.1 to 0.6 cm hr^{-1} at 50-100 cm depth, with no differences between land uses in the 15-50 cm depth. The reduction of Ks in the top 15 cm of the cropland soils was attributed to the higher surface BD that supported low TP while the increase of Ks in the 50-100 cm depth was associated with greater SOM at this depth following 65 years of chemical fertilizer application. Such contradictory findings suggest that the underlying processes that affect the response of soil physical properties following conversion of native forest are site-specific (Li et al., 2007; Strudley et al., 2008) and that they vary with soil properties, local climate and management (Shepherd et al., 2001; Kargas and Londra, 2015).

Furthermore, measurements of chemical and physical soil properties from native forest and grassland have been shown by many studies to be largely affected by the type of vegetation (Celik, 2005; Price et al., 2010; Agnese et al., 2011). In these studies, soils under natural forest appear to have lower BD and higher total organic carbon (TOC), Ks, FC and PWP than do soils under native grassland. This is mainly because soils under native forest are generally associated with diverse microbial communities that assist in the formation of stable aggregates, promoting better soil structure, water movement and reduced compaction compared to grassland soils (Leifeld et al., 2005; Kar et al., 2017). Apart from land use change studies, it is expected that native ecosystems may also have substantial variation in various water-related properties including BD due to different species composition, diversity and above and below-ground biomass (Everson, 2000; Snyman and du Preez, 2005; Agnese et al., 2011). Climate can also impose constraints on the processes that control BD, which may result in different changes of BD and other water-related properties under different environmental conditions (Evrendilek et al., 2004; Price et al., 2010). Such variability across soils and climates requires the need for better understanding of BD and other water-related properties under different native ecosystems and climates for the development of specific conservation initiatives for each vegetation type.

The sugar industry covers an area of approximately 380 000 ha and produces 2.2 million tonnes of sugar per year in South Africa (Nxumalo, 2015; Rhodes et al., 2018). Approximately 70 % of the sugarcane produced in South Africa is grown under rain-fed conditions on land formerly under native forest (van Antwerpen and Meyer, 1996), with a "significant proportion being grown on sandy loam humic soils derived from sandstone" (Meyer and van Antwerpen, 2010). These humic soils are well drained, with low base status and high OC (>1.8 %). Malepfane et al. (2022) found that conversion from forest to sugarcane plantation decreased SOC in the top 10 cm (from 6.1 % to 4.6 %) although the effects of this change on the BD and water-related properties of the different profile layers of these soils are not known. It is, however, important to understand the BD and water-related dynamics in relation to sugarcane cultivation practices to ensure sustainable management of both soil and water conservation. On the other hand, natural forest and grasslands cover approximately 0.56 and 29 % of South Africa's land mass, respectively (Everson, 2000). Despite clear evidence of human impact on plant-water use, biomass production (energy) and water use efficiency under native forest and grasslands in South Africa (Everson et al., 1998; Everson, 2000; Gush and Dye, 2009), very little research has been done on soil BD and other water-related properties in these ecosystems. Comparing natural ecosystems on similar soil type is important because it could provide information on different effects of distinct native ecosystems on hydrological processes (Everson, 2000).

The objectives of this study were thus to investigate (i) the effect of sugarcane cultivation compared to native forest on BD and some water-related properties of a sandy clay loam humic soil in northern KwaZulu-Natal, South Africa, and (ii) how different site conditions affect BD and some selected water-related properties in humic soils under native forest and grassland.

Such a study will (i) assist in the prediction of the effect of conversion to sugarcane production on the soil physical properties and the likelihood of successful cropping, especially in the drought-prone Zululand region of South Africa (SASA, 2015; Xulu et al., 2018) and (ii) provide insights on factors affecting soil physical properties other than human activities.

5.2 Materials and Methods

5.2.1 Description of sites

The same sites and soils described in Section 3.1 and Table 4.1 were used in this study. Some of the basic soil characteristics of the sugarcane site at Eshowe in northern KwaZulu-Natal are given in Table 5.1.

5.2.2 Field sampling protocol

The collection and analysis of soil samples was done following the protocols described in Sections 3.2 and 3.3.6, respectively.

5.2.3 Statistical analysis

Data were analysed with Genstat 18th edition (VSNInternational, 2015) by two-way analysis of variance ($p \leq 0.05$) to test the effects of (i) land use, soil depth, and their interaction between native forest and sugarcane at Eshowe (ii) site, soil depth, and their interaction between Eshowe native forest, Karkloof and Cedara grasslands. Duncan's multiple range test was used for multiple comparisons between the means at $p \leq 0.05$ level. Least significant differences (LSD) at $p = 0.05$ were also computed to separate treatment means for all the measured parameters. Pearson’s correlation coefficients (r) were determined for the correlation matrix of BD, TP, Ks, FC, PWP, AWC, TOC, clay, silt and sand for the 0-30 cm (topsoil), 30-100 cm (subsoil) and 0-100 cm (whole profile) depths. All results were based on the three replications collected at each depth in the field.

5.3 Results

5.3.1 Effects of sugarcane cultivation relative to native forest on some physical properties of humic soils

5.3.1.1 Basic properties of forest and sugarcane soils

The topsoil (0-30 cm) of the studied sites was a very dark brown (10YR 2/1 to 2/2), fine, subangular blocky humic A horizon, and the subsoil (30-100 cm) was dark yellowish brown (5YR to 7.5YR 4/6) with apedal to weak structure. The clay mineralogy of the soils under both land uses was dominated by kaolinite, with subsidiary interlayered chlorite, goethite and quartz. There were no significant differences in pH between land uses except that soil under forest had a higher pH in the 0-5 cm depth, and lower in the 60-80 and 80-100 cm depths, than that under sugarcane (Table 5.1).

There were no significant differences in sand content between land uses, except that in soil under sugarcane it was significantly ($p \leq 0.05$) higher in the 0-15 cm depth and lower at 50-60 and 80-100 cm than under forest (Table 5.1). The silt content under forest was significantly ($p \leq 0.05$) higher at 0-20, 30-40 and 60-80 cm but was lower at 80-100 cm than sugarcane, while no significant differences were observed at other depths (Table 5.1). The clay content was significantly higher under sugarcane than under forest, except at 0-5, 10-15 and 20-30 cm, where there were no differences.

The topsoil (0-30 cm) averaged 67 and 77 % sand, 19 and 7 % silt, and 14 and 16 % clay under forest and sugarcane, respectively. The subsoil under forest and sugarcane had average values of 72 and 68 % sand, 17 and 14 % silt, and 11 and 18 % clay, respectively. The overall (0-100 cm) textural class of the soils was Sandy Clay Loam.

Table 5. 1: pH and particle size distribution of humic soil profiles under native forest and sugarcane (n =3).

Land use	Depth	pH	Clay	Silt	Sand
	(cm)	(KCl)	(< 0.002 mm)	(0.002-0.05 mm)	(0.05-2 mm)
			………………(%)………………..		
Forest	0-5	4.75^{g}	16.5^{ghi}	19.2^{hjk}	64.3^{b}
Sugarcane		4.08^{ab}	16.5^{ghi}	3.2^{a}	80.3^{h}
Forest	5-10	4.18^{abcd}	12.5^{cde}	24.6^{l}	62.9^{b}
Sugarcane		4.09^{ab}	16.5^{ghi}	2.5^{a}	80.9^{h}
Forest	10-15	4.05^{ab}	16.5^{ghi}	20.5^{hjk}	63.0^{b}
Sugarcane		4.11^{abc}	15.8^{fgh}	9.9^{bc}	74.3^{fg}
Forest	15-20	4.03^{a}	11.2^{abc}	16.5^{efghi}	72.3^{defg}
Sugarcane		4.18^{abcd}	16.6^{ghi}	8.5^{b}	74.9^{g}
Forest	20-30	4.14^{abc}	14.4^{efg}	11.8^{bcd}	73.8^{efg}
Sugarcane		4.25^{cde}	14.6^{efg}	12.5^{cde}	72.9^{efg}
Forest	30-40	4.13^{abc}	11.9^{bcd}	17.2^{fghij}	70.9^{cde}
Sugarcane		4.16^{abc}	15.2^{fg}	13.2^{cde}	71.6^{cdef}
Forest	40-50	4.13^{abc}	13.8^{def}	14.5^{def}	71.7^{cdef}
Sugarcane		4.19^{bcd}	18.5^{ij}	12.6^{cde}	68.9^{c}
Forest	50-60	4.14^{abc}	9.9^{ab}	16.5^{efgh}	73.6^{efg}
Sugarcane		4.08^{ab}	17.8^{hij}	13.1^{cde}	69.1^{c}
Forest	60-80	4.14^{abc}	11.9^{bcd}	15.8^{efg}	72.3^{defg}
Sugarcane		4.34^{e}	19.9^{j}	10.5^{bcd}	69.6^{cd}
Forest	80-100	4.31^{de}	9.2^{a}	18.5^{ghij}	72.3^{defg}
Sugarcane		4.52^{f}	17.9^{hij}	22.5^{kl}	59.6^{a}

Means within the same column followed by different letters are significantly different ($p \leq 0.05$).

5.3.1.2 Bulk density and total porosity

The BD was higher ($p \leq 0.05$) under sugarcane (1.31 g cm^{-3}) than forest (0.93 g cm^{-3}) in the 0-5 cm layer, while in the 20-50 and 80-100 cm depths it was lower under sugarcane with no differences in other layers (Figure 5.1a). The BD increased significantly with depth from 0.93 g cm^{-3} (0-5 cm) to 1.65 g cm^{-3} (80-100 cm) under forest and from 1.31 to 1.50 g cm^{-3} under sugarcane ($p \leq 0.05$). The average (0-100 cm) BD was 4 % higher ($p \leq 0.05$) under forest (1.41 g cm^{-3}) than sugarcane (1.36 g cm^{-3}). As expected, the trend of TP results (Figure 5.1b) was the reverse of BD.

The TP decreased with depth from 65 to 38 % under forest and from 51 to 44 % under sugarcane ($p \leq 0.05$). Within the total profile depth, the average TP was 47 % under forest and 49 % under sugarcane ($p \leq 0.05$).

5.3.1.3 Water retention properties

The FC (Figure 5.2a) and PWP (Figure 5.2b) were significantly greater under sugarcane than forest ($p \leq 0.05$) in the top 15 cm while in the 15-20 cm layer PWP was lower under sugarcane than forest, with no differences in all other layers. The FC did not change with depth under both land uses ($p > 0.05$). Within the total profile depth (0-100 cm), the FC was higher under sugarcane (32 %) than forest (29 %) ($p \leq 0.05$). The PWP decreased with depth from 30 % to 20 % under sugarcane ($p \leq 0.05$) while no significant differences were observed with depth under forest. Similar to FC, the PWP within the total profile (0-100 cm) was higher ($p \leq 0.05$) under sugarcane (21 %) than under forest (17%). The AWC was similar between land uses at all depths except for the 30-40 cm layer where that under sugarcane (17 %) was higher ($p \leq 0.05$) than under forest (11 %) (Figure 5.2c). The AWC was also not significantly different between land uses in the whole profile depth (0-100 cm). The differences in FC and PWP at some of the sampled depths did not affect the AWC.

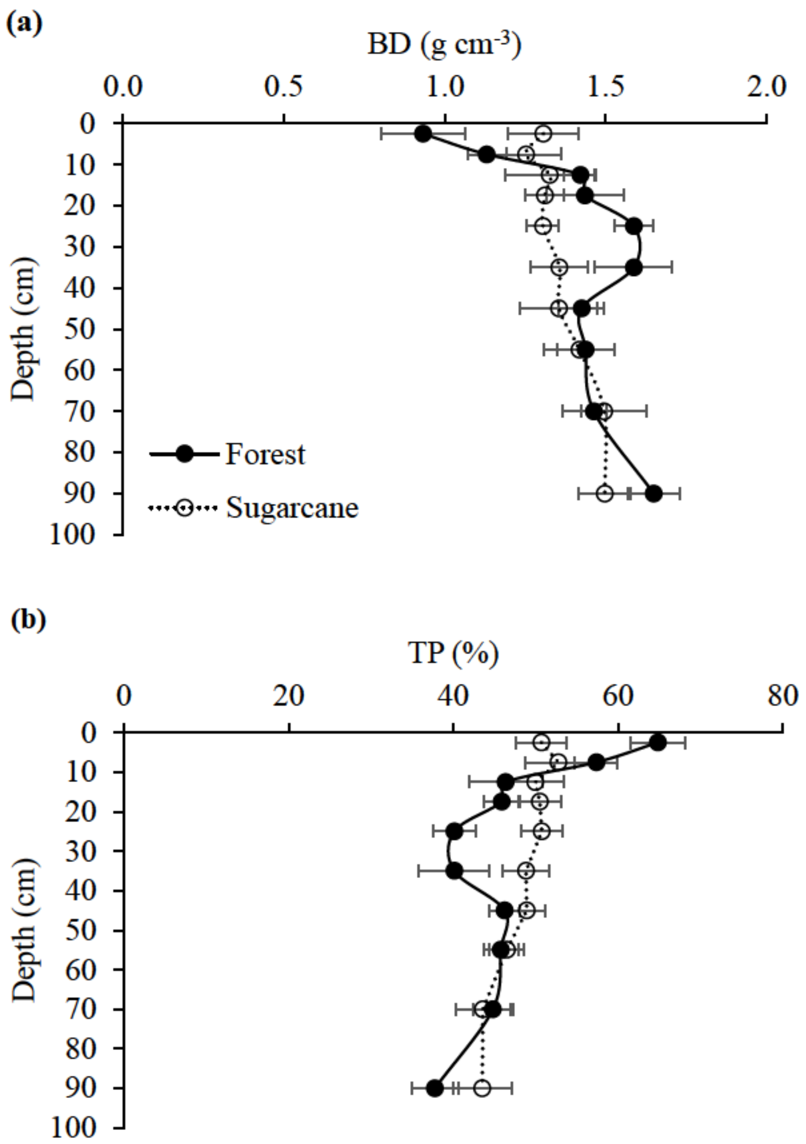

Figure 5. 1: The mean (±standard error) (a) bulk density (BD) and (b) total porosity (TP) measured in samples from 0-100 cm soil depth under native forest and sugarcane (n = 3).

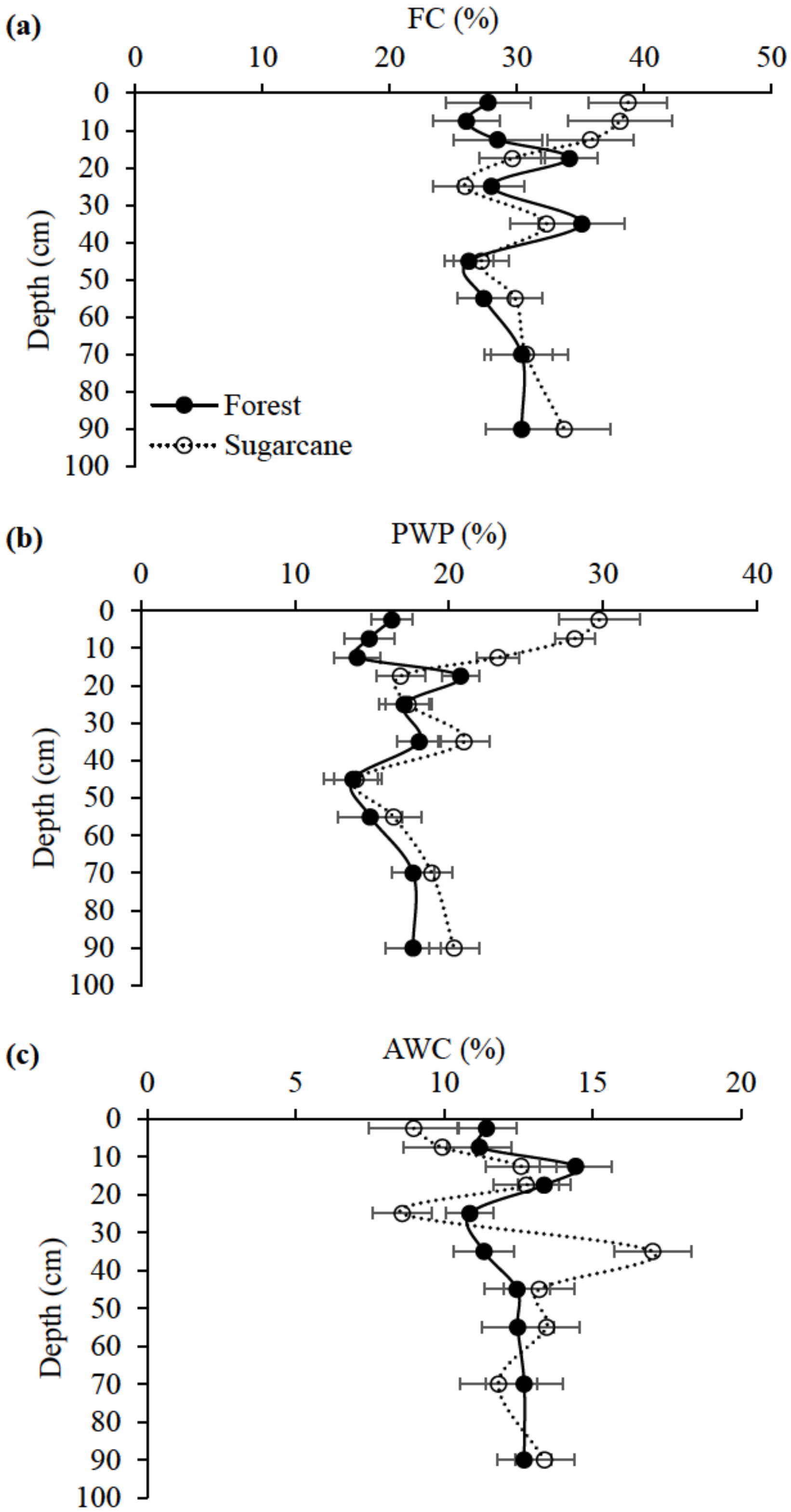

Figure 5. 2: The mean (±standard error) (a) field capacity (FC), (b) permanent wilting point (PWP) and (c) available water capacity (AWC) measured in samples from 0-100 cm soil depth under native forest and sugarcane (n = 3).

5.3.1.4 Saturated hydraulic conductivity

The Ks under sugarcane was significantly higher ($p \leq 0.05$) than forest at all depths except for the 15-20, 40-50 and 60-80 cm layers, which were not different between the land uses (Figure 5.3). The Ks decreased with depth (0-100 cm) from 33 to 23 cm hr^{-1} under forest and from 39 to 30 cm hr^{-1} under sugarcane. The average (0-100 cm) Ks was significantly higher ($p \leq 0.05$) under sugarcane (31 cm hr^{-1}) than under forest (26 cm hr^{-1}).

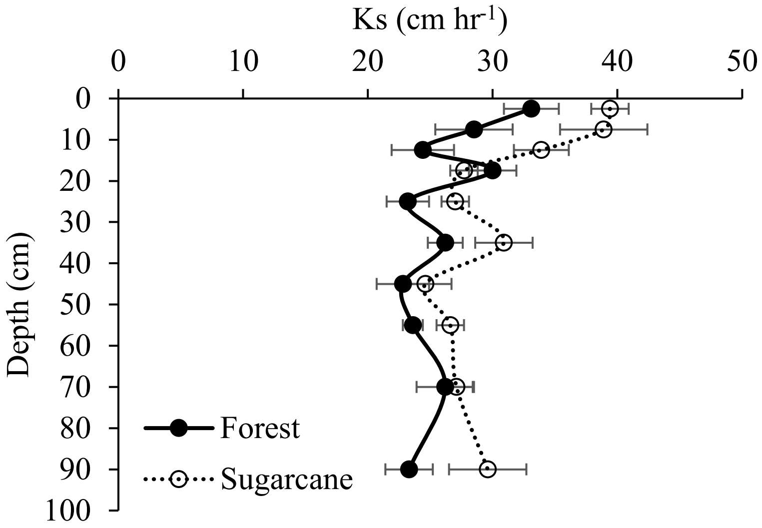

Figure 5. 3: The mean (±standard error) saturated hydraulic conductivity (Ks) measured in samples from 0-100 cm soil depth under native forest and sugarcane (n = 3).

5.3.1.5 Relationship between soil water content and some selected soil properties under sugarcane and native forest

The bulk soil properties considered are given on Table 5.1 (texture) and Table 5.2 (MWD, TOC, total Al and total Fe). The TOC correlated negatively with BD and positively with TP at all three depths while the same trend was observed between MWD and these variables in the 0-30 and 0-100 cm depths (Table 5.3). There were no significant correlations between TOC and the water retention characteristics (FC, PWP and AWC) of the soils (Table 5.3).

Table 5. 2: The mean weight diameter (MWD), total organic carbon (TOC), total aluminium (Al) and total iron (Fe) in bulk soils under native forest and sugarcane (n = 3).

Land use	Depth (cm)	MWD (mm)	TOC	Al	Fe
			………………. (g kg^{-1})………………..		
Forest	0-5	1.5^{d}	71.7^{j}	20.5^{a}	17.6^{a}
Sugarcane		1.2^{abc}	48.9^{i}	29.9^{de}	22.9^{fg}
Forest	5-10	1.5^{d}	49.1^{i}	19.6^{a}	17.4^{a}
Sugarcane		1.0^{ab}	42.8^{hi}	30.9^{ef}	22.9^{fg}
Forest	10-15	1.4^{cd}	40.8^{gh}	22.1^{a}	19.4^{abcd}
Sugarcane		1.2^{abc}	42.9^{hi}	31.2^{ef}	19.9^{abcde}
Forest	15-20	1.1^{ab}	35.3^{efg}	19.9^{a}	18.5^{abc}
Sugarcane		1.0^{a}	41.4^{gh}	30.1^{def}	22.5^{efg}
Forest	20-30	1.1^{ab}	30.1^{de}	25.2^{b}	17.7^{ab}
Sugarcane		1.1^{ab}	39.0^{fgh}	31.0^{ef}	23.6^{efg}
Forest	30-40	1.0^{a}	28.3^{cde}	26.5^{bc}	20.1^{abcde}
Sugarcane		1.2^{abc}	38.5^{fgh}	26.8^{bc}	20.9^{cdef}
Forest	40-50	0.9^{a}	26.7^{bcd}	30.7^{def}	21.3^{def}
Sugarcane		1.3^{abc}	33.4^{def}	28.1^{cd}	19.6^{abcd}
Forest	50-60	1.0^{a}	26.9^{bcd}	28.9^{cde}	17.7^{ab}
Sugarcane		1.2^{abc}	29.6^{de}	30.6^{def}	20.4^{bcdef}
Forest	60-80	1.1^{ab}	21.1^{ab}	29.7^{de}	21.9^{defg}
Sugarcane		1.2^{abc}	21.6^{abc}	31.4^{ef}	19.8^{abcd}
Forest	80-100	1.3^{abcd}	18.4^{a}	31.6^{ef}	23.7^{g}
Sugarcane		1.4^{bcd}	20.6^{ab}	32.9^{f}	20.5^{cdef}

The MWD, TOC, Al and Fe are discussed in detail in Chapter 7, however they are included here for completeness. Means within the same column followed by different letters are significantly different ($p \leq 0.05$).

The TOC (r = 0.37) and MWD (r = 0.39) were poorly correlated with Ks in the 0-30 and 30-100 cm depths, respectively (Table 5.3). Total Al (r = 0.46) and Fe (r = 0.40) were positively correlated to PWP in the top 30 cm depth. The silt correlated negatively with FC, PWP and Ks while the sand correlated positively with these parameters at 0-30 and 0-100 cm depths (Table 5.3). The Ks was strongly correlated to FC and PWP at all depths while the AWC was negatively correlated to PWP (r = -0.46) in the 0-100 cm depth. The FC and PWP were also positively correlated at all depths while the BD correlated negatively with TP at all depths.

Table 5. 3: Pearson's correlation coefficients between measured properties in the topsoil (0-30 cm), subsoil (30-100 cm) and whole profile (0-100 cm) under native forest and sugarcane (n = 6).

	BD	TP	Ks	FC	PWP	AWC
Parameter			Topsoil (0-30 cm)			
BD	1.00					
TP	**-0.99**	1.00				
Ks	-0.30	0.30	1.00			
FC	0.20	-0.20	**0.79**	1.00		
PWP	0.10	-0.10	**0.89**	**0.86**	1.00	
AWC	0.20	-0.20	-0.35	0.08	-0.44	1.00
MWD	**-0.50**	**0.50**	-0.13	-0.30	-0.27	0.02
TOC	**-0.90**	**0.90**	**0.37**	-0.09	-0.01	-0.15
Al	0.20	-0.20	0.32	0.39	**0.46**	-0.21
Fe	0.10	-0.10	0.33	0.38	**0.40**	-0.12
Clay	-0.10	0.10	0.23	-0.24	0.15	0.13
Silt	-0.20	0.20	**-0.48**	**-0.59**	**-0.64**	0.22
Sand	0.30	-0.30	**0.45**	**0.57**	**0.66**	-0.28
			Subsoil (30-100 cm)			
BD	1.00					
TP	**-1.00**	1.00				
Ks	-0.20	0.20	1.00			
FC	0.30	-0.30	**0.76**	1.00		
PWP	0.30	-0.30	**0.85**	**0.80**	1.00	
AWC	0.01	-0.10	-0.18	0.29	-0.35	1.00
MWD	-0.01	0.01	**0.39**	0.31	0.31	-0.01
TOC	**-0.60**	**0.60**	0.25	-0.07	-0.03	-0.06
Al	0.20	-0.20	-0.20	-0.17	-0.09	-0.13
Fe	0.30	-0.30	-0.03	0.13	0.13	-0.01
Clay	-0.30	0.30	0.33	0.02	0.12	-0.16
Silt	0.30	-0.30	0.08	0.30	0.15	0.23
Sand	0.01	0.01	-0.39	-0.31	-0.26	-0.07
			Whole soil profile (0-100 cm)			
BD	1.00					
TP	**-0.92**	1.00				
Ks	-0.10	0.10	1.00			
FC	0.30	-0.30	**0.86**	1.00		
PWP	0.20	-0.20	**0.93**	**0.95**	1.00	
AWC	0.10	-0.10	0.10	-0.47	-0.46	1.00
MWD	**-0.40**	**0.40**	0.11	-0.13	-0.09	-0.05
TOC	**-0.80**	**0.80**	0.10	-0.28	-0.19	-0.09
Al	0.36	-0.36	0.01	0.20	0.20	-0.04
Fe	0.20	-0.20	0.18	0.30	0.30	-0.06
Clay	0.20	-0.20	0.30	0.42	0.33	0.01
Silt	-0.30	0.30	**-0.70**	**-0.81**	**-0.80**	0.40
Sand	0.30	0.30	**0.70**	**0.80**	**0.81**	-0.41

BD: Bulk density; TP: Total porosity; Ks: Saturated hydraulic conductivity; FC: Field capacity; PWP: Permanent wilting point; AWC: Available water capacity; MWD: Mean weight diameter; TOC: Total organic carbon; Al: Total aluminium; Fe: Total iron. Correlation coefficients (r) in bold are statistically significant at $p \leq 0.05$.

5.3.2 Some physical properties of humic soils under native forest and grasslands

5.3.2.1 Bulk density and total porosity

The BD was lower ($p \leq 0.05$) at Karkloof than Cedara and Eshowe at 10-50 cm depth with no significant differences between sites at all other depths (Figure 5.4a). The BD increased significantly with depth (0-100 cm) from 0.89 to 1.65 g cm^{-3} at Eshowe, 0.86 to 1.51 g cm^{-3} at Cedara and from 0.86 to 1.61 g cm^{-3} at Karkloof. The average (0-100 cm) BD was lower ($p \leq 0.05$; Table 5.4) at Karkloof (1.11 g cm^{-3}) than Cedara (1.39 g cm^{-3}) and Eshowe (1.41 g cm^{-3}). As expected, the trend of TP results was the reverse to that of BD (Figure 5.4b). The TP decreased with depth from 65 to 38 % at Eshowe, 66 to 43 % at Cedara and from 68 to 39 % at Karkloof. Within the total profile depth, the average TP was higher ($p \leq 0.05$) at Karkloof (60 %) than Cedara (48 %) and Eshowe (47 %) (Table 5.4).

5.3.2.2 Water retention properties

The FC was significantly lower at Eshowe than the other sites at all depths with the exception of the 15-20 cm layer where Eshowe (34 %) was higher than Cedara (28 %) and the 5-10 cm layer which was not different between sites (Figure 5. 5a). At 0-5, 20-30 and 40-50 cm, Karkloof had significantly higher FC than Cedara, with no differences between these sites in all other layers. The FC did not change with depth at all sites ($p > 0.05$). Within the total profile depth (0-100 cm), the FC was higher ($p \leq 0.05$) at Karkloof (49 %) than Cedara (40 %) and Eshowe (29 %) (Table 5.4). The PWP was lower at Eshowe than the other sites at all depths with the exception of the 15-20 cm layer which was not different from Cedara (Figure 5.5b).

