GENESIS, CLASSIFICATION AND LAND USE PLANNING OF SOILS DEVELOPED FROM LIMESTONE GEOLOGICAL FORMATIONS IN CROSS RIVER STATE, NIGERIA

BY

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PG/Ph.D/17/00118

DEPARTMENT OF SOIL SCIENCE

FACULTY OF AGRICULTURE

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A THESIS SUBMITTED TO THE SCHOOL OF POSTGRADUATE STUDIES, UNIVERSITY OF NIGERIA, NSUKKA.
IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE AWARD OF DEGREE OF DOCTOR OF PHILOSOPHY IN SOIL GENESIS, CLASSIFICATION AND LAND USE PLANNING IN THE DEPARTMENT OF SOIL SCIENCE, UNIVERSITY OF NIGERIA, NSUKKA

SUPERVISORS: PROF. C. L. A. ASADU
PROF. P. I. EZEAKU

AUGUST, 2021

DECLARATION

I, Ofem, Kokei Ikpi, a Postgraduate Student in the Department of Soil Science with the Registration Number PG/Ph.D/17/00118, do hereby declare that the work embodied in this Thesis was written by me and is original, and has not been submitted in part or in full for any other Diploma or Degree of this or any other University.

APPROVAL PAGE

This thesis entitled, Genesis, Classification and Land Use Planning of Soils Developed from Limestone Geological Formations in Cross River State, Nigeria meets the regulations governing the award of Doctor of Philosophy of the University of Nigeria, Nsukka and it is approved for its contribution to knowledge and literary presentation.

…….

Prof. C. L. A. Asadu
Supervisor

……

Prof. P. I. Ezeaku

Supervisor

……..

Dr. C. M. Jidere
Ag. Head of Department

DEDICATION

This work is dedicated to the pillar of my strength and protector, the one who is always present at the point of my need, the Almighty God, and to my loving father of blessed memory.

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Ofem, Kokei Ikpi

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ABSTRACT

Soils developed from limestone geological formations in the humid tropics are unique in properties and differ from similar soils in areas with reduced rainfall. This research was to genetically investigate the soils developed from limestone geological formations in Cross River State regarding their morphological, physical and chemical properties and the forms of Fe and Al, elemental oxide content, and clay mineralogy in order to classify them and identify the dominant soil-forming processes. Soil mapping units were evaluated for oil palm cultivation, and indices of chemical weathering were used for pedogenetic evaluation. Google satellite imageries of the areas underlain by limestone were obtained and used as base maps. Digital elevation models (DEM) of the areas ranging from Ishibori to Agoi-Ibami and Mfamosing variously underlain by shale-limestone-sandstone (SLS), sandstone-limestone (SL), and shale-limestone with sandstone intercalation (SLSi), respectively were obtained from USGS Explorer SRTM Arc-second Global. Two elevation ranges representing mapping units in the SLS (IH1, IH2) and three elevation ranges each in SL (AI1, AI2, AI3) and SLSi (MF1, MF2, MF3) were identified. Two profile pits were sited randomly and dug in each mapping unit, while augered soils were used to confirm soil boundaries. Soil profiles were described, while core and profile samples were obtained and processed for laboratory analyses. Soil samples meant for elemental oxide content were obtained at a depth interval of 30 cm to the soil profile depth. The samples were further pulverized with an automatic mill and homogenized to 3-4 µm before analysis. Detailed soil maps were produced at 1:4000, 1:9500 and 1:15000 in SLS, SL and SLSi, respectively. The soils were 48 to 200 cm deep with epipedons ranging from 9 to 50 cm in thickness. Many to few Fe-Mn nodules and concretions were obtained in the endopedons of SLS and SL, while cracks of various widths were recorded in the endopedons of SL. Surface soil colours were in the range of dark brown (7.5YR 3/2) to black (5YR 2.5/1) in SLS, and very dark grey (7.5YR 3/1) and black (5YR 2.5/1) in SLSi. The soils were either loamy sand, sandy loam or sandy clay loam. However, more gravels were obtained in SLS. Bulk density ranged from 0.71 to 1.66 g/cm^3. Saturated hydraulic conductivity had range of 0.49-243.60 cm/h in the soils and correlated with bulk density (r= -0.712). Ranges of 5.2-7.3, 5.3-7.7 and 5.1-6.7 were obtained in SLS, SL and SLSi, respectively, for soil pH (H_2O). Organic C had ranges of 1.37-86.64, 1.03-22.16 and 0.69-46.34 g/kg for SLS, SL and SLSi, respectively, with comparatively higher values in the lowest elevation ranges. Calcium carbonate equivalent was 17.7-38.8, 15.8-30.0 and 6.1-45.7 g/kg in soils over SLS, SL and SLSi, respectively and correlated with clay (r= 0.387) and exchangeable Ca (r= 0.413) at $p< 0.05$. Dithionite extracted Fe ranged from means of 34.7 in SLS to 8.77 g/kg in SL, while dithionite Al ranged from 4.37 in SLSi to 1.53 g/kg in SL. Dithionite Fe and Al were higher than their oxalate counterpart in the studied soils. Crystalline forms of Fe and Al were highest in soils over SLS and SLSi. The highest values of immobile oxides such as TiO_2, MnO_2, Fe_2O_3 and ZrO_2 had means of 1.01, 0.12, 7.18 and 0.05 %, respectively in IH1P1. Ruxton index, weathering index and molar ratio identified soils developed from SL and SLSi as the most weathered. Quartz, kaolinite, montmorillonite, vermiculite and palygorskite are the most dominant and distributed clay minerals in the studied soils. Lessivation, calcification-decalcification, gleization, lateritization and latosolization, faunal pedoturbation and mineralization were the dominant soil forming processes. IH1P1 and IH1P2 were classified as Loamy Skeletal, Mixed, Superactive, Isohyperthermic, Typic Rhodustults (Ferric Rhodic Acrisols; Ochric) and Loamy Skeletal, Mixed, Superactive, Ishyperthermic, Typic Paleustult (Ferric Acrisols; Cutanic), respectively, while IH2P1 and IH2P2 were Fluvaquentic Humaquept (Gleyic Cambisol; Humic) and Humic Endoaquept (Stagnic Gleyic Cambisol; Humic), respectively. AI1P2 qualified as Sandy, Mixed, Superactive, Isohyperthermic, Rhodic Paleudults {Ferric Acrisols (Ochric)}, while AI1P2 was Fine loamy, Mixed, Superactive, Isohyperthermic, Plinthic Paleudults

{Plinthic Acrisols (Differentic, Hyperdystric, Ochric)}. AI3P1 qualified as Humic Endoaquept {Dystric Gleyic Cambisols (Humic, Loamic)}, while AI3P2 met the criteria for Typic Endoaquept {Dystric Cambisols (Arenic, Ochric)}. Currently, IH1, AI1, AI2, MF1 and MF2 were S2 for oil palm cultivation; however, only MF1P2 and MF2 were S1, potentially. In the land capability classification system, IH2, AI3 and MF3 were rated as class III soils, while IH1, AI1, AI2P1, MF1P1 and MF2 were rated as class II soils. However, AI2P2 and MF1P2 met the requirements for class I soils. Soils developed from SL and SLSi are more developed than those over SLS and were more suitable for oil palm cultivation, however, the entire soils are at the transitory stage from intermediate to an advanced stage of development.

CHAPTER ONE
INTRODUCTION

Soil genesis is the mode of origin of the soil with special reference to the factors and processes responsible for the development of the solum from unconsolidated parent materials (Soil Science Society of America, 2008). Parent materials have major influences on the properties of overlying soils and provide a starting material upon which other soil-forming factors such as; climate, living organisms, relief and time serve to modify the formation of soils. Fertile soil formation from inorganic bedrocks undergoes a complex network of biological, physical and chemical weathering processes and usually requires hundreds of years through complex interactions to convert the inorganic precursors to distinct soil horizons (Esu, 2010). The formation of such soil is a unique product of the combination of geologic materials, geomorphologic history, the presence and activity of biota, the history of land use and disturbance regime (Aislabie and Deslippe, 2013) as regulated by the factors of soil formation (Paul and Clark, 1996; Egli *et al*., 2011).

State factors, pedogenic processes and their interaction over time co-relate with landscape segments to improve soil mapping. According to Ezeaku (2011), detailed soil surveys are helpful in soil genesis and extrapolation of information and solve problems related to the distribution of soils. New trends in soil survey facilitate soil mapping through technologies such as remote sensing and geographic information system (GIS) to extrapolate points and provide approaches to meet the demand of resource-related modelling (Mermut and Eswaran, 2001; Salehi *et al*., 2003). The use of digital elevation models (DEM), satellite data and digital geological data to improve map quality has been investigated (Bayramin, 2001). Earlier, Vink (1970) opined that remote sensing techniques facilitate the process of soil surveys, but the technology should not replace but complement field surveys (Ezeaku, 2011). Through soil mapping, analysis and categorization put data obtained from resource inventory into a form that is useful to farmers by evaluating the mapping units for diverse land use types. Land use planning seeks to evaluate and assess land as a basis for decisions involving land use to reconcile competing demands for land and reduce incidences of soil degradation (Ameztegui *et al*., 2016).

The Geologic make-up of an environment determines soil type (Ibanga, 2006; Ekwueme, 2004). However, not all minerals in parent rocks are found in soil (Asadu *et al*., 2012). Nevertheless, the studied soils are likely to have alkaline to near alkaline properties, perhaps due to the presence of carbonates which influence soil genesis, especially in young soils (Ciolkosz *et al.,* 1995). In Turkey, Irmak *et al*. (2007) obtained Xeric Haplocalcid and

Lithic Xeric Haplocalcid, while Sevink and Verstraten (1979) obtained Rendollic Eutrochrepts in the Netherlands for soils developed from limestone. The increasing thickness of the solum and the progressive weathering of carbonate minerals with depth are obvious pieces of evidence of soil development in limestone regions (VandenBygaart and Protz, 1994; Carroll and Hathaway, 1953). Soils developed on limestone have variable depths but generally less than 4 m (Carroll and Hathaway, 1953; Hartemink and Bridges, 1995) and are exceptionally shallow (<30 cm) in some areas (Sedov *et al.,* 2008). The pH of the soils in Southeastern Nigeria ranges from 4.8 to 7.1 (Ibanga *et al*., 2005) and 6.8 in Tanzania (Hartemink and Bridges, 1995). Exchangeable Ca^{2+} is variably high in most limestone soils as Ibanga *et al*. (2005) obtained a range of 1.2-20 cmol/kg in the limestone soils of Southeast Nigeria.

Mineralogical studies of soils developed from limestone in Northern Yucatan (Estrada-Medina *et al.,* 2010), Northwest Spain (Castro *et al.,* 1999) and Western Thailand (Intamo *et al.,* 2014) indicate the presence of minerals at various percentage abundance and include; kaolinite, quartz, chlorite, vermiculite and illite, while feldspars, muscovite and oxides of iron are sparingly obtained. Other researchers have identified smectites as the most abundant clay mineral in $CaCO_3$ rich soils (Singer, 1989; Yilmaz, 1990; Irmak *et al*., 2007); however, Irmak *et al.* (2007) obtained palygorskite as second most abundant clay mineral. The proportion of $CaCO_3$ and non-carbonate accessories in limestone determines the rate of pedogenesis (Samonil, 2007). Common pedogenic processes in limestone areas are podzolization (Carroll and Hathaway, 1953), eluviation-illuviation (Zagorski, 2010), bioturbation (Sedov *et al.,* 2008) and fundamental pedogenic processes such as weathering, decarbonation and humification (Castro *et al.,* 1999).

There is a dearth of information regarding recent soil surveys in Cross River State, especially the use of remote sensing techniques such as satellite imageries and DEM for soil resources inventory in the limestone areas of the state. Also, there have been no identified published studies on the genesis and mineralogy as well as elemental oxide content of the limestone soils of the state. Previous studies by Ibanga *et al.* (2005), Aki and Antigha (2015) and Afu *et al*. (2017) focused on the morphological, physical and chemical properties as well as classification of the soils in the Southern rainforest of the state.

Despite its extensive occurrence, over the years, the industrial use of limestone in the state has masked the agricultural potentials of the soils overlying it, and so have not been fairly studied. This study seeks to bridge the gap, by undertaking a more detailed study of the soils and classifying same for appropriate land use recommendations and possible transfer for

use and management of similar soils elsewhere. The use of selected soil chemical indices of weathering to evaluate the genesis of the soils will aid the understanding of soil fertility status and development. This will further reconcile competing demands for land under various kinds of land use. Also, mining activities have led to ecological imbalances and influenced the agricultural land use of the soils. This study is therefore apt, as the presence or absence of weatherable minerals indicate the stage of soil development from which the fertility or infertility of the soil can be inferred and land use recommendations made; , especially as it concerns the capability and suitability of the soils for specified crop production.

The main objective of the research was to study the genesis of soils of limestone origin in Cross River State, Nigeria, to classify them and make land use recommendations.

Specific objectives included to:

i. carry out a detailed survey of some soils of limestone formations in Cross River State and study the properties, including morphological, physical and chemical properties;

ii. study the forms of Fe and Al, and elemental oxide content as well as clay mineralogy of the soils;

iii. identify the dominant soil-forming processes taking place and infer on the genesis of the soils;

iv. classify the soils according to the criteria of the USDA Soil Taxonomy and correlate with the World Reference Base for Soil Resources System (WRB); and

v. establish capability and suitability classes for oil palm cultivation.

CHAPTER TWO
LITERATURE REVIEW

Limestone is a sedimentary rock composed largely of calcite ($CaCO_3$), formed by either organic or inorganic processes (Serra, 2006). Such forms are of high quality and may contain over 80 % $CaCO_3$ (Fatoye and Gideon, 2013). The Calabar flank is the main carbonate province in Nigeria, with well-developed tropical karsts and caves (Reijer and Nwajide, 1998; Reijer and Petters, 1987), with the Odukpani limestone occurring in substantial quantity and thickness (Obaje, 2009).

2.1 Morphological Properties of Soils Developed from Limestone

Soil morphology describes the physical constitution of a soil profile as exhibited by the kinds, thickness, and arrangement of the horizons in the profile and the texture, structure, consistence, and porosity of each horizon (Soil Science Society of America, 2008; Ibanga, 2006). It describes the observable field attributes of the soil within the various horizons as well as the kinds and orientation of the horizons (Buol *et al.,* 2003). The above properties and how they are influenced by limestone are further elaborated below;

2.1.1 Horizonation

Soil horizonation constitutes the most significant contribution of soil genesis to soil surveys. Soil horizonation is the development of horizons in soil due mainly to a single or a combination of soil-forming processes (Soil Science Society of America, 2008). Therefore, active soil-forming processes in an environment are considered necessary for soil development as expressed by well-formed soil horizons. The soil layer is preferred if the soil properties are inherited from the parent material; on the other hand, horizons show the effects of pedogenesis such as the accumulation of illuvial clay, salts, plinthites etc. (Soil Science Division Staff, 2017). Also, the parent material weathers at the surface and works its way down. As a result, the upper most layers have been changed the most, while the deepest layers are most similar to the parent material, hence horizonation as influenced by soil-forming processes (Odoemena and Uchua, 2014).

According to Irmak *et al*. (2007), structural formation in limestone soils gave rise to the formation of cambic B and calcic C horizons (Ck). According to the authors, Ck indicates the accumulation of carbonates in the horizons. Buringh (1979), Buol *et al*. (1980) and Dinc *et al*. (1987) further stated that cambic B horizons are often developed alongside calcic C horizons (Ck) in limestone soils of arid and semi-arid regions. Earlier studies by Carroll and Hathaway (1953) indicated the occurrence of A, AB, B, BC and C horizon sequence with thicknesses of 0-10, 10-15, 15-21, 21-23 and 23-26 cm, respectively, in the Lenoir limestone

soils of Augusta county. However, in the tropical rainforest area of Cross River State, Afu *et al*. (2017) obtained A, B and C horizons in the limestone soils of Odukpani without Ck horizons. In the limestone soils of the Harran region in Turkey, A, Bw, and pronounced calcic horizons (Ck) were obtained with varying thicknesses (Irmak *et al*. 2007) and observed cambic B horizons, which they attributed to soil structural formation. Soil horizons in the Apodi plateau karst environment were very well defined and developed with clear and well-delineated A, B and C horizons; however, cambic and vertic horizons were also obtained (Girao *et al*., 2014). Singleton and Lavkulich (1987) observed that soil horizons along a chronosequence became progressively deeper and more differentiated with soil age, while VandenBygaart and Protz (1994) stated that iron and soil organic matter were likely transported to the B horizons where they were immobilized as coatings or cutans on sand grains in the limestone soils.

In the highly calcareous soils of Southern Iran, Emadi *et al*. (2008) observed developed A, B and C horizons with the B horizons dominated by carbonates; in addition, the argillic horizons had clay cutans and strong structures as primary characteristics. Calcic horizons are often traced to arid and semi-arid regions in carbonate and non-carbonate soils but develop as an illuvial horizon in carbonate-rich soils (Rabenhorst *et al*., 1991; Khormali *et al*., 2003). Calcic horizons limit root proliferation and water movement, affect CEC, soil buffering capacity, as well as nutrient fixation and availability.

The accumulation of organic matter in karstic plains encourages the development of mollic epipedons, leading to kastanozems in arid zones and phaeozems in areas with more moisture (Bautista *et al*., 2011). According to Folmer's model (Folmer, 1998), the nodulithic, lithic and Rendzic leptosols of tropical karst in Mexico were dominated by R, C/R, and A/C/R horizon sequences, respectively (Bautista *et al*., 2011); such horizon combinations are indicators of the youthfulness of the soils. However, Folmer (1998) obtained A/Bw/C/R and A/Bt/C/R horizon sequences in karstic depressions, while A/E/Bt1/Bt2/Ct/R horizon sequence was obtained in well-drained karstic soils. Such variations in horizon combinations or sequences in various climates is a direct result of variation in soil forming processes and soil development.

Luvisols with strong evidence of lessivation leading to the formation of BCt horizons are common in high rainfall areas and show the accumulation of secondary calcium carbonate translocated from one part of the soil to another with the resultant formation of BCtk horizons (Bautista *et al*., 2011). Horizon sequences of A/B, A/B/C and A/C were common in the soils containing carbonates in the Apodi plateau in Brazil with the accumulation of carbonates and

clay in the B horizons, which were considered as indicators of pedogenic processes in the soils except for the profile pit with A/C horizons (Ferreira *et al*., 2016). Sedov *et al*. (2008) obtained Ah/AC/C horizon sequence in an undisturbed forest tropical karst landscape in Yucatan, while soils in the depression were mainly A/Bg/C.

2.1.2 Horizon Thickness and Soil Depth

Soils may be shallow (less than 50 cm), moderately deep (50 - 100 cm) and deep (> 100 cm) to the consolidated parent rock (Osman, 2018). However, Soil Science Division Staff (SSDS) (2017) classifies depth to root restriction in the order < 25 cm, 25-50 cm, 50 – 100 cm, 100-150 cm and 150 cm or more as very shallow, shallow, moderately deep, deep and very deep, respectively. Variation in soil depth within short distances may result in complex soil depth definition, especially when tongues occur at horizon boundaries, thereby making it inadequate to describe soil boundary using the usual terms of topography (SSDS, 2017). Because of its impact on plant growth, internal water dynamics and land management, depth to bedrock is a crucial feature in agricultural soils, especially if encountered within 2 m of soil depth.

Soil depth is related to soil age and also depends on moisture as well as the parent material underlying the soil. There exist a positive, linear Chrono function of the thickness of B horizon with soil age; hence, as horizons increase in age, they are more easily identified than when they are in their early stages of development (SSDS, 2017; VandenBygaart and Protz, 1994). It is not surprising as carbonate weathering in limestone areas and the establishment of a pH gradient between the B and C horizons occur, resulting in the immobilization of Fe transported from overlying horizons (VandenBygaart and Protz, 1994). Continued soil development over time increases thickness of the solum. However, the thickness of the solum in the limestone area of Pinery provincial park, Ontario, ranged from 41 and 95 cm and, was described as shallow (VandenBygaart and Protz, 1994).

Pure limestone with trace amounts of insoluble residue will produce only thin soils, where such an area is influenced by erosion resulting in well-defined hill slopes, very little soil will remain, and the bedrock may be exposed (Carroll and Hathaway, 1953). However, if the limestone is impure and contains chert, sandy layers and clay bands, a thick soil may develop because the residuum is relatively insoluble and provides a body of material in which soil-forming processes can act (Carroll and Hathaway, 1953). This further affirms the fact that soils developed from $CaCO_3$ rich parent materials are poorly developed (Esu, 2010). Consequently, the Lenoir limestone soils in Augusta County were shallow and only 26 cm deep (Carroll and Hathaway, 1953), while studies by Afu *et al*. (2017) in the limestone soils

of Odukpani indicated that the studied soils had a minimum depth of 121 cm. The limestone soils of the Harran region in Turkey were shallow to moderately deep, and soil depth ranged from 60 to 110 cm (Irmak *et al*., 2007). Different soils developed from limestone may vary from a few centimetres to meters; this constitutes variations in limestone rock (Bautista *et al*., 2011) and can be considered a reason for the variation in soil depth from 1.9 m to over 4.3 m within a horizontal distance of 70 m (Girao *et al*., 2014).

Evans and Cameron (1979) found the depth of solum and the thickness of the Ah horizon to increase with soil age, while VandenBygaart and Protz (1994) observed that the most apparent evidence of soil development with age is the increasing thickness of the solum and the progressive weathering of carbonate minerals with depth. Aksoy *et al*. (2009) reported that soils overlying Neogene clay lime deposits in Northwest Bursa, Turkey, were shallow while a few others were very deep. According to them, shallow depths were found to restrict the agricultural potentials of the soils.

Depths of highly calcareous soils in southern Iran ranged from 85 to 150 cm; however, water tables were encountered at depths of 100, and 130 cm in two of the profiles studied (Emadi *et al*., 2008), while the depths of occurrence of pedogenic calcium carbonate in soil profiles was partly dependent on rainfall. In their study on the tropical karst areas of Mexico, Bautista *et al*. (2011) observed that soil depth or the amount of fine earth would increase in a given direction based on increasing rates of rock dissolution. Also, increase in soil depth and fine earth amount with increasing rainfall in the Northern Peninsula of Yucatan led to changes in soil development from Nodulithic Leptosols to Lithic Leptosols and later Rendzic Leptosols (Bautista *et al*., 2011).

All the soils containing carbonates in the Apodi plateau, Brazil, were above 95 cm deep except the one impeded by consolidated limestone deposit at 27 cm with thin Ap horizons of 3-10 cm and extensive B horizons (> 65 cm) (Ferreira *et al*., 2016). Limestone soils in Southeastern China were described as shallow, with low weathering status and mollic epipedons, quite different from profound red, deeply weathered lateritic soils on shale and Pleistocene sediments (Singer, 1989). A report by Sedov *et al*. (2008) indicates thin limestone soils of < 30 cm depth in the upland of the undisturbed forest; however, other upland soils were thinner with more stones; while soils in the depression were somewhat deeper. Consequently, wetland soils are often shallower than their upland counterparts (Sedov *et al*., 2008; Esu, 2010).

2.1.3 Soil Colour

Soil colour is usually influenced by soil moisture content, amount and state of organic matter, amount and state of iron oxide, and soil aeration. Though other morphological soil properties are as important as soil colour, it serves as the first criterion in the delineation of horizons within a given soil profile, especially when other morphological soil properties seem difficult to use as differentiae during delineation (Esu, 2010). Colour alone does not affect the soil, but it is often a reliable indicator of other soil properties. For instance, reddish colours indicate the presence of an oxidized form of iron, dark colour in the surface soils indicates a significant presence of organic matter, soils frequently saturated by water appear grey, blue or green because the minerals have been leached away and often exist in reduced conditions (Nsor and Ibanga, 2006).

Soil colour is an important aspect of soils developed from limestone (Aydinalp and Fitzpatrick, 2009). Reddish-brown colours dominated the cambic B horizons of the limestone soils of the arid region of Turkey (Irmak *et al*., 2007). The matrix soil colour in the Pinery dune soils overlying limestone was dominated by the hue of 10YR in the surface and subsurface soils and scantily 5YR, while its value ranged from 2 to 6 in the entire soils. Similarly, chroma of 1 dominated the surface soils while the subsurface soils ranged from 1 to 8 (VandenBygaart and Protz, 1994). Carroll and Hathaway (1953) obtained yellowish and brown to reddish-brown matrix soil colour in the Lenoir limestone soils in Augusta County, while the limestone soils of Harran region in Turkey had matrix soil colour notation of 7.5YR 4/6 and 5YR 4/6 in the surface soil and 5YR 4/4 to 7.5YR 4/6 in the subsurface soils, and were described as reddish-brown (Irmak *et al*., 2007).

Colour (moist) notation of soils developed from the karst environment of the Apodi plateau ranged from 5YR 3/4 to 7.5YR 4/3 to 10R 3/4 in the surface soils. In contrast, the subsurface soils had dominant hues of 2.5YR, 5YR and 10YR with a value ranging from 3 to 8 (Girao *et al*., 2014). According to them, the reddish soil colours were attributed to rapid precipitation of iron oxides due to high pH, favouring the precipitation of ferrihydrites, a precursor of hematite rather than goethite (SilvaNeto, 2008; Tardi and Nahon, 1985). Furthermore, VandenBygaart and Protz (1994) found that the darkening of the chroma in the B horizons was likely due to Fe–oxyhydroxides' pigmentation property and translocated organic matter. This combination forms dark coatings on mineral grains as a result of translocation from the eluvial horizons and through *in situ* weathering of Fe rich minerals. In the highly calcareous soils of southern Iran, the hue of 10R was obtained in the surface soils with the value ranging from 4 to 6, and chroma from 1, 3 to 4; the subsurface soils, however,

had notations from 7.5YR 4/4 to 10YR 8/3 with fine to medium and prominent to distinct mottles obtained in the soils of the area (Emadi *et al*., 2008).

The carbonate-rich soils of Apodi plateau in Brazil varied in soil colour notation (wet) from the surface to the subsurface soils; 5YR 4/3, 10YR 4/4, 5YR 3/3 and 10YR 3/4 were common in the surface soils, while the B horizons were dominated by 2.5YR (3/4, 4/4) while 5YR (4/4, 5/8, 6/8), 10YR (5/4, 5/8), 2.5Y 5/6 and 7.5YR 4/6 were obtained in the C horizons (Ferreira *et al*., 2016). Such colours were attributed to slow water permeability mainly due to prismatic soil structure, clayey textural class and proximity to the rock while reddish colours were traced to seasonal fluctuation of water. Iron released from the parent material in the dry season precipitates and gives rise to pedogenic hematite (Yaalon, 1997); yellowish colours however indicated the presence of goethite. Report by Sedov *et al*. (2008) on the soils overlying tropical karst in Yucatan revealed that colour notation varied from 5YR 3/3 (dark reddish brown) to 10YR 3/2 (very dark grayish brown) and 7.5YR 7/1 in the surface soils when dry with slight increase or decrease of 1 in value or chroma; while the subsurface soils were mainly 7.5YR 3/4 (dark brown).

2.1.4 Soil Texture

Soil texture is the relative proportions of the various soil separates in a soil sample as described by the classes of soil texture and shown in a soil textural triangle (Soil Science Society of America, 2008). Soil texture may be modified by the addition of suitable adjectives when rock or other fragments are present in substantial amounts; for example, “stony silt loam”. Soil texture influences a number of soil properties including bulk density, soil consistence, water and nutrient retention, and heat conductance, as well as soil drainage conditions, water holding capacity, amount and size of pores, and plant root development.

The Lenoir limestone soils in Augusta were sandy with some clay in the soil surface and less sandy with dense and clayey layer in the subsurface soils (Carroll and Hathaway, 1953). Sandy clay loam and loam textural classes were obtained in the surface soils while the subsurface soils were dominated by clay, especially in the middle slope and valley bottom of the Odukpani limestone soils of Cross River State (Afu *et al*., 2017). In the Harran limestone region of Turkey, soils overlying the limestone deposit had texture of clay loam in the surface soils and sandy clay to clay textural classes in the subsurface soils (Girao *et al*., 2014). Furthermore, Aksoy *et al*. (2009) reported clay, clay loam and sandy clay loam in the top 30 cm of the soils overlying Neogene clay lime in Northwest Bursa, Turkey. In the highly calcareous surface soils of Southern Iran, textural classes were mainly clay loam to loam and silty clay loam, and the underlying soils were dominated by clayey and loamy textural classes

(Emadi *et al*., 2008). Soil texture in the surface horizons of the carbonate rich soil of Apodi plateau in Brazil were mainly loam with variation due mainly to silt, clay and gravel content (Ferreira *et al*., 2016). The subsurface soils of this plateau were loamy, silty and clayey depending on the extent of occurrence of pedogenic processes in the area; however, clay and silty loam tend to dominate most subsurface horizons (Ferreira *et al*., 2016).

2.1.5 Soil Structure

Soil structure refers to the arrangement of primary soil particles into aggregates called peds. Aggregates are naturally occurring clusters or groups of soil particles in which the forces holding the particles together are stronger than the forces between the peds (Nsor, 2004). Soil structure is described in the field by using a combination of three qualities which form the name of the structure. The grade is written first and describes the degree of aggregation within and between aggregates. It may be weak, moderate or strong. The size of soil aggregates is described by the class of structure which can be very fine, fine, medium, coarse and very coarse. Finally, the type of soil structure describes the shape of the aggregates which includes granular, crumby, platy, blocky, subangular blocky, prismatic and columnar. These three qualities should make up a fully described soil aggregate (Esu, 2010).

Large amounts of coarse fragments tend to inhibit the formation of stable structural aggregates. Soil structural formation depends on the mineralogical composition, electrical properties, shape and orientation of solid particles; the nature and properties of soil water and its ionic composition; and the interaction of forces between soil particles, soil water and their adsorption complexes (Punmia *et al.*, 2005). Common agents of aggregation which are responsible for binding primary soil particles into peds include: colloidal clay minerals, sesquioxides, microbial gums, humus and carbonates (Esu, 2010).