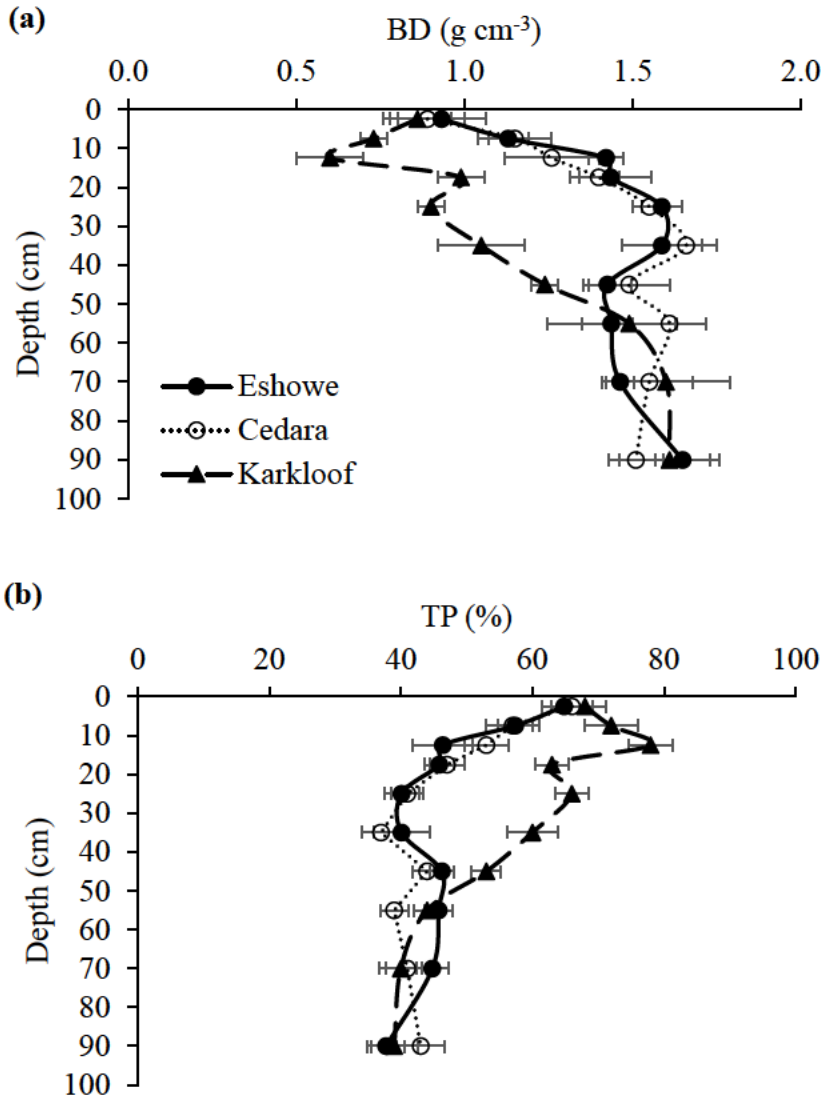

Figure 5. 4: The mean (±standard error) (a) bulk density (BD) and (b) total porosity (TP) measured in samples from 0-100 cm soil depth under Eshowe forest, Karkloof and Cedara grasslands (n = 3).

Table 5. 4: The bulk density (BD), saturated hydraulic conductivity (Ks), total porosity (TP), field capacity (FC), permanent wilting point (PWP) and available water capacity (AWC) of humic soil profiles (0-100 cm) under Eshowe forest, and Cedara and Karkloof grasslands (n = 3).

	BD	Ks	TP	FC	PWP	AWC
	(g cm^{-3})	(cm hr^{-1})	…………………….(%)………………………..			
Cedara	1.39^{a}	18.2^{a}	48.1^{a}	40.0^{b}	27.7^{b}	12.3^{a}
Karkloof	1.11^{b}	23.1^{ab}	59.8^{b}	49.2^{c}	36.8^{c}	12.4^{a}
Eshowe	1.41^{a}	26.4^{b}	47.3^{a}	29.5^{a}	16.6^{a}	12.9^{a}

Means within the same column followed by different letters are significantly different ($p \leq 0.05$)

At 0-5, 10-30 and 40-50 cm, Karkloof had significantly higher PWP than Cedara, with no differences between these sites in all other layers. The PWP did not change with depth at all sites ($p > 0.05$). Similar to FC, the PWP within the total profile (0-100 cm) was higher ($p \leq 0.05$) at Karkloof (37 %) than Cedara (28 %) and Eshowe (17 %) (Table 5.4).

The AWC was similar between sites at all depths except for the 0-5 cm layer where Cedara was higher than the other two sites and the 15-20 cm layer where Eshowe (13 %) was higher ($p \leq 0.05$) than Cedara (8 %) (Figure 5.5c). The AWC was also not significantly different ($p > 0.05$) between sites in the whole profile depth (0-100 cm) (Table 5.4). Although there were differences in FC and PWP for some of the sampled depths, these did not affect the AWC.

5.3.2.3 Saturated hydraulic conductivity

The Ks was higher at Eshowe than the other sites at all depths with the exception of the 5-15, 20-30 and 40-50 cm depths which were not different from Karkloof (Figure 5.6). The Ks was significantly higher ($p \leq 0.05$) at Karkloof than Cedara at all depths except for the 5-10 and 50-100 cm depths, which were not different between these sites. The Ks ranged with depth (0-100 cm) from 19 to 27 cm hr^{-1} at Karkloof, 14 to 27 cm hr^{-1} at Cedara and from 23 to 33 cm hr^{-1} at Eshowe. The average (0-100 cm) Ks was similar at Karkloof (23 cm hr^{-1}) and Eshowe (26 cm hr^{-1}) and higher ($p \leq 0.05$) than Cedara (18 cm hr^{-1}) (Table 5.4).

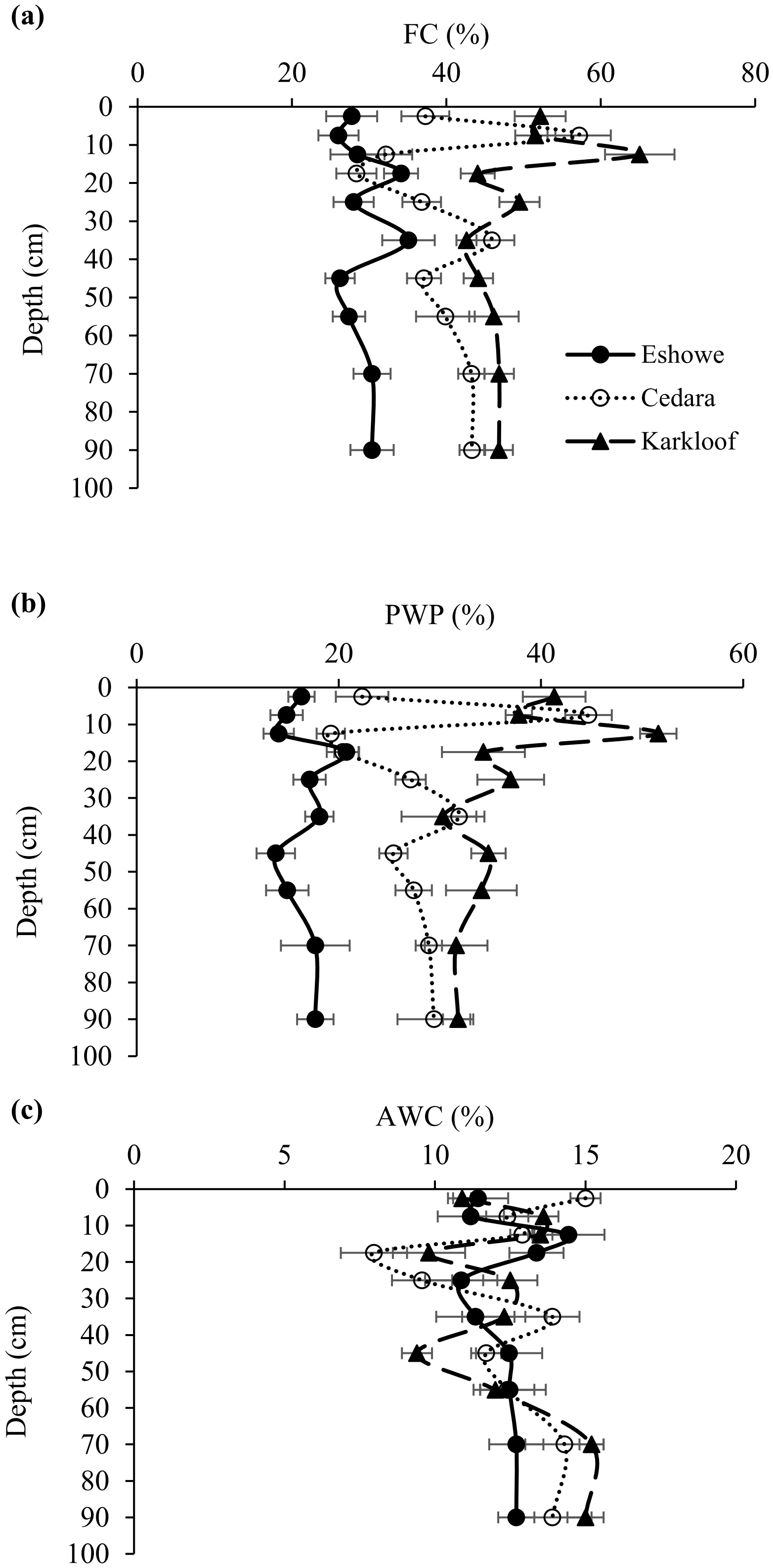

Figure 5. 5: The mean (±standard error) (a) field capacity (FC), (b) permanent wilting point (PWP) and (c) available water capacity (AWC) measured in samples from 0-100 cm soil depth under Eshowe forest, Karkloof and Cedara grasslands (n = 3).

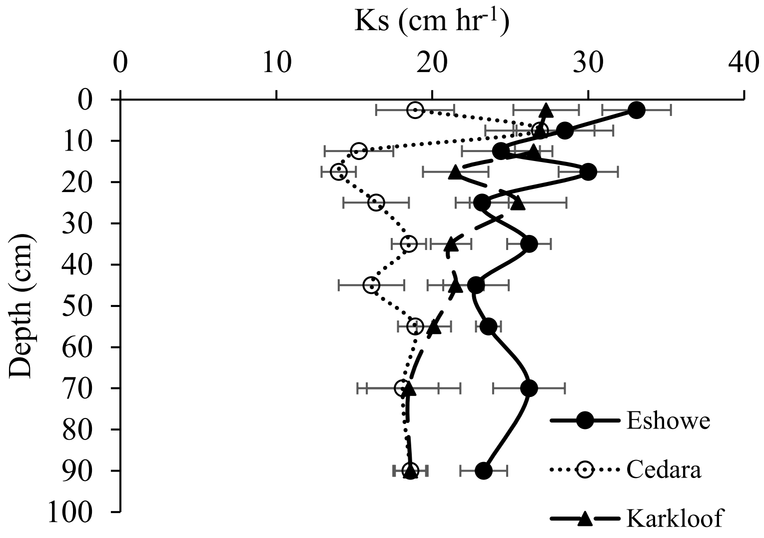

Figure 5. 6: The mean (±standard error) saturated hydraulic conductivity (Ks) measured in samples from 0-100 cm soil depth under Eshowe forest, Karkloof and Cedara grasslands (n = 3).

5.3.2.4 Relationship between water retention characteristics and some selected soil properties under native forest and grasslands

The bulk soil properties considered are given in Table 4.1. The relationship between MWD and Ks, FC and PWP was positive at all depths (Table 5.5). The TOC correlated positively with Ks (r = 0.62), FC (r = 0.50) and PWP (r = 0.49) in the top 30 cm depth. At all depths, the TOC correlated negatively with BD and positively with TP. The total Al and Fe were positively correlated with Ks, FC and PWP at all depths. The silt correlated positively with FC, PWP and negatively with Ks while the sand content correlated positively with Ks and negatively with FC and PWP at all depths. The Ks correlated positively with FC and PWP but negatively with clay at all depths while the FC was also positively correlated to PWP at all depths. At 0-30 and 0-100 cm depths, the Ks and MWD were negatively correlated to BD and positively to TP. The BD correlated negatively with TP at all depths (Table 5.5).

Table 5. 5: Pearson's correlation coefficients between measured properties in the topsoil (0-30 cm), subsoil (30-100 cm) and whole profile (0-100 cm) under Eshowe forest, Karkloof and Cedara grasslands (n = 9).

	BD	TP	Ks	FC	PWP	AWC
Parameter			Topsoil (0-30 cm)			
BD	1.00					
TP	**-1.00**	1.00				
Ks	**-0.70**	**0.70**	1.00			
FC	**-0.60**	**0.60**	**0.87**	1.00		
PWP	**-0.50**	**0.50**	**0.86**	**0.98**	1.00	
AWC	-0.30	0.30	0.19	0.32	0.12	1.00
MWD	**-0.70**	**0.70**	**0.60**	**0.46**	**0.48**	0.04
TOC	**-0.80**	**0.80**	**0.62**	**0.50**	**0.49**	0.14
Al	-0.40	0.40	**0.65**	**0.50**	**0.54**	-0.09
Fe	-0.30	0.30	**0.59**	**0.44**	**0.48**	-0.08
Clay	-0.20	0.20	**-0.47**	0.35	0.37	-0.03
Silt	**-0.70**	**0.70**	**-0.77**	**0.61**	**0.61**	0.15
Sand	**0.50**	**-0.50**	**0.50**	**-0.53**	**-0.55**	-0.07
			Subsoil (30-100 cm)			
BD	1.00					
TP	**-1.00**	1.00				
Ks	-0.40	0.40	1.00			
FC	-0.10	0.10	**0.93**	1.00		
PWP	-0.20	0.20	**0.97**	**0.96**	1.00	
AWC	0.40	-0.40	-0.30	-0.05	-0.32	1.00
MWD	-0.20	0.20	**0.80**	**0.76**	**0.78**	-0.23
TOC	**-0.50**	**0.50**	0.35	-0.41	0.43	-0.18
Al	-0.10	0.10	**0.84**	**0.80**	**0.78**	-0.10
Fe	-0.20	0.20	**0.83**	**0.78**	**0.75**	-0.06
Clay	0.01	-0.01	**-0.69**	**0.66**	**0.63**	-0.01
Silt	-0.30	0.30	**-0.60**	**0.57**	**0.57**	-0.09
Sand	0.10	-0.10	**0.53**	**-0.79**	**-0.77**	0.07
			Whole soil profile (0-100 cm)			
BD	1.00					
TP	**-1.00**	1.00				
Ks	**-0.60**	**0.60**	1.00			
FC	-0.40	0.40	**0.87**	1.00		
PWP	-0.40	0.40	**0.89**	**0.97**	1.00	
AWC	0.10	-0.10	-0.02	0.20	-0.04	1.00
MWD	**-0.50**	**0.50**	**0.68**	**0.54**	**0.57**	-0.11
TOC	**-0.80**	**0.80**	0.42	0.25	0.28	-0.10
Al	-0.10	0.10	**0.62**	**0.55**	**0.57**	-0.04
Fe	-0.10	0.10	**0.61**	**0.52**	**0.54**	-0.03
Clay	-0.01	0.02	**-0.52**	0.43	0.45	-0.01
Silt	**-0.50**	**0.50**	**-0.70**	**0.58**	**0.58**	0.03
Sand	0.30	-0.30	**0.55**	**-0.60**	**-0.61**	-0.01

BD: Bulk density; TP: Total porosity; Ks: Saturated hydraulic conductivity; FC: Field capacity; PWP: Permanent wilting point; AWC: Available water capacity; MWD: Mean weight diameter; TOC: Total organic carbon; Al: Total aluminium; Fe: Total iron. Correlation coefficients (r) in bold are statistically significant at $p \leq 0.05$.

The silt correlated negatively to BD but positively to TP in the 0-30 and 0-100 cm depths while the sand correlated positively to BD (r = 0.50) and negatively to TP (r = -0.50) in the top 30 cm (Table 5.5).

5.4 Discussion

5.4.1 Effects of sugarcane cultivation relative to native forest on some selected soil properties of humic soils

5.4.1.1 Basic soil properties of the bulk soils

The lower pH in the 0-5 cm depth of the sugarcane soils could be attributed to the regular addition of acidifying fertilizers (Du Toit, 1993; Hartemink, 1998; Dominy and Haynes, 2002; Wang et al., 2016). In most cropping practices, soil acidification problems are related to (i) the use of ammoniacal fertilizers which encourage the displacement of basic cations by NH_4^+, (ii) nitrification (2 mol of H^+ is produced per mole of NH_4^+) and (iii) toxicity of Al and other metals (Fey et al., 1990; Hartemink, 1998; Jim, 2003). The higher total Al in the soils under sugarcane suggests that its hydrolysis in these acidic soils could further lower soil pH (Hartemink, 1998). Although the soils under sugarcane are also limed, the application of 1 to 10 t lime ha^{-1} once every eight years may not be sufficient to neutralize the acidity produced by annual fertilization and nitrification, together with potential hydrolysis of Al. Texturally, the soils from both land use types were similar (sandy clay loams). In the topsoils, the average clay content under forest (13 %) was slightly lower than that under sugarcane (16 %) and both were lower than those reported by Gubevu (1997) for shallow (< 30 cm) humic soils in the Ngome forest (30 %) and sugarcane (42 %) plantations in KwaZulu-Natal. This difference may be because the soils in the Gubevu (1997) study were derived from dolerite under higher rainfall (>1500 mm p.a.) and higher altitude (1300 m a.s.l) conditions compared to those of the present study area. The predominance of sand under both land uses was a result of inheritance from the sandstone parent material, while the highly weathered state of the soils was indicated by the dominance of kaolinite in the clay fraction.

5.4.1.2 Water related properties of the bulk soils

The increased BD and reduced TP in the top 5 cm of the soil under sugarcane, compared to forest, are indicative of soil compaction. This observation could be related to the reduced TOC found at this depth (Malepfane et al., 2022). Numerous studies (e.g. van Antwerpen and Meyer, 1996; Qongqo and van Antwerpen, 2000; Nxumalo, 2015) have reported significant increases in topsoil (0-20 cm) BD following 20 to 50 years of sugarcane cultivation on soils that were previously under native forest. In South Africa, Meyer and van Antwerpen (2010) reported that in-field traffic and harvesting of sugarcane under wet conditions generally cause an increase in topsoil compaction that if not managed poses a great risk of runoff and erosion under high intensity rainfall (Hammad and Dawelbeit, 2001; Rawls et al., 2003). The higher average BD, indicating greater soil compaction, obtained under forest (1.41 g cm^{-3}) than sugarcane (1.36 g cm^{-3}), may once again be associated with the lower average TOC found under forest (34.8 g C kg^{-1}) in contrast to sugarcane (35.9 g C kg^{-1}) (Table 3.1). This is supported by the negative correlation between BD and TOC as well as MWD at all depths. This confirms that TOC enhances soil aggregation, consequently increasing TP and reducing BD (Kay and VandenBygaart, 2002; Bescansa et al., 2006; Strudley et al., 2008; Chen et al., 2009; Li et al., 2021).

These results could also be reflecting compaction from tree roots and overburden pressure as well as the lack of tillage of the forest soil (Igwe, 2005; Barger et al., 2011; Li et al., 2021). Wakgari et al. (2020) also reported higher average BD under a native forest (1.54 g cm^{-3}) than sugarcane (1.51 g cm^{-3}) on clayey Luvisols on the Finchaa Sugar Estate, Ethiopia. This was attributed to either the loosening of the soils under long term (> 20 years) sugarcane cultivation or cane residues left on the surface after harvesting of the cultivated fields and the use of agricultural additives (e.g. filter cake) during cultivation. Barzegar et al. (2000) reported that sugarcane residues reduced BD following the long-term (36 years) cultivation of a clay loam soil in Khuzestan Province of Iran. The overall (0-100 cm) BD value for the sugarcane soil is, however, within the optimum BD range of 1.30 to 1.40 g cm^{-3} for sugarcane production as suggested by Ridge (2013) in Australia and lower than limiting thresholds of 1.50 to 1.70 g cm^{-3} that were suggested by Yang (1974) in Taiwan.

The higher BD and lower TP account for the lower Ks (0-100 cm) observed under forest (26 cm hr^{-1}) than under sugarcane (31 cm hr^{-1}) but these results differ from many other reported studies. For instance, Celik (2005) studied the effects of changes in land use on the physical properties of a silty clay loam Typic Haploxeroll (0-20 cm) in a Mediterranean highland region of Turkey and found Ks to be 25 % higher under forest (15 cm hr^{-1}) than wheat (*Triticum aestivum*) (12 cm hr^{-1}). Compared to native forest (28 cm hr^{-1}), Gol (2009) found a 180 % average decrease in Ks in soils of three textural classes (clay loam, sandy clay loam and silty clay loam) under corn (*Zea mays*) (10 cm hr^{-1}) in the Dagdami river catchment, Turkey. In a shallow (0-30 cm) silty loam soil in Germany, Wahren et al. (2009) found that Ks was three times higher under ancient forest (50 cm hr^{-1}) than under cropland (15 cm hr^{-1}). In all these studies, higher values of Ks in forest soils were explained by the higher proportion of macro-pores because of decaying roots and faunal activity. The difference between these studies and the current study may be a result of the differences in the levels of SOC involved (Strudley et al., 2008; Blanco-Canqui et al., 2017; Fu et al., 2021), with the humic soils having > 4 % C in the top 15 cm and >1.8 % even at 100 cm depth. Furthermore, soils under sugarcane at the study site are only cultivated at replanting, which equates to about every eight years in northern KwaZulu-Natal while the other studies made comparisons with soils under annual crops. Soils under sugarcane have been shown to have higher biomass input than that of annual crops (van Antwerpen and Meyer, 1996; Meyer and van

Antwerpen, 2010), which might improve C inputs to the soil (Shwetha and Varija, 2015) and consequently increase Ks (Iversen et al., 2001; Chen et al., 2009).

The decrease in Ks with depth was accompanied by a decrease in TP and an increase in BD as has been commonly reported (Haynes and Naidu, 1998; Cousin et al., 2003; Evrendilek et al., 2004; Fu et al., 2021). The negative correlation between Ks and silt in the 0-30 ($r = -0.48$) and 0-100 cm depths ($r = -0.70$) suggests that the higher silt in the soils under forest could have masked the impacts of any increase in Ks caused by SOC in the studied soils (Tomasella and Hodnett, 1998). These results further indicate that when the heterogeneity of the silt grain distribution increases, porosity decreases, leading to an increase in resistance to penetration and, consequently, a reduction in Ks (Igwe et al., 2013). In the current study, soils under both land uses exhibit high Ks (> 25 cm hr^{-1}) (Pam and Brain, 2007), suggesting that runoff is likely to be minimal due to enhanced infiltration potential (Bauer and Black, 1992; Haynes and Naidu, 1998; Rawls et al., 2003).

The higher water content at FC and PWP in the top 15 cm under sugarcane than forest could partly be associated with the dominance of smaller-sized and lower water conducting pores under sugarcane (Strudley et al., 2008; Gol, 2009). Evidence from other sugarcane studies indicates a reduction in the macro-pore connectivity following smearing from tillage thus lowering aeration and increasing water content at FC and PWP (Barzegar et al., 2000; Wakgari et al., 2020). The lower average water content at FC and PWP in soils under forest than sugarcane could again be attributed to their higher silt content (Igwe et al., 2013). This is supported by the negative correlations between silt and FC and PWP in the 0-30 and 0-100 cm depths which again could be indicating heterogeneity in the size distribution of the silt grains (Igwe et al., 2013), with coarse silt inducing lower porosity and water retention at FC and PWP than fine silt (Tomasella and Hodnett, 1998). The weak but positive relationship between Al and Fe and PWP at the 0-30 cm depth might again be indicating a slight dominance of micropores following micro-aggregation by these elements in the studied soils (Duiker et al., 2003). The lower average soil water content at PWP under forest (17 %) than sugarcane (21 %) indicates that the forest soils have a slightly lower micro-porosity than sugarcane thereby reducing the ease of water availability to plant roots (Iversen et al., 2001; Kargas and Londra, 2015; Yost and Hartemink, 2018).

The higher AWC under sugarcane (17 %) than forest (11 %) at 30-40 cm may be related to SOM which was higher under sugarcane than forest at that depth (Malepfane et al., 2022). This is supported by other reported work showing that SOM enhances AWC (Bescansa et al., 2006; Kargas and Londra, 2015; Blanco-Canqui et al., 2017). However, the AWC results of the current study are not in agreement with those of Bescansa et al. (2006) who reported higher AWC in undisturbed than in cultivated loamy soils in Spain. In that study, the higher AWC was associated with higher TOC content and changes in the pore size distribution of the undisturbed soils. The differences in the results may perhaps be due to the TOC status of the study areas. The significantly higher TOC content in the undisturbed than cultivated soils in Spain increased AWC under this land use, while in the present study the TOC was significantly higher under sugarcane than forest only in the 30-40 cm depth, suggesting that its effects on AWC were similar between these land uses (Hammad and Dawelbeit, 2001; Fu et al., 2021). The similarity of AWC in soils under sugarcane and forest could be because the lower SOC under sugarcane reduced the water content at both FC and PWP resulting in minimal effects on the difference between these two parameters (Minasny and McBratney, 2017).

The lack of relationships between TOC and the FC, PWP and AWC in the present study may be due to a number of reasons. Firstly, the formation of organo-mineral complexes with different species of Al and Fe, mediated by organic matter, is a topsoil characteristic of humic soils (Soil Classification Working Group, 2018; Malepfane et al., 2022) that could affect the ability of the C to influence soil water retention characteristics (Lal, 1978; Rawls et al., 2003; Yost and

Hartemink, 2018). Secondly, the soils in the present study have low clay (up to 20 %) and high sand (81 %) contents. The high correlation of sand with FC and PWP in the top 30 cm, where sugarcane soils had higher sand content, suggests that the larger portions of the organic matter in this soil could be in more particulate form, storing more water at the two threshold water potentials.

Soil clay content was not correlated with any of the soil water-related parameters in the current study, suggesting that the influence of the clay fraction and its predominant kaolinitic mineralogy in such loamy humic soils may be minimal. Previous studies have shown that the more clayey soils tend to have higher OM than sands, causing strong relationships between clay content, SOC and water retention characteristics (Cousin et al., 2003; Rawls et al., 2003; Minasny and McBratney, 2017). All these studies indicated that SOC, clay and water retention characteristics are likely to influence each other synergistically.

On the other hand, the positive relationship between Ks, FC and PWP at all depths and between Ks and AWC for the 0-100 cm depth is consistent with Igwe et al. (2013) who reported a close association between these properties in sandstone-derived Inceptisols and Ultisols. In addition, the positive correlations between FC and PWP in the 0-100 and 0-30 cm depths of the present soils indicate that these parameters are affected by the same factors, including SOC and texture. The results confirmed those of Minasny and McBratney (2017) who reported that the shape, size, degree of weathering and geological origin of the mineral soil component can strongly affect the hydraulic conductivity and water retention characteristics of the soil.

5.4.2 Some physical properties of humic soils under native forest and grasslands

The lower average (0-100 cm) BD at Karkloof could be explained by the higher TOC content at this site (45 g kg^{-1}) than at Eshowe (36 g kg^{-1}) and Cedara (27 g kg^{-1}) (Malepfane et al., 2022). This is supported by the negative correlation between TOC and BD at all depths confirming that

OC improves soil aggregation and, consequently, increases TP and reduces BD (Hudson, 1994; Rawls et al., 2003; Celik, 2005). The positive correlation between TP and TOC at all depths and the higher average TP at Karkloof than Cedara and Eshowe agrees with the findings of Leifeld et al. (2005) who concluded that the main effects of OC were to decrease the BD and increase the TP following aggregate stability development in clayey textured grassland soils (0-100 cm) in Switzerland. The increase in BD with depth at all the studied sites is commonly observed (e.g. Park, 1971; Celik, 2005; Wang et al., 2009) and attributed to (i) more clay in the subsoil and (ii) the slight compaction of the subsurface layers due to less OM build-up thereby resulting in poor aggregation in this part of the profile.

The greater average (0-100 cm) Ks at Eshowe (26 cm hr^{-1}) and Karkloof (23 cm hr^{-1}) than Cedara (18 cm hr^{-1}) may again be explained by better soil aggregation due to higher TOC content, in addition to the higher sand content in the soil derived from the Natal Group Sandstone compared to those from dolerite. This enhanced aggregation promotes the dominance of macro-pores, thereby encouraging preferential flow and increasing Ks (Leifeld et al., 2005; Yost and Hartemink, 2018). The positive association between Ks and MWD at all depths, which indicates reduced slaking under saturated conditions (Le Bissonnais et al., 2002; Lado et al., 2004), supports this. Park (1971) studied the variation in soil water associated with aggregation in humic gley podzols (0-35 cm) under a native silver beech forest in the Tararua Mountains, New Zealand. The results showed that when MWD was high (1.90 mm), the Ks was 6.70 cm hr^{-1}, but when the MWD was low (0.71 mm), the Ks was reduced to 1.20 cm hr^{-1}. It was concluded that soil structure is important in influencing the movement and availability of water in the soil. The positive correlation between Ks and TOC in the top 30 cm corroborates most previous studies on native forest and grasslands on soils of a wide range of textures (Hudson, 1994. Leifeld et al., 2005; Wang et al., 2009; Jarvis et al., 2013; Kar et al., 2017; Yost and Hartemink, 2018). While all these studies have concluded that OC is usually considered to improve soil structure, which would imply a positive correlation with Ks, this relationship varies with the quality and quantity of C inputs. Other studies (e.g. Golchin et al., 1994; Nemes et al., 2005; Li et al., 2007; Wang et al., 2016) have reported a negative correlation between Ks and SOC due to the reduction in the rate of wetting and retention of water in soil caused by the presence of hydrophobic coatings on soil particles.

The positive relationship between Ks and Al and Fe observed at all depths is a result of a two-fold indirect effect i.e., Al and Fe have strong flocculating characteristics which improve soil structure (Beare et al., 1994; Bronick and Lal, 2005), and this further increases the mean pore size thereby increasing the Ks (Snyman and du Preez, 2005; Igwe et al., 2013). Shainberg et al. (1987) also reported a positive relationship between Ks and Al and Fe in Alfisols (0-50 cm) in California. It was concluded that the presence of Al and Fe in soils has a favourable effect on physical properties, increasing aggregation, porosity, permeability and Ks while reducing BD and clay dispersion. The positive relationship between Ks and the sand fraction observed at all depths was also expected since an increase in sand content leads to greater porosity thus increasing Ks (Saxton and Rawls, 2006; Wang et al., 2016). The negative relationship between Ks and silt and clay contents at all depths is in agreement with studies by Seema et al. (2019) carried out to investigate the hydraulic properties of soils (0-30 cm) varying in texture (sand, loam, sandy clay loam, silty clay and sandy clay) and OM under native forest and grassland in Haryana State, India. They concluded that Ks decreased with an increase in silt and clay contents due to the reduction in pore size diameter of the soils.