Parent materials have profound influence on soil structure. For instance, Webster and Wilson (1980) stated that parent materials high in ferromagnesian minerals will weather to give soils high in iron and of good structure while granite low in iron but high in quartz (SiO_2) will weather into weakly structured soil. Soils developed on parent materials rich in sesquioxides such as; goethite, dolerite, limonite, gibbsite, and lepidocrocites increase aggregate stability and structure whereas parent materials with high relative sodium levels give rise to soils with agriculturally poor physical properties including dispersed soil particles and poor soil structure (Gray and Murphy, 1999).

Soils overlying limestone in the Harran region of Turkey had moderate and strong medium angular and subangular blocky structures in the surface soils while the subsurface soils were moderate medium granular and angular blocky to massive structural types (Irmak

et al., 2007). Similarly, soil structure in the karst environment of the Apodi plateau was dominated by various grades of granular and subangular blocky structures; while the subsurface soils were moderate, medium and coarse subangular and angular blocky with prismatic structures at the C horizons (Girao *et al*., 2014). Furthermore, the highly calcareous soils in Southern Iran were mainly angular blocky in structure for the surface soils and subangular blocky, angular blocky and massive in the subsurface soils (Emadi *et al*., 2008), while findings by Ferreira *et al*. (2016) indicates that soil structure in the carbonate soils of Apodi plateau, Brazil were weak to moderate and strong; coarse, medium, fine and very fine granular, subangular blocky and angular blocky structural types (Ferreira *et al*., 2016). Fine granular soil structure was obtained by Sedov *et al*. (2008) in a tropical karst landscape in Yucatan while the B horizon had subangular blocky structure. From the foregoing, variation in the structural type of limestone soils is certain and depends on the amount of clay and carbonates which dominate soil cementing agents.

2.1.6 Soil Consistence

Soil consistence is the physical condition of a soil at various moisture contents (wet, moist and dry) as evidenced by the behavior of that soil towards manipulations (Esu, 2010; Ibanga, 2006). It describes the degree of resistance of peds to rupture and is often described under moist, wet and dry soil moisture conditions. Sandy soils exhibit minimal cohesive and adhesive properties and are so easily deformed that automobiles easily get stuck (Esu, 2010). Consistence depends on the soil water state and its description has little or no meaning unless the water state class is specified; this is however different for soil consistency (SSDS, 2017).

The surface soils of the highly calcareous soils of Southern Iran were friable and firm in consistence while the subsurface soils ranged from friable, firm and slightly hard (Emadi *et al*., 2008). Loose, soft and very hard consistencies (dry) were obtained in the surface soils while very friable, friable and very firm consistencies were obtained when moist, and non-plastic, slightly plastic and very plastic consistencies (wet) in the surface of the soils of the Apodi plateau in Brazil (Ferreira *et al*., 2016). The subsurface soils of the Apodi plateau were hard, slightly hard, extremely hard and soft in consistencies (dry) and friable, very friable and extremely firm (moist) while consistencies under wet condition indicated slightly plastic, plastic and very plastic.

2.1.7 Soil Nodules and Concretions

Soil nodule is a cemented concentration of a chemical compound such as calcium carbonate, silica or iron oxide that can be removed from the soil intact and has no orderly internal organization. On the other hand, a cemented concentration of a chemical compound,

such as calcium carbonate or iron oxide, that can be removed from the soil intact and that has crude internal symmetry organized around a point, line, or plane is called a concretion. Example is Fe-Mn concretion (Soil Science Society of America, 2008). Nodules have also been described as uniform concentrations of oxides or carbonates while concretions are concentric rings or admixtures of usually two or more materials. One may therefore have Fe or Mn or $CaCO_3$ nodules as against Fe-Mn or Mn-$CaCO_3$ concretions (Esu, 2010).

According to Irmak *et al*. (2007), leaching and accumulation of carbonates were indicated by the presence of secondary carbonate nodules. The carbonate nodules dominated the Ck horizons at depths of 55-95 cm and 30-48 cm. Oliveira *et al*. (2000) described cambisols developed from limestone in the north of the state of Minas Gervais and highlighted the presence of nodules and concretions of iron – manganese in the profiles. Studies in the karst environment of the Apodi plateau region have shown variation in the presence of iron nodules (Alencar, 2002; Mota *et al*., 2007), while Yaalon (2009) reported that nodules of various specifications in terms of size, hardness and distribution were common in soils developed from carbonate rocks. Furthermore, Jimenez-Millan and Nieto (2008) reported that Fe-Mn concretions in soils overlying carbonate rocks in Spain were not only common, but related to differences in faces in the source material. Moreover, developed horizons and profiles had reduced diameter of iron nodules; an evidence that the process of degradation has been experienced and weathering process has reached advanced stage.

Few to common carbonate nodules and filamentous carbonate between the nodules in the highly calcareous soils in semi-arid regions of southern Iran were obtained; according to the authors, calcite was found as powders, filaments, nodules and concretions cemented in petrocalcic horizons (Emadi *et al*. 2008). Climatic changes and low soil permeability between seasons led to redox reactions and the formation of Fe-Mn concretions as well as petroplinthites (Schwertmann and Taylor, 1989). Furthermore, the presence of Fe-Mn concretions associated with $CaCO_3$ were identified in the dry climate of Minas Gervais (Oliveira *et al*., 2000). Nodules and concretions of secondary carbonates are common in Vertisols from calcareous rocks in areas that encourage the buildup of Ca^{2+} and Mg^{2+} in the soil solution (Hendricks, 1991). Further in the process of forming nodules and concretions, primary carbonates are altered by dissolution, translocation and then precipitation to form secondary calcium carbonate concretions in the form of calcite (Rabenhorst *et al*., 1991). Report by Alencar (2002) attributed the reddish soil colour of the Apodi plateau to quartz and ferruginous concretions in sand fractions. Calcite nodules identified in calcic horizons by Shankar and Achyuthan (2007) where quartz grains cemented by micrite while Sedov *et al*.

(2008) obtained Fe nodules in the Bg horizon of soils in the tropical karst landscape in Yucatan.

2.1.8 Other Morphological Soil Properties

a. Strong to violent effervescence was produced when the highly calcareous surface and subsurface soils of southern Iran were reacted with dilute HCl (Emadi *et al*., 2008). However, effervescence with 20 % H_2O_2 in the calcareous soils of Apodi plateau in Brazil indicated the presence of manganese oxides (Ferreira *et al*., 2016).

b. Soil pores ranged from many, fine to very fine pores in the surface soils, and few to many and fine to very fine pores in the subsurface soils of the highly calcareous soils of Southern Iran (Emadi *et al*., 2008); with size and number decreasing with increase in soil depth.

c. Cutans are modifications of the fabric of natural surfaces in soil materials due to concentrations of clay, sesquioxides, salts and humus. The presence of argillans on the ped faces in the B horizons indicates an argillic horizon while sesquans and organs in the B horizons indicate the occurrence of a spodic horizon (Esu, 2010). Few thin clay cutans were obtained on ped faces of the Bt horizons of the highly calcareous soils of Southern Iran (Emadi *et al*., 2000). However, Ballagh and Rung (1970) had earlier attributed continuous clay cutans on ped faces to clay accumulation resulting from its translocation from overlying horizons. Similarly, Sedov *et al*. (2008) obtained thin clay cutans over soil aggregates in the tropical karst landscape region in Yucatan.

2.2 Soil Physical Properties

Characteristics, processes or reactions of a soil that are caused by physical forces and that can be described by, or expressed in physical terms or equations are referred to as soil physical properties (Soil Science Society of America, 2008). Some soil physical properties and how they are influenced by limestone are described below:

2.2.1 Saturated Hydraulic Conductivity of Soils

Hydraulic conductivity is a measure of a soil's ability to transmit water (Amoozegar and Warrick, 1986). Saturated hydraulic conductivity is a quantitative measure of a soil's ability to transmit water when subjected to a hydraulic gradient (SSDS, 2017). It is an important parameter in all aspects of soil water and solute movement. Hydraulic conductivity in clayey soils is lower than that of sandy soils because the pore size distribution in sandy soil favours large pores, even though sandy soils have higher bulk densities and lower total porosities than clayey soils (SSDS, 2017).

The measurement of hydraulic conductivity of saturated soils is based on the direct application of Darcy's equation to a saturated soil column of uniform cross sectional area and is given by:

$Ks = VL/At(H2-H1)$

Where V = volume of water that flows through the sample of cross sectional area A in time (t) and hydraulic head difference (H2-H1) imposed across the sample of length (L).

The K value of a saturated soil is the average hydraulic conductivity which depends on particle size distribution, shape and particle arrangement as well as the temperature, viscosity and density of the water (Sarki *et al*., 2014). Soil saturated hydraulic conductivity is directly related to its effective porosity (Ahuj*a et al*., 1984) and determined by soil structure and indirectly affected by clay mineralogy, particle size distribution and organic carbon content (Sarki *et al*., 2014). The coefficient of variability of saturated hydraulic conductivity may exceed 100 %; this variability is associated with the total porosity and pore size distribution of the soil (SSDS, 2017). Classes of saturated hydraulic conductivity are such that; ≥ 36, 3.6 to < 36, 0.36 to < 3.6, 0.036 to < 0.36, 0.0036 to < 0.036 and < 0.0036 cm/h represent very high, high, moderately high, moderately low, low and very low, respectively (SSDS, 2017).

2.2.2 Pore size distribution

Pore size distribution is the relative abundance of each pore size in a representative volume of soil (Nimmo, 2013). Pore size distribution is related to other soil properties in a complex but useful manner, it indicates greater complexity of soil structure than porosity alone. Pore size can permit distinction between micro- and macro-pores. Pore size distribution may be inferred from particle size distribution, particularly for mineral soils as their particles consist of readily definable individual grains (Rezanezhad *et al*., 2009). Weathering increases porosity and reduces water holding capacity and bulk density (SSDS, 2017).

When a saturated soil sample is successively drained and the volume of water that is removed between successive steps is measured, the volume of water removed may be equal to the soil pore volume drained. The pore size distribution is then determined if the size range of drained pores is calculated from:

$r = 2T/\rho gh.$

Theoretically, the largest pores are drained first, before successively smaller pores (Danielson and Sutherland, 1986), possibly by smaller pressures which may then be increased

to drain smaller pores. Pore space has influence on soil forming processes, just as soil forming processes affect it and affect the movement of water, air and other fluids in the soil as well as chemical reactions and the life span of soil organisms (Florian *et al*., 2011). It also plays a major role in the prediction of soil hydraulic properties (Nimmo, 1997; Ragab *et al*., 1982) and its variation has been attributed to Na^+ accumulation, organic matter deposition and clay dispersion or expansion (Jnad *et al*., 2001).

Soil structure is an important property of the soil especially in the retention and transport of solutions, gases and heat but influenced by the amount, size, configuration and distribution of soil pores (Ringrose-Voase and Bullock, 1984). According to Esu (2010), good aeration makes soil forming processes to operate faster. The pore spaces, rather than the soil particles are most useful in characterizing the soil as a medium for plant growth or other uses (Danielson and Sutherland, 1986); however, researchers and consultants of farm related projects focus more on the distribution of soil particle sizes. According to Florian *et al*. (2011), the distribution of pore sizes is more important than total porosity and often results from the settlement of soil primary particles and structural aggregates (Florian *et al*., 2011). The complex inter and intra aggregate spaces in the soil vary in size, shape, tortuosity and continuity which make quantification of the soil pores difficult or sometimes impossible mainly due to the lack of standards in classification into distinct size ranges. Soil pores larger than 50 µm allow for the circulation of water and air while those less than 50 µm retain soil water (Florian *et al*., 2011). The equivalent diameters of the pores are estimated from liquid retention and capillary pressure, while the total pore space has been determined by calculation and direct evaluation by gas pycnometer. On assumption, the above parameters of size distribution of soil pores can be made with useful accuracy. A common method to measure total porosity and pore size distribution is to use suction curve S (θ), where S is the soil suction and θ is the volumetric water content (Kosugi, 1994; Lety *et al*., 2000; Statescu *et al*., 2010; Statescu and Zauca, 2008).

In a study, it was found that irrigation with treated wastewater modifies the pore size distribution by increasing macro-pore diameter (>50 µm) and decreasing micro-pore diameter (<50 µm) (Goncalves *et al*., 2010).

Larger pores do not necessarily mean higher hydraulic conductivity of bulk soil samples, however, the saturated hydraulic conductivity of soils is affected by biogenic gas bubbles (Baird *et al*., 2004). The determination of the pore size diameters of meso- and macro-pores is essential in characterizing both saturated and unsaturated hydraulic conductivity of soils (Carey *et al*., 2007).

Coarse textured soils are less porous than fine textured soils, however, the average size of individual pores is higher in coarse soils than in fine textured soils (Florian *et al*., 2011). Organic matter also plays a key role in soil porosity. For instance; decomposed and mineralized organic matter loses its capability as a cementing agent and the pore walls which it acted as a binding agent collapses and closes the pore (Pagliai and Vignozzi, 2002), resulting in a dead end. Porosity and pore size distribution of a soil characterizes its pore space which affects and is affected by the movement of water, air and other fluids; the transport and the reaction of chemicals; and the residence of roots and other biota (Nimmo, 2013). Furthermore, soil processes can synthesize and destroy pores and bring about changes in their properties. According to Nimmo (2013), soil processes that more often affect soil pore size distribution are shrinkage, swelling, mechanical compression, disturbance from digging or plowing, biological activity and chemical activity.

The capillarity theory relates r and S by:

$$r = \frac{-2 \infty \cos \alpha}{S}$$

Where ∞ is the surface tension and α is the angle of contact (Carrillo *et al*., 1999; Watson and Lety, 1970). According to the micromorphometric method, total porosity of a soil can be classified as <5, 5-10, 10-25, 25-40 and >40 % to represent very compact, compact, moderately porous, highly porous and extremely porous soils, respectively (Pagliai, 1988).

2.2.3 Particle Size Distribution

Particle size is the effective diameter of a soil particle measured by sedimentation or sieving. The fractions of the various soil particle sizes or separates in a soil sample, expressed as mass percentages is what is described as particle size distribution and it is determined by particle size analysis (Soil Science Society of America, 2008).

Particle size distribution is an important soil property in pedogenesis, as the development of clay bulge in the B horizon is diagnostic in the identification of argillic, kandic or natric endopedons and is also important in the determination of particle size classes at the family level of Soil Taxonomy (Udo *et al*. 2009). Lithologic discontinuities are detected based on shifts in sand, silt or clay contents in adjacent horizons (Esu, 2010). Also, old parent materials have silt/clay ratio < 0.15, while the index > 0.15 is indicative of young parent material and value of 0.15 is indicative of moderately weathered soils (Ayolagha, 2001).

Findings by Olson (1958), Rickert and Tedrow (1967), Cowie (1968) and Thompson (1981) indicate that total silt and clay in soil profiles show straight line relationships with soil age; such increase may be due to the weathering of coarser fractions or influx of finer particles from wind-blown materials. However, VandenBygaart and Protz (1994) later noted that soil weathering is the main contributor to the decreasing particle size; consequently, other researchers emphasized that larger particles weather physically to smaller particle sizes with soil age (Birkeland, 1984; Protz *et al*., 1984; Singleton and Lavkulich, 1987). Soil particle size fractions and how they are influenced by limestone are further elaborated below:

a. Sand Fraction

Total sand content ranged from 13.24 to 42.72 % in the soils formed on massive limestone in Pivska Planina, Serbia (Petar *et al*., 2016), while in the Pinery dune limestone soils of Ontario, medium and fine sand dominated the particle size distribution with values of total sand ranging from 67.0 to 99.2 % in the A horizons while subsurface soil horizons ranged in values from 80.3 to 99.2 % (VandenBygaart and Protz, 1994). According to the authors, such high values indicate that the dune soils have been deposited on the limestone material over the years and attributed the sharp decrease in fine sand to a change in deposition from dune sand to beach sand. However, Carroll and Hathaway (1953) noted that the total sand content of the Lenoir limestone soils in Augusta County was dominated by very fine sand content with values ranging from 3.9 to 12.7 % and decreased with soil depth. In Odukpani limestone soils of Cross River State, sand fraction ranged from 43 to 68 % in the surface soils and 27 to 54 % in the subsurface soils (Afu *et al*., 2017). Furthermore, in the limestone soils of Harran region of Turkey, Irmak *et al*. (2007) obtained 23.48 to 34.36 % in the soil surface of the limestone soils and 22.11 to 46.52 % in the subsurface soils. Studies by Girao *et al*. (2014) in the karst region of the Apodi plateau reveals joint contribution of the particle sizes to textural classes with sand fraction ranging from 42.8 to 54.0 % in the surface soils and from 15.8 to 56.1 % in the subsurface soils. Findings by Aksoy *et al*. (2009) revealed values of sand ranging from 24.6 % in Typic Calcixeverts to 54.20 % in Typic Xerochrepts in the Neogene clay lime soils of northwest Bursa, Turkey.

Sand content varied from 10 to 43 % in the surface of highly calcareous soils of Iran and from 6 to 55 % in the subsurface soils (Emadi *et al*., 2008). Results obtained by Ferreira *et al*. (2016) in the Apodi plateau in Brazil indicated that total sand ranged from 21.3 to 36.3 % in the surface soils, while values in the subsurface soils ranged from 3.5 to 42.6 %, however, the soil's total sand content was dominated by coarse sand fraction.

b. Silt fraction

Silt content ranged from 8.76 to 38.28 % in the soils formed on massive limestone in Pivska Planina, Serbia (Petar *et al*., 2016). Consequently, straight line relationships were obtained between total silt and soil age, and suggest that soil weathering is a contributor to decreasing particle size; however, its values ranged from 0.4 to 12.4 % in the surface soils and from 0.4 to 9.0 % in the subsurface soils (VandenBygaart and Protz, 1994). Silt fraction in the Lenoir limestone soils in Augusta County dominated the particle size distribution with values ranging from 33.6 to 57.7 % with a regular decrease with soil depth (Carroll and Hathaway, 1953). Nevertheless, the value of silt in the Odukpani limestone soils was obtained as the least in the particle size distribution with ranges of 8.7–36.7 % and 4.7-28.7 % in the surface and subsurface soils, respectively (Afu *et al*., 2017); however, they obtained relatively high values in the surface soils of the middle slope. In the limestone of Harran in Turkey, ranges of 35.31- 41.81 % and 20.42-36.68 % were obtained in the surface and subsurface soils, respectively (Irmak *et al*., 2007). Girao *et al*. (2014) while studying soils on the karst environment of the Apodi plateau, obtained values of 12.5 – 19.0 % in the surface soils while values in the subsurface soils ranged from 8.9 to 47.1 %. Aksoy *et al*. (2009) obtained values of silt with ranges from 23.9 in Typic Calcixeverts to 30.2 % in Pachic Calcixerolls in the soils overlying Neogene clay lime in Northwest Bursa, Turkey.

Further studies by Emadi *et al*. (2008) obtained 28-56 % and 27-47 % as ranges for the surface and subsurface soils, respectively in the highly calcareous soils of Southern Iran. Findings by Khresat and Qudah, (2006) attributed silt accumulation in the surface horizons of the soils to continuous addition by wind. Studies by Ferreira *et al*. (2016) in the Apodi plateau of Brazil indicated a range from 37.4 to 56.7 % in the surface soils while the subsurface soil values ranged from 31.6 to 86.3 %. This soil particle fraction dominated most of the soils in the area; however, no definite trend down the soil profiles was traced to the size fraction.

c. Clay fraction

The translocation of clay may sometimes be aided by organic matter (Tan, 1976). The carboxylic functional group in the organic acid reacts with clay surface. The surface potential of the clay-organic complex appears greater than that of clay, hence a larger electro kinetic potential. As a clay-organic complex, the clay is suspended for a longer time and moves

downwards with percolating water, a factor that is responsible for clay movement and initiates flocculation of clay when it stops (Tan, 1998).

Clay content in the soils formed on massive limestone in Pivska Planina, Serbia varied from 26.1 to 78.0 % (Petar *et al*., 2016). Using chronofunctions, VandenBygaart and Protz (1994) obtained straight line relationships between clay and soil age, and suggested that soil weathering was a major contributor as against influx from wind-blown materials to decreasing particle size. This agrees with findings by earlier studies by Birkeland (1984), Protz *et al*. (1984), Singleton and Lavkulich (1987), that larger particles weather physically to smaller particle sizes with soil age. Clay content in the Pinery dune soils overlying limestone was quite low and ranged from 0.4 to 25.2 % in the surface soils while values in the subsurface soils ranged from negligible to 10.8 % (VandenBygaart and Protz, 1994). In the Lenoir limestone soils in Augusta County, clay content increased regularly with soil depth and ranged from 17.1 to 61.7 % (Carroll and Hathaway, 1953). Also, in the Odukpani limestone soils, clay had a range of 20.3 – 23.3 % in the surface soils and dominated the subsurface soils with range of 28.3-66.3 %. Clay content in the surface soils of the limestone of Harran region in Turkey ranged from 30.33 to 34.71 % and from 33.06 to 53.62 % in the subsurface soils (Irmak *et al*., 2007). In the karst environment of Apodi plateau, values of clay in the soil ranged from 28.3 to 43.2 % in the surface soils and from 12.6 to 69.9 % in the subsurface soils (Girao *et al*., 2014). Such high values of clay are evidences of clay illuviation, a process also noticed by Delgado *et al*. (2003) as a determining factor in the genesis of soils on karst regions in Spain.

Values of clay in the soils overlying Neogene clay lime deposits in Northwest Bursa, Turkey ranged from 21.40 % in the Typic Xerochrepts to 51.50 % in the Typic Calcixeverts (Aksoy *et al*., 2009). Similarly, in the highly calcareous soils of Southern Iran, clay content varied from 21 to 39 % in the surface soils and from 11 to 64 % in the subsurface soils (Emadi *et al*., 2008); such increasing clay content with soil depth was attributed to active illuviation process.

Clay has important influence on soil formation. For instance, the Arenosols and Regosols are weakly developed soils with shallow A horizon, good internal drainage, and low CEC because of low clay content (Bautista *et al*., 2011). Clay formation in the surface soils of Apodi plateau, Brazil ranged from 18.7 to 41.4 % with values bulging at the B horizon of only one profile pit; consequently, subsurface soil values ranged from 8.9 to 52.2 % (Ferreira *et al*., 2016). Accumulation of secondary carbonates in the subsurface soils encouraged clay flocculation; consequently, clay increased especially in the B horizon giving rise to a Bt

horizon (Ferreira *et al*., 2016). High clay content was reported in the A horizons of the upland Rendzinas in the tropical karst landscape of Yucatan with values greater than 70 % in the Phaeozems and about 50 % in the Leptosols (Sedov *et al*., 2008).

2.2.4 Gravel Content

Gravel contents of over 20 % in the top soil invariably means decrease in clay or fine silt which hold essential nutrients for plants and may affect plant growth and development, and influence nutrient supply, water storage and root penetration especially in shallow soils like those developed from limestone. According to the Soil Science Division Staff (SSDS) (2017), stone lines that exhibit a different lithology from the bedrock are evidence that the soils did not form entirely from the residuum.

According to Carroll and Hathaway (1953), the Lenoir limestone soils in Augusta had a range of 0.3-14.4 % for gravel with values decreasing from 14.4 % in the surface soil to 0.3 % in the C horizon. Report by Ferreira *et al*. (2016) indicates that coarse fraction ranged from 0.3 to 43.7 % in the surface soils and from 0.5 to 80.9 % in the subsurface soils.

2.2.5 Bulk density

Bulk density is the weight per unit volume of a soil. Undisturbed core, paraffin coated clod, excavation and gamma radiation densitometry techniques are some methods often used for its determination (Blake and Hartge, 1986). According to Esu (2010), soil bulk density values greater than 1.8 g/cm^3 are associated with fragipans, duripans and petrocalcic horizons, while values less than 0.90 g/cm^3 indicate the presence of volcanic ash in mineral soils low in organic matter. According to the author, bulk density values of 1.6-1.8 g/cm^3 may indicate that air and water movement are low for optimum crop growth. Its values depend on the use pattern and they generally increase with soil depth (Esu, 2005); consequently, low bulk density values (<1.3 g/cm^3) generally indicate a porous soil condition. Sands and sandy loam soils may vary in bulk density from 1.2 to 1.8 g/cm^3, whereas silt loam, clay loam, and clay soils may have bulk densities ranging from 1.0 to 1.6 g/cm^3 (Brady and Weil, 2002). Soil bulk density depends greatly on the mineral make up of soil and the degree of compaction.

The Pinery dune soils overlying limestone in Ontario had bulk density values ranging from 0.20 to 1.52 g/cm^3 in the surface soils and from 0.45 to 1.58 g/cm^3 in the subsurface soils (VandenBygaart and Protz, 1994). Soils overlying Neogene clay lime in Northwest Bursa, Turkey had bulk density values of 1.24-1.42 g/cm^3 (Aksoy *et al*., 2009).

2.3 Chemical Properties of Limestone Soils

2.3.1 Soil pH

Soil pH has been described as a measure of soil reaction by Tan (1998) and indicates the acid-base reactions in soils. According to Udo *et al*. (2009), pH value of <3.5 indicates the presence of acid sulphate while significant amounts of exchangeable Al^{3+} are indicated by pH of 4.5-5.5, and free carbonates are indicated by values of 8.0-8.5. Weathering of primary soil minerals is promoted by low soil pH (Tan, 1998); this translates to the fact that limestone soils which are often high in soil pH will discourage the weathering of primary minerals. During hydrolysis, H^+ ions are consumed while OH^- ions are produced resulting in more basic soil solution (Wild, 1993). Alumina is more soluble than silica at soil pH of <4.0; however, alumina becomes insoluble at pH range of 5-9 while silica becomes more soluble (Tan, 1998). This results in the leaching of more soluble silica and formation of kaolinite, gibbsite and bauxite (Ollier, 1975). Soil pH values that encourage the occurrence of silica advances the process of smectification (Tan, 1998). Soil pH appears to be the most important factor influencing crop performance (Agbede, 1984) as it controls the rate of organic matter decomposition, activities of microorganisms, nutrient availability and uptake by crops (Agbede, 2009).

In the soils formed from massive limestone in Pivska Planina, Serbia, soil pH (H_2O) ranged from 5.28 to 6.16, while pH (KCl) ranged from 3.88 to 4.62 (Petar *et al*. 2016). The Pinery dune soils overlying limestone in Ontario had pH ($CaCl_2$) range of 4.4-7.2 in the surface soils and 3.5-7.6 in the subsurface soils (VandenBygaart and Protz, 1994). Surprisingly, Carroll and Hathaway (1953) obtained a range of 4.3-4.6 as pH in the Lenoir limestone soils in Augusta County whereas, Afu *et al*. (2017) obtained a range of 6.3-7.9 in the limestone soils of Odukpani, Cross River State of Nigeria. Furthermore, the soils developed from limestone in Harran region of Turkey had ranges of 7.3-7.4 and 7.4-7.6 in the surface and subsurface soils, respectively (Irmak *et al*., 2007). The authors attributed the moderately alkaline pH and increasing values with soil depth to $CaCO_3$ accumulation with depth. Findings by Girao *et al*. (2014) in the karst environment of Apodi plateau results in soil pH (H_2O) ranging from 7.0 to 8.0 in the surface soils and from 5.2 to 8.8 in the subsurface soils. The trend of results obtained by VandenBygaart and Protz (1994) negates those obtained in previous studies, as soil pH decreased with soil age in the B and C horizons. This was attributed to the weathering of carbonate minerals and leaching of released ions (Ca^{2+} and Mg^{2+}), being replaced by H^+.

Findings by Aksoy *et al*. (2009) in the soils overlying Neogene lime deposits in Northwest, Turkey shows that pH ranged from 6.40 in the Vertic Xerochrepts to 7.92 in the Chromic Haploxeverts.

Emadi *et al*. (2008) recorded pH values from 7.2 to 7.8 in the surface soils and from 7.45 to 8.20 in the subsurface soils of the highly calcareous soils in semi-arid regions of Southern Iran. They further pointed out that higher pH values in the subsurface soils is not surprising for soils developed from parent materials with free carbonates. Soil pH (H_2O) in the surface soils of the Apodi plateau in Brazil ranged from 7.8 to 8.6 and from 7.4 to 8.9 in the subsurface soils with values generally increasing as the Bk or Ck horizons were approached (Ferreira *et al*., 2016). In the same soils, values of pH($CaCl_2$) were lower with range of 6.2-7.9 in the entire soils. Consequently, negative Delta pH was obtained in all the soils; an indication that net negative charges predominate the soil particle surface. Reports by Khormali *et al*. (2003) indicates a range of 7.1-8.2 (H_2O) in calcareous soils in Iran while pH of 7.0 (H_2O) and 6.5 (KCl) were obtained in some limestone soils in Spain (Taboada and Silva, 1999). While working on the uniqueness of limestone soil forming substrate of the Bohemian karst, Samonil (2007) obtained pH (H_2O) of 4.70 to 8.03. Soil pH values in a tropical karst landscape in Yucatan ranged from 6.3 to 7.6 (Sedov *et al*., 2008). The dissolution of limestone gives rise to $CaCO_3$ that increases pH values close to neutral and avoids the problems associated with tropical soils, especially excessive acidity and aluminum toxicity.

2.3.2 Organic Carbon

According to Udo *et al*. (2009), the differentiation of the eight epipedons in Soil Taxonomy is mainly based on differences in the amount of organic carbon (OC). For instance, histic epipedon contains 120-180 g/kg OC while mollic epipedon contains > 6 g/kg. However, mollic and umbric epipedons contain at least 6 g/kg while an ochric epipedon contains < 6 g/kg OC. On the other hand, fluvents contain > 2 g/kg OC or decrease irregularly with soil depth. Low content of organic carbon may be attributed to long arid periods and poor vegetation.