The negative correlation between Ks and BD at the 0-30 and 0-100 cm depths suggests that the increase in BD results in the reduction of mean pore size and consequently a decrease in Ks (Hudson, 1994; Kar et al., 2017). The positive relationship between Ks and FC and PWP at all depths corroborates the findings of Cousin et al. (2003) who indicated that, depending on the number and diameter of pores in the soil, Ks is a function of BD, TP, FC, PWP and AWC among other soil properties. The higher water content at FC and PWP at Karkloof reflects the higher TOC and silt contents at this site (Table 4.1; Malepfane et al., 2022).

This is supported by the positive correlation between these variables (FC and PWP) and TOC in the top 30 cm and with silt at all depths (Table 5.5). These trends are comparable to what has been reported in other studies on grassland and forest soils. For example, Jamison and Kroth (1958) reported a linear relationship between silt and water content at FC and PWP in 271 soil samples from Missouri suggesting that the release of water from silt over the suction range of availability is higher than that from clay and sand particles. However, if a clay soil has fine stable micro-structure with most of the aggregates in the size ranges of very fine sand and silt, it will retain a large portion of its storage water after the wilting point is reached. In another study, Seema et al. (2019) reported a positive correlation between silt content and FC ($R^2 = 0.85$) and PWP ($R^2 = 0.87$). These results were attributed to the increase in volume of micro-pores capable of holding water at FC and increased specific surface area contributing to adsorption of more water at PWP. In India, Shewtha and Varija (2015) also reported a similar relationship between SOC and FC and PWP which was attributed to the higher volume of water holding pores at FC while the higher soil moisture at PWP was associated with the capacity of SOC to increase specific surface area. According to Hudson (1994) and Rawls et al. (2003), the quantity of water which can be stored at FC and PWP varies mainly with silt plus clay contents while the organic matter content can also contribute because much of the water stored in the organic fraction is held at tensions above wilting.

The negative correlation observed between sand and FC and PWP at all depths is common (Jamison and Kroth, 1958; Nemes et al., 2005; Saxton and Rawls, 2006) and indicates that water contents at FC and PWP are higher when sand content is less. Generally, the sand particles reduce pore water retention by reducing the total volume of small pores, regardless of the SOC and clay contents (Hudson, 1994; Rawls et al., 2003; Yost and Hartemink, 2018). The positive relationship between Al and Fe and FC and PWP at all depths in the present study confirmed the findings of Duiker et al. (2003) and Igwe et al. (2013) who reported that the interaction of Al and Fe with kaolinite can synergistically encourage micro-aggregation thereby increasing water storage, but not necessarily plant-available water. The BD correlated negatively with FC (r=-0.60) and PWP (r=-0.50) in the 0-30 cm depth. In agreement with these findings, Seema and Phogat (2019) attributed the decrease in soil water at FC with increasing BD to the effect of the coarse fraction on micro-pores accountable for retaining water at FC while the PWP trend was linked with a decrease in specific surface area of soils, predominantly due to a higher sand fraction.

5.5 Conclusions

The purpose of this study was to determine (i) how sugarcane production affected BD and some water-related properties of a sandy clay loam humic soil in northern KwaZulu-Natal, South Africa, and (ii) how different site conditions affected BD and some selected water-related properties in humic soils under native forest and grassland. The findings from the first objective of this study showed that sugarcane cultivation results in higher BD, FC and PWP and lower TP in the topsoil than under forest. Moreover, the sandy clay loam humic soils under sugarcane cultivation showed increased Ks compared to the forested soils. Differences between the surface layers can be mostly ascribed to changes in BD following a decrease in TOC after cultivation of these soils. While TOC content is relatively high in the humic soils (> 4 % C in the top 15 cm and >1.8 % even at 100 cm depth), it does not fully explain the water retention characteristics (FC, PWP and AWC) throughout the soil profile. Texture, specifically silt and sand content, plays a significant role in influencing the water retention characteristics of these humic soils. The Al and Fe also showed a weak but positive correlation to PWP in the studied soils. Given the decrease in TOC following cultivation, and its influence on soil properties, implementing practises that enhance OM content in the soil is crucial. This can be achieved through practises such as cover cropping, crop residue management, and the incorporation of organic amendments. Increasing OM content will help improve soil structure, reduce BD, and enhance water retention characteristics in the drought-prone Zululand region of South Africa.

The results of the second objective revealed lower BD and greater TP, FC, and PWP at Karkloof in the 0-100 cm depth when compared to Cedara and Eshowe. This pattern is attributed to the higher TOC content observed at Karkloof site. The higher TOC at Karkloof also played a significant role in the higher Ks observed compared to Cedara. Additionally, the higher sand content at Eshowe site emerged as a crucial parameter associated with higher Ks at that site compared to Cedara. The TOC primarily influences Ks in the top 30 cm depth, while Al, Fe and sand have an influence at all depths. The water contents at FC and PWP are influenced by TOC in the 0-30 cm depth, and clay in the 30-100 cm depth, and Al, Fe and silt at all depths. The sand correlated negatively with FC and PWP at all depths indicating the reduction of TP by sand particles following a decrease in the total volume of small pores. The BD also correlated negatively with Ks while AWC was not significantly different between the three sites. With regards to Cedara findings, it is recommended that site specific soil management strategies are implemented. These may include proper grazing management such as rotational grazing that can allow for adequate plant regrowth to preserve soil structure, reduce BD, and enhance water infiltration and retention. In order to gain knowledge about the water quality and nutrient availability under these humic soils, a study of the distribution of TOC, Al, and Fe in aggregate size fractions is necessary. This will allow the creation of site-specific management techniques for maintaining soil health and production.

CHAPTER 6: TOTAL ORGANIC CARBON, ALUMINIUM AND IRON IN AGGREGATE SIZE FRACTIONS OF HUMIC SOILS FROM THREE SITES UNDER NATIVE VEGETATION

6.1 Introduction

Both native forest and grassland ecosystems are essential components of the biogeochemical carbon (C) cycle as their combined total organic carbon (TOC) storage is more than 10 % of the global total (Conant et al., 2001; Briske et al., 2005). The TOC in these ecosystems is important for vegetation productivity and soil structural stability among many other functions (Chaplot et al., 2011; Dlamini et al., 2016). There is generally a mutual relationship between vegetation and aggregate stability (AS) in that soil structure is the framework for bulk density and porosity, and influences water, air and nutrient flow to the vegetation (Stone and Buttery, 1989; Perfect et al., 1990). In turn, the vegetation supplies the soil with fresh organic residues and roots for aggregate structural development (Feller and Beare, 1997). The nature and properties of aggregates are thus determined by the quantity and quality of vegetation root residues (Lal, 2004) and by the degree of their interaction with soil particles (Tisdall and Oades, 1982; Chenu et al., 2006).

In addition, internal soil climate has been widely reported to impose constraints on the processes that control TOC accumulation thereby resulting in changes of TOC distribution under different environmental conditions (Callesen et al., 2003; Amelung et al., 1998; Virto et al., 2012; Mureva et al., 2018). In spite of these research efforts, less attention has been given to climate-related effects on organo-mineral associations, which are also crucial for organic carbon (OC) accumulation in some soils (Torn et al., 1997; Ko¨gel-Knabner et al., 2008). Numerous studies (Torn et al, 1997; Filimonova et al., 2016; Malepfane et al., 2022) on highly leached soils have indicated the importance of different aluminium (Al) and iron (Fe) species to soil aggregation and OC storage potential, but did not explain the distribution of these elements within aggregates. Generally, the role of these elements differs between aggregates of different particle sizes (Zhang and Horn, 2001; Igwe et al., 2009; Inagaki et al., 2019). The Al and Fe minerals provide surfaces for adsorption and/or complexation of OC (Beare et al. 1994; Hobley et al., 2016) resulting in different C residence times between macro-aggregates (> 250 µm) and micro-aggregates (< 250 µm) (Blanco-Canqui and Lal, 2004). In addition, some studies of microenvironments in well-drained tropical and subtropical soils have revealed that the OC (Beare et al., 1994), Al and Fe within aggregates respond more quickly to the environmental conditions than those in the bulk soil (Barthès et al., 2008). The bulk soils at Karkloof and Eshowe had higher average TOC than those at Cedara, and the bulk soils' average Al and Fe contents were higher under grasslands than in forests (Chapter 4). It is important to understand how these elements would be distributed in aggregate-sized fractions of these humic soils with different textures and clay mineral compositions under various climatic conditions.

Although humic soils occur under a wide range of land uses and climatic conditions in South Africa, the relationship between precipitation, temperature and TOC, Al and Fe in different aggregate size fractions in these soils is not clear. While studies on most soils, especially in the temperate region, show that OC is increased by biomass addition, where nutrients are not limiting, humic soils are highly leached, with low base status and phosphorus availability is limited by high acidity. Understanding the dynamics of carbon storage/loss, determining the effect on the physical environment of the soil, assessing biomass productivity, and directing ecosystem management and conservation efforts all require research into the distribution of SOC, Al, and Fe in various aggregate size fractions within native grassland and forest ecosystems. This information gives policymakers an evidence-based foundation for implementing sustainable land management techniques and enhancing ecosystem services (Inagaki et al., 2019). Therefore, the main objectives of this study were to investigate how different site conditions affect (i) AS and size distribution, and (ii) the distribution of total Al, Fe and OC within different aggregate size fractions in some humic soils under native forest and grassland.

6.2 Materials and methods

6.2.1 Description of soils and sites

The soils from the grassland sites at Cedara (Section 3.1.1) and Karkloof (Section 3.1.2) and from the forest site at Eshowe (Section 3.1.3) were used for this study. The soil sampling procedure was described in Section 3.2.

6.2.2 Aggregate stability

The method used has been described in Section 3.3.2. The TOC and total Al and Fe in the aggregates were analysed as described in Sections 3.3.3 and 3.3.4.

6.3 Statistical analysis

Data were analysed with Genstat 18th edition (VSN International, 2015) by a two-way analysis of variance ($p \leq 0.05$) to test the effects of site, depth, and their interaction for individual aggregate size fractions. Differences between the means of the significant factor were assessed with Duncan's multiple range test ($p \leq 0.05$). Least significant differences (LSD) at $p = 0.05$ were computed to separate treatment means for all properties. Linear regression analyses between TOC and total Al or Fe were also performed separately for the forest soils at Eshowe and those under grassland at Cedara and Karkloof. All results were based on three replications in the field.

6.4 Results

6.4.1 Aggregate size distribution

The distribution of the aggregate size fractions differed depending on site and depth ($p \leq 0.05$) (Table 6.1; Appendix 6.1). The proportion of the large macro-aggregate (LM) fraction was higher at Karkloof than Cedara and Eshowe at all depths. Although the LM proportion showed no significant differences with depth at Cedara, this fraction was greater in the 0-80 cm depth at

Karkloof and in the top 15 cm at Eshowe. The proportion of the small macro-aggregates (SM) was higher at Cedara than Karkloof at all depths. With the exception of the 20-30 and 80-100 cm layers at Karkloof, no significant differences were found with depth in the SM fraction under both grasslands. At Eshowe, the SM fraction increased with depth to about 80 cm and then suddenly declined. The proportion of SM was, however, higher at Eshowe than Cedara and Karkloof at 30-60 cm depth with no differences at all other depths.

The proportion of micro-aggregates (M) was higher ($p \leq 0.05$) at Cedara than Karkloof at all depths with the exception of the 80-100 cm layer. The M fraction was higher at Cedara than Eshowe in the 5-15 cm depth while Eshowe was higher than Cedara in the 30-100 cm depth with no differences in other layers. The M proportion did not change with depth under the grasslands but increased with depth at Eshowe (native forest).

The Cedara grassland soil had a significantly higher proportion of the silt + clay (SC) fraction than the other sites at 0-10 and 50-80 cm depths with no differences in other layers. The SC fraction did not change with depth at all sites. The average (0-100 cm) aggregate size distribution followed the order of LM (75 %) > SM (17 %) > M (4 %) = SC (4 %) at Karkloof, LM (42 %) > SM (38 %) > M (14 %) > SC (7 %) at Cedara, and SM (44 %) > LM (32 %) > M (19 %) > SC (5 %) at Eshowe. The mean weight diameter (MWD) was higher ($p \leq 0.05$) at Karkloof than the other sites at all depths with the exception of the 80-100 cm layer where Cedara was not significantly different from Karkloof (Table 6.1).

Table 6. 1: Distribution of water stable aggregates (LM: large macro-aggregates, SM: small macro-aggregates, M: micro-aggregates, SC: silt + clay) and the mean weight diameter (MWD) of humic soil profiles (0-100 cm) from Eshowe forest, and Cedara and Karkloof grasslands (n = 3).

Site	Depth (cm)	Aggregate size distribution ……………………. (Mass %)……………...				MWD (mm)
		LM	SM	M	SC	
		> 2000 μm	250-2000 μm	63-250 μm	< 63 μm	
Cedara	0-5	48.4^{fgh}	31.9^{cdef}	14.3^{defgh}	5.4^{bcdef}	1.4^{efg}
Karkloof		83.6^{no}	10.3^{a}	5.6^{ab}	0.9^{a}	1.8^{kl}
Eshowe		65.7^{jkl}	19.8^{ab}	11.8^{abcdef}	2.7^{ab}	1.5^{ghij}
Cedara	5-10	39.7^{defg}	34.7^{cdefg}	18.9^{hi}	6.7^{efg}	1.3^{ef}
Karkloof		85.4^{o}	8.3^{a}	2.6^{a}	3.7^{abcde}	1.7^{l}
Eshowe		58.4^{hijk}	29.4^{bcde}	8.4^{abc}	3.8^{abc}	1.5^{ghij}
Cedara	10-15	41.1^{defg}	38.3^{defg}	15.6^{fghi}	5.3^{bcdef}	1.4^{efg}
Karkloof		80.6^{mno}	11.6^{a}	3.1^{a}	4.7^{bcdef}	1.8^{kl}
Eshowe		53.4^{ghij}	33.6^{cdef}	8.7^{abcde}	4.3^{abcd}	1.4^{fgh}
Cedara	15-20	36.2^{cdefg}	43.7^{fgh}	14.4^{defgh}	5.7^{bcdef}	1.3^{efg}
Karkloof		69.1^{klmn}	19.7^{ab}	4.9^{ab}	6.3^{defg}	1.7^{jkl}
Eshowe		31.8^{cd}	46.7^{cdefg}	15.0^{cdefgh}	6.5^{bcde}	1.1^{abc}
Cedara	20-30	35.6^{cdef}	40.6^{fg}	16.4^{ghi}	7.4^{fg}	1.3^{a}
Karkloof		65.5^{jklm}	25.6^{bc}	5.0^{ab}	3.9^{abcde}	1.7^{ijkl}
Eshowe		23.7^{bc}	48.5^{ghi}	22.2^{ijk}	5.6^{bcdefg}	1.1^{abc}
Cedara	30-40	40.0^{defg}	38.5^{efg}	15.0^{efghi}	6.5^{defg}	1.4^{efg}
Karkloof		75.6^{lmno}	19.5^{ab}	3.8^{a}	1.1^{a}	1.7^{kl}
Eshowe		11.6^{ab}	57.5^{i}	25.3^{jkl}	5.6^{bcdefg}	1.0^{ab}
Cedara	40-50	43.5^{defgh}	35.4^{cdefg}	14.5^{fghi}	6.6^{bcdefg}	1.4^{efg}
Karkloof		76.0^{lmno}	17.2^{ab}	2.9^{a}	3.9^{abcde}	1.8^{kl}
Eshowe		7.1^{a}	55.9^{i}	29.1^{l}	7.9^{fg}	0.9^{a}
Cedara	50-60	43.3^{defgh}	36.0^{cdefg}	14.4^{defgh}	6.3^{cdefg}	1.4^{efg}
Karkloof		77.5^{lmno}	17.6^{ab}	3.9^{a}	1.0^{a}	1.8^{kl}
Eshowe		11.3^{ab}	56.7^{i}	26.3^{kl}	5.7^{bcdef}	1.0^{ab}
Cedara	60-80	40.0^{defg}	42.7^{fgh}	9.4^{bcdefg}	7.9^{g}	1.5^{fgh}
Karkloof		76.2^{lmno}	17.3^{ab}	2.7^{a}	3.8^{abcde}	1.8^{kl}
Eshowe		22.5^{bc}	52.6^{hi}	21.0^{ijk}	3.9^{abcde}	1.1^{bcd}
Cedara	80-100	48.0^{efgh}	37.5^{cdefg}	8.3^{abcd}	6.1^{cdefg}	1.5^{fgh}
Karkloof		62.6^{ijkl}	25.8^{bcd}	5.7^{ab}	5.9^{cdefg}	1.6^{hijk}
Eshowe		31.0^{cde}	42.9^{fgh}	19.4^{hij}	6.7^{egf}	1.3^{cde}
Cedara	0-100	41.6^{ab}	37.8^{b}	14.1^{ab}	6.5^{b}	1.38^{a}
Karkloof		75.2^{b}	17.3^{a}	4.0^{a}	3.5^{a}	1.75^{b}
Eshowe		31.6^{a}	44.4^{b}	18.7^{b}	5.3^{ab}	1.20^{c}

Means within the same column followed by different letters are significantly different ($p \leq 0.05$).

The average (0-100 cm) MWD values followed the order: Karkloof (1.75 mm) > Cedara (1.38 mm) > Eshowe (1.20 mm) ($p \leq 0.05$).

6.4.2 Relationship between mean weight diameter and some selected soil properties in bulk soils

Table 6.2 gives the correlation coefficients between MWD and some bulk soil properties at 0-30 cm (topsoil), 30-100 cm (subsoil) and 0-100 cm (whole soil profile) depths across the studied sites. The MWD correlated positively with TOC (r=0.67) in the top 30 cm and with clay (r= 0.58) in the 30-100 cm depth. The MWD correlated positively with silt and exchangeable acidity at all depths and with both Al and Fe in the 30-100 and 0-100 cm depths in which it was negatively correlated with sand.

Table 6. 2: Pearson's correlation coefficient between mean weight diameter (MWD) and some soil properties at 0-30, (topsoil), 30-100 (subsoil) and 0-100 cm (whole soil profile) depths across the three sites (n =3).

Parameter	MWD		
	0-30 cm	30-100 cm	0-100 cm
Ph	-0.29	-0.11	-0.19
Clay	0.18	**0.58**	0.40
Silt	**0.71**	**0.76**	**0.73**
Sand	-0.42	**-0.80**	**-0.62**
TOC	**0.67**	-0.44	0.29
Al	0.42	**0.88**	**0.61**
Fe	0.34	**0.83**	**0.57**
Exchangeable Ca	-0.18	-0.05	**-**0.06
Exchangeable Mg	-0.22	-0.05	-0.08
Exchangeable K	-0.37	0.03	-0.14
Exchangeable acidity	**0.62**	**0.42**	**0.51**

TOC: total organic carbon; Al: total aluminium; Fe: total iron; Ca: calcium; Mg: magnesium; K: potassium. Correlation coefficients in bold are statistically significant at $p \leq 0.05$.

6.4.3 Total organic carbon in the aggregate size fractions

The distribution of TOC in different aggregate fractions differed significantly ($p \leq 0.05$) between sites at all depths (Table 6.3; Appendix 6.2). The TOC content of the LM fraction was higher at Karkloof than the other sites at all depths with the exception of the 60-100 cm depth where Eshowe was the highest. Karkloof also had higher TOC in the SM fraction at all depths with the exception of the 40-100cm depth where Eshowe was higher than the other sites.

Table 6. 3: Total organic carbon (g kg^{-1}) in different aggregate size fractions (LM: large macro-aggregates; SM: small macro-aggregates; M: micro-aggregates; SC: silt +clay) at each of the sampled depths under Eshowe forest, and Cedara and Karkloof grasslands (n = 3).

Aggregate size fraction	Site	0-5	5-10	10-15	15-20	20-30	30-40	40-50	50-60	60-80	80-100	0-100
		……………………………………..……………… (cm)……………… ……………………….......										
LM (> 2000 μm)	Cedara	10.9^{b}	6.4^{a}	4.1^{a}	2.5^{a}	3.1^{a}	2.1^{a}	1.6^{a}	1.4^{a}	1.8^{ab}	1.1^{bc}	3.5^{a}
	Karkloof	43.7^{g}	37.3^{f}	28.3^{f}	28.3^{g}	19.1^{e}	10.3^{e}	8.6^{e}	4.0^{d}	2.1^{bc}	0.4^{ab}	18.2^{b}
	Eshowe	30.8^{f}	21.6^{e}	20.2^{e}	15.8^{e}	10.5^{cd}	8.2^{d}	7.1^{d}	6.4^{e}	4.0^{d}	3.3^{d}	12.8^{ab}
SM (250-2000 μm)	Cedara	13.9^{bc}	15.9^{d}	8.9^{bc}	6.0^{bc}	7.5^{abc}	3.8^{b}	3.0^{b}	2.3^{ab}	2.2^{bc}	0.3^{ab}	6.7^{ab}
	Karkloof	28.6^{f}	21.0^{e}	28.5^{f}	18.2^{f}	14.9^{de}	9.9^{e}	4.8^{c}	3.6^{cd}	2.3^{bc}	0.2^{a}	13.2^{ab}
	Eshowe	17.9^{cd}	12.4^{bc}	8.0^{b}	6.1^{bc}	5.4^{abc}	5.2^{bc}	6.1^{d}	6.1^{e}	4.3^{d}	1.5^{c}	7.3^{ab}
M (63-250 μm)	Cedara	14.8^{bc}	14.7^{cd}	7.9^{b}	6.1^{bc}	10.0^{bcd}	5.9^{c}	6.2^{d}	5.8^{e}	7.0^{e}	4.4^{e}	8.3^{ab}
	Karkloof	22.9^{e}	22.2^{e}	16.1^{d}	15.3^{e}	16.4^{e}	12.3^{f}	8.8^{e}	5.6^{e}	2.8^{c}	0.3^{ab}	12.3^{ab}
	Eshowe	6.2^{a}	4.8^{a}	3.9a	4.6^{b}	4.5^{ab}	4.9^{bc}	4.8^{c}	6.0^{e}	4.6^{d}	4.7^{e}	4.9^{ab}
SC (< 63 μm)	Cedara	20.2^{de}	20.9^{e}	10.7^{c}	7.5^{cd}	5.7^{abc}	5.8^{c}	4.2^{bc}	2.7^{bc}	1.3^{a}	0.4^{ab}	7.9^{ab}
	Karkloof	5.7^{a}	5.7^{a}	4.7^{a}	4.7^{b}	5.3^{abc}	4.1^{b}	3.0^{b}	2.2^{ab}	1.3^{a}	0.1^{a}	3.7^{a}
	Eshowe	16.6^{cd}	10.3^{b}	8.6^{bc}	8.8^{d}	9.7^{bcd}	10.0^{e}	8.7^{e}	8.5^{f}	8.2^{f}	8.9^{f}	9.8^{ab}

Values followed by different letters in a column are statistically different for each fraction-site treatment per depth at $p \leq 0.05$.

The TOC in the M fraction was higher at Karkloof to a depth of 50 cm while Cedara was higher in the 60-80 cm layer and similar to Eshowe in the 80-100 cm layer. The SC fraction had a higher TOC content in the 0-20 cm depth at Cedara while Eshowe had higher values than the other sites at 30-100 cm depth. The TOC declined with depth within all the aggregate size fractions at all sites ($p \leq 0.05$; Table 6.3). The average (0-100 cm) TOC results were in the order of M = SC (8 g kg^{-1}) > SM (6 g kg^{-1}) > LM (4 g kg^{-1}) at Cedara, LM (18 g kg^{-1}) > SM (13 g kg^{-1}) = M (12 g kg^{-1}) > SC (4 g kg^{-1}) at Karkloof, and LM (13 g kg^{-1}) > SC (10 g kg^{-1}) > SM (7 g kg^{-1}) > M (5 g kg^{-1}) at Eshowe (Table 6.3).

6.4.4 Total aluminium and iron in the aggregate size fractions

The distribution of Al in the different aggregate size fractions differed significantly between the sites and at all depths (Table 6.4; Appendix 6.2). Eshowe had lower Al content than Cedara and Karkloof at all depths and within all aggregate size fractions. The Al content of the LM fraction was higher at Karkloof at all depths with the exception of the 0-10 cm depth where Cedara was higher and the 10-20 cm depth that was not different between the two grassland sites. Karkloof had higher Al in the SM fraction at all depths with the exception of the 0-5 cm layer where Cedara was higher and the 5-10 and 80-100 cm layers which were similar at these two sites. In the 10-15, 20-30, 50-60, and 80-100 cm layers, the Al content in the M size fraction was higher at Karkloof than at Cedara. However, there were no significant differences between these sites in other depth ranges. The Al associated with SC size fraction was not different between the grassland sites at all depths with the exception of the 10-40 cm depth where Karkloof was higher than Cedara. The Al increased with depth in all the aggregate size fractions at all sites ($p \leq 0.05$; Table 6.4). The average (0-100 cm) Al content was not significantly different between aggregate size fractions at all the studied sites (Table 6.4).

Table 6. 4: Total aluminium (g kg^{-1}) in different aggregate size fractions (LM: large macro-aggregates; SM: small macro-aggregates; M: micro-aggregates; SC: silt +clay) at each of the sampled depths under Eshowe forest, and Cedara and Karkloof grasslands (n =3).

Aggregate size fraction	Site	0-5	5-10	10-15	15-20	20-30	30-40	40-50	50-60	60-80	80-100	0-100
		……………………………………					(cm)	…………………………………				
LM (> 2000 µm)	Cedara	14.2^{gh}	15.4^{f}	15.9^{d}	16.9^{de}	17.3^{f}	19.6^{d}	21.5^{e}	17.3^{cde}	18.1^{c}	17.9^{d}	17.4^{cd}
	Karkloof	12.6^{e}	12.9^{d}	15.9^{d}	18.1^{ef}	20.1^{g}	22.7^{e}	24.5^{f}	24.2^{g}	24.2^{f}	24.8^{g}	20.0^{d}
	Eshowe	7.8^{a}	7.1^{a}	8.0^{a}	9.6^{b}	12.4^{bc}	12.4^{a}	14.5^{a}	14.7^{b}	15.1^{b}	15.7^{b}	11.7^{ab}
SM (250-2000 µm)	Cedara	17.6^{i}	16.8^{g}	16.9^{d}	16.6^{d}	17.4^{f}	17.8^{c}	19.4^{d}	17.7^{de}	20.1^{d}	20.9^{f}	18.1^{cd}
	Karkloof	14.9^{h}	16.7^{g}	18.7^{e}	19.4^{g}	22.5^{h}	23.6^{e}	22.2^{e}	22.4^{f}	22.7^{e}	21.7^{f}	20.1^{d}
	Eshowe	10.3^{c}	10.6^{c}	12.5^{b}	10.2^{b}	11.1^{ab}	11.8^{a}	14.1^{a}	15.1^{b}	15.1^{b}	16.8^{c}	12.8^{abc}
M (63-250 µm)	Cedara	13.9^{fgh}	14.3^{e}	13.9^{c}	16.2^{d}	15.8^{e}	18.2^{cd}	17.4^{bc}	15.8^{bc}	18.2^{c}	16.9^{cd}	16.1^{bcd}
	Karkloof	14.3^{gh}	13.9^{de}	16.2^{d}	17.0^{de}	21.8^{h}	17.7^{c}	16.6^{b}	18.3^{e}	19.2^{cd}	19.1^{e}	17.4^{cd}
	Eshowe	9.1^{b}	8.9^{b}	9.1^{a}	7.7^{a}	10.2^{a}	11.2^{a}	12.9^{a}	8.5^{a}	10.1^{a}	9.6^{a}	9.7^{a}
SC (< 63 µm)	Cedara	13.0^{ef}	13.2^{de}	14.0^{c}	14.0^{c}	13.1^{cd}	14.2^{b}	17.3^{bc}	17.8^{de}	18.2^{c}	19.7^{e}	15.5^{bcd}
	Karkloof	13.4^{efg}	13.8^{de}	16.9^{d}	18.6^{fg}	22.9^{h}	19.2^{cd}	18.7^{cd}	18.9^{e}	19.2^{cd}	19.0^{e}	18.1^{cd}
	Eshowe	11.4^{d}	10.5^{c}	12.0^{b}	10.2^{b}	13.9^{d}	14.7^{b}	16.5^{b}	16.3^{bcd}	15.9^{b}	17.5^{cd}	13.9^{abc}

Values followed by different letters in a column are statistically different for each fraction-site treatment per depth at $p \leq 0.05$.

The distribution of Fe in different aggregate size fractions also differed significantly between sites at all depths (Table 6.5; Appendix 6.2). With the exception of the 60-100 cm depth in the SC size fraction, the Fe content was lower at Eshowe than Cedara and Karkloof at all depths and in all fractions. The Fe content of the LM was higher at Cedara than Karkloof at 0-15 and 60-100 cm depths while Karkloof was higher in the 30-40 cm layer. The Karkloof site had higher Fe in the SM at 30-60 and 80-100 cm depths while Cedara was higher at 0-10 cm. The Fe in the M size fraction was higher at Karkloof at 10-60 cm depth while Cedara was higher at 0-10 and 40-50 cm depths. Karkloof had higher Fe associated with SC at 10-40 cm depth while Cedara was higher than the other sites at 60-100 cm depth with no significant differences between sites at all other depths. Except for the M size fraction at Eshowe, the Fe increased with depth in all the aggregate fractions and at all sites ($p \leq 0.05$) (Table 6.5). The average (0-100 cm) Fe content was in the order of LM (15 g kg^{-1}) > SM (12 g kg^{-1}) = M (11 g kg^{-1}) = SC (10 g kg^{-1}) at Cedara with no differences in other sites (Table 6.5).