VandenBygaart and Protz (1994), obtained a logarithmic relationship between the mass of organic matter and soil age. Organic matter ranged from 1.0 to 557 g/kg in the surface of the Pinery dune soils over limestone in Ontario while the subsurface soils ranged from negligible values to 186 g/kg (VandenBygaart and Protz, 1994); however, lower values were obtained by Afu *et al*. (2017) in the Odukpani limestone soils with ranges of 8.0-23.0 g/kg and 3.0-9.0 g/kg in the surface and subsurface soils, respectively. Organic matter of the

soils developed from limestone in the Harran region of Turkey ranged from 1.84 to 2.97 g/kg in the surface soils and from 1.05 to 2.32 g/kg in the subsurface soils (Irmak *et al*., 2007). Soil organic carbon content in the Apodi plateau karst environment had ranges of 10.0-18.81 g/kg in the surface soils and 1.0-8.25 g/kg in the subsurface soils (Girao *et al*., 2014). Protz *et al*. (1988) earlier found that the depth at which leaching occurred was proportional to soil age and organic matter increased to a maximum at about 2,000 years of soil development. While studying soils overlying the Neogene clay lime deposits of Northwest Bursa, Turkey, Aksoy *et al*. (2009) found that, values of organic matter ranged from 6.5 g/kg in the Typic Xerorthents to 25.0 g/kg in the Pachic Calcixerolls. These values were described as low and moderate in newly deforested lands; and decreased with soil depth.

Values of organic carbon in the highly calcareous soils of Southern Iran ranged from 6 to 8 g/kg in the surface soils, and from 1.0 to 6.0 g/kg in the subsurface soils (Emadi *et al*., 2008). The low organic carbon in the area was linked to rapid organic matter mineralization in the arid region. Total organic carbon ranged from 10.9 to 23.3 g/kg in the surface soils of the Apodi plateau in Brazil while the subsurface soil values ranged from 1.2 to 20.6 g/kg and decreased with increasing soil depth in all the soil profiles studied (Ferreira *et al*., 2016).

Organic carbon content in a tropical karst landscape in Yucatan ranged from 6.8 to 153 g/kg in the entire soils with generally higher values in the surface soils. In the upland Rendzinas, organic carbon was > 80 g/kg while in the karst depression of the Leptic gleyic Phaeozems, organic carbon was greater than 150 g/kg. The accumulation and stabilization of humus as reflected in the organic carbon content was due to the near neutral pH resulting from the dissolution of $CaCO_3$ in the soils (Sedov *et al*., 2008).

2.3.3 Total nitrogen

Soil nitrogen exists in organic and inorganic forms. Most of the nitrogen in soils occur in organic form while the inorganic form constitutes only < 2 % of total N and predominantly occurs as NH_4^+ and NO_3^- for the available forms and NO^{2-} which rarely exists in soils (Udo *et al*., 2009).

The red and brown bauxite soils of Jamaica occurring on hard, pure, white limestone have not only been described as being low in terms of general fertility but the N in it has been described as being deficient (Ahmad *et al*., 1966), while ranges of 0.7-1.9 g/kg and 0.2-0.8 g/kg were obtained in the surface and subsurface soils, respectively in the Odukpani limestone soils (Afu *et al*., 2017). Samonil (2007) obtained values of 0.7-3.12 g/kg in the Bohemian karst region.

2.3.4 Available phosphorus

The determination of soil phosphorus status centers on the determination of total P, organic P and available P indices, and the fractionation of the inorganic P into saloid – bound P, Al-P, Fe-P, reductant soluble P, occluded Fe and Ca-P. The distribution and relative occurrence of the various forms are of importance in soil chemistry, fertility and genesis (Udo *et al*., 2009). Consequently, mollic and anthropic epipedons are separated on the basis that an anthropic epipedon has equal to or greater than 1500 mg/kg of available P.

In the soils developed from limestone in Odukpani, available P had ranges of 20.8-29.6 mg/kg and 18.0-39.6 mg/kg in the surface and subsurface soils, respectively (Afu *et al*., 2017). The decrease in available P with depths was attributed to the increase in clay with soil depth, mainly as a result of fixation of P in the subsurface soils. In carbonate containing soils in the Apodi plateau, Brazil, available P determined by Olsen method ranged from 3.0 to 18 mg/kg in the surface soils and from 1.0 to 19.0 mg/kg in the subsurface soils with the A/C horizon sequence profile having very high values (Ferreira *et al*., 2016). High total P is often recorded in calcareous soils due to the presence of low – solubility Ca phosphate (Pansu and Gautheyrou, 2006); however, the available P may be low. In a tropical karst landscape in Yucatan, available P ranged from 1.12 to 10.10 mg/kg in the entire soils with values decreasing as soil depth increased (Sedov *et al*., 2008).

2.3.5 Exchangeable Bases

Quantitative determination of exchangeable cations on the soil adsorption complex gives an insight of soil forming processes as well as the fertility status of the soils (Markoski *et al*., 2018). Exchangeable bases occupy charged sites on the surface of soil particles that can be readily replaced by another cation in salt solution. Cations in exchangeable forms are described as positively charged ions held on the surface of soil particles by a negative surface and which may be replaced by other positively charged ions in the soil solution. These cations are represented mainly by Ca^{2+}, Mg^{2+}, K^{+} and Na^{+} and are discussed below based on limestone influence on them;

a. Exchangeable calcium

Many authors (Spalevic *et al*., 2017., Mukaetov *et al*., 2000 and Djordjevic, 1993) had pointed out that exchangeable Ca dominates the adsorption complex of soils developed on limestone and dolomite. The cation is involved in the creation of stable granular structures and creates insoluble Ca-humates from neutralized minerals and humic acids (Markoski *et al*., 2018). Limestone soils of the Harran region in Turkey had exchangeable Ca and Mg

values ranging from 38.84 to 45.75 cmol/kg in the surface soils and from 22.86 to 54.25 cmol/kg in the subsurface soils. These values accounted for > 95 % of the soil exchange complex (Irmak *et al*., 2007). The high values were traced to high contents of Ca and Mg in the chemical composition of the parent material due mainly to the dissolution of carbonates and weathering of feldspars and soil ferromagnesian minerals. In the karst environment of the Apodi plateau, exchangeable Ca ranged from 3.1 to 6.8 cmol/kg in the surface soil and had a range of 1.5 – 7.5 cmol/kg in the subsurface soil (Girao *et al*., 2014). Furthermore, in the soils overlying Neogene clay lime in Northwest Bursa, Turkey Aksoy *et al*. (2009) obtained exchangeable Ca and Mg values ranging from 20 cmol/kg in Typic Xerochrepts to 38.5 cmol/kg in Pachic Calcixerolls; again, these values occupied over 95 % of the soil exchange complex.

Exchangeable Ca values ranged from 6.7 to 25.6 cmol/kg in the surface soils of the Apodi plateau, Brazil while subsurface soil values ranged from 4.7 to 24.4 cmol/kg (Ferreira *et al*., 2016). In the soils developed over a massive limestone in Marcedona, exchangeable Ca had means of 33.68 and 27.39 cmol/kg in the surface and subsurface soils, respectively (Markoski *et al*., 2018).

b. Exchangeable magnesium

Afu *et al*., (2017) obtained ranges of 2.2-3.2 cmol/kg and 1.8-3.2 cmol/kg in the surface and subsurface soils, respectively for the Odukpani limestone soils and rated the values moderate for the soils. However, values obtained by Samonil (2007) negates those of Afu *et al*. (2017) as exchangeable Mg ranged from 28 to 245.4 cmol/kg in the soils developed on the Bohemian karst. The surface soils of the karst environment of the Apodi plateau ranged from 1.6 to 3.3 cmol/kg while the subsurface soils ranged from 0.9 to 10.2 cmol/kg (Girao *et al*., 2014). In a related study, Ferreira *et al*. (2016) obtained values that ranged from 1.4 to 3.3 cmol/kg in the surface soils of the Apodi plateau in Brazil while subsurface values ranged from 0.9 to 7.5 cmol/kg. Similarly, Markoski *et al*. (2018) obtained mean values of 4.58 and 4.10 cmol/kg in the surface and subsurface, respectively in the soils developed on a massive limestone deposit in Macedonia.

c. Exchangeable potassium

The forms of K present in soil are always considered on the basis of the degree of availability as; readily available K, slowly available K and difficultly available K (Udo *et al*., 2009). Readily available K is K in soluble and exchangeable form and constitutes between 1-2 % of total K; while slowly available K is K fixed in biotite mica, illite and vermiculite and

forms about 2-10 % of total K in mineral soils. The K in primary minerals such as orthoclase, feldspar and muscovite mica is the difficultly available or relatively unavailable K and constitutes about 95-98 % of total K. Besides fertilizer effects, high K in farmlands may be linked to the presence of mica and parent materials rich in Ca and Mg; furthermore, the effects of mica on exchangeable K was also recorded by Khresat and Taimeh (1998) in Vertisols of limestone origin.

Low K content in adsorbed form in limestone soils is as a result of its content in the parent material and the smaller capacity of adsorption in terms of Ca and Mg, and so it is often easily expelled from the soil adsorption complex and leached away from plant root zone (Markoski *et al*., 2018). Exchangeable K in the Odukpani limestone soils had ranges of 6.0 – 8.2 cmol/kg and 5.4-8.0 cmol/kg in the surface and subsurface soils, respectively and rated moderate for the soil (Afu *et al*., 2017), while a range of 42.3-235.7 mg/kg was obtained in the Bohemian karst (Samonil, 2007). Red and brown bauxite soils of Jamaica overlying hard and pure white limestone was reportedly low in soil fertility properties and very deficient in K (Ahmad *et al*., 1966). Soils developed from limestone in the Harran region of Turkey, had ranges of 0.35-1.36 cmol/kg and 0.49-1.17 cmol/kg for exchangeable K in the surface and subsurface soils, respectively (Irmak *et al*., 2007). Exchangeable K in the soils overlying the karst environment in Apodi plateau, Brazil ranged from 0.1 to 1.1 cmol/kg in the surface soils while the subsurface soils ranged from negligible values to 2.2 cmol/kg (Girao *et al*., 2014). Similarly, Ferreira *et al*. (2016) reported values with ranges of 0.23-0.74 cmol/kg and 0.03-0.95 cmol/kg in the respective surface and subsurface soils of the plateau, while Markoski *et al*. (2018) obtained mean values of 0.51 and 0.49 cmol/kg, respectively in the surface and subsurface soils, respectively of a massive limestone deposit in Macedonia.

Exchangeable K decreased gradually with soil depth in the limestone soils of the Harran region in Turkey (Irmak *et al*., 2007) while its values in the soils developed from limestone in the Odukpani area ranged from 0.08 to 0.11 cmol/kg in the surface and subsurface soils, and rated low in exchangeable K (Afu *et al*., 2017). In the soils overlying Neogene clay lime in Northwest Bursa, Turkey values of exchangeable K ranged from 0.27 cmol/kg in the Typic Xerochrepts to 1.44 cmol/kg in Typic Calcixeverts (Aksoy *et al*., 2009).

d. Exchangeable sodium

The low amount of Na in the adsorption complex of most soils of limestone origin is due to the ease of leaching of the cation as well as low amount of silica residuum which is released by dissolving the limestone and dolomite (Markoski *et al*., 2018). According to Afu

et al. (2017), exchangeable sodium in the surface and subsurface soils had ranges of 0.06-0.9 cmol/kg and 0.05-0.08 cmol/kg, respectively in the limestone soils of Odukpani in Cross River State. These values were rated low. In soils overlying Neogene clay lime in Northwest Bursa, Turkey the values ranged from 0.3 cmol/kg in Typic Xerorthents to 1.5 cmol/kg in Vertic Xerochrepts (Aksoy *et al*., 2009), while exchangeable Na and K were low in the limestone soils of the Harran region in Turkey (Irmak *et al*., 2007). In the Bohemian karst, exchangeable Na ranged from 6.2 to 22.3 mg/kg (Samonil, 2007); furthermore, the limestone soils of the Harran region of Turkey, had values of exchangeable Na ranging from 0.28 to 0.31 cmol/kg in the surface soils and from 0.28 to 0.42 cmol/kg in the subsurface soils. However, values of exchangeable Na in the surface soils of the karst environment of the Apodi plateau, Brazil were somewhat lower as ranges of 0.1-0.4 cmol/kg and 0.1-0.8 cmol/kg were obtained in the surface and subsurface soils, respectively (Girao *et al*., 2014). On the other hand, Ferreira *et al*. (2016) reported values ranging from 0.06 to 0.61 cmol/kg in the surface soils of the carbonate rich Apodi plateau in Brazil, while the subsurface soils values ranged from 0.06 to 0.68 cmol/kg. Similarly, Markoski *et al*. (2018) obtained mean values of 0.17 and 0.15 cmol/kg in the surface and subsurface soils, respectively of a massive limestone deposit in Marcedona.

2.3.6 Cation exchange capacity (CEC)

Cation exchange capacity is the sum of exchangeable cations in soil. The completeness of replacement depends on the salt solution used since the ease of replacement of exchangeable cations varies (Udo *et al*., 2009).

Kandic and Oxic subgroups of Soil Taxonomy are differentiated on the basis of their low CEC/kg of clay. Often < 16 cmol/kg at pH 7.0 or < 12 cmol/kg by ECEC in their kandic, argillic or Oxic horizons (Esu, 2010; SSS, 2014a). Cation exchange capacity values are useful in evaluating the capacity of the soils to retain cations, their degree of weathering and general chemical reactivity (Buol *et al*., 1997). The cation exchange capacity determined by NH_4OAc at pH 7.0 to clay ratio is sometimes used to assess the probable contribution of clay to soil exchange capacity and soil solution chemistry (Esu, 2010). Such data serve as an internal check of clay mineralogy and is used to estimate clay mineralogy when mineralogy data is lacking (Rhoades, 1982).

Percent base saturation is an important parameter used in separating Ultisols from Alfisols; base data for this procedure is provided by the CEC. Furthermore, the natric diagnostic subsurface horizon is identified among other criteria by an ESP value of >= 15 % in the argillic horizon.

The CEC values of the soils developed on limestone in the Harran region of Turkey ranged from 14.81 cmol/kg to 56.66 cmol/kg (Irmak *et al*., 2007); such high values were attributed to high clay content in the soils. Effective CEC was generally low in the surface and subsurface soils of the Odukpani limestone soils and had values ranging from 8.91 to 14.49 cmol/kg (Afu *et al*., 2017) while the limestone soils of the Bohemian karst region had values of CEC ranging from 121 to 493 cmol/kg in the studied soils (Samonil, 2007) and ranges of 40-47.39 cmol/kg and 23.77-55.72 cmol/kg were obtained in the surface and subsurface soils, respectively in the limestone of Harran region in Turkey (Irmak *et al*., 2007). Values of CEC in the Harran region were quite higher than those of other karst regions.

The CEC in the surface soils developed from the Apodi plateau karst environment ranged from 5.8 to 11.7 cmol/kg and from 3.9 to 18.6 cmol/kg in the subsurface soils (Girao *et al*., 2014). Furthermore, its values in soils overlying the Neogene clay lime deposits of northwest Bursa province, Turkey ranged from 20.86 to 40.64 cmol/kg and were rated high because of high clay amounts and type (Aksoy *et al*., 2009). In the highly calcareous soils of Southern Iran, CEC values ranged from 19.3 to 29.9 cmol/kg in the surface soils and from 18.0 to 32.5 cmol/kg in the subsurface soils (Emadi *et al*., 2008); they attributed the increasing CEC with soil depth to the transformation of illite to smectite, which of course has higher surface area and adsorption capacity. Furthermore, soils in the Apodi plateau, Brazil had values ranging from 11.5 to 30.0 cmol/kg in the surface soils and from 8.2 to 29.2 cmol/kg in the subsurface soils (Ferreira *et al*., 2016).

2.3.7 Exchangeable Acidity

Exchangeable acidity is the total titrable acidity of soils and represents acidity due to exchangeable hydronium or aluminum ions (Udo *et al*., 2009). As Al and Fe are released during acidification or weathering processes, they become more accessible on cation exchange sites, in solution or simply on exposed surfaces. Both ions react readily with phosphates, forming relatively insoluble compounds through a process known as phosphate fixation. This process is a problem in tropical soils that are high in Fe because their ‘appetite’ for phosphates can sometimes become almost insatiable (Harter, 2007). This identifies Al toxicity as a major limiting factor in plant growth in acid, highly weathered tropical soils. According to the percentage abundance, the exchangeable acidic cations (H^+, Al^{3+}) followed Ca and Mg cations and their presence on the adsorption complex in the massive limestone deposit in Macedonia was affected by climate, vegetation, altitude, exposition and soil development (Markoski *et al*., 2018).

In the soils formed on massive limestone in Pivska Planina, Serbia, exchangeable H^+ ranged from 16.11 to 39.47 cmol/kg (Petar *et al*., 2016). Soil exchangeable acidity was very low in the soils and had soil surface values ranging from negligible values to 1.0 cmol/kg, while the subsurface values ranged from negligible values to 2.2 cmol/kg in the karst environment of the Apodi plateau (Girao *et al*., 2014). Out of four limestone soil profiles studied, Ferreira *et al*. (2016) reported that three of them had exchangeable acidity values of zero in all horizons except the third with values that ranged from 0.7 to 1.0 cmol/kg. Values of exchangeable Al ranged from 0.16 to 0.48 cmol/kg in the surface and, 0.32 and 0.84 cmol/kg in the subsurface of the soils developed from limestone in Odukpani area (Afu *et al*., 2017). Exchangeable acidity (Al^{3+} and H^+) in the Bohemian karst region ranged from 0.60 to 94.50 cmol/kg (Samonil, 2007); such values were described as high and rather surprising especially in a limestone region. Exchangeable H^+ ranged from 0.2 to 0.92 cmol/kg in the surface soils and from 0.28 to 0.92 cmol/kg in the subsurface soils of the Odukpani limestone area (Afu *et al*., 2017). In a massive limestone deposit in Macedonia, Markoski *et al*. (2018) obtained mean values of 14.51 and 15.46 cmol/kg (Al^{3+} + H^+) in the surface and subsurface soils, respectively.

2.3.8 Base Saturation

In the limestone soils of Odukpani in Cross River State, Afu *et al*. (2017) reported base saturation values ranging from 89 to 92 % in the surface soils and from 86 to 93 % in the subsurface soils. These high values were attributed to high clay content and the presence of nutrient reserve in the soil exchange complex. Base saturation in the Bohemian karst region ranged from 21.70 to 100 % (Samonil, 2007), while values in the karst environment of the Apodi plateau in Brazil were quite high and had values ranging from 87 to 100 % in the surface soils and from 64 to 100 % in the subsurface soils (Girao *et al*., 2014). Also, values were very high and approached 100 % in the Neogene clay lime soils of Northwest Bursa, Turkey (Aksoy *et al*., 2009). Report from studies by Ferreira *et al*. (2016) in the Apodi plateau, Brazil indicates values of base saturation ranging from 94 to 100 % in the surface soils and from 92 to 100 % in the subsurface soils.

2.3.9 Calcium Carbonate Equivalent (CCE)

Calcium carbonate equivalent refers to the amount of $CaCO_3$ in soil (SSDS, 2017). It determines the carbonate content in calcareous soils and simulates a situation where $CaCO_3$ is the only carbonate in the soil. Calcium accumulation in the soils and subsequent formation of calcic and petrocalcic endopedons begin with the translocation of bicarbonates in the soils

solution and their precipitation depends on climate and soil relief (Bachman and Machette, 1977).

Soils overlying Neogene clay lime in Northwest Bursa, Turkey were variable in $CaCO_3$ content with values ranging from 0.2 to 37.0 % (Aksoy *et al*., 2009). Calcium carbonate equivalent in the highly calcareous soils of Southern Iran ranged from 21.4 to 57.1 % in the surface soils and 30.6-91.2 % in the subsurface soils (Emadi *et al*., 2008). According to them, the increase in $CaCO_3$ with soil depth is an indication that $CaCO_3$ is lost from the epipedon during wet periods and opined that $CaCO_3$ accumulates deeper in the soil with reduced physiographic condition. Calcium carbonate equivalent in the soils containing carbonate on the Apodi plateau, Brazil had a range of 9.1-849.1 g/kg in the surface and subsurface soils with clear accumulation in the B and C horizons (Ferreira *et al*., 2016). They further identified the precipitation of secondary carbonates as the most remarkable morphological feature in the carbonate soils especially upon the accumulation of $CaCO_3$ in the subsurface soils.

2.4 Forms of Fe and Al Oxides and Ratios

Free oxides and hydroxides of soils include crystalline compounds of silicon, aluminum, iron, titanium and manganese. The term free refers to compounds having mainly a single species of coordinating cation such as oxides of silicon or aluminum; while combined oxides contain two or more species of coordinating cations; aluminosilicates and ferrialuminosilicates are examples (Jackson *et al*., 1986). The oxides of Fe and Al are collectively called sesquioxides; their content and distribution influence pedogenetic processes as well as physical and chemical soil properties (Schwertmann and Taylor, 1989; Jelic *et al*., 2011). It has also been used to predict the direction, degree and stage of pedogenesis (Durn *et al*., 2001; Igwe, 2001; Osedeke *et al*., 2005) as well as the type of soil horizon and soil drainage (Ibia, 2002; and Essoka and Esu, 2003). However, methods of extraction do not give absolute contents of elements as results obtained for a given method are required for soil comparison (Niskanen, 1989). Most of the forms of Fe subjected to translocation emanate from the weathering of biotite and ferromagnesian minerals (Tan, 1998).

The percentage of free iron increases with increasing weathering and soil age, and decreases with increasing poor drainage (Udo *et al*., 2009, and Dolui and Bera, 2001) and reprecipitates as oxides and hydroxides of Fe and Al (Seal *et al*., 2006). The profile distribution of oxides of extractable Fe and Al indicate the stage and degree of soil development (Mahaney and Fahey, 1988). These oxides contribute to greater soil aggregate

stability and as stated earlier, act as an index of soil genesis. Dithionite – citrate extractable iron and aluminum are used at the family level of soil classification and in the recognition of spodic endopedons (SSS, 2014b).

Dithionite extractable Fe quantifies total amount of Fe oxides (Cornell and Schwertmann, 2003). On the other hand, aqua regia extraction is the most suitable for extracting heavy metals (Lima *et al*., 2016). It is adequate for analyzing total recoverable heavy metals in soils. Elements that are not released by this digestion procedure are unimportant for estimating the behaviour of such elements (Niskavaara *et al*., 1997).

The red and brown bauxite soils of Jamaica overlying pure white limestone were very high in Fe with almost all of it being present in dithionite form (Ahmad *et al*., 1966). According to Seal *et al*. (2006), oxalate Fe and Al are more in illuvial horizons of soils and indicate translocation of Fe and Al, and were attributed to the translocation of Al-fulvate complex, proto-imogolite and hydroxyl polymer of Al in the horizons (Farmer *et al*., 1980). According to Dolui and Bera (2001), and Dolui and Chakraborty (1998), the difference between oxalate and pyrophosphate extractable Fe and Al is the amorphous inorganic Fe and Al. Furthermore, Dolui *et al*. (1988), and Doliu and Chakraborty (1998), stated that the difference between values obtained by oxalate and dithionite methods represent the amount of Fe and Al present in definite crystalline form.

Co-migration of Fe and Al with clay is evident by the ratio of clay/Fe_d and clay/Al_d. Within a soil profile where horizon formation is well expressed and established, the ratio increases with depth (Doliu and Bera, 2001). The formation of crystalline Fe (Fe_d-Fe_o) is often facilitated by high temperature and prolonged dry season (4-5 months); also, the ratio Fe_o/Fe_d and Al_o/Al_d in the solum is low when less than one is obtained (Seal *et al*., 2006). Low values further indicate that free iron and aluminum oxides in the soil are at an advanced stage of crystallinity or aging (Mahaney *et al*., 1991). With increasing soil age, the crystalline iron and Al oxides increase at the expense of poorly crystalline forms (Seal *et al*., 2006). According to Mahaney *et al*. (1991), soils with high ratios were younger soils; whereas low ratios indicate older soils. The ratio Fe_o/Fe_d was obtained as 0.27-0.42 while Al_o/Al_d was 0.26-0.42 and both described as low and suggest that weathering has progressed to an advanced stage or soils are older (Seal *et al*., 2006).

Soils developed from limestone in the Harran region of Turkey had ranges of 0.76-1.84 % in the surface soils for extractable Fe_2O_3 and 0.48-1.9 % in the subsurface soils, while total Fe_2O_3 had ranges of 4.14-4.78 % in the surface soil and 2.06-5.46 % in the subsurface soils (Irmak *et al*., 2007). On the other hand, extractable Al_2O_3 in the soils developed from

limestone in the Harran region of Turkey had ranges of 0.12-0.21 % in the surface soils and 0.07-0.27 % in the subsurface soils, while total Al_2O_3 had ranges of 6.24-7.11 % and 2.08-8.15 % in the surface and subsurface soils, respectively (Irmak *et al*., 2007).

Dithionite iron (Fe_d) ranged from 20.27 to 38.88 g/kg in the surface soils of the Apodi plateau karst environment, while subsurface soil values ranged from 2.57 to 43.91 g/kg. On the other hand, oxalate iron (Fe_o) ranged from 2.23 to 3.31 g/kg in the surface soils and from 0.39 to 4.35 g/kg in the subsurface soils of the same karst environment (Girao *et al*., 2014). The ratio of Fe_o to Fe_d indicates the degree of crystallinity of the oxides. The degree of crystallinity is higher when the ratio of oxalate to dithionite Fe is lower (Osayande *et al*., 2013). Values of Fe_o/Fe_d were generally low and ranged from 0.04 to 1.14 in the entire soils; however, relatively lower values were obtained in the B horizons (Girao *et al*., 2014), while Fe_d/clay ratio had a range of 0.03-0.10 with comparatively lower values in the B horizons (Girao *et al*., 2014).

The contents of crystalline and amorphous forms as well as their various ratios have been used to assess the degree of soil weathering and development (Udo *et al*., 2009). The crystalline forms indicate advanced stage of soil weathering while the non-crystalline or amorphous forms are mobile and associated with soil organic matter (Osayande *et al*., 2013; Obi *et al*., 2009).

2.5 Elemental Oxides of Limestone Soils Determined by X-ray Fluorescence (XRF)

X-ray fluorescence is the emission of secondary X-rays from a material that has been excited by bombarding with high energy X-rays (De Viguerie *et al*., 2009). The technique has been widely used in studies of several branches of science for identification and quantification of chemical elements in diverse materials and has advantages over other methods because of its speed and does not generate chemical residues (Stockmann *et al*., 2016; Weindorf *et al*., 2014).

Various techniques have been used in elemental oxide analysis in solid and liquid samples including X-ray Fluorescence (XRF) and Wavelength Dispersive X-Ray Fluorescence; however, XRF has advantages in the speed of analysis, ease of sample preparation, high stability and precision as well as wider range or levels of detection of elements in samples and does not generate chemical residues (TFS, 2012., Stockmann *et al*., 2016., Weindorf *et al*., 2014). Recently, Silva *et al*. (2018) used portable XRF for soil characterization; the equipment has also been found to contribute to the description of pedogenic processes (Weindorf *et al*. 2014 and Stockmann *et al*. 2016), while Weindorf *et al*. (2012) used the technique to differentiate between albic and spodic endopedons. According

to Silva *et al*. (2018), the technique is widely used for studies involving soil genesis as it aids the derivation of indices that reflect the degree of soil weathering and has been recently found to have potentials in soil mapping and surveys as it helps in separating soil classes (Silva *et al*., 2018 and Silva *et al*., 2016). Success has also been recorded in the use of XRF to investigate soil profile development, while Weindorf *et al*. (2015) determined lithological continuity of soils using XRF. There is no doubt saying that XRF is a formidable technique to reckon with in pedogenetic studies.

The following are elemental oxides as found in some soils developed from limestone:

a. MgO: In the Harran region of Turkey, total MgO contents of the soils on $CaCO_3$ rocks ranged from 0.211 to 0.263 % (Irmak *et al*., 2007), while soils on the Lenoir limestone soils found in Augusta County had a range of 0.44-1.04 % for MgO (Carroll and Hathaway, 1953).

b. CaO: In the Harran region of Turkey, total CaO in the soils on $CaCO_3$ rocks ranged from 2.047 to 14.99 % and was comparatively higher than values in soils developed from other parent materials (Irmak *et al*., 2007). The higher contents of CaO were traced to the chemical composition of the $CaCO_3$ rock on which the soils were developed. Similarly, the Lenoir limestone soils found in Augusta County had a range of 3.10-9.60 % for CaO (Carroll and Hathaway, 1953).

c. K_2O: In the limestone soils of Harran region, K_2O ranged from 0.269 to 0.855 % (Irmak *et al*., 2007), while the Lenoir limestone soils found in Augusta County; had a range of 0.91-3.7 % for K_2O (Carroll and Hathaway, 1953).

d. Al_2O_3: According to Ahmad *et al*. (1966), the red and brown bauxite soils of Jamaica occurring over pure white limestone were very high in Al_2O_3 and low in silica. In the Lenoir limestone soils found in Augusta County; 23.02-29.15 % was obtained for Al_2O_3 (Carroll and Hathaway, 1953). They further attributed the variation in the ratio of SiO_2:Al_2O_3 in the profile pits to podzolic weathering and explained that silica was present in excess of the requirement for the formation of kaolinite, hydrous mica and chlorite. The soils overlying the karst environment of Apodi plateau, Brazil had ranges of 137.6-179.9 g/kg and 28.2-260.7 g/kg for Al_2O_3 in the surface and subsurface soils, respectively (Girao *et al*., 2014).

e. SiO_2: In the Lenoir limestone soils found in Augusta County, SiO_2 ranged from 42.19 to 48.79 % (Carroll and Hathaway, 1953), while in the karst environment of the Apodi plateau, Girao *et al*. (2014) obtained a range of 155.4-187.4 g/kg in the surface soils and 7.0-323.3 g/kg in the subsurface soils for SiO_2. Extractable SiO_2 ranged from 0.043 to 0.083 % in

the surface soil and 0.058-0.087 % in the subsurface soils of the limestone area in Harran region of Turkey (Irmak *et al*., 2007); such low values were attributed to quartz weathering.

f. Fe_2O_3: In the karst environment of the Apodi plateau, Girao *et al*. (2014) obtained ranges of 67.9-87.8 g/kg and 10.1-99.6 g/kg in the surface and subsurface soils, respectively for Fe_2O_3.

g. MnO: Soils overlying the karst environment of Apodi plateau, Brazil had a range of 0.57-1.04 g/kg in the surface soils and 0.15-1.95 g/kg in the subsurface soils of the karst environment of the Apodi plateau for MnO (Girao *et al*. 2014).

h. TiO_2: The relevance of Ti in soil studies is related to soil genesis, it is less mobile in soils and vary in soils depending on soil parent material reflecting the occurrence of the metal in the parent rock (Curi and Franzmeier, 1987; Moreira and Oliveira, 2008). In the soils overlying the karst environment of Apodi plateau, Brazil values of TiO_2 ranged from 5.02 to 6.37 g/kg in the surface soils and from 1.33 to 7.26 g/kg in the subsurface soils of the karst area (Girao *et al*., 2014).