6.4.5 Relationships between carbon, aluminium and iron

There were few significant relationships between the measured variables and only the most significant are discussed.

6.4.5.1 Between bulk soils and aggregate size fractions

The TOC in bulk soils (Section 4.4.1; Table 4.1) was positively correlated with the proportion of LM at Karkloof (R^2=0.47) and Eshowe (R^2=0.68) (Figure 6.1a) but negatively with the SM proportion at Cedara (R^2= -0.47) and Eshowe (R^2= -0.58) at 0-30 cm depth (Figure 6.1d). The Al (R^2= -0.63; Figure 6.1b) and Fe (R^2= -0.71; Figure 6.1c) were negatively correlated with LM in the 0-30 cm depth at Karkloof but positively correlated to SM at the same site and soil depth (Figures 6.1e and f). The Al also correlated positively with SM in the 0-30 cm depth at Cedara (R^2=0.60; Figure 6.1e). No relationships observed for LM and SM below 30 cm at all sites.

Table 6. 5: Total iron (g kg^{-1}) in different aggregate size fractions (LM: large macro-aggregates; SM: small macro-aggregates; M: micro-aggregates; SC: silt +clay) at each of the sampled depths under Eshowe forest, and Cedara and Karkloof grasslands (n = 3).

Aggregate size fraction	Site	0-5	5-10	10-15	15-20	20-30	30-40	40-50	50-60	60-80	80-100	0-100
		……………	……………	……………	……………	……………	……………	(cm)……	……………	……………	……………	……………
LM (> 2000 µm)	Cedara	16.7^{i}	16.1^{g}	14.9^{g}	13.9^{e}	13.3^{d}	12.1^{d}	13.9^{e}	12.9^{ef}	18.0^{f}	19.2^{i}	15.1^{e}
	Karkloof	9.3^{fg}	10.8^{f}	11.5^{f}	13.0^{e}	15.0^{e}	17.8^{g}	14.7^{e}	14.6^{f}	14.5^{e}	14.6^{gh}	13.6^{de}
	Eshowe	4.4^{a}	4.2^{a}	5.3^{a}	5.4^{a}	5.3^{a}	6.9^{a}	6.8^{a}	6.8b	7.6b	8.3b	6.1^{a}
SM (250-2000 µm)	Cedara	10.3^{h}	9.9^{e}	10.5^{def}	10.5^{d}	12.5^{cd}	11.5^{cd}	12.6^{d}	10.9^{cd}	14.5^{e}	13.7^{fg}	11.7^{cd}
	Karkloof	8.6^{ef}	8.9^{d}	10.0^{cd}	13.2^{e}	12.8^{cd}	14.6^{f}	14.0^{e}	14.4^{f}	14.8^{e}	14.8^{h}	12.6^{cde}
	Eshowe	7.5^{cd}	7.3^{c}	7.6^{b}	6.6^{b}	6.1^{a}	6.7^{a}	6.9^{a}	6.5^{b}	6.8^{ab}	8.2^{b}	7.0^{ab}
M (63-250 µm)	Cedara	10.2^{gh}	10.4^{ef}	10.2^{cde}	9.9^{d}	11.7^{c}	13.3^{e}	12.7^{d}	10.4^{cd}	13.1^{d}	12.1^{de}	11.4^{cd}
	Karkloof	9.2^{f}	9.1^{d}	11.7^{f}	15.2^{f}	18.1^{g}	14.6^{f}	10.4^{bc}	11.6^{de}	12.4^{d}	12.9^{ef}	12.5^{de}
	Eshowe	6.1^{b}	6.2^{b}	5.9^{a}	6.1^{ab}	5.1^{a}	6.8^{a}	7.1^{a}	3.7^{a}	6.3^{a}	5.8a	5.9^{a}
SC (< 63 µm)	Cedara	8.2^{de}	8.3^{d}	8.9^{c}	10.8^{d}	9.6^{b}	10.7^{c}	10.9^{c}	11.6^{de}	12.2^{d}	12.9^{ef}	10.4^{bcd}
	Karkloof	7.9^{cde}	9.0^{d}	11.3^{ef}	12.9^{e}	16.4^{f}	13.3^{e}	10.9^{c}	10.7^{cd}	10.7^{c}	10.9^{c}	11.4^{cd}
	Eshowe	7.1^{c}	7.2^{c}	8.8^{c}	8.4^{c}	8.8^{b}	8.3^{b}	9.7^{b}	9.4c	10.6^{c}	11.5cd	8.9^{abc}

Values followed by different letters in a column are statistically different for each fraction-site treatment per depth at $p \leq 0.05$.

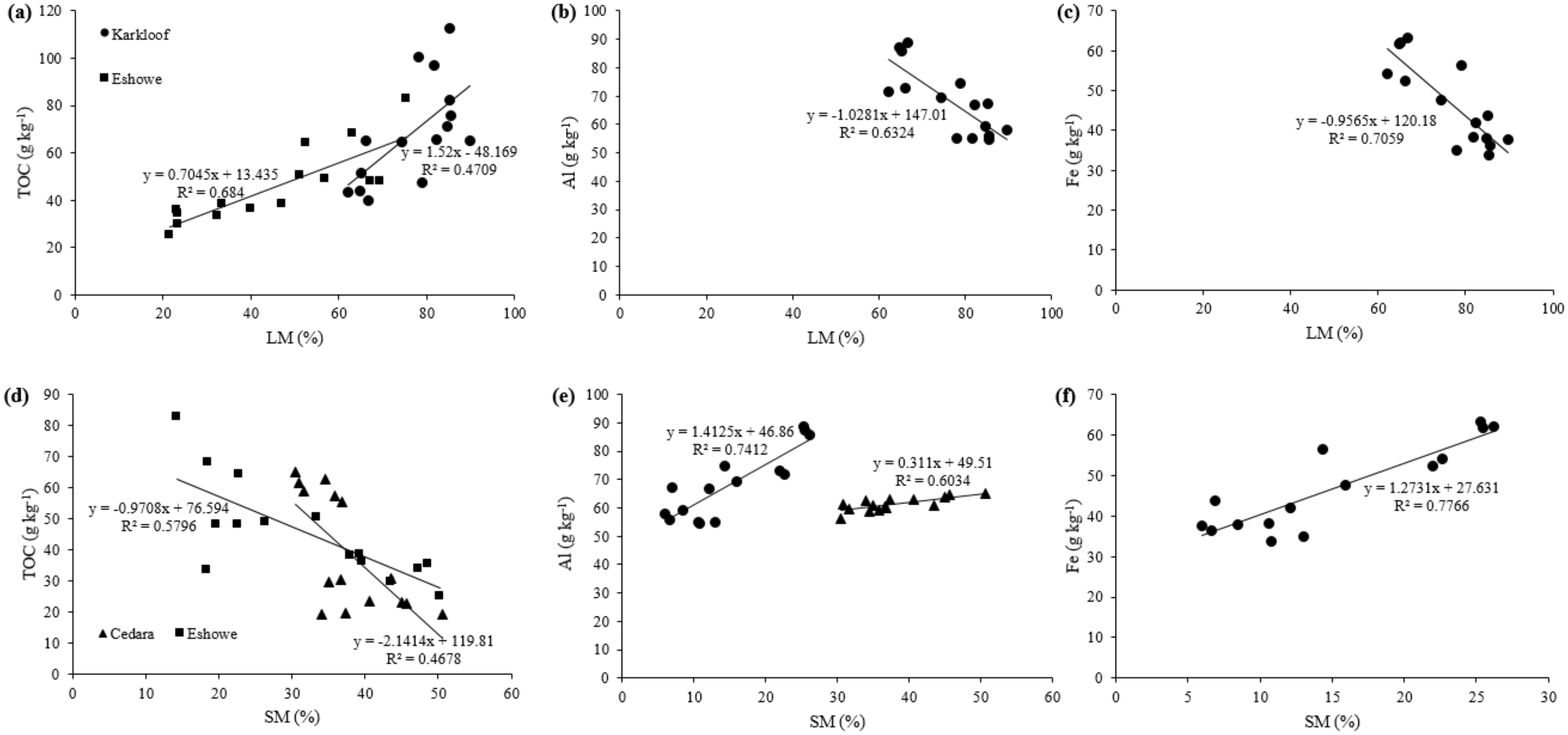

Figure 6. 1: Relationship between large macro-aggregates (LM) and (a) total organic carbon (TOC), (b) total aluminium (Al) and (c) total iron (Fe) and between small macro-aggregates (SM) and (d) TOC, (e) total Al and (f) total Fe in the humic topsoils (0-30 cm) under Eshowe forest, and Karkloof and Cedara grasslands.

At 30-100 cm the TOC was positively correlated with M at Cedara (R^2=0.49) and Eshowe (R^2=0.57) (Figure 6.2a) while Fe was negatively correlated with M at the same depth at Cedara (R^2= 0.48; Figure 6.2b). The TOC (R^2= 0.48; Figure 6.2c) correlated negatively with the SC in the top 30 cm at Karkloof.

6.4.5.2 Within the aggregate size fractions

Within the LM size fraction, the TOC correlated negatively with Al at 0-30 cm depth at all sites (Figure 6.3a) and with Fe at 0-30 cm depth at Cedara (R^2=0.66; Figure 6.3b). The correlation between Al and Fe was positive at 0-30 cm depth at Karkloof (R^2=0.88; Figure 6.3c) but negative at the same depth at Cedara (R^2= 0.79; Figure 6.3c) and at 30-100 cm depth at Karkloof (R^2= 0.77; Figure 6.3e).

The relationship between TOC and Al was negative at the 0-30 cm depth (R^2= 0.50; Figure 6.4a) in the SM fraction but positive at the 30-100 cm depth (R^2=0.60; Figure 6.4d) at Karkloof. Within the SM size fraction, the TOC correlated negatively with Fe in the top 30 cm at Karkloof (R^2= 0.52) and Eshowe (R^2= 0.78) (Figure 6.4b). The relationship between Al and Fe in the SM was positive in the top 30 cm at Karkloof (R^2=0.63) and at 30-100 cm depth at Cedara (R^2=0.59) and Eshowe (R^2=0.56) (Figure 6.4e).

In the M fraction, the TOC correlated negatively with Fe in the 0-30 cm depth at Karkloof (R^2= 0.54; Figure 6.5a) while the correlation between Al and Fe was positive at the same depth at Karkloof (R^2= 0.85; Figure 6.5b) and at the 30-100 cm depth at Cedara (R^2=0.56) and Eshowe (R^2=0.62) (Figure 6.5c). The TOC correlated negatively with Fe in the top 30 cm depth at Cedara (R^2= 0.61; Figure 6.6a) in the SC fraction. The Al correlated positively with Fe in the top 30 cm at Karkloof (R^2=0.96) and Eshowe (Figure 6.6b) in the SC fraction and at the 30-100 cm depth at Cedara (R^2=0.62) and Eshowe (R^2=0.56) in the same size fraction (Figure 6.6e). At Cedara (30-100 cm), the TOC correlated negatively with Al (R^2= 0.78; Figure 6.6c) and Fe (R^2= 0.73; Figure 6.6d) in the SC fraction.

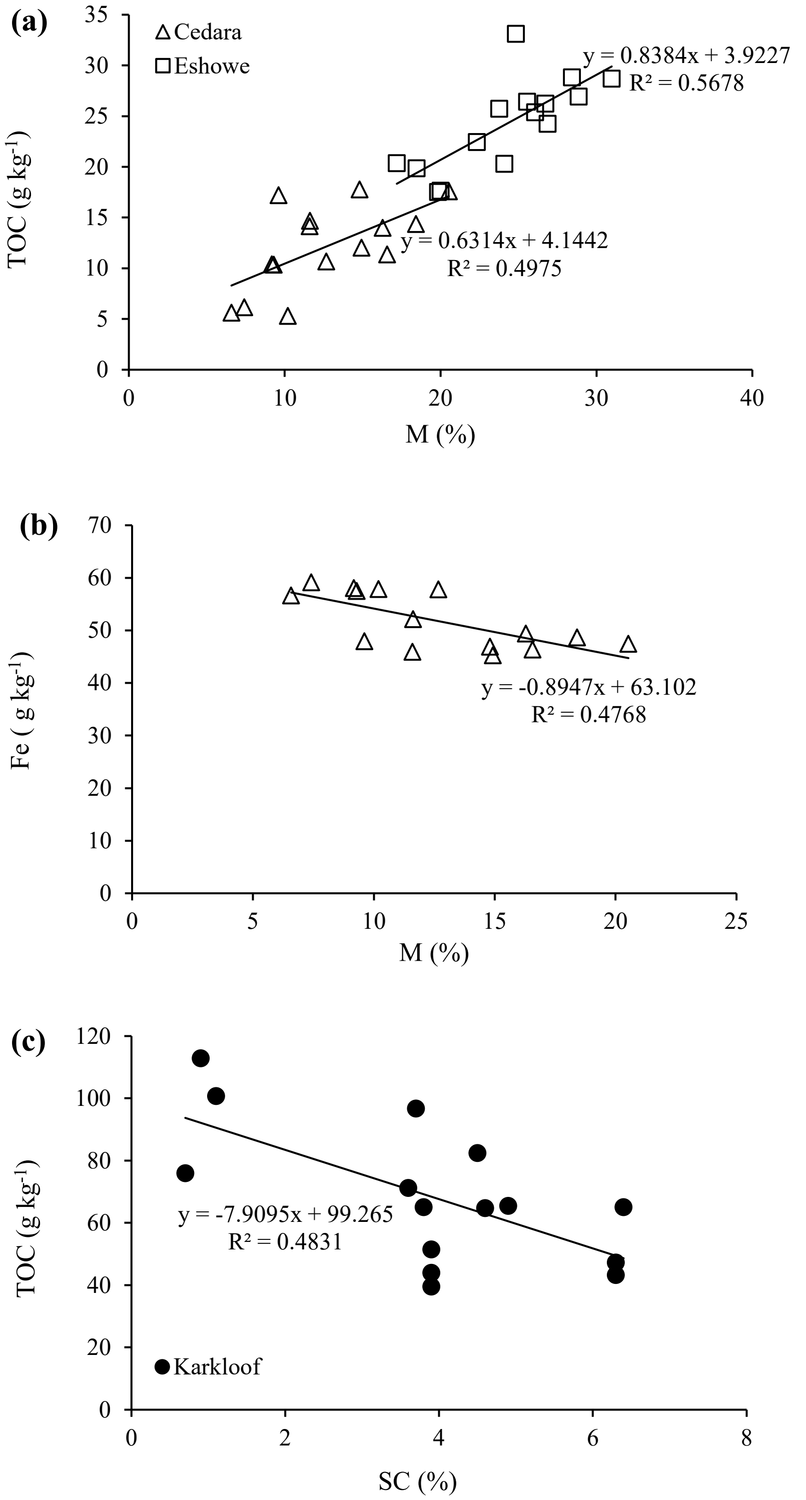

Figure 6. 2: Relationship between micro-aggregates (M) and (a) total organic carbon (TOC) and (b) total iron (Fe) in subsoils (30 - 100 cm) at Eshowe and Cedara and (c) TOC and the silt + clay (SC) fraction in the Karkloof topsoil (0-30 cm).

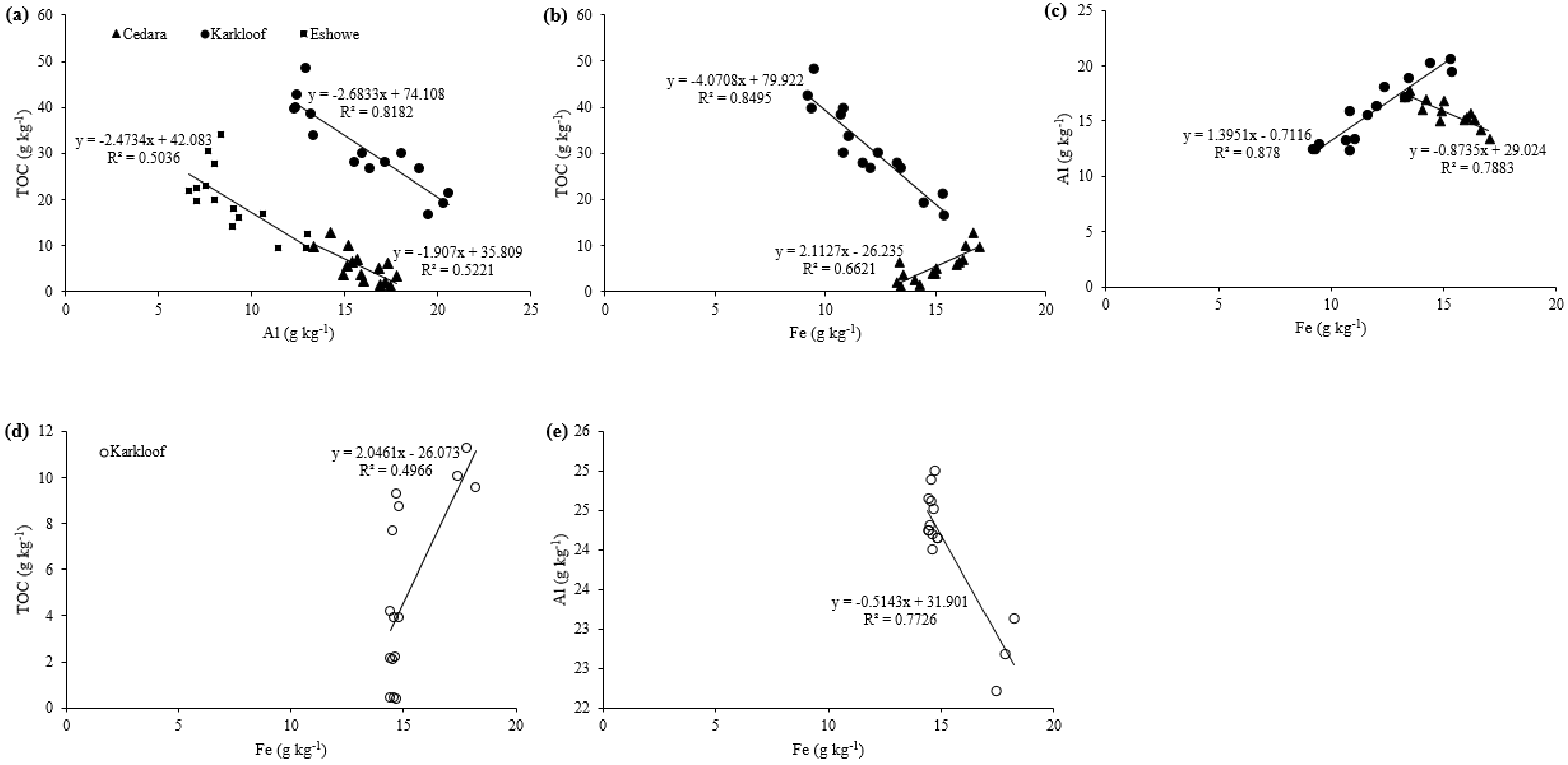

Figure 6. 3: Relationship between total organic carbon (TOC) and (a) total aluminium (Al) and (b) total iron (Fe), (c) total Al and total iron (Fe) in the humic topsoils (0-30 cm) and (d) TOC and total Fe and (e) total Al and total Fe in the subsoils (30-100 cm) of the large macro-aggregates under Eshowe forest, and Karkloof and Cedara grasslands.

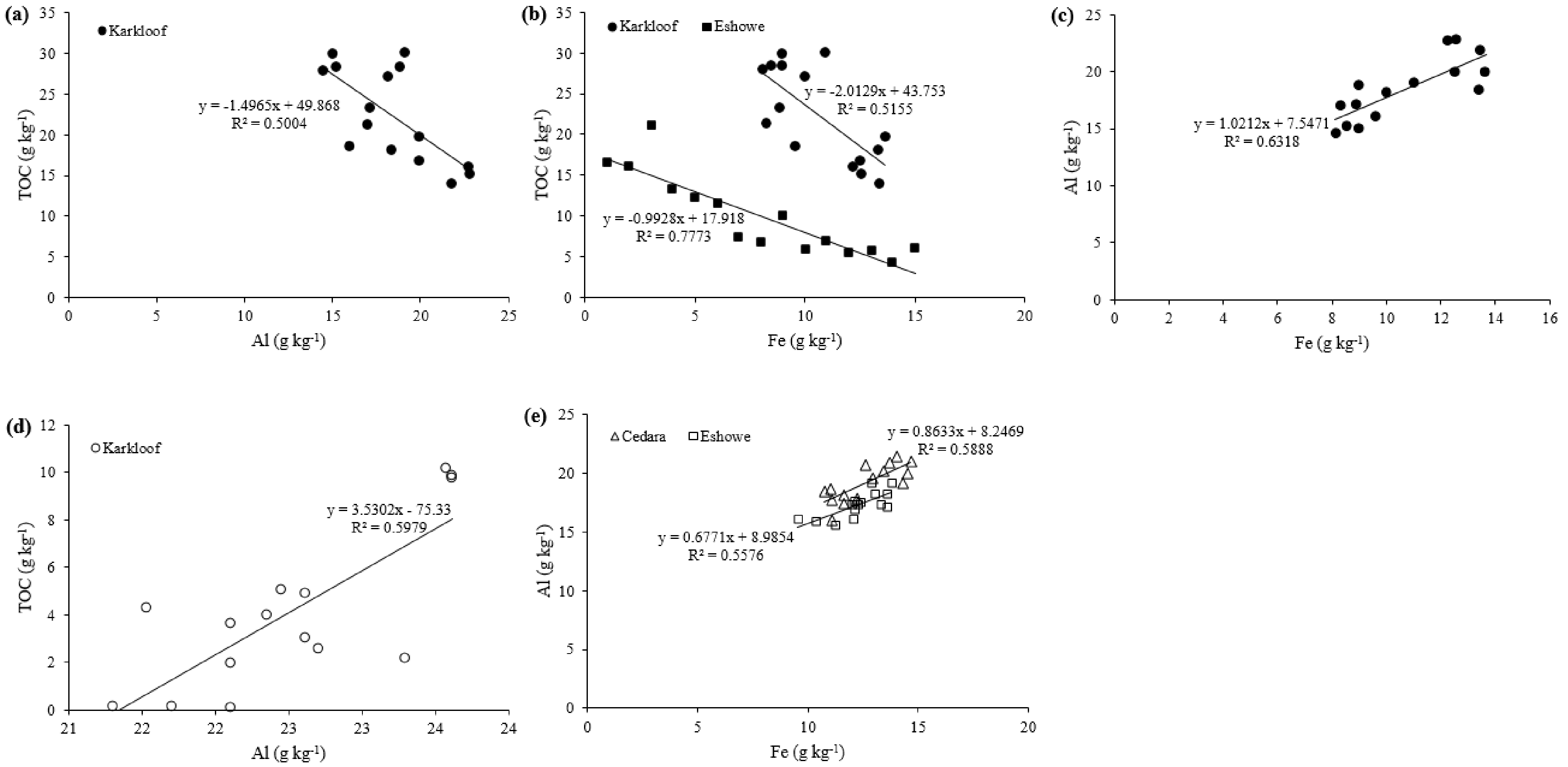

Figure 6. 4: Relationship between total organic carbon (TOC) and (a) total aluminium (Al) and (b) total iron (Fe), (c) total Al and total iron (Fe) in the humic topsoils (0-30 cm) and (d) TOC and total Al and (e) total Al and total Fe in the subsoils (30-100 cm) of the small macro-aggregates under Eshowe forest, and Karkloof and Cedara grasslands.

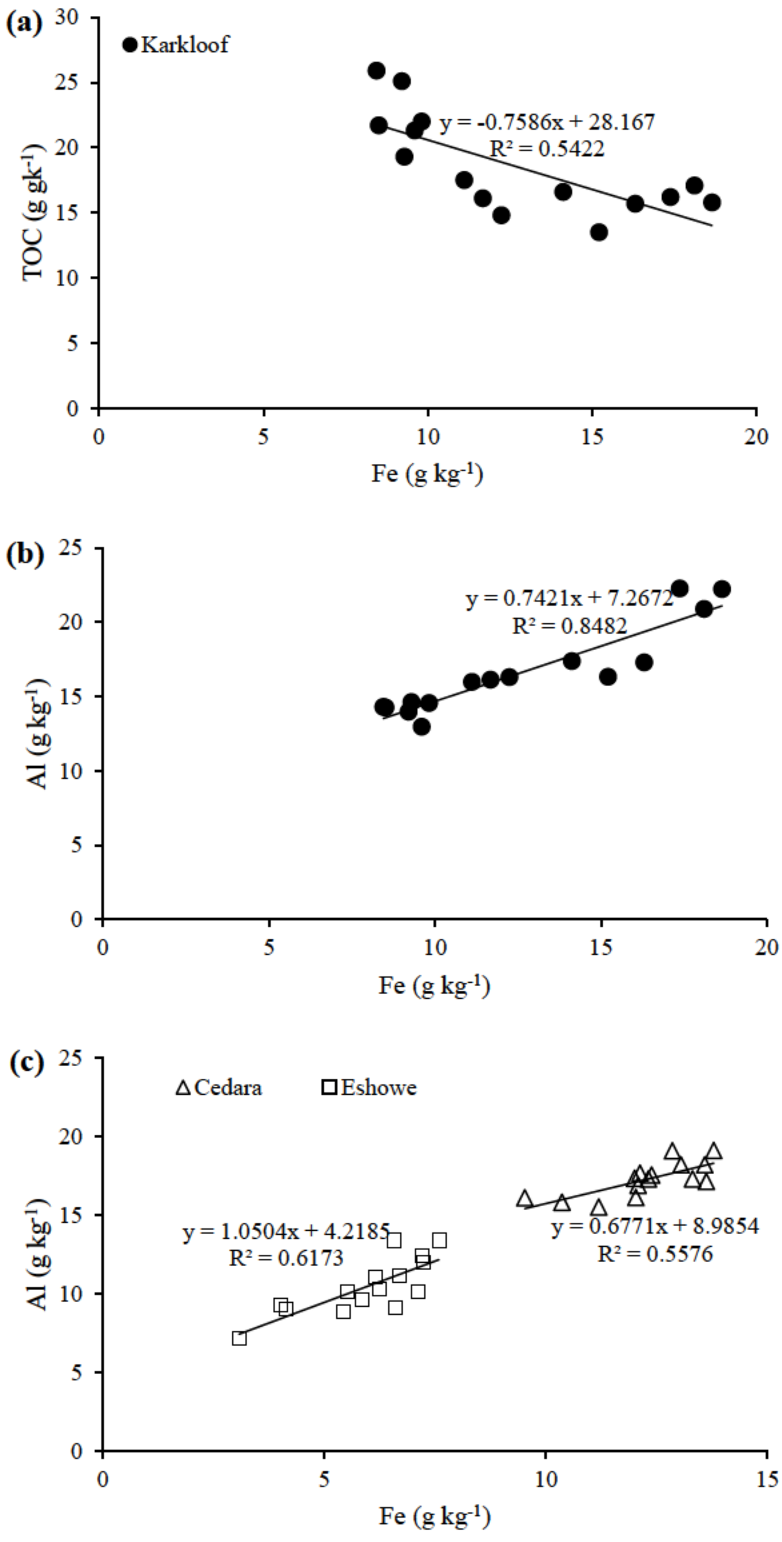

Figure 6. 5: Relationship between (a) total organic carbon (TOC) and total iron (Fe) and (b) total aluminium (Al) and total Fe in the humic topsoils (0-30 cm) and (c) total Al and total Fe in the subsoils (30-100 cm) of the micro-aggregates under Eshowe forest, and Karkloof and Cedara grasslands.

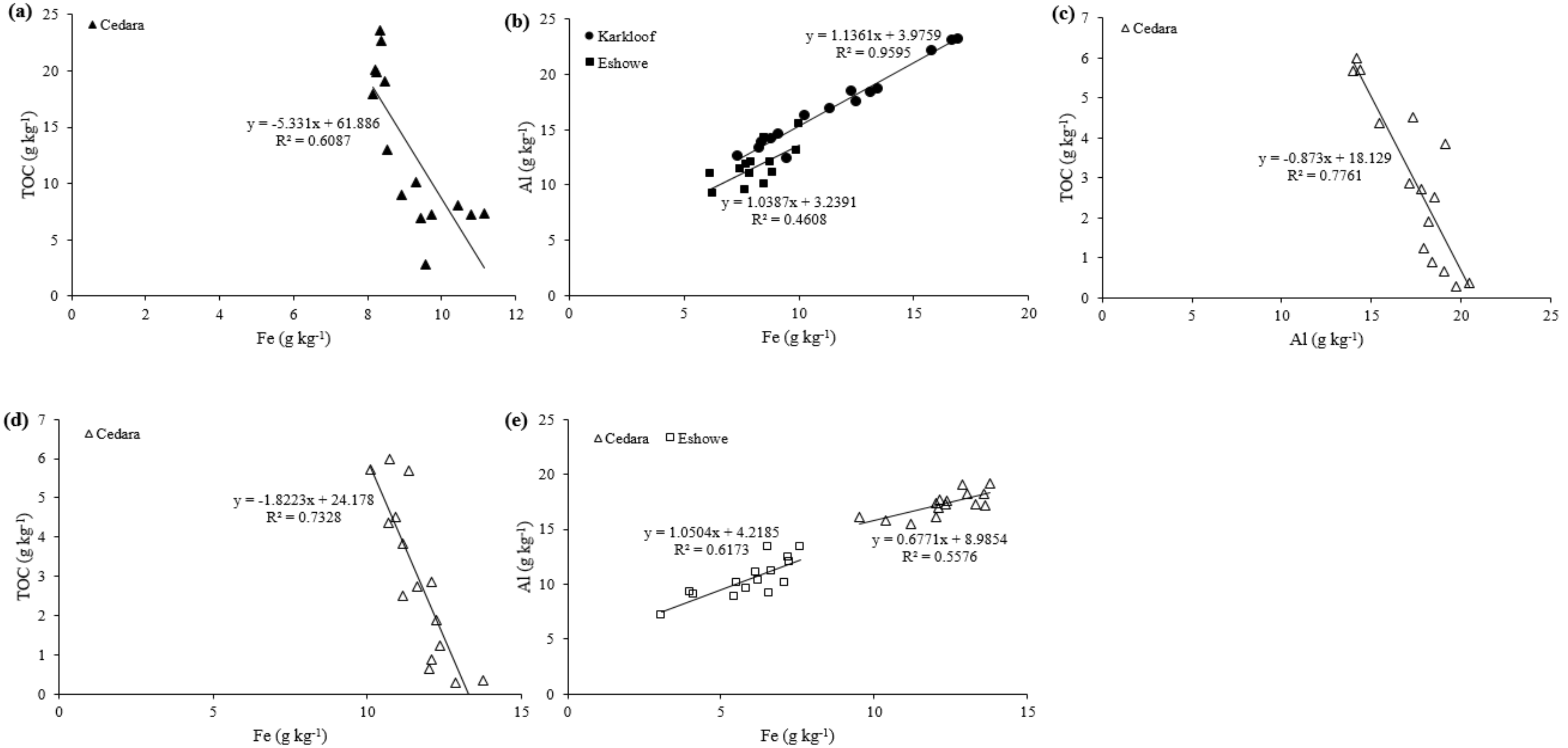

Figure 6. 6: Relationship between (a) total organic carbon (TOC) and total iron (Fe) and (b) total aluminium (Al) and total Fe in the humic topsoils (0-30 cm), TOC and (c) total Al and (d) total Fe, and (e) total Al and total Fe in the subsoils (30-100 cm) of the silt + clay fraction under Eshowe forest, and Karkloof and Cedara grasslands.

6.5 Discussion

The higher MWD at Karkloof and Cedara than Eshowe is probably due to the higher clay content at 30-100 cm depth at these sites. This is supported by the positive relationship between MWD and clay (r=0.58) at this depth, and with Al and Fe in the 30-100 and 0-100 cm depths. Generally, clay acts as an aggregating agent by binding particles together, including Fe and Al, which, in turn, increases soil AS (Wakindiki and Ben-Hur, 2002). In agreement with these findings, Evrendilek et al. (2004) reported higher MWD in soils under grassland (3.50 mm) than forest (2.89 mm) in a Typic Haploxeroll (0-20 cm) in the Taurus Mountains of the southern Mediterranean region of Turkey. This was attributed to the faster growth rate of grasses than trees that resulted in higher organic matter turnover and stability of aggregates in grassland soils as a result of a protective water-repellent lattice of long-chain polymethylene compounds around the soil aggregates. According to Blanco-Canqui and Lal (2004), grass roots enmesh fine particles into stable aggregates by (i) drying the soil environment around roots, (ii) reorienting clay particles parallel to the axis of the roots, (iii) supplying decomposable organic residues to the soil, (iv) releasing polyvalent cations and (v) supporting a large microbial population in the rhizosphere.

On the other hand, the lower MWD at Eshowe might be a result of the low clay content. In agreement with these findings, Wuddivira et al. (2010) showed a decrease in MWD with a decrease in clay content in 23 Trinidadian soil profiles. It was concluded that clay and aggregate stability are closely related, with decreased clay having a detrimental effect on aggregate stability. The relationship between these variables, however, is not always linear because clay type plays a significant role in aggregation (Nciizah and Wakindiki, 2014, Mbanjwa et al., 2022). Even though the average MWD values are significantly different between Karkloof (1.75 mm) and Cedara (1.38 mm), both are within the 1.3 to 2.0 mm range that indicates a stable soil, while the MWD of 1.20 mm at Eshowe is indicative of a partly stable soil (Le Bissonnais et al., 2002). Such stable structure under grassland favours gas and water transfer in soil, crop rooting and reduces susceptibility to erosion (Haynes, 1999; Fattet et al., 2011; Podwojewski et al., 2014).

The positive correlation between MWD and TOC in the top 30 cm (Table 6.2) is not surprising given the higher OC in the soil surface layers at all sites (Table 4.1) due to litter presence (Bronick and Lal, 2005). A similar relationship has been reported in previous studies (Chenu et al., 2000; Blanco-Canqui and Lal, 2004; Ayoubi et al., 2012) in a wide range of soils and climatic conditions and has often been attributed to the central role of SOC in the formation of stable aggregates. The positive correlation between MWD and total Al and Fe below 30 cm extends the work of Oades and Waters (1991) and Barthès et al. (2008), which showed that aggregation is controlled by both Al and Fe through cationic bridging and the formation of organo-metallic compounds and gels in acidic soils with low SOC to soils with higher OC. The positive correlation between MWD and silt at all depths and with clay below 30 cm is also common and is often attributed to the higher specific surface area of these fine particles that enables them to flocculate and bind with sand to form stable aggregates (Chenu et al., 2000; Six et al., 2000b). The positive correlation between MWD and exchangeable acidity at all depths was probably a result of the greater ability of Al to complex with soil components at low pH (Wesselink et al., 1996).

The aggregate size distribution was dominated by LM and SM fractions at the expense of the SC size fraction at all sites. This could be due to the permanent vegetation cover at the studied sites, which increased biomass addition and particulate organic carbon (POC) content, thus enhancing the formation of macro-aggregates (Cambardella and Elliot, 1992; Schwendenmann and Pendall, 2006; Fattet et al., 2011). This is supported by the positive relationship between TOC and LM in the top 30 cm at Karkloof (R^2=0.48) and Eshowe

(R^2=0.68) (Figure 6.1a) as well as higher TOC associated with the LM fraction at these sites (Table 6.4). The higher LM proportion at Karkloof than the other sites may be explained by a higher biomass production coupled with slower decomposition due to the cooler climatic conditions at this site. There is evidence (Elliot, 1986; Evrendilek et al., 2004; Rasmussen et al., 2006; Schwendenmann and Pendall, 2006; Dlamini et al., 2016) that the increased canopy and residue cover in high rainfall environments increases the POC component of the soil and ultimately increases the LM proportion. In addition, the aggregate conceptual model of Tisdall and Oades (1982) postulated that OC occluded in macro-aggregates has a short turnover time (1 to 10 years) because of the biological binding agents (polysaccharides and root hyphae) in this fraction while persistent binding agents (oxides, organic polymers and polyvalent cations) in micro-aggregates and silt plus clay fractions can last for decades (10-1000 years). Based on this model, the higher TOC associated with LM fraction at Karkloof and Eshowe has a limited longevity because the biological agents tend to rapidly decompose resulting in the breakdown of macro-aggregates which, in turn, releases the encapsulated SOC (Six et al., 2000a; Blanco-Canqui and Lal, 2004; Bronick and Lal, 2005).

The negative relationship between SM and TOC in the top 30 cm at Cedara (R^2= -0.47) and Eshowe (R^2= -0.47) (Figure 6.1d) indicates that TOC is not responsible for the stabilization of SM in these soils (Bronick and Lal, 2005). The positive relationship between SM and Al at the same depth at Cedara (R^2= 0.60; Figure 6.1e), and Karkloof (R^2= 0.74; Figure 6.1e) and between SM and Fe at Karkloof (R^2= 0.78; Figure 6.1f) suggests that, as opposed to TOC, the presence of both Al and Fe has an influence on the stabilization of the SM fraction at these sites (Igwe et al., 2009; Inagaki et al., 2019) while other factors such as silt and indirectly exchangeable acidity through Al might be responsible for stabilizing this fraction at Eshowe. At 30-100 cm depth, the TOC correlated positively with Al (R^2= 0.60; Figure 6.4d) in the SM fraction and positively with Fe (R^2= 0.50; Figure 6.3d) in the LM fraction at Karkloof, indicating the role of these cementing substances in aggregates of different sizes (Eusterhues et al., 2005; Barthès et al., 2008; Peng et al., 2015).

The higher M proportion at Cedara than Karkloof could be a result of the higher clay content at this site. A greater M proportion in fine textured soils has been found by Bronick and Lal (2005) and Rasmussen et al. (2018), and is attributed to the greater stabilizing influence of clay-sized particles because of their large surface area. Moreover, the higher TOC associated with M and SC size fractions at Cedara suggests that these aggregate sizes are critical for medium to longer term organic matter turnover due to a greater degree of TOC protection from microbial and enzymatic degradation (Tisdall and Oades, 1982). Conversion of the land use from native vegetation to cropping on these humic soils may affect the concentration of stored TOC in the different aggregate fractions. In agreement with these findings, Beare et al. (1994) reported higher SOC concentration in the micro-aggregate fraction (106-250 µm) than in both large and small macro-aggregate size-classes in a sub-tropical Ultisol under grassland in Georgia, USA. It was concluded that soil aggregation and SOC accumulation due to physical protection are intrinsically linked phenomena. On the other hand, the negative relationship between TOC and Fe (R^2= -0.54; Figure 6.5a) in the M fraction in the top 30 cm at Karkloof may possibly be indicating the limited influence of Fe on the stabilization of C in the M size fraction at this site (Amelung et al., 1998; Eusterhues et al., 2005; Kaiser and Guggenberger, 2007).

The lack of correlation between SC and both Al and Fe at all sites is in contrast with the findings of Igwe et al. (1999), who reported a strong correlation between SC and Fe (R^2= 0.86) and Al (R^2= 0.71) in Nigeria. The contradiction in the findings could be explained by the differences in the forms of Al and Fe studied, as the current study also included the forms which may be less active than the fractions that are soluble, which Igwe et al. (1999) focussed on. Generally, particular fractions of Al and Fe may be more active than others in stabilizing aggregates (Ko¨gel-Knabner et al., 2008; Igwe et al., 2009; Inagaki et al., 2019). The notable increase in Al and Fe with depth in all aggregate fractions and at all sites (except for Fe in the M size fraction at Eshowe) might be indicating the strong weathering of the parent materials, and the release of Fe from primary minerals to form secondary oxides (Rasmussen et al., 2006; Igwe et al., 2013; Inagaki et al., 2019). A positive relationship between Al and Fe within all aggregates is commonly observed (Igwe et al., 1999; Igwe et al., 2013; Peng et al., 2015) and is often attributed to the release of the two elements from the weathering of primary minerals and are related to the control of aggregation by both Al and Fe in acidic soils following flocculation at low pH.

6.6 Conclusions

The primary goals of this study were to investigate the effects of different site conditions on (i) aggregate size distribution and (ii) the distribution of total Al, Fe and OC within different aggregate size fractions in some humic soils under native forest and grassland. The study reveals that different site conditions influence AS and size distribution in humic soils under native forest and grassland. The MWD is higher at Karkloof and Cedara compared to Eshowe, primarily due to the higher clay content in the dolerite than sandstone-derived soils. The positive correlation observed between MWD and OC in the top 30 cm, exchangeable Al and silt at all depths, and with clay and total Fe and Al below 30 cm indicates the involvement of different cementing agents in aggregation formation. These agents appear to vary with depth, indicating complex interactions between vegetation, parent material, and environmental factors such as precipitation and temperature.

The aggregate size distribution at all sites is dominated by large and small macro-aggregates, with a lower representation of the SC size fraction. This dominance is attributed to the presence of permanent vegetation cover in these ecosystems. The concentration of total Al and Fe is generally higher in macro-aggregates than micro-aggregates at Karkloof and Cedara. In contrast, the SC fraction shows a higher concentration of these elements at Eshowe. This indicates that the distribution of Al and Fe within aggregates varies with factors such as vegetation type, parent material, and gradients of precipitation and temperature. This information can aid in the improvement of the understanding of soil dynamics and the creation of specific soil management techniques for various ecosystems. Studying the TOC, Al, and Fe in bulk samples and aggregate size fractions after the studied humic soils are converted to arable agriculture is, however, crucial for making informed decisions about the long-term productivity, fertility, and sustainability of these soils under cultivation.

CHAPTER 7: TOTAL ORGANIC CARBON, ALUMINIUM AND IRON IN BULK SAMPLES AND AGGREGATE SIZE FRACTIONS OF A SANDY CLAY LOAM HUMIC SOIL UNDER SUGARCANE RELATIVE TO NATIVE FOREST

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7.1 Introduction

Aggregate stability (AS) is an important soil structural component (Feller and Beare, 1997; Amèzketa, 1999; Six et al., 2000b; Bronick and Lal, 2005) as it controls the dynamics of soil organic matter (SOM) by protecting it within stable aggregates (Six et al., 2000a; Bronick and Lal, 2005; Zhao et al., 2017). Numerous authors have discussed soil aggregate formation processes (Feller and Beare, 1997; Six et al., 2000a; Bronick and Lal, 2005; Peng et al., 2015; Wang et al., 2016). Evidence from such studies suggests that AS is driven by (i) internal factors such as clay mineralogy, organic matter, aluminium (Al) and iron (Fe) contents, and exchangeable cations, and (ii) external factors including climate, soil formation processes, land use and land management.

The external factors alter the internal factors in a direct or indirect manner (Amézketa, 1999; Krull et al., 2013). For example, some studies (Six et al., 2000b; Lal, 2005; Chivenge et al., 2007; Chaplot et al., 2010; Rabbi et al., 2014; Wu et al., 2017) have shown that most of the negative effects on AS, following the conversion of natural ecosystems to arable agriculture, are largely a result of (i) a decreased supply of inputs due to management practices such as stubble burning, (ii) export of carbon (C) through the harvesting of plant matter, and (iii) higher rates of loss and reduction in soil organic carbon (SOC) with cultivation. Grohmann (1960) reported that intensive cultivation of natural forest soils reduced the percentage of aggregates larger than 2 mm by about half in both Oxisols and Ultisols from Brazil. Tisdall and Oades (1982) found that micro-aggregates (< 0.25 mm) were less affected by cropping and management than macro-aggregates. Beare et al. (1994) made a similar observation in a wide range of soils under different climates in the USA.

It is widely considered that the interaction of positively charged elements such as Fe and Al with clay or SOM can synergistically promote aggregation in soils, thereby improving structural stability through cationic bridging and formation of organo-mineral complexes (Igwe et al., 1995; Igwe et al., 1999; Dominy and Haynes, 2002; Pulleman and Marinissen, 2004; Denef and Six, 2005; Igwe et al., 2005; Kaiser and Guggenberger, 2007; Six and Paustian, 2014; Peng et al., 2015). The Al and Fe may naturally be distributed unevenly in different size fractions of soil aggregates (Tisdall and Oades, 1982; Six et al., 2004), and may be affected by a variety of land use management activities, including cultivation (Six et al., 2000b; Denef and Six, 2005). While the influence of land management practices on AS is well documented, much of the published research on the total Al, Fe and C distribution within different aggregate size classes has focused on the top 30 cm of soil profiles (Igwe et al., 1995; Zhang and Horn, 2001; Barthès et al., 2008).

The organic matter in surface layers is affected by tillage operations and processes of addition and decomposition due to the availability of oxygen (Pulleman and Marinissen, 2004; Torres-Sallan et al., 2017; Torres-Sallan et al., 2018). In deeper layers a lower supply of oxygen may limit decomposition (Jobbágy and Jackson, 2000; Six and Paustian, 2014; Olson and Al-Kaisi, 2015; Wiesmeier et al., 2019). The decline in total C with depth provides an opportunity to study the relationships between C, Al and Fe and their contribution to aggregation. These relationships and effects of management are not clearly understood for humic soils which are physically very stable due to their strong micro-aggregation (Fey, 2010). Many humic soils have been converted to various agricultural uses and the effects of such changes on AS have not been well studied.

The primary objectives of this study were, thus, to determine the effects of land use change from native forest to sugarcane farming on (i) AS and size distribution, and (ii) the distribution of total Al, Fe and C within different aggregate size fractions to a 1m depth in some humic soils. This study is important for the management of humic soils, as the results will indicate the long-term structural effects of putting these highly weathered, acid soils with high SOC under agricultural production.

7.2 Materials and methods

7.2.1 Site description

The study was conducted near Eshowe in northern KwaZulu-Natal, South Africa and the site characteristics have been outlined in Section 3.1.3. Some of the basic soil characteristics of the sugarcane and forest sites at Eshowe in northern KwaZulu-Natal were given in Table 5.1.

7.2.2 Soil sampling, preparation and analysis

The collection and analysis of soil samples were done following the protocols described in Sections 3.2 and 3.3, respectively. The methods used for aggregate size separation and the determination of AS were given in Section 3.3.2.

7.2.3 Statistical analysis

Data were analysed using Genstat 18 (Payne et al., 2011) by a two-way analysis of variance ($p \leq 0.05$) to test the effects of (i) land use, soil depth, and their interaction for bulk soils, and (ii) land use, aggregate size fraction, and their interaction for individual depths. Differences between the means of the significant factor were assessed with Duncan's multiple range test ($p \leq 0.05$). Least significant differences (LSD) at $p = 0.05$ were computed to separate treatment means for all properties. Linear regression analyses between total organic carbon (TOC) and total Al or Fe were also performed separately for forest soils and those under sugarcane. All results were based on three replications in the field.

7.3 Results

7.3.1 Properties of the forest and sugarcane soils

The TOC was significantly lower in the top 5 cm and higher at 20-40 cm depth under sugarcane than under forest ($p \leq 0.05$), with no significant differences at other depths between land uses (Figure 7.1a). The TOC concentration significantly decreased with depth under both land uses and the overall (0-100 cm) TOC was not significantly different between the land uses ($p > 0.05$).

Total Al was significantly higher under sugarcane than under forest ($p \leq 0.05$) in the top 30 cm, with no differences in deeper layers (Figure 7.1b). The concentration of total Al increased with depth under forest but not under sugarcane. Similarly, total Fe was significantly greater ($p \leq 0.05$) under sugarcane than under forest in the top 30 cm, except at 10-15 cm (Figure 7.1c). Below 30 cm, the total Fe was not significantly different between land uses except in the 80-100 cm layer where it was lower under sugarcane.

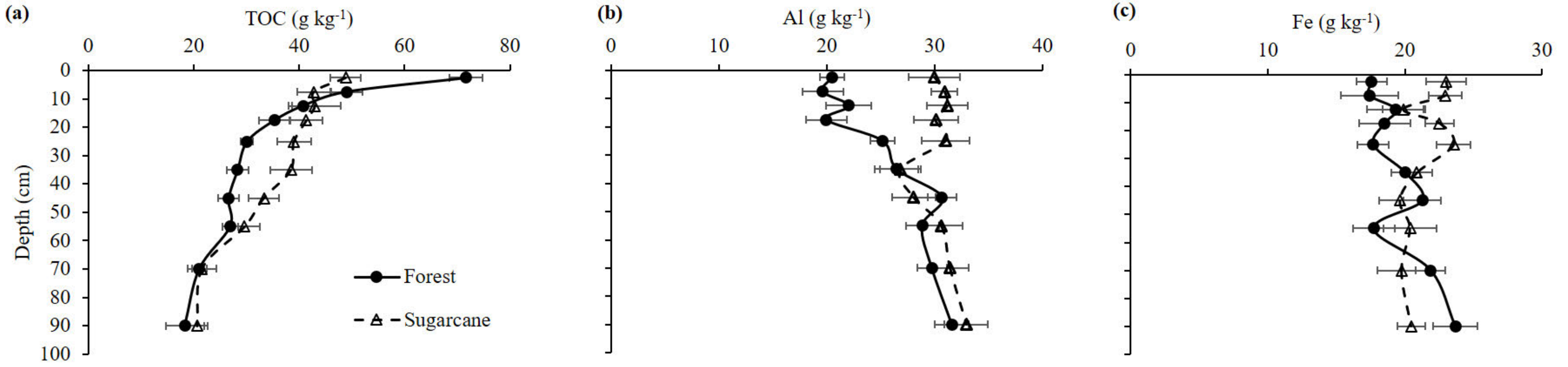

Figure 7. 1: The concentration (±standard error) of (a) total organic carbon (TOC), (b) total aluminium (Al) and (c) total iron (Fe) with depth (0-100 cm) in bulk soil under native forest and sugarcane (n = 3).

7.3.2 Soil aggregate stability and size distribution

The distribution of the aggregate size fractions differed depending on land use and depth ($p \leq 0.05$; Table 7.1). The proportion of large macro-aggregates (LM) was higher under forest in the top 15 cm, with no significant differences between land uses at other depths. The LM proportion did not change with depth under sugarcane, while it decreased under forest. On the other hand, the proportions of small macro-aggregates (SM) and micro-aggregates (M) under forest were not significantly different to those under sugarcane except in the top 10 cm where these fractions were lower under forest. The SM proportion did not change with depth under sugarcane, while it increased under forest. The SM generally made up a higher proportion of the total than the LM at all depths except in the top 15 cm of the forest soil (Table 7.1). The proportion of M was generally lower than that of the SM, while the silt + clay (SC) fraction had the lowest proportion irrespective of land use and soil depth. The proportion of SC was higher under sugarcane than under forest in the top 20 cm, with no differences in deeper layers and did not change with depth under sugarcane, whereas it increased under forest. There were no significant differences in the mean weight diameter (MWD) between forest and sugarcane soils at all depths, except at 0-10 cm, where the forest soil had higher values. The MWD did not change with depth under sugarcane, while it decreased under forest from an average of 1.32 in the topsoil to 1.06 mm in the 30-100 cm depth (Table 7.1).

Table 7. 1: Distribution of water stable aggregates and the mean weight diameter (MWD) of humic soil profiles under native forest and sugarcane (n = 3).

Land use	Depth (cm)	Aggregate size distribution ……………………. (Mass %)…………………				MWD (mm)
		LM* > 2000µm	SM 250-2000 µm	M 63-250 µm	SC < 63 µm	
Forest	0-5	65.7^{e}	19.8^{a}	11.8^{abc}	2.7^{a}	1.5^{d}
Sugarcane		28.8^{abc}	45.0^{cdefg}	20.2^{abcdef}	5.9^{cdef}	1.2^{abc}
Forest	5-10	58.4^{e}	29.4^{ab}	8.4^{a}	3.8^{ab}	1.5^{d}
Sugarcane		16.1^{abc}	48.0^{defgh}	25.6^{def}	10.3^{g}	1.0^{ab}
Forest	10-15	53.4^{de}	33.6^{bc}	8.7^{ab}	4.3^{abc}	1.4^{cd}
Sugarcane		30.8^{abc}	42.9^{cde}	16.9^{abcde}	9.4^{ef}	1.2^{abc}
Forest	15-20	31.8^{bc}	46.7^{defgh}	15.0^{abcd}	6.5^{abcd}	1.1^{ab}
Sugarcane		13.6^{abc}	54.3^{efgh}	24.4^{def}	7.7^{efg}	1.0^{a}
Forest	20-30	23.7^{abc}	48.5^{defgh}	22.2^{bcdef}	5.6^{bcde}	1.1^{ab}
Sugarcane		25.1^{abc}	48.1^{defgh}	19.7^{abcdef}	7.1^{cdef}	1.1^{ab}
Forest	30-40	11.6^{ab}	57.5^{h}	25.3^{def}	5.6^{cdef}	1.0^{a}
Sugarcane		23.1^{abc}	49.1^{efgh}	19.4^{abcdef}	8.4^{fg}	1.2^{abc}
Forest	40-50	7.1^{a}	55.9^{fgh}	29.1^{f}	7.9^{efg}	0.9^{a}
Sugarcane		27.9^{abc}	47.3^{defgh}	19.3^{abcdef}	5.5^{abcde}	1.3^{abc}
Forest	50-60	11.3^{ab}	56.7^{gh}	26.3^{ef}	5.7^{abcde}	1.0^{a}
Sugarcane		29.4^{bc}	42.9^{cdef}	21.2^{bcdef}	6.5^{abcde}	1.2^{abc}
Forest	60-80	22.5^{abc}	52.6^{efgh}	21.0^{bcdef}	3.9^{abcd}	1.1^{ab}
Sugarcane		29.3^{bc}	42.9^{cde}	22.9^{cdef}	4.9^{abcde}	1.2^{abc}
Forest	80-100	31.0^{bc}	42.9^{cdef}	19.4^{abcdef}	6.7^{def}	1.3^{abcd}
Sugarcane		29.4^{bc}	50.3^{efgh}	14.9^{abcde}	5.4^{abcde}	1.4^{bcd}

* LM: large macro-aggregates; SM: small macro-aggregates; M: micro-aggregates; SC: silt + clay
Means within the same column followed by different letters are significantly different ($p \leq 0.05$).

7.3.3 Total organic carbon in the aggregate size fractions

The distribution of TOC within the aggregate size fractions differed depending on land use and aggregate size fraction ($p \leq 0.05$; Figure 7.2a). The TOC concentration in the LM and SM fractions in the top 15 cm was lower under sugarcane than forest. The TOC contents of the SM and M at 40-50 cm and the SC fraction at 30-50 cm were higher under sugarcane compared to forest (Figure 7.2a). The greatest decline of TOC concentration with depth occurred in the top 15 cm below which no consistent TOC trend was observed in most aggregate size fractions under sugarcane.

Within the total profile depth (0-100 cm), the TOC in the LM (13 and 7 g C kg^{-1}) and SM (7 and 6 g C kg^{-1}) fractions was 85 and 17 % higher under forest than sugarcane, respectively. On the other hand, the TOC in the M (9 and 5 g C kg^{-1}) and SC (13 and 10 g C kg^{-1}) fractions was 80 and 13 % higher under sugarcane than forest, respectively. The TOC in aggregates under forest was generally higher at 0-15 cm and lower at 20-50 cm when compared to sugarcane, with the exception of the SC fraction.

7.3.4 Total aluminium and iron in the aggregate size fractions

At all depths the Al concentration was significantly higher ($p \leq 0.05$) in all aggregate size fractions under sugarcane compared to forest (Figure 7.2b). The average (0-100 cm) Al content was 167, 57, 180, and 129 % greater under sugarcane than forest in the LM (16 and 6 g Al kg^{-1}), SM (11 and 7 g Al kg^{-1}), M (14 and 5 g Al kg^{-1}) and SC (16 and 7 g Al kg^{-1}), respectively.

In the top 10 cm, total Fe concentration in the LM, M and SC fractions was significantly higher ($p \leq 0.05$) under sugarcane compared to forest with no differences in the SM fraction (Figure 7.2c). At 20-30, 50-60 and 60-80 cm, the Fe concentration was higher in the LM and M under sugarcane than forest with no differences in the SM and SC fractions. At 80-100 cm, total Fe was higher in the M fraction under sugarcane (12 g Fe kg^{-1}) than forest (6 g Fe kg^{-1}) with no significant differences in the other size fractions. The average (0-100 cm) Fe content was 50 % higher under sugarcane than forest in the LM and M (9 and 6 g Fe kg^{-1}), and 11 % higher in the SC (10 and 9 g Fe kg^{-1}) fractions.

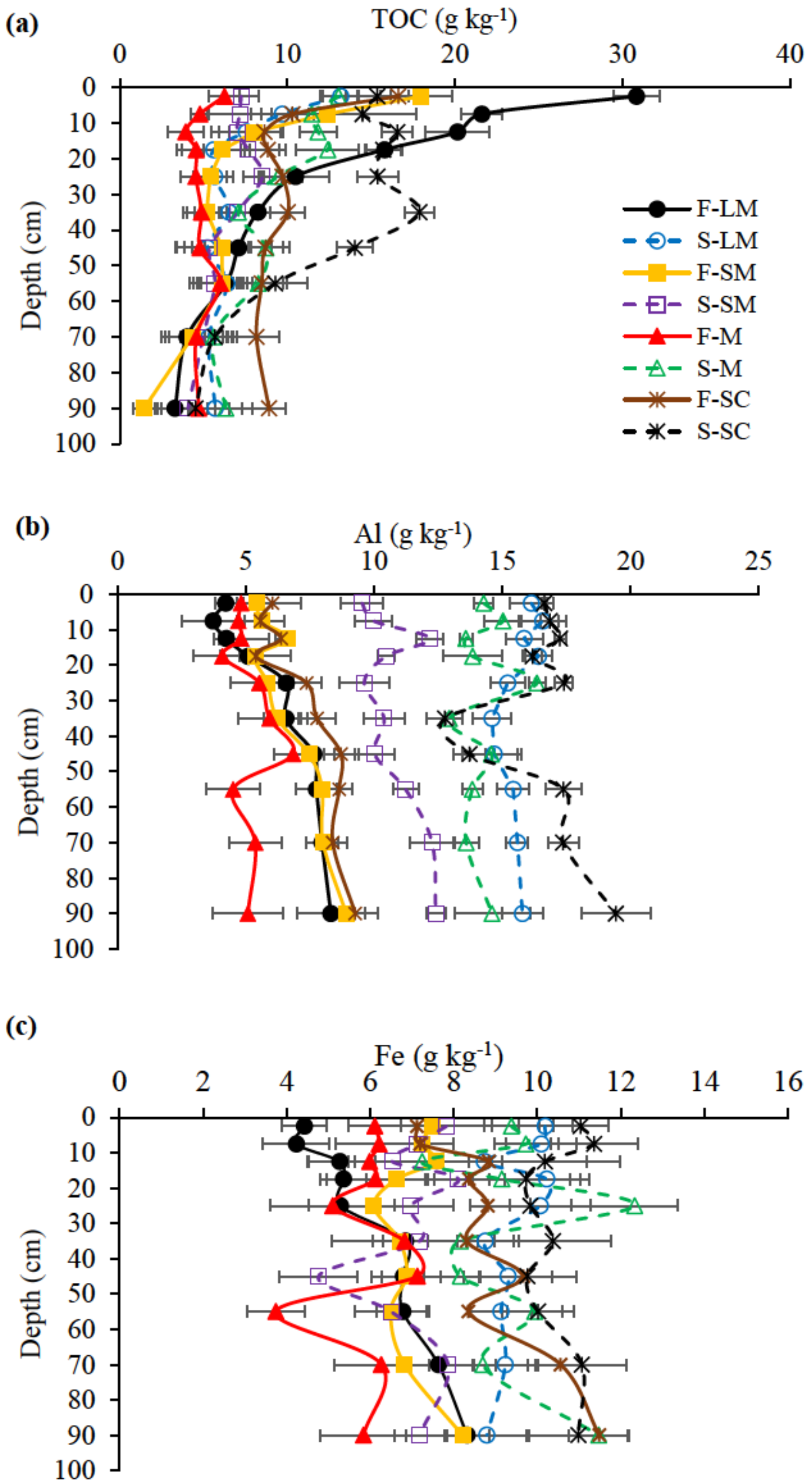

Figure 7. 2: The concentration (±standard error) of (a) total carbon (TOC), (b) total aluminium (Al) and (c) total iron (Fe) with depth in large macro-aggregates (LM), small macro-aggregates (SM), micro-aggregates (M) and the silt + clay fraction (SC) of humic soils under native forest (F) and sugarcane (S) (n = 3).