2.6 Mineralogical Properties of the Soils Developed on Limestone

Mineralogy data generated from fine sand fraction is often used in the Taxonomic classification of soils at the family level (SSS, 2014b) as well as in understanding the weathering potential and stage of weathering in a soil profile (Esu, 2010). He further opined that the presence of weatherable minerals in soil is an important index of the stage of weathering and soil development from which the fertility status of the soil can be assessed.

Mineralogical transformation in carbonaceous rocks begins with the dissolution of calcium carbonate by percolating rain water when enriched by carbon dioxide and plant exudates. In such a situation, the less soluble $CaCO_3$ in the parent rock reacts with carbonic acid and is transformed into easily soluble $Ca(HCO_3)_2$ and then evacuated by the drainage water. This process leaves behind a weathering product with a variable amount of free $CaCO_3$. The process also depends on the nature and hardness of the parent rock and on the amount and chemical aggressiveness of the water that percolates through the soil and rock (Verheye and de la Rosa, 2005). Carroll and Hathaway (1963) suggested the following order of events for clay mineral weathering in limestone soils:

Detrital volcanic minerals → montmorillonite → gibbsite → boehmite

Incorporated soil minerals → halloysite → gibbsite → boehmite

According to Cady *et al*. (1986), knowledge about the nature and conditions of the minerals in the fine sand fraction provides information on the source of parent materials and the degree of weathering as key to the history, genetic processes and fertility reserve of the

soils. The higher the degree of weathering, the lower the ratio of weatherable to resistant minerals. The cause of clay mineral formation and soil development is suggested based on the type of primary mineral present in the soil, and the higher the content of weatherable minerals in a soil, the higher the soil fertility reserve.

Limestone contains mainly fine grained calcite and some mica, especially glauconite as its prominent minerals; however, most sedimentary rocks vary in composition due to variation in the source or origin of their cementing agents (Esu, 2010). In the soils developed from limestone in the Harran region of Turkey, X-ray diffraction analysis showed that smectite dominated the clay mineralogy of the soils (Irmak *et al*., 2007) and further stated that the presence of smectite was in agreement with the CEC and swelling properties of the soils. According to them, the level to gently undulating landscape, semi-arid condition, high pH and saturation of the soils with water may have encouraged the accumulation of bases and therefore, smectification and the formation of other 2:1 clay minerals (Buol *et al*., 1980; Dinc *et al*., 1987; Yesilsoy, 1994). Though some researchers have claimed that excess Ca^{2+} content in arid soils would increase the formation of smectite but decrease the formation of kaolinite (Kapur, 1975; Yesilsoy, 1994); many more have identified smectite as the abundant clay mineral in most of the soils formed on parent materials rich in $CaCO_3$ (Singer, 1989; Yilmaz, 1990). In the karst environment of the Apodi plateau, Brazil, palygorskite was the second most abundant mineral after smectite (Irmak *et al.*, 2007). The presence of smectite and palygorskite in the soil can be traced to a common origin of parent material (Kapur, 1975; Singer, 1989; Dixon and Weed, 1989; Irmak *et al*., 1991). According to Bigham *et al*., (1980), the coexistence of the two minerals in soils shows that palygorskite may have been transformed to smectite during weathering; however, others opined that palygorskite has formed in the porous $CaCO_3$ grains and discharged into the soil by dissolution (Kapur, 1975; Yilmaz, 1990; Stolt *et al*., 1991; Sumner, 2000). Report by Irmak *et al*. (2007) indicates that low amounts of kaolinite and illite also occurred in the soils of the karst environment in Apodi plateau, Brazil.

Studies by Ahmad *et al*. (1966) on bauxite soils overlying pure white limestone shows the presence of Al_2O_3 as gibbsite and boehmite in the ratio 2:1. The soils were reported to have less than 10 % kaolinite with traces of chlorite and montmorillonite in deeper layers. The presence of rutile and anatase, though in small amounts indicates advanced weathering, this was also reported by Esu (2010). According to Ahmad *et al*. (1966), low contents of silicates in the bauxite is associated with low SiO_2 and possibly traced to the genesis of bauxite from underlying limestone. Furthermore, Petar *et al*. (2016) revealed that quartz,

muscovite, K-feldspar and hematite were the main minerals obtained in the sand fraction of limestone soils, however, the content of these minerals varied from one profile to another and was attributed to the dissolution of carbonates, which increased the content of insoluble minerals, hence the quartz and muscovite content or the deposition of materials by wind which led to the concentration of quartz and mica (Egli *et al*., 2008). Mineral content in the silt fraction varies in terms of prevalence (Petar *et al*., 2016) as illite dominated the fraction while muscovite increased with a resultant decrease in quartz and feldspar compared to the sand fraction of soils formed on massive limestone in Pivska Planina. Petar *et al*. (2016) however, observed a direct relationship between quartz content and coarse sand. In the soils formed from massive limestone in Pivska Planina, the clay fraction was dominated by kaolinite with other minerals like vermiculite, smectite, mixed layer silicates, chlorites and low content of illite (Petar *et al*., 2016). The prevalence of kaolinite over illite was described by a geochemical transformation of kaolinite into illite instead of sedimentation or deposition of materials from other sources (Moresi and Mongelli, 1988).

Soils overlying limestone and influenced by aeolian dust in Southwestern Japan were dominated by amphiboles and pyroxenes which decreased with soil depth (Miura *et al*., 1988) in the sand fraction. These groups of minerals are dark coloured ferromagnesian inosilicates. Other minerals include muscovite and Opaque which increased with soil depth while olivine and epidote were obtained in traces. Clay mineralogy of soils overlying limestone in Southwestern Japan was indicated by abundance of Al-vermiculite which decreased at the subsurface. The subsurface soil was ranked ‘many’ in terms of the relative abundance of kaolinite and illite/vermiculite; however, illite and vermiculite were ‘few’ in the surface soils and ‘common’ in the subsurface soils (Miura *et al*., 1988). Findings by Ross and Hendricks (1945) reveals kaolinite as an end product of clay mineral development in limestone soils, consequently, chert commonly acts as a skeleton holding up the weathering rock and the developing soil (Carroll and Hathaway, 1953). In the clay fraction of the Lenoir limestone soils in Augusta County, kaolinite was present in ‘medium’ amounts while quartz occurred in ‘small’ amounts in the entire profile depth. Chlorite was also present in ‘small’ amounts in the solum and possibly present in the C horizon; however, montmorillonite was absent in the soil while hydrous mica was either present in small amount or possibly present in the B or C horizon. The limestone soils show the presence of hydrous mica with some montmorillonite and kaolinite which were sparingly obtained in the soils (Carroll and Hathaway, 1953).

According to studies by Carroll and Hathaway (1953), the original clay minerals in limestone soils are hydrous mica and montmorillonite which are changed during soil profile development under leaching conditions to kaolinite and chlorite.

Smectite was 'dominant' in the Ap and Bw horizons of soils developed from limestone in the Harran region of Turkey and 'fair' in the Ck horizons, while palygorskite was 'much' and illite 'fair' in all horizons of the two pedons studied; however, kaolinite varied down the soil profiles between 'fair' and 'much' in abundance (Irmak *et al*., 2007). Based on crystallization of clay minerals, kaolinite and smectite were 'good' while palygorskite and illite were 'poor' (Irmak *et al*., 2007). According to Yesilsoy (1994), excess Ca^{2+} content in arid soils increase the formation of smectite but reduce the formation of kaolinite.

In Brazil, hematite, goethite, kaolinite and mica were common minerals in soils developed from calcareous materials (Girao *et al*., 2014) while Evans and Cameron (1979) found that illite weathered to vermiculite and aluminous vermiculite with advancement in soil age. The presence of mafic silicate minerals and alkaline feldspars may indicate that the source of sedimentary rocks was subjected to immature weathering which inhibited the formation of mature soil profile (Dhannoun *et al*., 2010). The highly calcareous soils of Southern Iran were dominated by illite and chlorite as well as palygorskite and smectite with abundances varying from 10 to 40 % while kaolinite in the B horizon ranged from 10 to 20 % (Emadi *et al*., 2008). Vermiculite was very low and had a range of 1-6 % in the calcareous soils of Southern Iran.

According to Khormali *et al*., 2003, in calcareous environments, high Mg and Si mobility, low activity of K^+ and Al^{3+} may encourage the transformation of illite to smectite in the soil surface. Very low vermiculite in limestone soils has also been attributed to high soil pH and low activity of Al in the soils (Emadi *et al*., 2008). According to the authors, illite and palygorskite were precursors for smectite formation in soil surface. Consequently, the origin of palygorskite has been traced to its inheritance from parent materials (Badraoui *et al*., 1992) and through pedogenesis (Monger and Daugherty, 1991). Furthermore, Khademi and Mermut (1999) reported the entrapment of eluviated palygorskite in the Bk horizons. Nevertheless, Emadi *et al*. (2008) traced the origin of palygorskite in limestone soils to inheritance, Neoformation and transformation from other minerals. Palygorskite, smectite, chlorite, illite, kaolinite and vermiculite were obtained as the dominant clay minerals in calcareous soils (Khormali *et al*., 2003 and Monger and Daugherty, 1991).

Smectite in Leptosols of the karst plains of Mexico were formed in shallow saline coastal swamps and directly from the weathering of bedrocks (Bautista *et al*., 2011). According to Isphording and Wilson (1974), smectite require moderate drainage and abundance of Si and Al ions to form. In the contrary, Deepthy and Balakrishnan (2005) observed that smectite is scarce in soils formed from karst in older platforms with good drainage while sub humid tropical climate reflects the influence of rainfall on the weathering of smectite. Kaolinite and aluminum (Boehmite), titanium (anatase) and iron oxides (hematite) were present in cambisols, Leptosols and Calcisols of karstic plains (Bautista *et al*., 2011) while quartz, hematite, anatase and kaolinite were more in soils of older platforms. Cabadas-Baez *et al*. (2010) reported vermiculite, kaolinite and quartz in the pedo-sediments of karstic sinkholes of Northeast Yucatan, Mexico; while Sedov *et al*. (2008) reported vermiculite in Phaeozems from tectonic karstic plains but was not recorded by Bautista *et al*. (2011) in the tropical karst soils of the Peninsula of Yucatan, Mexico. Traces of palygorskite were therefore obtained by Bautista *et al*. (2011) in the tropical karst of Yucatan, Mexico and abundantly by Krekeler *et al*. (2010), and Krekeler and Kearns (2009). Goethite and hematite were found in red karstic soils such as cambisols and Luvisols, similar to reports by Bautista *et al*. (2003 and 2004) in the karstic plains of Yucatan; while kaolinite was more common in soils with long exposure on karstic landscapes (Bautista *et al*., 2011). The presence of abundant anatase in the clay fraction of Luvisols is an evidence of advanced soil development (Allen and Hajek, 1995).

According to Bautista *et al*. (2011), it is assumed that part of the kaolinite in soils is inherited from limestone, while anatase in the clay fraction is authigenic and formed from sphene, ilmenite, pyroxene and biotite weathering (Allen and Hajek, 1995). Mota *et al*. (2007) found quartz, goethite, hematite and magnetite in the sand fraction of soils from Apodi plateau; while Moreira *et al*. (2000) obtained quartz, orthoclase, pyroxene, ilmenite and small amounts of kaolinite, hematite and goethite in an Inceptisol; however, Lemos *et al*. (1997) found quartz, feldspar and calcite in cambisols formed from limestone.

Intense peaks of quartz in all horizons characterized fine sand mineralogy. Carbonates from the parent material as well as accumulated secondary $CaCO_3$ were obtained in the Apodi plateau, while mineralogy of the coarse sand indicates small amounts of feldspars (Ferreira *et al*., 2016) as mica was commonly obtained in calcareous materials (Oliveira *et al*., 2004). Calcareous soils change in mineralogy with depth as observed by Shankar and Achyuthan (2007), who identified chlorite, quartz, hornblende, garnet, hematite, calcite and feldspars in the coarse sand fraction of limestone soils in India. The dry climate in Apodi

plateau supports secondary $CaCO_3$ accumulation in soils from the parent material; however, quartz in limestone depends on rock purity (Ferreira *et al*., 2016).

The presence of smectite in the clay fraction may be by inheritance from the parent material or from lithologic discontinuity (Ferreira *et al*. 2016); similarly, Correa *et al*. (2003) had traced its occurrence to dry climates and limited moisture which reduces desilication. On the other hand, kaolinite was identified as the main clay mineral in the carbonate rich soils of Apodi plateau in Brazil (Ferreira *et al*., 2016). High temperatures favoured hematite formation while low pH as well as abundant water and organic matter content aided the formation of goethite (Kampf and Curi, 2012); hence the dominance of these oxides in the tropical soils. Clay mineralogy in the tropical karst landscape of the Yucatan was dominated by 2:1 mineral and Al-hydroxide interlayered vermiculite (Sedov *et al*., 2008).

2.7 Soil Classification

Soil classification is the systematic arrangement of soils into groups based on their characteristics (Esu, 2010). It helps in knowledge organization about soils, provides basis for a common language among the scientific community and contributes in understanding the relationships among soil individuals. A Soil Scientist however, requires knowledge about the genesis of the soil to classify it (Soil Science Division Staff, 2017). With soil classification, the properties of soils are more easily remembered and predictions made about the behavior of soils. Soil classification begins with examination of soil profiles. During the examination, subsoil horizons are given greater emphasis than surface horizons which are often easily influenced by anthropogenic activities so that they hardly bear any relationship with genetic processes (Kang and Tribathi, 2000). Although, modal profiles of two adjacent soil series may be distinctly different, there is usually a gradual transition in the field, between one series and another (Peterson and Calvin, 1986).