7.3.5 Relationship between carbon, aluminium and iron in bulk soils and aggregate size fractions

There were few significant relationships between the measured variables and only the most significant are discussed. The relationship of TOC to Fe under forest (R^2=0.42; Figure 7.3a) and to Al under sugarcane (R^2=0.47; Figure 7.3b) in bulk subsoils were negative. Under forest, TOC in bulk soils was positively correlated with the proportion of LM in the topsoil (R^2=0.68; Figure 7.4a) but negatively with the SM fraction (R^2=0.58; Figure 7.4b). In the subsoil, the TOC was positively correlated with the M fraction (R^2=0.57; Figure 7.4c). No significant relationships were observed between the proportion of aggregate size fractions and TOC in the soils under sugarcane. There were also no significant relationships between total Al or Fe and the proportions of any of the aggregate size fractions under both land uses. Under forest, a negative relationship was observed between TOC and Al in the LM fraction of the topsoil (R^2=0.50; Figure 7.5a). A stronger relationship was observed between Al and Fe in the M fraction of the subsoil (R^2=0.62; Figure 7.5b). The relationships between these variables were also positive but weaker for both the topsoil (R^2=0.37) and subsoil (R^2=0.46) in the SC fraction under forest (Figure 7.5c). Under sugarcane, the relationship between Al and Fe was strong in the M fraction of the topsoil (R^2=0.75) but weaker for the subsoil (R^2=0.44) (Fig. 7.6a). The relationship between TOC and Al in the SC fraction of the subsoil was strong and negative (R^2=0.61; Figure 7.6b).

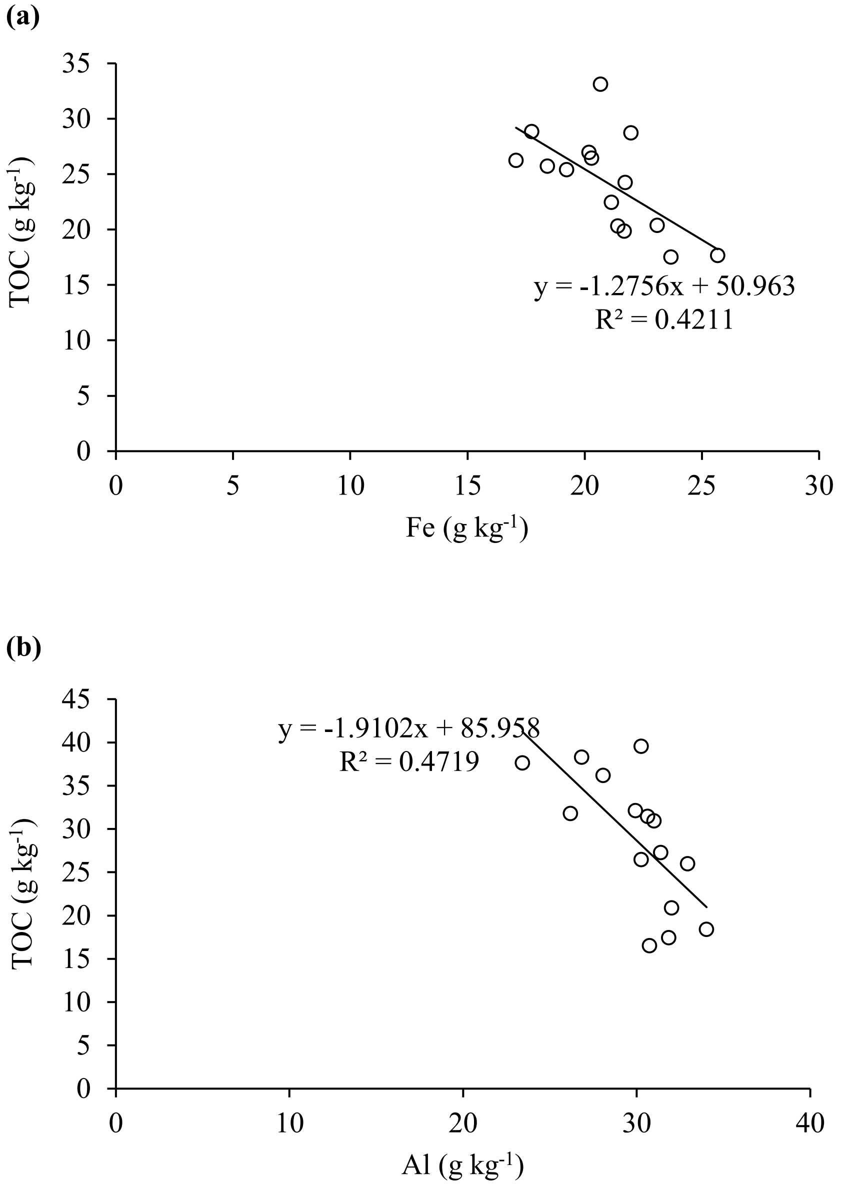

Figure 7. 3: Relationship between total organic carbon (TOC) and (a) total iron (Fe) under native forest and (b) total aluminium (Al) under sugarcane in the humic subsoils (30-100 cm).

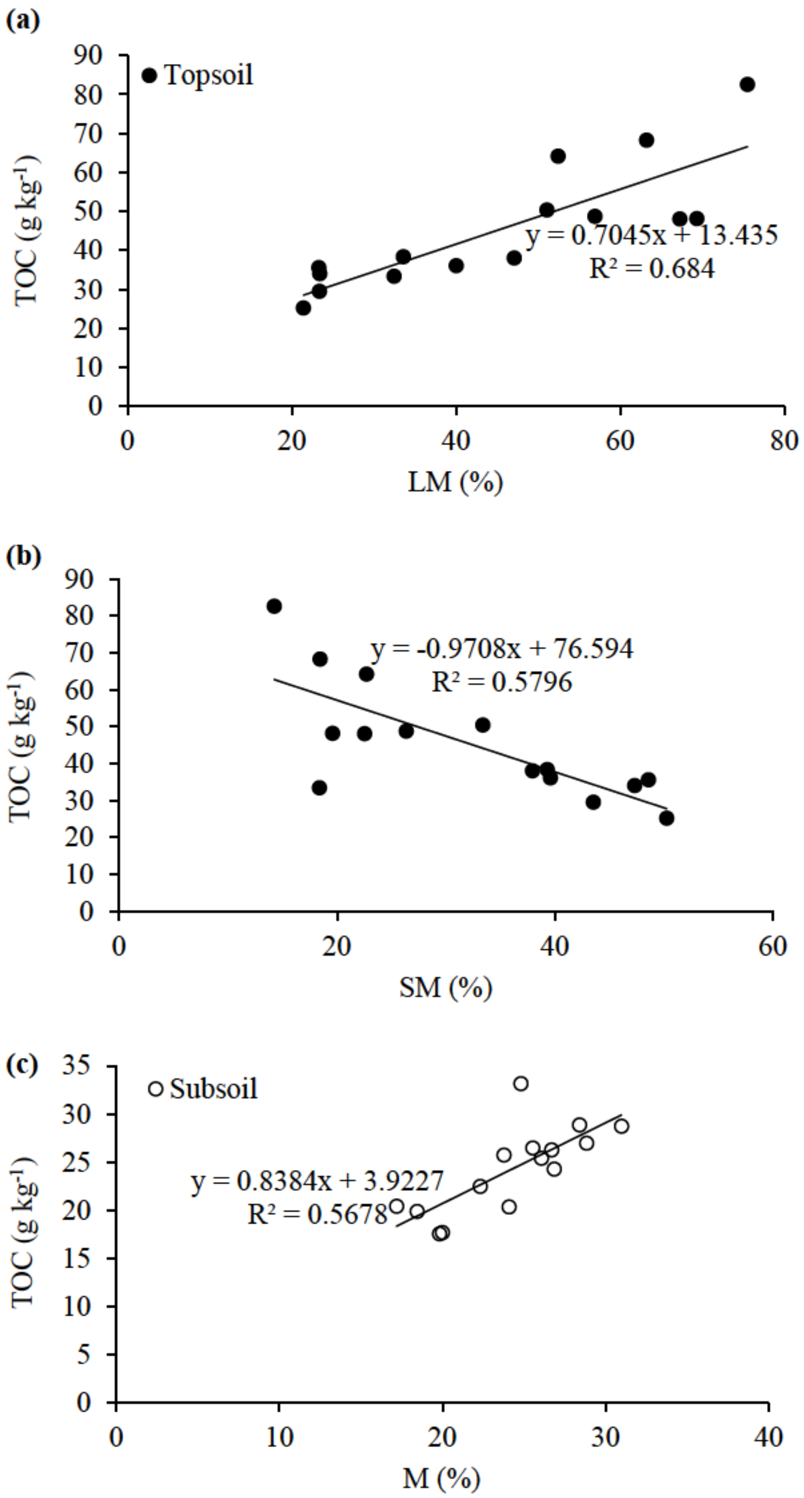

Figure 7. 4: Relationship between total organic carbon (TOC) and (a) large macro-aggregates (LM), (b) small macro-aggregates (SM) and (c) micro-aggregates (M) in the humic topsoils (0-30 cm) and subsoils (30-100 cm) under native forest.

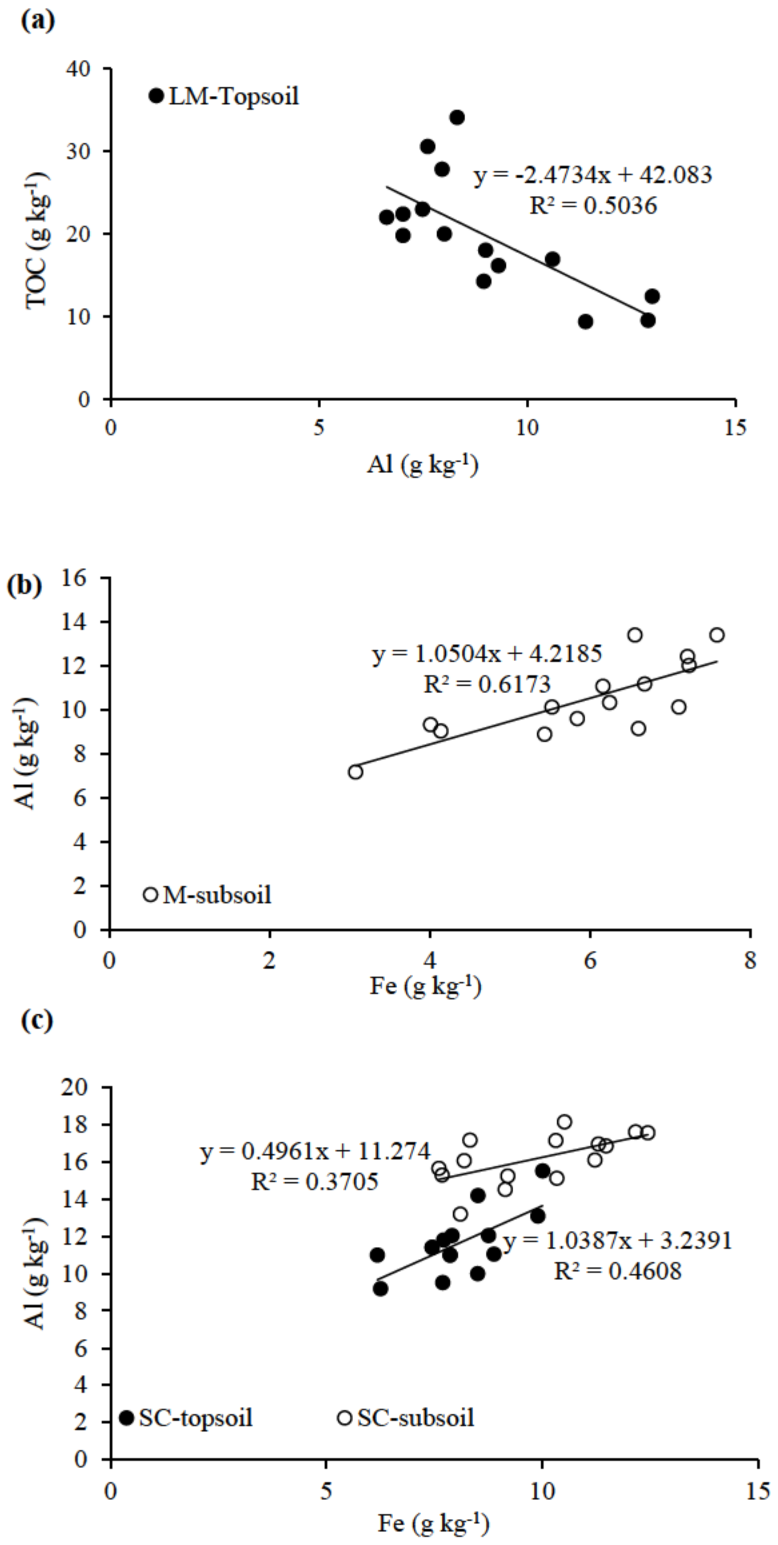

Figure 7. 5: Relationship between (a) total organic carbon (TOC) and total aluminium (Al) in large macro-aggregates (LM), (b) total Al and total iron (Fe) in micro-aggregates (M) and (c) total Al and total Fe in the silt + clay (SC) fraction in the humic topsoils (0-30 cm) and subsoils (30-100 cm) under native forest.

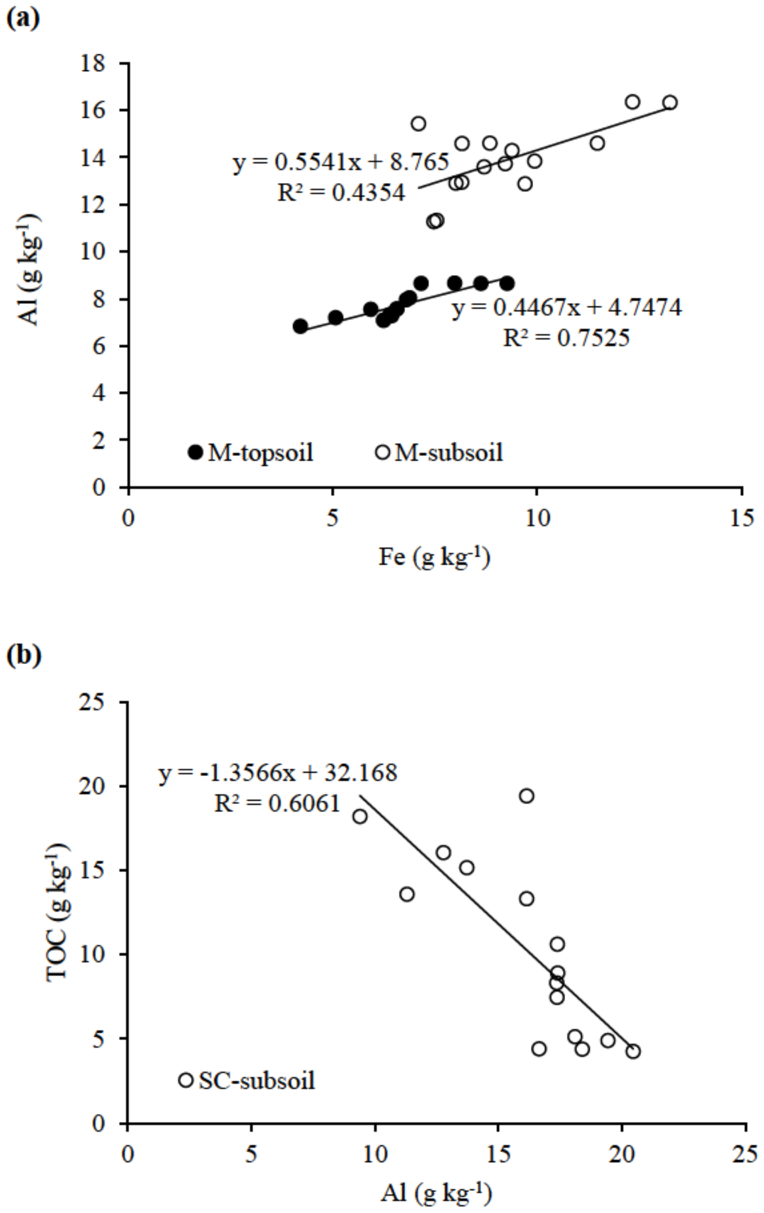

Figure 7. 6: Relationship between (a) total aluminium (Al) and total iron (Fe) in micro-aggregates (M) of the humic topsoils (0-30 cm) and subsoils (30-100 cm), and (b) total organic carbon (TOC) and total Al in the silt +clay (SC) fraction of the subsoils under sugarcane.

7.4 Discussion

7.4.1 Carbon, aluminium and iron in bulk soils

The TOC in the bulk soils supports results from the large number of studies that have found strong evidence for a decline of 30 to 80 % in C content when forests are converted to arable agriculture (van Antwerpen and Meyer, 1996; Hartemink, 1998; Six et al., 2000b; Dominy and Haynes, 2002; Blanco-Canqui and Lal, 2004; Zhao et al., 2005; Chivenge et al., 2007; Zhao et al., 2017). Based on a meta-analysis using data from 74 publications from around the world, Guo and Gifford (2002) reported a 42 % average loss of the antecedent TOC pool from native ecosystems on conversion to croplands although the authors did not indicate whether sugarcane was among the crops investigated. In all these studies, the loss of C has often been attributed to (i) erosion, (ii) lower C inputs, (iii) a reduced stabilization of SOM due to reduced aggregation, and (iv) subsequent mineralization promoted by increased soil temperature and aeration. As expected, the TOC decreased from the topsoil to subsoil layers due to the continuous aboveground C input by vegetation residues (Jobbágy and Jackson, 2000; Olson and Al-Kaisi, 2015) and the lack of soil disturbance more especially under native ecosystems (Hartemink, 1998; Six et al., 2004; Chaplot et al., 2010). The higher TOC under sugarcane than forest in the 20-40 cm depth may possibly be a result of the translocation of C to lower depths (Balabane and Balesdent, 1996; Hartemink, 1998; Jobbágy and Jackson, 2000; Dominy and Haynes, 2002; Olson and Al-Kaisi, 2015) and contribution of root biomass (Bronick and Lal, 2005).

The higher Al and Fe contents in the bulk sugarcane topsoils than those under forest could possibly be a result of the dilution by high organic matter in the forest soil, as SOC was not removed before the analysis of Al and Fe in these soils. Using such samples would have masked the concentration of Al and Fe in the soil that would otherwise be observed if the

SOC was removed prior to the analysis of these elements (Howard and Howard, 1990; Gubevu, 1997; Klein and Dutrow, 2007). Compared to forest (6 g Fe kg^{-1}) soil, Yost and Hartemink (2019) found four times higher total Fe content in a bulk sandy clay loam soil (0-100 cm) under agriculture (23 g Fe kg^{-1}) in the Central Sand Plains of Wisconsin. Other researchers (Du Toit, 1993; van Antwerpen and Meyer, 1996) have reported an increase in the total Fe and Al content of tropical soils following application of N and P inorganic fertilizers at planting. In both these studies, Al was found to be chemically stabilized by interaction with organic matter such that the reduction in organic matter following cultivation increased the concentration of this element in the soil. The notable increase in Fe and Al content in the subsoils again corresponds with lower TOC, thus lower dilution (Gubevu, 1997; Klein and Dutrow, 2007).

7.4.2 Soil aggregate stability and size distribution and organic carbon in aggregate fractions

The higher MWD in the forest soil surface layers may possibly be a result of the minimal soil disturbance and higher TOC, thus favouring the formation of more stable aggregates (Le Bissonnais et al., 2002; Blanco-Canqui and Lal, 2004). Blair (2000) also reported a significant reduction in the MWD of sugarcane soils compared to undisturbed grasslands in Australia. The lower MWD values measured in the subsurface layers of the forest soils may not only be a result of lower TOC but could also be due to the lower clay (11 %) content than in the sugarcane subsoil (18 %) (Oades and Waters 1991; Bronick and Lal, 2005). Aggregate stability has generally been found to increase with increasing clay content (Dominy and Haynes, 2002; Chaplot et al., 2010; Wang et al., 2016) especially in soils with non-expanding, crystalline clays, such as kaolinite, that are less dispersive (Wakindiki and Ben-Hur, 2002). Denef and Six (2005) suggested that clay minerals may interact with organic matter through the formation of organo-mineral assemblages which, in turn, affect aggregation. The lack of significant differences in the MWD between the forest and sugarcane soils was presumably because inputs of organic matter and decomposition rates are quite similar under both land use systems since soils under sugarcane are only disturbed at replanting which occurs about every eight years at the locality.

The higher proportion of the LM in the surface layers of the forest soils than under sugarcane could be attributed to higher SOC, which could include live and decaying plant roots, fungal hyphae, and casts of earthworms and termites, which are rapidly destroyed by cultivation (Oades and Waters, 1991; Bronick and Lal, 2005). This finding is similar to that of Roth et al. (1991), who reported a higher proportion of LM in the surface layer (0-10 cm) of virgin forest soils compared to sugarcane for a similar soil type at Londrina, Brazil. The substantial loss of TOC in the LM and SM fractions of the sugarcane soils was expected as the break-up of macro-aggregates and increased aeration caused by soil disturbance both favour decomposition of SOM (Beare et al., 1994; Paustian et al., 1997), thereby reducing the TOC concentration (Balabane and Balesdent, 1996; Six et al., 2000a; Schmidt et al., 2011). The addition of inorganic fertilizers and lime to the soils under sugarcane could have increased microbial activity to the extent that SOM was decomposed more rapidly, lowering TOC. Castro Filho et al. (2002) found the TOC content to be three times greater in macro-aggregates under forest (39 g C kg^{-1}) compared to sugarcane (13 g C kg^{-1}) in a Rhodic Ferralsol from southern Brazil.

With the exception of the Fe in the SM fraction, the higher total Al and Fe in all aggregate size fractions under sugarcane than in the forest soils, suggests that both are involved at various levels in the aggregation hierarchy of these soils, and are affected by changes in land-use and soil management (Amezketa, 1999; Barthès et al., 2008; Totsche et al., 2018). The decrease in the proportion of LM with depth under forest may be due to the lower SOM content at greater depth (Roth et al., 1991; Paustian et al., 1997; Jobbágy and Jackson, 2000; Le Bissonnais et al., 2002). On the other hand, the higher proportion of SM in the surface layers of sugarcane soils than under forest suggests that tillage and cropping result in the mineralization of the organic C from larger aggregates causing the breakdown of LM to SM (Blanco-Canqui and Lal, 2004).

The effects of different land use and management were less pronounced for the M and SC fraction, possibly due to similarities in TOC, Al and Fe contents (Tisdall and Oades, 1982; Blanco-Canqui and Lal, 2004; Yost and Hartemink, 2019). These results are, however, in contrast with the findings of Zhang et al. (2012), who reported an increase of the M and SC fractions 20 years after native grassland was converted to maize farming in China. These contradictory findings could be associated with the crop species. Sugarcane fields are not ploughed as often as maize fields, which generally results in aggregates that are more resistant to change than the more frequently ploughed maize soils (Balabane and Balesdent, 1996; Hartemink, 1998; Castro Filho et al., 2002).

7.4.3. Relationship between carbon, aluminium and iron in bulk soils and within aggregate size fractions

The negative relationship observed between TOC and Fe ($R^2 = 0.42$) or Al ($R^2 = 0.47$) in bulk subsoils (Figures 7.3a-b) indicates the poor association of SOC with total Al and Fe in these soils (Igwe et al., 1999). The positive relationship between TOC and LM ($R^2 = 0.68$) in the forest topsoil (Figure 7.4a) was expected as the SOM maintains the stability of larger soil aggregates (> 250 µm) (Roth et al., 1991; Paustian et al., 1997). The negative ($R^2 = 0.58$) relationship between TOC and SM in the topsoil (Figure 7.4b) may be indicating the importance of a different organic matter fraction that could be making up only a small proportion of the total but still involved in aggregate formation and stabilization (Jobbágy and Jackson, 2000; Totsche et al., 2018) while the positive ($R^2 = 0.57$) relationship between TOC and M in the subsoil (Figure 7.4c) may possibly be a result of C translocation to lower depth (Lugato et al., 2010; Olson and Al-Kaisi, 2015).

Similar to the trend in bulk soil samples, the negative relationship between TOC and Al in the LM ($R^2 = 0.50$) in the forest topsoil (Figure 7.5a) and in the SC fraction ($R^2 = 0.61$) in the sugarcane subsoil (Figure 7.6b) again suggests that the protection of TOC in aggregates is not explained by total Al and Fe (Igwe et al., 1999; Zhao et al., 2005). Similar results were reported by Oades and Waters (1991), Dalal and Bridge (1996) and Zhang and Horn (2001) in Alfisols, Entisols and Ultisols, respectively. However, Malepfane et al. (2022) reported positive correlations between TOC and Mehlich 3 extractable Fe and Al, and concluded that this fraction of Al and Fe contributes to the stabilisation of OC in humic soils that included the ones used in this study. Given that the current study also included less reactive forms than the fractions that are soluble and those in amorphous oxides that Malepfane et al. (2022) focused on, the contradiction in the results could be explained by changes in the forms of Al and Fe investigated. Other studies have indicated that low-crystalline Fe oxides or oxyhydroxides stabilize SOM more efficiently than crystalline ones because they have a greater specific surface area and density of hydroxyl sites than crystalline ones which improves their ability to chelate (Wen et al., 2019; De Mastro et al., 2020). The stabilisation of TOC by Fe and Al in humic soils may, therefore, depend on the form in which these elements occur.

The positive relationship between Al and Fe in the M fraction in the subsoil (Figure 7.5b) and the SC fraction under forest (Figure 7.5c) as well as in the M fraction under sugarcane (Figure 7.6a) indicates the enrichment of M and SC fractions with these elements (Totsche et al., 2018). These findings are consistent with a number of studies that have found positive correlations between these elements in a wide range of soils (Igwe et al., 1995; Zhang and

Horn, 2001; Castro Filho et al., 2002; Jim, 2003; Kaiser and Guggenberger, 2007; Peng et al., 2015). According to Dalal and Bridge (1996) and Shepherd et al. (2001), the M and SC fractions are formed with either Al and Fe or phyllosilicate clays serving as their nucleus. Oades and Waters (1991) and Alekseeva (2007) call attention to the fact that M and SC fractions seem to be stabilized mostly by short-range van-der-Waals forces and electrostatic binding largely involving Al and Fe.

7.5 Conclusions

The study aimed to assess the effects of converting native forest to sugarcane cultivation on (i) AS and size distribution, and (ii) the distribution of total Al, Fe, and C within different aggregate size fractions in some humic soils. Sugarcane cultivation resulted in (i) a lower proportion of LM in the top 15 cm and a significant increase in the SC fraction in the top 20 cm, (ii) reduced average TOC, (iii) increased total Al and Fe, (iv) reduction of the AS in the top 10 cm, and (v) redistribution of Al, Fe and TOC within the different aggregate size fractions. Total Al and Fe were higher in the LM, M and SC fractions under sugarcane than under forest. Sugarcane soils contained less TOC in the macro-aggregates at all depths compared to forest soils. However, TOC increased in the M and SC size fractions as these are less affected by cultivation. The negative relationship between Al and TOC in LM of the topsoil under forest and in the SC fraction of the subsoil under sugarcane indicated that total Al and Fe do not necessarily explain the protection of TOC in aggregates in humic soils.

Collectively, these results further indicated that humic soils are potentially productive due to their high TOC content, but intensive production may reduce TOC, weakening the AS in the long term. The adoption of practices inclined to improve or maintain their TOC status is, therefore, highly recommended especially when these soils are under intensive agricultural production.

CHAPTER 8 CHAPTER 8: GENERAL DISCUSSION, CONCLUSIONS AND RECOMMENDATIONS

8.1 Introduction

Humic soils occur particularly on old land surfaces in KwaZulu-Natal and along the Pondoland coast and the eastern escarpment of Mpumalanga (Macvicar et al., 1984). Within these well-watered regions, there is a diversity of land uses including native forests, grasslands and sugarcane and they all provide critical services to human survival (van Antwerpen and Meyer, 1996). This chapter syntheses the scientific insights obtained during the study to better understand some selected soil properties related to humic soils, thereby providing evidence-based sustainable future management of these soils in the South African context. The four main objectives of the study described in Chapter 1 are discussed together in this chapter and then conclusions, recommendations and some research gaps are given.

8.2 Discussion

All of the soils used in this study were acidic, with kaolinite and quartz as dominant minerals in the clay fraction at all the sites as a result of intensive weathering (Marshall and von Brunn, 1999; Wakindiki and Ben-Hur, 2002) and leaching of exchangeable bases, which are replaced by Al (Fey et al., 1990; Inagaki et al., 2019). The dominance of clay in the dolerite-derived soils in the Midlands region (Cedara and Karkloof) and sand from the sandstone-derived soils in the North Coast region (Eshowe) reflects the fact that texture is an inherent soil factor influenced by parent material rather than land use and management practice (Bronick and Lal, 2005). When compared to the other sites, the combined effect of higher rainfall, composition of the parent material and vegetation resulted in higher organic matter (OM) accumulation in soils at Karkloof. The higher total aluminium (Al) and iron (Fe) contents in soils at Karkloof and Cedara is related to the higher clay content at these sites than at Eshowe. Aluminium and Fe oxides, according to Duiker et al. (2003), act as cementing agents between the surfaces of clays and as charged discrete particles in many highly weathered, acidic soils which these humic soils studied are.

The higher average (0-100 cm) mean weight diameter (MWD) of soils at Karkloof and Cedara than those at Eshowe were consistent with the findings of Duiker et al. (2003). In agreement with these findings, Leifeld et al. (2005) concluded that the attraction between Al and negatively charged clay exchange sites under grasses promotes flocculation to form stable aggregates. Another crucial factor is that grasses grow quicker than trees, resulting in higher OM turnover which enhances diverse microbial activity and consequently aggregate stability in grassland soils (Kaiser et al., 2002). Forest soils can host a variety of microbial communities, however because of the shade, litter layer, and various microclimate conditions, the results suggest that there may be less overall biological activity and diversity (Del Galdo et al., 2003). Grazing by large herbivores, which can stimulate the growth of belowground biomass and turnover and encourage soil aggregation, is another natural disturbance that can affect grasslands (Snyman and du Preez, 2005).

The results of correlations among various measured parameters revealed that the organic materials are the primary binding agents in the top 30 cm depth of these soils while Al, Fe and clay are important for aggregation below 30 cm depth. The aggregate size distribution further revealed that large macro-aggregate (LM) and small macro-aggregate (SM) size fractions were dominant at all sites at the expense of the silt + clay (SC) size fraction, which result could be attributed to permanent vegetation cover and increased litter retention due to the lack of soil disturbance at the studied sites (Haynes and Naidu, 1998; Chenu et al., 2000). According to Blanco-Canqui and Lal (2004), the release of cementing substances such as polysaccharides and organic acids during organic source decomposition plays an important role in the stability of macro-aggregates. The higher OC content of the micro-aggregates at Cedara is, however, highly stable, with a relatively longer turnover time than the OC stored in the macro-aggregates at Karkloof and Eshowe (Bronick and Lal, 2005). In addition, total Al and Fe were higher in macro-aggregates at Karkloof and Cedara while the SC fraction showed a higher concentration of these elements at Eshowe probably due to the effect of different climatic conditions on mineral associations within soils of similar mineralogy (Igwe et al., 2005). The lower bulk density (BD), moisture retention capacity, and hydraulic qualities of the soils from Karkloof and Cedara in comparison to those from Eshowe may be attributed to their greater TOC, clay content, and aggregate stability (Shepherd et al., 2001; Lado et al., 2004).

The higher MWD at Karkloof resulted in lower BD, but higher hydraulic conductivity (Ks), total porosity (TP), field capacity (FC), and permanent wilting point (PWP) in the 0-100 cm depth than at Cedara and Eshowe. According to Rawls et al. (2003) and Seema et al. (2019), there is a strong relationship between SOC content, MWD, BD, Ks and FC and PWP, with high SOC, MWD, Ks and FC observed more frequently in soils with low BD. The TOC was found to be positively correlated to Ks, FC and PWP in the top 30 cm while BD correlated negatively with these variables at the same depth and with Ks in the 0-100 cm depth. The Ks further correlated positively with Al, Fe and sand but negatively with silt and clay at all depths, indicating the variation of hydraulic properties with the pore size diameter of the soils (Celik, 2005; Seema et al., 2019). The FC and PWP also correlated positively with Al, Fe and silt, indicating the increase in the volume of micro-pores capable of retaining water at these suctions (Duiker et al., 2003). The AWC was not different between the studied sites confirming the findings of Haynes and Naidu (1998), who reported that if an increase in TOC causes an increase in moisture content at both FC and PWP, the net result on AWC might not be greatly affected. While AWC did not significantly differ between native grassland and forest soils from the different sites, conversion of forest to cropping/sugarcane production could have more significant effects on these properties.

Sugarcane cultivation (i) reduced TP and (ii) increased BD, FC and PWP in the top 30 cm compared to under forest. This was likely due to the in-field traffic and harvesting of sugarcane under wet conditions (Meyer and van Antwerpen, 2010) which, in turn, reduced the pore size diameter thereby increasing water content at FC and PWP (Haynes and Naidu, 1998; Bescansa et al., 2006). Although the higher average Ks observed in the 0-100 cm depth under sugarcane than forest is not in agreement with results from a number of other studies (e.g. Gol, 2009; Wahren et al., 2009), the differences may be a result of the levels of SOC involved (Blanco-Canqui et al., 2017; Fu et al., 2021), with the humic soils used in this study having > 4 % C in the top 15 cm and >1.8 % even at 100 cm depth. Such high levels of SOC might have reduced BD thereby improving TP which, in turn, increased Ks under sugarcane (Lie et al., 2021). The lower average water content at FC and PWP in soils under forest than sugarcane could be attributed to the heterogeneity of silt. Other studies (e.g. Jamison and Kroth, 1958; Tomasella and Hodnett, 1998) have indicated the possibility of the coarse silt inducing lower porosity and water retention at FC and PWP than fine silt. The higher AWC at 30-40 cm under sugarcane than forest reflects higher SOC found at this depth. Even though the TOC could not explain the water retention characteristics (FC, PWP and AWC) of the whole soil profile, the negative correlations between silt content and Ks, FC and PWP (and corresponding positive correlations with sand content) suggested that texture plays an important role in affecting the water retention characteristics of the humic soils studied. With regards to PWP, Al and Fe seem to be equally important.

Bulk soils under sugarcane were lower in macro-aggregates and TOC but higher in the SC fraction and total Al and Fe while TOC in the macro-aggregates was lower than forest soils at all depths. These results support the hypothesis that the transient binding agents of macro-aggregates are disrupted by tillage-induced disturbance thereby favouring SOM decomposition which, in turn, reduces the SOC concentration (Chenu et al., 2000; Chivenge et al., 2007). The TOC, however, increased in the micro-aggregate (M) and SC size fractions as this is less affected by cultivation (Bronick and Lal, 2005). The negative relationship between Al and TOC in LM of the top 30 cm depth under forest and in the SC fraction of the 30-100 cm depth under sugarcane indicated that total Al and Fe do not necessarily explain the protection of TOC in aggregates in humic soils. The MWD was statistically not different between the forest and sugarcane soils due to the lack of frequent soil disturbance since sugarcane is only disturbed at replanting about every eight years at the locality (Meyer and van Antwerpen, 2010). The notable increase in total Al and Fe contents in bulk soils and within aggregate size fractions under sugarcane may be associated with the application of N and P inorganic fertilizers at planting (van Antwerpen and Meyer, 1996). Such application can increase the concentration of Al and Fe in the soil through various mechanisms, including changes in soil pH, P fixation, and P cycling (Chivenge et al., 2007; De Mastro et al., 2020). However, other studies (Gubevu, 1997; Klein and Dutrow, 2007) have attributed this to the dilution by high organic matter in the forest soil, as SOC was not removed before the analysis of Al and Fe. According to Howard and Howard (1990), this would have masked the concentration of Al and Fe in the soil that would otherwise be observed if the SOC was removed prior to the analysis of these elements.

8.3 Conclusions

The chemical and physical properties of the humic soils studied generally conform to the properties of other highly weathered soils reported in the literature. The high TOC makes humic soils unusual. The properties that are related to other highly weathered soils include the low pH, low exchangeable base cations, high Al and Fe content and predominance of kaolinite as well as quartz in the clay–size fraction, and these properties are affected by site factors and land use.

The results from the first objective in the present study showed (i) lower pH and higher TOC at Karkloof than other sites, (ii) higher Al, Fe, clay and silt contents at Cedara and Karkloof than at Eshowe, and (iii) higher sand fraction at Eshowe than Cedara and Karkloof. The differences observed between the sites can be explained by a combination of soil mineralogy, topographical features, climatic factors, and the characteristics of the parent material. The soils at Cedara and Karkloof share a similar parent material. The increased precipitation at Karkloof is the determining factor leading to more acidic soils compared to those at Cedara. The second objective showed that (i) 30 years of sugarcane cultivation resulted in an increase of BD and Ks on the sandy clay loam humic soil, (ii) the TOC could not explain the water retention characteristics (FC, PWP and AWC) of the whole soil profile, indicating that C is not the main driver but rather texture, especially silt, may be influencing the water retention in these soils, (iii) higher TOC and MWD at Karkloof resulted in lower BD and higher TP thereby increasing Ks, FC and PWP more at this site than Cedara and Eshowe, and (iv) the TOC could only explain Ks and water retention characteristics (FC and PWP) in the top 30 cm depth, indicating that C is not the main driver at 30-100 cm depth but rather texture may be influencing the Ks and water retention in these humic soils under native vegetation.

The results under the third objective showed that (i) MWD is higher at Karkloof and Cedara than at Eshowe, (ii) while exchangeable acidity (and hence Al) and silt are responsible for aggregation at all depths, TOC is responsible for aggregation, particularly the formation of large and small aggregates, in the upper 30 cm, with clay and total Fe and Al predominating below the 30 cm depth in all aggregates. The aggregate size distribution results further revealed that (i) both LM and SM dominate all sites at the expense of the SC size fraction, (ii) the M size fraction has a higher TOC content at Cedara while LM and SM have a higher concentration at Karkloof and Eshowe, and (iii) the concentration of total Al and Fe is higher in both LM and SM than M at Karkloof and Cedara while the SC fraction has a higher concentration of these elements at Eshowe. The findings of the fourth objective revealed that sugarcane cultivation reduced aggregate stability and TOC in the LM and SM, increased Al and Fe content in all aggregates, and that total Al and Fe did not necessarily explain the protection of TOC in the aggregates in the soils studied.

Although these humic soils have the potential to be extremely productive due to their high C sequestration capacity, limitations relating to an increase in total Al and Fe contents when these soils are cultivated are typical, necessitating liming to ameliorate soil acidity increase and increase Ca and Mg contents. The results of the current study also showed that TOC is not the main driver of AS and that it varies in its importance with depth in the profile. For instance, the high TOC content in these soils promotes the formation of stable aggregates thereby influencing BD, TP, Ks, FC and PWP, more especially in the topsoil.

8.4 Recommendations

The higher sand content accompanied by lower AS in the Eshowe soils make them more vulnerable to erosion (wind and water), crusting and subsequent impeded germination compared to the other sites. It is critical to protect these soils from raindrop impact and micro-cracking through the implementation of sustainable ecological practices and management. These include reducing tillage, managing water, balancing nutrients, and improving organic matter content. These practices will help maintain the integrity, resilience, and biodiversity of the forest ecosystem, resulting in higher vegetation density and improved soil structure. As a consequence, this will lead to AS improvement, reducing the risk of soil erosion while increasing water intake rates and rainfall use efficiency. In addition, the capacity of humic soils to promote C sequestration within both aggregates and bulk soil when placed under cultivation can be improved by reducing aggregate breakdown, for example through employing minimum tillage practices on these specific soils. The adoption of practices inclined to increase or maintain their TOC status such as adding organic matter, reducing tillage, using cover crops, rotating crops, maintaining soil moisture, and applying lime is, therefore, highly recommended since the stability of aggregates against disaggregation is of paramount importance in the prevention of soil degradation. Such practices will not only arrest and ameliorate the negative effects of sugarcane mono-cropping on soil degradation, but will also improve the soil microbial community and diversity thereby improving the overall fertility spectrum of these soils through diverse microbial functional processes.

Under native vegetation, the SC showed no correlation with either total Al or total Fe at all sites which may be reflecting the inclusion of the less reactive forms in the current study than the fractions that are soluble. Moreover, the negative relationship between Al and TOC in LM of the topsoil under forest and in the SC fraction of the subsoil under sugarcane indicated that total Al and Fe do not necessarily explain the protection of TOC in these aggregate sizes in humic soils. Therefore, it is recommended that further experiments are carried out using extractions designed specifically to estimate the types of Al and Fe oxides because the stabilisation of TOC by Fe and Al in humic soils may depend on the specific forms in which these elements occur. Such studies would help provide a better understanding of the specific Al and Fe fractions that are important in order to develop C sequestration strategies that may help to mitigate any OC losses following cultivation.

In addition, despite the high TOC content of the humic soils studied, this parameter could not explain the water retention characteristics (FC, PWP and AWC) of the whole soil profile under forest and sugarcane, suggesting that some particular fraction of the TOC may be influencing the water retention in these soils. Further work is thus required to separate the TOC into discrete pools to account for the value of each pool to water retention. Only through such separation will it be possible to (a) determine which carbon pool/s contribute most to each of the hydraulic properties and (b) how much of each carbon pool is required to ensure sustainable management of humic soils.

REFERENCES

Abu, S.T., 2013. Evaluating long-term impact of land use on selected soil physical quality indicators. Soil Res. 51, 471–476.

Agnese, C, Bagarello, V., Baiamonte, G., Iovino, M., 2011. Comparing physical quality of forest and pasture soils in a Sicilian watershed. Soil Sci. Soc. Am. J. 75, 1958–1970.

Alekseeva, T.V., 2007. Soil Microstructure and Factors of its Formation. Eur. J. Soil Sci. 40 (6), 649–659.

Amelung, W., Zech, W., Zhang, X., Follett, R. F., Tiessen, H., Knox, E., Flach, K.W., 1998. Carbon, Nitrogen, and Sulfur Pools in Particle-Size Fractions as Influenced by Climate. Soil. Sci. Soc. Am. J. 62, 172–181.

Amézketa, E., 1999. Soil aggregate stability: a review. J. Sustain. Agric. 14, 83–151.

Arshad, M.A., Franzluebbers, A.J., Azooz, R.H., 2004. Surface-soil structural properties under grass and cereal production on a Mollic Cyroboralf in Canada. Soil Tillage Res.77, 15–23.

Aucamp, A.J., 2008. Linking cultivated pastures with rangelands. Newsletter of the Grassland Society of Southern Africa, 8, 110–115.

Ayoubi, S., Karchegani, P.M., Mosaddeghi, M.R., Honarjoo, N., 2012. Soil aggregation and organic carbon as affected by topography and land use change in western Iran. Soil Tillage Res. 121, 18–26.

Balabane, M., Balesdent, J., 1996. Major contribution of roots to soil carbon storage inferred from maize cultivated soils. Soil Biol. Biochem. 28 (9), 1261–1263.

Baldock, J.A., Skjemstad, J. O., 2000. Role of the soil matrix and minerals in protecting natural organic materials against biological attack. Org. Geochem. 31, 697–710.

Barger, N.N., Archer, S.R., Campbell, J.L., Huang, C., Morton, J.A., Knapp, A.K., 2011. Woody plant proliferation in North American drylands: a synthesis of impacts on ecosystem carbon balance. J. Geophys. Res. Biogeosci. 116, 1–17.

Bartoli, F., Burtin, G., Guerif, J., 1992. Influence of organic matter on aggregation in Oxisols rich in gibbsite or in goethite. II. Clay dispersion, aggregate strength and water-stability. Geoderma 54, 259–274.

Barthès, B.G., Kouakoua, E., Larré-Larrouy, M.C., Tantely M.R, Edgar, F. de Luca, Azontonde, A., Carmen S.V., Neves, J., Pedro L. de Freitas, Christian, L. F., 2008. Texture and sesquioxides effects on water-stable aggregates and organic matter in some tropical soils. Geoderma 143, 14–25.

Barzegar, A., Asoodar, M., Ansari, M., 2000. Effectiveness of sugarcane residue incorporation at different water contents and the Proctor compaction loads in reducing soil compactibilty. Soil Tillage Res. 57, 167–172.

Blair, N., 2000. The impact of cultivation and sugarcane trash management on soil carbon fractions and aggregate stability. Soil Tillage Res. 55, 183–191.

Beare, M.H., Hendrix, P.F., Coleman, D.C., 1994. Water-stable aggregates and organic matter fractions in conventional and no-tillage soils. Soil Sci. Soc. Am. J. 58, 777–786.

Bescansa, P., Imaz, M.J., Virto, I., Enrique, A., Hoogmoed, W.B., 2006. Soil water retention as affected by tillage and residue management in semiarid Spain. Soil Tillage Res. 87, 19–27.

Bouyoucos, G.J., 1962. Hydrometer method improved for making particle size analyses of soils. J. Agron. 54, 464–465.

Blanco-Canqui, H., Lal, R., 2004. Mechanism of carbon sequestration in soil aggregates. Crit. Rev. Plant Sci. 72, 693–701.

Blanco-Canqui, H., Wienhold, B.J., Jin, V.L., Schmer, M.R., Kibet, L.C., 2017. Long-term tillage impact on soil hydraulic properties. Soil Tillage Res. 170, 38–42.

Briske, D.D., Fuhlendorf, S.D., Smeins, F.E., 2005. State-and-transition models, thresholds, and rangeland health: a synthesis of ecological concepts and perspectives. Rangel. Ecol. Manag. 58, 1–10.

Bronick, C. J., Lal, R., 2005. Soil structure and management: A review. Geoderma 124, 3–22.

Bühmann, C., Escott, B.J., Hughes, J.C., 2004. Soil mineralogy research in South Africa, 1978 to 2002 - a review. S. Afr. J. Plant Soil, 21, 316–329.

Bullinger-Weber, G., Bayon, R.L., Guenat, C., Gobat, J., 2007. Influence of some physicochemical and biological parameters on soil structure formation in alluvial soils. Eur. J. Soil Biol. 43, 57–70.

Calhoun, F. G., Hammond, L.C., Caldwell, R.E., 1973. Influence of particle size and organic matter on water retention in selected Florida soils. Procs Soil Crop Sci Soc Fla. 32, 111–113.

Callesen, I., Liski, J., Raulund-Rasmussen, K., Olsson, M.T., Tau-Strand, L., Vesterdal, L., Westman, C.J., 2003. Soil carbon stores in Nordic well-drained forest soils - relationships with climate and texture class. Global Change Biol. 9, 358–370.

Cambardella, C.A., Elliott, E.T., 1992. Particulate soil organic matter. Changes across a grassland cultivation sequence. Soil Sci Soc Am J. 56, 777–783.

Cambardella, C.A. Elliott, E.T., 1994. Carbon and nitrogen dynamics of soil organic matter fractions from cultivated grassland soils. Soil Sci Soc Am J. 58, 123–130.

Carter, M.R., 1992. Influence of reduced tillage systems on organic matter, microbial biomass, macro-aggregate distribution and structural stability of the surface soil in a humid climate. Soil Tillage Res. 23, 361–372.

Carter, M.R., Skjemstad, J.O., MacEwan, R.J., 2002. Comparison of structural stability, carbon fractions and chemistry of krasnozem soils from adjacent forest and pasture areas in south-western Victoria. Aust. J. Soil Res. 40, 283–297.

Castro Filho, C., Lourenço, A., Guimarãesde, M.F., Fonseca, I.C.B., 2002. Aggregate stability under different management systems in a red Latosol in the State of Paraná, Brasil. Soil Tillage Res. 65, 45–51.

Celik, I., 2005. Land-use effects on organic matter and physical properties of soil in a southern Mediterranean highland of Turkey. Soil Tillage Res. 83, 270–277.

Cerdá, A., 1998. Soil aggregate stability under different Mediterranean vegetation types. Catena, 32, 73–86.

Ćirić, V., Manojlovic, M., Nesic, Lj., Belic, M., 2012. Soil dry aggregate size distribution: effects of soil type and land use J. Soil Sci. Plant Nutr. 12, 689–703.

Conant, R.T., Paustian, K., Elliot, E.T., 2001. Grassland management and conversion into grassland: effects on soil carbon. Ecol. Appl. 11, 343–355.

Cousin, I., Nicoullaud B., Coutadeur, C., 2003. Influence of rock fragments on the water retention and water percolation in a calcareous soil. Catena 53, 97–114.

Chaplot, V., Bouahom, B., Valentin, C., 2010. Soil organic carbon stocks in Laos: spatial variations and controlling factors. Glob. Change Biol. 16 (4), 1380–1393.

Chaplot, V., Brown, J., Dlamini, P., Eustice, T., Janeau, J.L., Jewitt, G., Lorentz, S., Martin, L., Mchunu, C. N., Oakes, E., Podwojewski, P., Revil, S., Rumpel, C., Zondi. N., 2011. Rainfall simulation to identify the storm scale mechanisms of gully bank retreat. Agric. Water Manage. 98, 1704–1710.

Chen, X., Zhang, Z., Chen, X., Shi, P., 2009. The impact of land use and land cover changes on soil moisture and hydraulic conductivity along the karst hillslopes of southwest China. Environ. Earth Sci. 59, 811–820.

Chenu, C., Le Bissonnais, Y., Arrouays, D., 2000. Organic matter influence on clay wettability and soil aggregate stability. Soil Sci. Soc. Am. J. 64, 1479–1486.

Chenu, C., Plante, A.F., Puget P., 2006. Organo-mineral relationships. encyclopedia of soil science doi:10.1081/E-ESS-120006622.

Chivenge, P., Murwira, H., Giller, K., Mapfumo, P., Six, J., 2007. Long-term impact of reduced tillage and residue management on soil carbon stabilization: implications for conservation agriculture on contrasting soils. Soil Tillage Res. 94, 328–337.

Churchman, G.J., Skjemstad, J.O., Oades, J.M., 1993. Influence of clay minerals and organic matter on effects of sodicity on soils. Aust. J. Soil Res. 31, 779–800.

Dalal, R.C., Bridge, B.J., 1996. Aggregation and organic matter storage in sub-humid and semi-arid soils. In: Carter, M.R., Stewart, B.A. (Eds.), Structure and Organic Matter Storage in Agricultural Soils. CRC Press, Boca Raton, FL, pp. 263–307.

De Mastro, F., Traversa, A., Cocozza, C., Pallara, M., Brunetti, G., 2020. Soil Organic Carbon Stabilization: Influence of Tillage on Mineralogical and Chemical Parameters. Soil Syst. 4(3), 58.

Del Galdo, I., Six, J., Peressotti, A., Cotrufo, M.F., 2003. Assessing the impact of land-use change on soil carbon sequestration in agricultural soils by means of organic matter fractionation and stable carbon isotopes. Glob Chang Biol. 9, 1204–1213.

Denef, K., Six, J., 2005. Clay mineralogy determines the importance of biological versus a biotic process for macro-aggregates formation and stabilization. Eur. J. Soil Sci. 65, 468–479.

Diego, C., Claire, C., Yves, L.B., 2006. Aggregate stability and microbial community dynamics under drying-wetting cycles in a silt loam soil. Soil Biol. Biochem. 38, 2053–2062.

Dlamini, P., Chivenge, P., Chaplot, V., 2016. Overgrazing decreases soil organic carbon stocks the most under dry climates and low soil pH: a meta-analysis shows. Agric. Ecosyst. Environ, 221, 258–269.

Dominy, C., Haynes, R., 2002. Influence of agricultural land management on organic matter content, microbial activity and aggregate stability in the profiles of two Oxisols. Biol. Fert. Soils 36, 298–305.

Duiker, S.W., Rhoton, F.E., Torrent, J., Smeck, N.E., Lal, R., 2003. Iron (hydrous) oxide crystallinity effects on soil aggregations. Soil Sci. Soc. Am. J. 67, 606–611.

Du Toit, B., 1993. Soil acidification under forest plantations and the determination of the acid neutralising capacity of soils. MSc Thesis, University of Natal: Pietermaritzburg, South Africa.

Elliott, E., 1986. Aggregate structure and carbon, nitrogen, and phosphorus in native and cultivated soils. Soil Sci. Soc. Am. J. 50, 627–633.

Emerson, W.W., McGarry, D., 2003. Organic carbon and soil porosity. Aust. J. Soil Res. 41, 107–118.

Eusterhues, K., Rumpel, C., K¨ogelknabner, I., 2005. Organo-mineral associations in sandy acid forest soils: importance of specific surface area, iron oxides and micropores. Eur. J. Soil Sci. 56, 753–763.

Everson, C.S., Molefe, G.L., Everson, T.M., 1998. Monitoring and Modelling Components of the Water Balance in a Grassland Catchment in a Summer Rainfall Area of South Africa. WRC Report No. 493. Water Research Commission, Pretoria, South Africa.

Everson, C.S., 2000. The water balance of a first order catchment in the montane grasslands of Natal. J. Hydrol. 24, 110–123.

Evrendilek, F., Celik, I., Ertekin, C., 2004. Changes in soil organic carbon and other physical soil properties along adjacent Mediterranean forest, grassland, and cropland ecosystems in Turkey. J. Arid Environ. 59, 743–752.

Fattet, M., Fu, Y., Ghestem, M., Ma, W., Foulonneau, M., Nespoulous, J. Le Bissonnais, Y., Stokes, A., 2011. Effects of vegetation type on soil resistance to erosion: Relationship between aggregate stability and shear strength. Catena 87, 60–69.

Feller, C., Beare, M.H., 1997. Physical control of soil organic matter dynamics in the tropics. Geoderma 79, 69–116.

Fey, M.V., 2010. Soils of South Africa. Cambridge University Press, Cape Town, South Africa, pp. 25–34.

Fey, M.V., Manson A.D., Schutte, R., 1990. Acidification of the pedosphere. South Afr. J. Sci. 86, 403–406.

Filimonova, S., Kaufhold, S., Wagner, F.E., Hausler, W., Kögel-Knabner, I., 2016. The role of allophane nano-structure and Fe oxide speciation for hosting soil organic matter in an allophanic Andosol. Geochim. Cosmochim. Acta 180, 284–302.

Franzluebbers, A.J., Arshad, M.A., 1996. Water-stable aggregation and organic matter in four soils under conventional and zero tillage. Can. J. Soil Sci. 76, 387–393.

Franzluebbers, A.J., Langdale, G.W., Schomberg. H.H., 1999. Soil carbon, nitrogen, and aggregation in response to type and frequency of tillage. Soil Sci Soc Am J. 63, 349–355.

Fu, Z., Hu, H., Beare, M., Thomas, S., Carrick, S., Dando, J., Langer, S., Müller, K., Baird, D., Lilburne, L., 2021. Land use effects on soil hydraulic properties and the contribution of soil organic carbon. J. Hydrol. 602, 126741.

Gijsman, A.J., Thomas, R.J., 1995. Aggregate size distribution and stability of an Oxisol under legume-based and pure grass pastures in the eastern Colombian savannas. Aust. J. Soil Res. 33, 153–65.

Golchin, A., Oades, J.M., Skjemstad, J.O., Clarke, P., 1994. Soil structure and carbon cycling. Aust. J. Soil Res. 32, 1043–1068.

Grohmann, F., 1960. "Distribucao de tamanho de poros entres tipos de solos do Estado ae Sao Paulo". Bragantia 19, 319–328. [Cross Ref]

Gubevu, S.J., 1997. A study of the pH-dependent charge of a selection of highly weathered humic soils. PhD Thesis, University of Cape Town, South Africa.

Guo, L.B., Gifford, R.M., 2002. Soil carbon stocks and land use change: a meta-analysis. Global Change Biol. 8, 345–360.

Gush, M.B., Dye, P.J., 2009. Water use efficiency within a selection of indigenous and exotic tree species in South Africa as determined using sap flow and biomass measurements. Acta Hort. 846 323–330.

Hammad, E.A., Dawelbeit, M.I., 2001. Effect of tillage and field condition on soil physical properties, cane and sugar yields in vertisols of Kenana sugar estate, Sudan. Soil Tillage Res. 62, 101–109.

Hartemink, A.E., 1998. Changes in soil fertility and leaf nutrient concentration at a sugar cane plantation in Papua New Guinea. Commun. Soil Sci. Plant Anal. 29, 1045–1060.

Haynes, R.J., 1999. Labile organic matter fractions and aggregate stability under short-term, grass-based leys. Soil Biol Biochem. 31, 1821–1830.

Haynes, R.J., 2000. Interactions between soil organic matter status, cropping history, method of quantification and sample pretreatment and their effects on measured aggregate stability. Biol. Fertil. Soils, 30, 270–275.

Haynes, R.J., Beare, M.H., 1997. Influence of six crop species on aggregate stability and some labile organic matter fractions. Soil Biol. Biochem. 29, 1647–1653.

Haynes, R.J., Naidu, R., 1998. Influence of lime, fertilizer and manure applications on soil organic matter content and soil physical conditions: a review. Nutr. Cycl. 51, 123–137.

Haynes, R.J., Swift, R.S., 1990. Stability of soil aggregates in relation to organic constituents and soil water content. J. Soil Sci. 41, 73–83.

Hernánz, J.L., López, R., Navarrete, L., Sánchez-Girón, V., 2002. Long-term effects of tillage systems and rotations on soil structural stability and organic carbon stratification in semi-arid central Spain. Soil Tillage Res. 66, 129–141.

Hillier, S., 2003. Quantitative analysis of clay and other minerals in sandstones by X-ray powder diffraction (XRPD). Int. Assoc. Sediment. Spec. Publ. 34, 213–251.

Howard, P.J.A., Howard, D.M., 1990. Use of organic carbon and loss-on-ignition to estimate soil organic matter in different soil types and horizons. Biol. Fert. Soils, 9, 306–310.

Hobley, E., Willgoose, G.R., Frisia S, Jacobsen G., 2016. Stability and storage of soil organic carbon in a heavy-textured Karst soil from south eastern Australia. Soil Res. 49, 1015–1026.

Hudson, B.D., 1994. Soil organic matter and available water capacity. J Soil Water Conserv. 49, 189–194.

Hunter, A.H., 1974. Tentative ISFEI soil extraction procedure. International Soil Fertility Evaluation and Improvement Project. N.C. State University, Raleigh, N.C. USA.

Igwe, C.A., 2005. Soil physical properties under different management systems and organic matter effects on soil moisture along soil catena in southeastern Nigeria. Trop. Subtrop. Agroecosystems 5, 57–66.

Igwe, C.A., Akamigbo, F.O.R., Mbagwu, J.S.C., 1995. Physical properties of soils of south-eastern Nigeria and the role of some aggregating agents in their stability. Soil Sci.160, 431– 441.

Igwe, C.A., Akamigbo, F.O.R., Mbagwu, J.S.C., 1999. Chemical and mineralogical properties of soils in south-eastern Nigeria in relation to aggregate stability. Geoderma 92, 111–123.

Igwe, C.A., Zarei, M., Stahr K., 2009. Colloidal stability in some tropical soils of southeastern Nigeria as affected by iron and aluminium oxides. Catena 77, 232–237.

Igwe, C.A., Zarei, M., Stahr, K., 2013. Soil hydraulic and physico-chemical properties of Ultisols and Inceptisols in south-eastern Nigeria. Arch. Agron. Soil Sci. 59, 491–504.

Inagaki, T.M., Mueller, C.W., Lehmann, J., Kögel-Knabner, I., 2019. Andosol clay re-aggregation observed at the microscale during physical organic matter fractionation. J. Plant Nutr. Soil Sci. 182, 145–148.

IUSS Working Group WRB. 2014. World Reference Base for Soil Resources: International Soil Classification System for Naming Soils and Creating Legends for Soil Maps; World Soil Resources Reports No. 106. FAO, Rome.

Iversen, B.V., Moldrup, P., Schjonning, P., Loll, P., 2001. Air and water permeability in differently textured soils at two measurement scales. Soil Sci. 166, 643–659.

Jacinthe, P.A., Lal, R., Kimble, J.M., 2001. Organic carbon storage and dynamics in croplands and terrestrial deposits as influenced by subsurface tile drainage. Soil Science, 166, 322–335.

Jamison, V.C., Kroth, E.M., 1958. Available moisture storage capacity in relation to textural composition and organic matter content of several Missouri soils. Procs Soil Sci Soc Am J. 22, 189–192.

Jarvis, N., Koestel, J., Messing, I., Moeys, J. Lindahl, A., 2013. Influence of soil, land use and climatic factors on the hydraulic conductivity of soil. Hydrol. Earth Syst. Sci. 17, 5185–5195.

Jastrow, J.D., Miller, R.M., Lussenhop, J., 1998. Contributions of interacting biological mechanisms to soil aggregate stabilization in restored prairie. Soil Biol. Biochem. 30, 905–916.

Jim, C.Y., 2003. Conservation of soils in culturally protected woodlands in rural Hong Kong. For. Ecol. Man. 175, 339–353.

Jobbágy, E.G., Jackson, R.B., 2000. The vertical distribution of soil organic carbon and its relation to climate and vegetation. Ecol. Appl. 10, 423–436.

Kaiser, K., Eusterhues, K., Rumpel C., Guggenberger, G., Ko¨gel-Knabner, I., 2002. Stabilization of organic matter by soil minerals – investigations of density and particle-size fractions from two acid forest soils. J. Plant. Nutr. Soil Sci. 165, 451–459.

Kaiser, K., Guggenberger, G., 2007. Sorption stabilization of organic matter by microporous goethite: Sorption into small pores vs. surface complexation. Eur. J. Soil Sci. 58, 45–59.

Kavvadias, V.A., Alifragis, D., Tsiontsis, A., Brofas, G., Stamatelos, G., 2001. Litter fall, litter accumulation and litter decomposition rates in four forest ecosystems in northern Greece. For. Ecol. Manag. 144, 113–127.

Kar, G., Kumar, A., Panigrahi, S., Dixit, P.R., Sahoo, H.N., 2017. Particle size distribution, soil organic carbon stock and water retention of some upland use system of Odisha and assessing their interrelationship. J. Indian Soc. Soil Sci. 65, 48–53.

Kargas, G., Londra, P.A., 2015. Effect of tillage practices on the hydraulic properties of a loamy soil. Desalin. Water Treat. 54, 2138–2146.

Kemper, W.D., Rosenau, R.C., 1986. Aggregate stability and size distribution. In: Klute, A. (Ed), Methods of soil analysis: part 1 physical and mineralogical methods, 5.1, second edition. American Society of Agronomy, Madison, WI. pp. 837–871.

Keren, R., Singer, M.J., 1990. Effect of pH on permeability of clay-sand mixture containing hydroxy polymers. Soil Sci Soc Am J. 54, 1310–1315.

Klein, C., Dutrow, B., 2007. Mineral science. John Wiley and Sons, Inc., Hoboken, NJ. pp 675.

Klute, A., 1986. Water retention: Laboratory Methods analysis. In: Klute, A. (Ed), Methods of Soil Analysis: Part 1 Physical and Mineralogical Methods, 5, American. Society of Agronomy Madison, WI. pp: 635–662.

Ko¨gel-Knabner, I., Guggenberger, G., Kleber, M., Kandeler, E., Kalbitz, K., Scheu S., Eusterhues, K., Leinweber, P., 2008. Organo-mineral associations in temperate soils: Integrating biology, mineralogy, and organic matter chemistry. J. Plant Nutr. Soil Sci. 171, 61–82.

Kornecki, T.S., Fouss, J.L., 2011. Sugarcane residue management effects in reducing soil erosion from quarter drains in southern Louisiana. Am. Soc. Agric. Biol. Eng. 27, 597–603.

Krull, E.S., Baldock, J.A., Skjemstad, J.O., 2013. Functions of Soil Organic Matter and the Effect on Soil Properties. Grains Research & Development Corporation Final Report CSO 00029.

Lado, M., Ben-Hur, M., 2004. Soil mineralogy effects on seal formation, runoff and soil loss. Appl. Clay Sci. 24, 209–224.

Lado, M., Paz, A., Ben-Hur, M., 2004. Organic matter and aggregate size interactions in infiltration, seal formation and soil loss. Soil Sci Soc Am J. 68, 935–942.

Lal, R., 1978. Physical properties and moisture retention characteristics of some Nigerian soils. Geoderma 21, 209–223.

Lal, R., 2002. Soil carbon dynamics in cropland and rangeland. Environ. Pollut. 116, 353–362.

Lal, R., 2004. Soil carbon sequestration impacts on global climate change and food security. Science 11, 1623–1627.

Lal, R., 2005. Forest soils and carbon sequestration. For. Ecol. Man. 220, 242–258.

Lal, R., Kimble, J.M., 1997. Conservation tillage for carbon sequestration. Nutr. Cycling Agroecosyst. 49, 243–253.

Lal, R., Wilson, G.F., Okigbo, B.N., 1979. Changes in properties of an alfisol produced by various crop covers. Soil Science, 127, 377–382.

Lawes, M.J., Midgley, J.J., Chapman, C.A., 2004. South Africa's forests: the ecological and sustainable use of indigenous timber resources. In: Lawes MJ, Eeley, HAC, Shackleton CM, Geach BGS (Ed) Indigenous forests and woodlands in South Africa: Policy, people and practice. Scottville, University of KwaZulu-Natal Press, pp 31–63.

Le Bissonnais, Y., 1996. Aggregate stability and assessment of soil crustability and erodibility: I. Theory and methodology. Eur. J. Soil Sci. 47, 425–437.

Le Bissonnais, Y., Duval, O., Gaillard, H., 2002. Measurement of aggregate stability for the evaluation of the sensitivity to sealing and crusting. INRA Orlèans Unitè de Science du Sol.

Leifeld, J., Bassin, S., Fuhrer, J., 2005. Carbon stocks in Swiss agricultural soils predicted by land-use, soil characteristics, and altitude. Agric. Ecosyst. Environ. 105, 255–266.

Liu, H., Lei, T.W., Zhao, J., Yuan, C.P., Fan, T.Y., Qu, L.Q., 2011. Effects of rainfall intensity and antecedent soil water content on soil infiltrability under rainfall conditions using the run off-on-out method. J. Hydrol. 396, 24–32.

Li, X.G., Li, F.M., Zed, R., Zhan, Z.Y., Bhupinderpal, S., 2007. Soil physical properties and their relations to organic carbon pools as affected by land use in an alpine pastureland. Geoderma 139, 98–105.

Li, H., Shen, H., Chen, L., Liu, T., Hu, H., Zhao, X., Zhou, L., Zhang, P., Fang, J., 2016. Effects of shrub encroachment on soil organic carbon in global grasslands. Sci. Rep. 6, 28974.

Li, H., Yao, Y., Zhang, X., Zhua, H., Wei, X., 2021. Changes in soil physical and hydraulic properties following the conversion of forest to cropland in the black soil region of Northeast China. Catena 198, 104986.

Lugato, E., Simonetti, G., Morari, F., Nardi, S., Berti, A., Giardini, L., 2010. Distribution of organic and humic carbon in wet-sieved aggregates of different soils under long-term fertilization experiment. Geoderma 157, 80–85.

Macvicar, C. N., Fitzpatrick, R. W., Sobczyk, M. E., 1984. Highly weathered soils in the east coast hinterland of Southern Africa with thick, humus-rich A1 horizons. J. Soil Sci. 35, 113–115.

Madikizela, S., 2014. The effects of burning or mulching of harvest residues on selected soil properties in a Eucalyptus plantation in northern KwaZulu-Natal. MSc thesis. University of KwaZulu-Natal.

Malepfane, N.M., Muchaonyerwa, P., Hughes, J.C., Zengeni, R., 2022. Land use and site effects on the distribution of carbon in some humic soil profiles of KwaZulu-Natal, South Africa. Heliyon, 8, e08709.

Marshall, C.G.A., von Brunn, V., 1999. The stratigraphy and origin of the Natal Group. South Afr. J. Geol. 102(1), 15–25.

Matejovic, I. 1996. The application of Dumas method for determination of carbon, nitrogen, and sulphur in plant samples. Rostlinna Vyroba 42, 313–316.

Mbanjwa, V.E., Hughes, J.C, Muchaonyerwa, P., 2022. Organic Carbon and Aggregate Stability of Three Contrasting Soils as Affected by Arable Agriculture and Improved Pasture in Northern KwaZulu-Natal, South Africa. J. Soil Sci. Plant Nutr. https://doi.org/10.1007/s42729-022-00815-x

McBride, E.F., 1963. A classification of common sandstones. J. Sediment. Petrol. 33, 664–669.

McBride, R.A., MacIntosh, E. E., 1984. Soil survey interpretations from water retention data: 1. Development and validation of a water retention model. Soil Sci Soc Am J. 48, 1338–1343.

Meyer, J.H, van Antwerpen, R., 2010. Advances in sugarcane soil fertility research in Southern Africa. S. Afr. J. Plant Soil, 27, 19–31.

Mills, A.J., Fey, M.V., 2004. Declining soil quality in South Africa: effects of land use on soil organic matter and surface crusting. S. Afr. J. Sci. 99, 429–436.

Milne, R.M., Haynes, R.J., 2002. Comparative effects of annual permanent dairy pastures on soil physical properties in the Tsitsikamma region of South Africa. MSc thesis. University of KwaZulu-Natal.

Minasny, B., McBratney, A.B., 2017. Limited effect of organic matter on soil available water capacity. Eur. J. Soil Sci. 69, 39–47.

Motavalli, P.P., Palm, C.A., Parton, W.J., Elliott, E.T., Frey, S.D., 1995. Soil pH and organic C dynamics in tropical forest soils: evidence from laboratory and simulation studies. Soil Biol. Biochem. 27, 1589–1599.

Mthimkhulu, S.S., 2011. The effects of trees and grass on soil aggregate stability in Potshini catchment, KwaZulu-Natal, South Africa. MSc thesis. University of KwaZulu-Natal.

Mubarak, A.R., Elshami, O.M.E., Azhari, A.A., 2005. Long- and short-term effects of cultivation on properties of a vertisol under sugarcane plantation. Soil Tillage Res. 84, 1–6.

Mucina, L., Rutherford, M.C., 2006. The vegetation of South Africa, Lesotho and Swaziland. South African National Biodiversity Institute. Pretoria.

Mureva, A., Ward, D., Pillay, T., Chivenge, P., Cramer, M., 2018. Soil organic carbon increases in semi-arid regions while it decreases in humid regions due to woody-plant encroachment of grasslands in South Africa. Sci. Rep. 8, 15506.

Nciizah, A.D., Wakindiki, I.I.C., 2014. Rainfall pattern effects on crusting, infiltration and erodibility in some South African soils with various texture and mineralogy. Water S. Afr. J. 40, 57–64.

Nemes, A., Rawls, W.J., Pachepsky, Y.A., 2005. Influence of organic matter on the estimation of saturated hydraulic conductivity. Soil Sci. Soc. Am. J. 69, 1330–1337.

Nxumalo, B.N.G., 2015. Growth and yield responses of commercial sugarcane cultivars to residue mulching in different environments, MSc Dissertation, University of Pretoria, South Africa.

Oades, J.M., 1984. Soil organic matter and structural stability: mechanisms and implications for management. Plant Soil, 76, 319–337.

Oades, J.M., Waters, A.G., 1991. Aggregate hierarchy in soils. Aust. J. Soil Res. 29, 815–828.

Olson, K.R., Al-Kaisi, M.A., 2015. The importance of soil sampling depth for accurate account of soil organic carbon sequestration, storage, retention and loss. Catena 125, 33–37.

Pam, H., Brain, M., 2007. ‘Interpreting soil test results. What do the numbers mean?’ NSW Department of Natural Resources. (CSIRO Publishing: Melbourne).

Paustian, K., Collins, H.P., Paul, E.A., 1997. Management controls on soil carbon. In: Paul EA, Paustian, K, Elliot ET, Cole CV (Eds) Soil organic matter in temperate agroecosystems. CRC Press, Boca Raton, FL. pp 15–49.

Park, G.N., 1971. Changes in the structure and composition of the vegetation/soil system, Maymorn Ridge, Southern Tararua Range. M.Sc. Thesis. Victoria University of Wellington, New Zealand.

Payne, R.W., Murray, D.A., Harding, S.A., Baird, D.B., 2011. Soutar, D. M. GenStat for Windows, 18th Edition; VSN International, Hemel Hempstead, UK.

Peng, X., Yan, X., Zhou, H., Zhang, Y.Z., Sun, H., 2015. Assessing the contributions of sesquioxides and soil organic matter to aggregation in an Ultisols under long-term fertilization. Soil Tillage Res. 146, 89–98.

Perfect, E., Kay, B.D., 1990. Relations between aggregate stability and organic components for a silt loam soil. Can. J. Soil Sci. 70, 731–735.

Perfect, E., Kay B.D., van Loon, W.K.P., Sheard, R.W., Pojasok, T., 1990. Factors influencing soil structural stability within a growing season. Soil Sci. Soc. Am. J. 54, 173–179.

Podwojewski, P., Grellier, S., Mthimkhulu, S., Titshall, L., 2014. How tree encroachment and soil properties affect soil aggregate stability in an eroded grassland in South Africa. Soil Sci. Soc. Am. J. 78, 1753–1764.

Podrázský, V., Holubik, O., Vopravil, J., Khel, T., Moser, W.K., Prknova, H., 2015. Effects of afforestation on soil structure formation in two climatic regions of the Czech Republic. J. For. Sci., 55, 299–305.

Price, K., Jackson, C.R., Parker, A.J., 2010. Variation of surficial soil hydraulic properties across land uses in the southern Blue Ridge Mountains, North Carolina, USA. J. Hydrol. 383, 256–268.

Pulleman, M.M., Marinissen, J.C.Y., 2004. Physical protection of mineralizable C in aggregates from long-term pasture and arable soil. Geoderma 120, 273–282.

Qongqo, L.I., van Antwerpen, R., 2000. Effect of long-term sugarcane production on physical and chemical properties of soils in KwaZulu-Natal. Procs S. Afr. Sugar Technol. Assoc. 74, 114–121.

Rabbi, S.M.F., Wilson, B.R., Lockwood, P.V., Daniel, H., Young, I.M., 2014. Soil organic carbon mineralization rates in aggregates under contrasting land uses. Geoderma 216, 10–18.

Ramos, M.C., Nacci, S., Pla, I., 2003. Effect of raindrop impact and its relationship with aggregate stability to different disaggregation forces. Catena 53, 365–376.

Rasmussen, C., Southard, R.J., Horwath W.R., 2006. Mineral control of organic carbon in a range of temperate conifer forest soils. Global Change Biology. 12, 834–847

Rasmussen, C., Heckman, K., Wieder, W.R., Keiluweit, M., Lawrence, C.R., Berhe, A.A., Blankinship, J.C., Crow, S.E., Druhan, J.L., Hicks Pries, C.E., Marin-Spiotta, E., Plante, A.F., Schädel, C., Schimel, J.P., Sierra, C.A., Thompson, A., Wagai, R., 2018. Beyond clay: towards an improved set of variables for predicting soil organic matter content. Biogeochemistry 137, 297–306.

Rawls, W.J., Pachepsky, Y.A., Ritchie, J.C., Sobecki, T.M., Bloodworth, H., 2003. Effect of soil organic carbon on soil water retention. Geoderma 116, 61–76.

Ridge, R., 2013. Sugarcane fertilizing for high yield. IPI Bulletin No. 21. International Potash Institute, Coral Cove, Australia.

Rhodes, R., Miles, N., Hughes, J.C., 2018. Interactions between potassium, calcium and magnesium in sugarcane grown on two contrasting soils in South Africa. Field Crops Res. 223, 1–11.

Roth, C.H., Castro Filho, C., de Mediros, G.B., 1991. "Análise de fatores f´ısicos e qu´ımicos relacionados com a agregação de um Latossolo Roxo distrófico". R. Bras. Ci. Solo 15, 241–248. [Cross Ref]

SASA (South African Sugar Association)., 2015. The South African Sugar Industry 2014/2015 progress report. Available at www.sugar.org.za/sasri. [Accessed on 23/07/2022].

Saxton, K.E., Rawls, W.J., 2006. Soil water characteristic estimates by texture and organic matter for hydrologic solutions. Soil Sci Soc Am J. 70, 1569–1578.

Schmidt, M.W.I., Torn, M.S., Abiven, S., Dittmar, T., Guggenberger, G., Janssens, I.A., Kleber, M., Kogel-Knabner, I., Lehmann, J., Manning, D.A.C., Nannipieri, P., Rasse, D.P., Weiner, S., Trumbore, S.E., 2011. Persistence of soil organic matter as an ecosystem property. Nature. 478, 49–56.

Schulze, R.E., 1997. South African atlas of agrohydrology and climatology. Rep.TT 82/96. Water Resource Commission, Pretoria, South Africa.

Schwendenmann, L., Pendall. E., 2006. Effects of forest conversion into grassland on soil aggregate structure and carbon storage in Panama: Evidence from soil carbon fractionation and stable isotopes. Plant Soil 10, 1110–1115.

Seema, D., Dahiya, R., Phogat, V.K., 2019. Investigation of Hydraulic Properties of Soils Varying in Texture, Organic Carbon and Soluble Salt Contents of Arid and Semi-arid Regions. Int. J. Curr. Microbiol. App. Sci. 8(4), 2827–2838.

Selim, H.M., Newman, A., Zhang, L., Arceneaux, A., Tubaña., B, Gaston, L.A., 2016. Distributions of organic carbon and related parameters in a Louisiana sugarcane soil. Soil Tillage Res.155, 401–11.

Shainberg, I., Singer, M.J., Jynitzky, P., 1987. Effect of aluminium and iron oxides on hydraulic conductivity of sandy loam soil. Soil Science Society of America Journal 51, 1283–1287.

Shainberg, I., Levy, G.J., 1994. Organic polymers and soil sealing in cultivated soils. Soil Science, 158, 267–273.

Shepherd, T.G., Saggar, S., Newman, R.H., Ross, C.W., Dando, J.L., 2001. Tillage induced changes in soil structure and soil organic matter fractions. Aust. J. Soil Res. 39, 465–489.

Shwetha, P., Varija, K., 2015. Soil water retention curve from saturated hydraulic conductivity for sandy loam and loamy sand textured soils. Aquatic Procedia 4, 1142 – 1149.

Singer, M.J., Southard, R.J., Warrington, D.N., Janitzky, P., 1992. Stability of synthetic sand-clay aggregates after wetting and drying cycles. Soil Sci Soc Am J. 56, 1843–1848.

Six, J., Elliott, E.T., Paustian, K., Doran, J.W., 1998. Aggregation and soil organic matter accumulation in cultivated and native grassland soils. Soil Sci Soc Am J. 62, 1367–1377.

Six, J., Elliott, E.T., Paustian, K., 2000a. Soil macroaggregate and microaggregate formation: a mechanism for C-sequestration under no tillage agriculture. Soil Biol. Biochem. 32, 2099–2103.

Six, J., Paustian, K., Elliott, E.T., Combrink, C., 2000b. Soil structure and soil organic matter: I. Distribution of aggregate-size classes and aggregate associated carbon. Soil Sci Soc Am J. 64, 681–689.

Six, J., Bossuyt, H., Degryze, S., Denef, K., 2004. A history of research on the link between (micro) aggregates, soil biota, and soil organic matter dynamics. Soil Tillage Res. 79, 7–31.

Skjemstad, J.O., Clarke, P., Taylor, J.A., Oades, J.M., McClure, S.G., 1996. The chemistry and nature of protected carbon in soil. Aust. J. Soil Res. 34, 251–271.

Snyman, H.A., du Preez, C.C., 2005. Rangeland degradation in an semi-arid South Africa – II: Influence on soil quality. J. Arid Environ. 60, 483–507.

Soil Classification Working Group. 2018. Soil Classification: A Natural and Anthropogenic System for South Africa. ARC-Institute for Soil, Climate and Water, Pretoria, South Africa.

Stone, J.A., Buttery, B.R., 1989. Nine forages and the aggregation of a clay loam soil. Can. J. Soil Sci. 69, 165–169.

Strudley, M.W., Green, T.R., Ascough, J.C., 2008. Tillage effects on soil hydraulic properties in space and time: state of the science. Soil Tillage Res. 99, 4–48.

Sumner, M.E., 1957. The physical and chemical properties of Tall Grass Veld soils of Natal in relation to their erodibility. MSc thesis. University of Natal, Pietermaritzburg.

Tayel, M.Y., Abdel-hady, M., Eldardiry, E.I., 2010. Soil structure affected by some soil characteristics. Am Eurasian J Agric Environ Sci. 7, 705–712.

Tisdall, J.M., Oades, J.M., 1982. Organic matter and water stable aggregates in soils. J. Soil Sci. 33, 141–163.

Tomasella, J., Hodnett, M.G., 1998. Estimating soil water retention characteristics from limited data in Brazilian Amazonia. Soil Science, 163, 190–200.

Torn, M.S., Trumbore S.E., Chadwick, O.A., Vitousek, P.M., Hendricks, D.M., 1997. Mineral control of soil organic carbon storage and turnover. Nature 389, 170–173.

Totsche, K.U., Amelung, W., Gerzabek, M.H., Guggenberger, G., Klumpp, E., Knief, C., Lehndorff, E., Mikutta, R., Peth, S., Prechtel, A., Ray, N., Ko¨ gel-Knabner, I., 2018. Microaggregates in soils. J. Plant Nutr. Soil Sci. 181, 104–136.

Torres-Sallan, G., Rachel, E.C., Lanigana, G.J., Reidy, B., Kenneth, A.B., 2018. Effects of soil type and depth on carbon distribution within soil macro-aggregates from temperate grassland systems. Geoderma 313, 52–56.

Torres-Sallan, G., Schulte, R.P.O., Lanigan, G.J., Byrne, K.A., Reidy, B., Simó, I., Six, J., 2017. Clay illuviation provides a long-term sink for C sequestration in subsoils. Scientific Reports 7, 45635.

Unger, P.W., 1995. Organic matter and water-stable aggregate distribution in ridge-tilled surface soil. Soil Sci Soc Am J. 59, 1141–1145.

US Salinity Laboratory Staff., 1954. The diagnosis and improvement of saline and alkali soils. Soil and Water Conservation Research Branch, Agricultural Research Service, Agriculture Handbook No. 60.

van Antwerpen, R., Meyer, J.H., 1996. Soil degradation under cultivation in Northern KwaZulu-Natal. Procs S. Afr. Sugar Technol. Assoc. 70, 29–33.

Vazquez, F.M., 1981. Formation of gibbsite in soils and saprolites of temperature-humid zones. Clay Minerals, 16, 43–52.

Velde, B., Barré, P., 2010. Soils, Plants and Clay Minerals: Mineral and Biologic Interactions. Eur. J. Soil Sci. 61, 6.

Villarreal, R., Lozano, L.A., Salazar, M.P., Bellora, G.L., Melani, E.M., Polich, N., Soracco, C.G., 2020. Pore system configuration and hydraulic properties. Temporal variation during the crop cycle in different soil types of Argentinean Pampas Region. Soil Tillage Res. 198, 104528.

Virto, I., Barré, P., Burlot, A., Chenu, C., 2012. Carbon input differences as the main factor explaining the variability in soil organic C storage in no-tilled compared to inversion tilled agrosystems. Biogeochemistry 108, 17–26.

VSN International, 2015. GenStat for Windows, 18th ed. VSN International: Hemel Hempstead, UK.

Wahren, A., Feger, K.H., Schwärzel, K., Münch, A., 2009. Land use effects on flood generation-Considering soil hydraulic measurements in modelling. Adv. Geosci. 21, 99–107.

Wakindiki, I.I.C., Ben-Hur, M., 2002. Soil mineralogy and texture effects on crust micromorphology, infiltration and erosion. Soil Sci. Soc. Am. J. 66, 897–905.

Wakgari, T., Kibret, K., Bedadi, B., Temesgen, M., Erkossa, T., 2020. Effects of long term sugarcane production on soils physicochemical properties at Finchaa sugar estate. J. Soil Sci. Environ. Manage. 11, 30–40.

Wang, T., Wedin, D., Zlotnik, V. A., 2009. Field evidence of a negative correlation between saturated hydraulic conductivity and soil carbon in a sandy soil, Water Resour. Res., 45, W07503.

Wang, J.G., Yang, W., Yua, B., Yang, W., Li, Z.X., Cai, C.F., 2016. Estimating the influence of related soil properties on macro- and micro-aggregate stability in ultisols of south-central China. Catena 137, 545–55.

Wen, Y., Xiao, J., Goodman, B., He, X., 2019. Effects of Organic Amendments on the Transformation of Fe (Oxyhydr) Oxides and Soil Organic Carbon Storage. Front. Earth Sci.7, 2019.

Wiesmeier, M., Urbanskia, L., Hobleya, E., Langc, B., von Lützowa, M., Marin-Spiottad, E., van Wesemaele, B., Rabotf, E., Ließ, M., Garcia-Francoa, N., Wollschlägerf, U., Vogelf, H.J., Kögel-Knabnera, I., 2019. Soil organic carbon storage as a key function of soils - A review of drivers and indicators at various scales. Geoderma 333, 149–162.

Wilson, M.J. 1999. The Origin and Formation of Clay Minerals in Soils: Past, Present and Future Perspectives. Clay Minerals, 34, 7–25.

Wu, X., Wei, Y., Wang, J., Wang, S., Wang, J., Cai, C., 2017. Effects of soil physicochemical properties on aggregate stability along a weathering gradient. Catena 156, 205–215.

Wuddivira, M.N., Ekwue, E.I., Stone, R.J., 2010. Modelling slaking sensitivity to assess the degradation potential of humid tropic soils under intense rainfall. Land Degrad. Dev. 21, 48–57.

Yang, S.J., 1974. Compaction studies on mechanised cane field soils. Influence of soil texture and moisture content on soil compaction. Taiwan Sug. Res. Inst. Rep. 64, 11–22.

Yeomans, J.C., Bremner, J.M., 1988. A rapid and precise method for routine determination of organic carbon in soil. Commun. Soil Sci. Plant Anal. 19, 1467–1476.

Yost, L.J., Hartemink, A.E., 2018. Effects of carbon on moisture storage in soils of the Wisconsin Central Sands, USA. Eur. J. Soil Sci. 70, 565–577.

Yost, L.J., Hartemink, A. E., 2019. Soil organic carbon in sandy soils: A review. Adv. Agron. https//: 10.1016/bs.agron.2019.07.004

Yost, L.J., Roden, E.E., Hartemink, A. E., 2019. Geochemical Fingerprint and Soil Carbon of Sandy Alfisols. Soil Syst. 3, 59.

Xulu, S., Peerbhay, K., Gebreslasie, M., Ismail, R., 2018. Drought influence on forest plantations in Zululand, South Africa, using MODIS time series and climate data. Forests. 9, 528.

Zhang, M.K., He, Z.L., Chen, G.C., Huang, C.Y., Wilson M.J., 1996. Formation and water stability of aggregates in red soil as affected by organic matter. Pedosphere, 6, 39–45.

Zhang, B., Horn, R., 2001. Mechanisms of aggregate stabilization in ultisols from subtropical China. Geoderma 99, 123–145.

Zhang, S., Li, Q., Zhang, X., Wei, K., Chen, L., Liang, W., 2012. Effects of conservation tillage on soil aggregation and aggregate binding agents in black soils of Northeast China. Soil Tillage Res. 124, 196–202.

APPENDICES

Appendix 6. 1: The summary of statistics for the aggregate size distribution and mean weight diameter (MWD) under Eshowe forest, Cedara and Karkloof grasslands (n = 9).

	> 2000	250-2000	63-250	<63	MWD
Source of variation	………………..……........	(µm)	....………………		(mm)
Site	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Depth	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Site x Depth	**<.001**	**<.001**	**<.001**	**.03**	**<.001**

All values are statistically significant at $p \leq 0.05$.

Appendix 6. 2: The summary of statistics for total organic carbon (TOC), total aluminium (Al), and total iron (Fe) in different aggregate size fractions at the sampled depths under Eshowe forest, Cedara and Karkloof grasslands (n = 9).

	0-5	5-10	10-15	15-20	20-30	30-40	40-50	50-60	60-80	80-100
Source of variation	………………………………Depth (cm)…………..…………………………..									
TOC										
Site	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Fraction	**<.001**	**<.001**	**<.001**	**<.001**	**.042**	**.007**	**<.001**	**<.001**	**<.001**	**<.001**
Site x Fraction	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Al										
Site	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Fraction	**<.001**	**<.001**	**<.001**	**<.001**	.071	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Site x Fraction	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Fe										
Site	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Depth	**<.001**	**<.001**	**.003**	.184	**.017**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**
Site x Fraction	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**	**<.001**

Values in bold are statistically significant at $p \leq 0.05$.