Among the many systems of soil classification, the USDA System of Soil Taxonomy and the World Reference Base for Soil Resources classification system are most widely used (Kang and Tribathi, 2000; Esu, 2010).

According to the Canadian system of soil classification, soils developed on calcareous materials in Ontario were classified as Orthic Regosols, Eluviated Eutric Brunisol and Eluviated Dystric Brunisol (VandenBygaart and Protz, 1994). Limestone soils with the A-C horizons in the Bohemian karst region were classified as Rendzic Leptosols, Lithic Leptosols or Hyperskeletic Leptosols, while those with A-B-C horizons in the same region were classified as Eutric cambisols or Calcaric Cambisol; however, recent soils with the A-E-B-C horizons were classified as Vertic Luvisols or Luvic Cambisols (Samonil, 2007).

Soils developed from limestone in the Harran region of Turkey were classified as Xeric Haplocalcid and Lithic Xeric Haplocalcid at the subgroup category of the USDA and correlated with Haplic Calcisol in the WRB for soil resources (Irmak *et al*., 2007). Silva *et al*. (2013) obtained red Argisols and Haplic Cambisols in the karst region of Mato Grosso do sul; while Lynch (2009) obtained Chernosols, Nitosols and Latosols in the soils of the karst region of Planaltina region of the state of Goias. Six pedons overlying the Neogene clay lime deposits of Northwest part of Bursa province, Turkey were classified by the USDA system and correlated with WRB for soil resources system as follows: Typic Xerorthents (Eutric Leptosol), Pachic Calcixerolls (Haplic Calcisol), Vertic Xerochrepts (Haplic Calcisol), Typic Xerochrepts (Calcaric Cambisol), Chromic Haploxererts (Eutric Vertisol) and Typic Calcixererts (Calcic Vertisol) (Aksoy *et al*., 2009). Highly calcareous soils in the semi-arid regions of southern Iran were classified at the family level as Fine, Mixed, Active, Thermic Calcic Haploxeralf and Fine, Mixed, Superactive, Typic Calcixerepts in the low lands; while those in the Piedmont were Fine, Silty, Mixed, Superactive, Thermic, Typic Calcixerepts and Fine Silty, Mixed, Active, Thermic Petrocalcic Calcixerepts (Emadi *et al*., 2008).

The accumulation of calcium carbonates in endopedons as calcic or petrocalcic horizons is an important morphogenic marker for soil classification (Wilding *et al*., 1990). Lithologic discontinuities are common on all karstic plains; Bautista *et al*. (2011) identified lithic contacts in two Leptosols and Cambisols obtained from the dissolution of red limestone. According to Shinzato (1998), soils developed from limestone are classified in a system equivalent to Mollisols, Vertisols, Vertic or Carbonatic Inceptisols, Alfisols and Ultisols, while Bautista *et al*. (2011) found Luvisols on toeslopes in the undulating karstic plains and hills of Peninsula of Yukatan, Mexico. Soils in karstic regions have carbonate or hypocarbonate properties and calcic or petrocalcic diagnostic horizons (Santos *et al*., 2013). Accumulation of carbonates in such horizons influences chemical characteristics such as pH, soil buffering, cation exchange capacity and plant nutrient availability (Khresat, 2001).

Based on the criteria of the Brazilian system of classification (Santos *et al*. 2013), the carbonate rich soils of Apodi plateau were classified as Cambissolo Haplico Carbonatico Vertissolico, Luvissolo Cromico Palico Petroplintico and Neossolo Litolico Carbonatico Tipico (Ferreira *et al*., 2016). Apart from young Entisols and Mollisols formed on limestone, such soils at advanced stages of development like Alfisols and Ultisols have been documented (Bruce, 1983), while the deeply weathered ferrallitic profiles are known to have been formed on limestone (Ortega, 1984) and Oxisols associated with the karstic bauxites in Jamaica (Scholten and Andriesse, 1986). Sedov *et al*. (2008) classified the soils on a tropical

karst landscape in the Yucatan as chromic cambisol, Leptic Gleyic Phaeozem, Leptic calcisol and Leptic Calcaric Phaeozem; in general, the upland soils were associated with the concept of Rendzina which translates to shallow humus-rich soils on calcareous parent materials.

2.8 Concept of Soil Genesis, Development and Forming Processes in Limestone Areas

2.8.1 Soil Genesis and Development

Pedogenesis is the process of soil formation as regulated by the effects of place, environment and history (Buol *et al*. 1973). Bedrock undergoes various changes as it is weathered beginning with the gradual but consistent removal of readily weathered minerals; hence, evaluating which minerals are present may indicate the degree of weathering that the rock or soil has undergone (SSDS, 2017; Coleman and Dethier, 1986). The point where rock weathering ends and soil development begins is not clear; it may be consecutive or overlapping (SSDS, 2017), however, weathering results in the removal of mineral constituents and accumulation of fabric without significant loss in volume (Pavitch, 1986).

The starting point of soil development is weathering of freshly accumulated parent materials as conditioned by other factors of soil formation (Jenny, 1994). The weathering process brings about the release of nutrients which are used as sources of energy by soil microbes. These microbes produce acids (oxalate) which contribute to weathering and also leave behind organic residues. In this manner, soils increase in depth with time and support higher organisms starting with pioneer species (Fungi) and proceeds to more complex organisms. These complex organisms encourage the accumulation of organic matter and the top soil deepens through soil mixing. As leaching and organic matter accumulation take place, soil layers develop which marks the beginning of soil profile formation. A rate of Pedogenesis of 1/10mm per year is in order of magnitude of the global soil production estimates (Scalenghe *et al*. 2016 and Wilkinson and Humpreys, 2005).

In 1883, the Russian geologist and father of pedology, V. V. Dokuchaev stated that soil formation occurs over time under the influence of climate, topography, vegetation and parent material. He further demonstrated this in 1898 (Jenny, 1980) using the primary soil forming equation: S = f(cl, o, p)tr; where, cl = climate, o = organisms, p = parent material, tr = relative time. Later in 1941, the American Soil Scientist, Hans Jenny, considered time as a factor of soil formation and left the ellipsis open for factors to be added as our understanding of soils become more refined:

Soil = f(cl, o, r, p, t …).

Esu (2010) observed soil formation to advance through two phases. The changes from a consolidated mass not capable of supporting plant growth to the development of unconsolidated layers of materials that can support plants if climate is favourable and moisture is available. Second, the changes occurring within the loose material as time progresses. The later phase leads to soil horizonation and development as soil forming processes act on the loose material. Local soil differences may occur in areas due to variation in one or more factors of soil formation, however, the soils in some areas have some properties in common as a result of the influences imposed by factors of soil formation. For instance, low base status may be traced to most humid regions or areas with underlying acid rocks; this contrasts soils in arid regions or calcareous soils which are typically high in bases (SSSD, 2017). Consequently, the less developed a soil is the greater will be the effect of parent material on the soil properties (Esu, 2010). Earlier, Jenny (1941) had studied changes in soil properties with time, especially as it concerned toposequences, biosequences, chronosequences, lithosequences and climosequences with their functions.

Though in predictive capacity, the degree of soil development was assessed quantitatively by Bockheirn (1980) by the use of chronofunctions. Evans and Cameron (1979) observed through pedological approach to soil development that, depth of solum and thickness of the epipedon increased with soil age. In their study, Evans and Cameron (1979) observed that illite weathered to vermiculite and aluminous vermiculite as time proceeded. Singleton and Lavkulich (1987) observed that horizons progressively deepened and became more differentiated with soil age while soil surface organic matter increased. In their studies on the rates of soil development on Podzols, Protz *et al*. (1984) observed age related trends as Ah horizon developed within 750 years and B horizons required between 1893 and 2300 years to form. In a related study by Protz *et al.* (1988) in Ontario, it was found that the depth of leaching was proportional to soil age while organic matter increased at about 2000 years of soil development. Years later, VandenBygaart and Protz (1994) observed that the most obvious evidence of soil development with age is the increasing thickness of the solum and the progressive weathering of carbonate minerals with depth in the limestone soils of Ontario. They further opined that the darkening of the chroma in the B horizons of the limestone soils of Pinery Provincial Park, Ontario may be due to the pigmentation properties of Fe-oxyhydroxides which may have been co-migrated with organic matter.

The Fe oxides and organic matter translocated from the eluvial horizon, and *in situ* weathering of ferromagnesian minerals, form dark coatings on mineral grains (as cutans). VandenBygaart and Protz (1994) further stated that the lowering of soil pH with age was due

to the weathering of carbonate minerals and leaching thereafter of Ca^{2+} and Mg^{2+} being replaced by H^{+}. However, according to Broak (2008), time available for soil development as well as the geological condition of the area are important factors in karst development.

The origin of limestone soils has given rise to diverse views. Ahmad and Jones (1969) considered the residual origin of soils on the Pleistocene limestone of Barbados, while Borg and Banner (1996) emphasized non-regolith aeolian components in their parent material. The absence of deeply weathered ferrallitic soils in the tropical karst landscape in Yucatan may be due to high intensity of karstification process (Sedov *et al*., 2008).

The development of soil formed on limestone is usually seen in the continual dissolution of the parent rock and in the surface accumulation of soil consisting of transformed clay (Saly, 1978). The implication of this concept is that, soil cannot develop on absolutely pure limestone and in the absence of *ex situ* materials (Samonil, 2007). Nevertheless, the accumulation of organic matter through leaf fall from vegetation and its subsequent conversion would be sufficient for initial soil development (Nemecek *et al*., 2001). Nemecek *et al*. (1990) informed that soils developed *in situ* from limestone are very shallow and aligns with studies made by Samonil (2007) in the Bohemian karst. Consequently, the proportion of $CaCO_3$ and non-carbonate accessories in limestone determine the rate of Pedogenesis as well as the physical and chemical makeup of the developing soils (Samonil, 2007). Kuzvart *et al*. (1992) observed that the amount of non-carbonate accessories is variable and may differ within a single limestone region. Also, the uniqueness of soils on carbonate rock deepens on the amount and nature of material *ex situ* or on the presence of aeolian material.

Sedov *et al*. (2008) described the fertility of thin calcareous soils of Yucatan, which is, the Leptic Phaeozems and Rendzic Leptosols as extraordinary and stated that it is quite surprising to have such thin but fertile soils as they lack the disadvantages associated with the thick but extremely leached humid tropical soils.

2.8.2 Processes of Soil Formation in Limestone Areas

Soil forming processes are generalized processes which cause horizon differentiation in most soils (Simonson, 1959; Esu, 2010). They are often put into four categories including; addition, losses, transformation and translocation processes (Esu, 2010; Asadu *et al*., 2012). Many a time, Soil Scientists during soil characterization studies tend to emphasize the properties of soil than the processes of soil formation. This is probably due to the assumption that soil properties result from soil forming processes and are more readily quantifiable than soil processes (Arnold, 1983). Soil forming processes are internal soil building processes

which act on weathered soil parent materials giving rise to soil which later results in different pedogenic horizons. Usually, they do not occur singly but as a series of complex processes. In most cases, more than two processes may occur simultaneously on the soils overlying the same parent material within the same period. Pidwirny (2006) had identified key soil forming processes particularly important to macro scale patterns of soil formation to include lateritization, podsolization, calcification, salinization and gleization.

The dominant soil forming processes in the tropics have been outlined by Esu (2010) and Ofem *et al*. (2015) as ferralization/lateritization, salinization and desalinization, allitization and deallitization, alkalization and dealkalization, leaching and enrichment, gleization, ferrolysis, pedoturbation, eluviation and illuviation, mineralization and surficial erosion. Other processes that may occur are calcification and decalcification, solonization and solodization, lessivage and leucinization.

While studying the soils formed from massive limestone in Pivska Planina, Serbia, Petar *et al*. (2016) identified two stages of soil formation in the area; calcomelanosol and calcocambisol, and opined that dissolution and leaching of alkali earth carbonates, slow mineralization and accumulation of organic matter and denudation processes occur in the soils. Pedogenesis varies with parent materials; for instance, pedogenesis on carbonate substrate differs significantly from silicate rocks (Sedov *et al*., 2008). Soil forming processes observed in the limestone soils of Bohemian karst were; organic matter accumulation and conversion, decalcification, calcification, illimerization, braunification, rubefication and gleization (Samonil, 2005). The development of calcic endopedons and the eluviation and illuviation of clay were identified as the most important pedogenic processes in the highly calcareous soils of Southern Iran (Emadi *et al*., 2008).

The dissolution of limestone has been considered the main process that produces lime free residues (Scholten and Andriesse, 1986), the contribution of allochtonous materials through aeolian deposits in the Mediterranean region (Durn, 2003; Muhs *et al*., 2010) and Mexico (Cabadas-Baez *et al*., 2010) as well as through residual dissolution of red limestone formed by debris mud, ash or aeolian dust in parts of Australia (Merino and Banerjee, 2008). However, soil age and rainfall strongly influence the process of limestone dissolution and depletion of soil carbonates in the tropical karst areas of Mexico (Bautista *et al.*, 2011) as well as in the karst region of Yucatan (SGM, 2007). The dissolution and precipitation of calcite is affected by soil pH, temperature and evapotranspiration, as well as biological activity which may drive the reaction in the opposite direction (Kampf and Curi, 2012). Consequently, Chadwick *et al*. (1995) opined that, pedogenic processes which occur under

extreme climatic conditions lead to the formation of soils with pedogenetic features. According to reports by Ferreira *et al.* (2016), pedogenesis in the Apodi plateau limestone soils was affected by climate, surface relief and stratification of weathered parent material leading to varied degrees of soil development. In their study, argilli-pedoturbation indicated by the Vertic character, large prismatic structures, extremely hard consistence, eluviation-illuviation that resulted in clay accumulation as well as calcification indicated by the accumulation of $CaCO_3$ and formation of plinthic endopedons were dominant processes of soil formation in the Apodi plateau.

Soil genesis on calcareous parent materials differ from that of soils developed on silicate minerals. Soils formed from carbonate rich materials in the Mediterranean and temperate regions (Terra Rossa, Terra Fusca and Rendzinas) have been well studied; there are however limited studies of such soils in the tropics (Sedov *et al*., 2008). The authors further described bioturbation in the area as being strong. Elevated land surface in the Yucatan area provided a leaching environment while the lower wetlands provided a suitable condition for the accumulation of materials, consequently, the poor internal drainage in the wetlands did not encourage *in situ* mineral alteration (Sedov *et al*., 2008). The mass of $CaCO_3$ in the Pinery dune soils overlying limestone in Ontario decreased linearly with soil age as weatherable carbonate minerals are progressively weathered. The depth of carbonate leaching also showed a positive linear relationship with soil age (VandenBygaart and Protz, 1994), and suggested that precipitation may be a main influence on the rate of carbonate weathering and thus, the depth of leaching of carbonate rich soils of Ontario.

2.8.3 Chemical Weathering and Indices of Measurement

Chemical weathering indices also called indices of alteration (Price and Velbel, 2003). They are put to use by plotting a particular index against soil depth which shows the weathering profile of the parent rock. Such changes with soil depth are gradual, steady and systematic for homogeneous parent rocks (Sutton and Maynard, 1992); and reflects the leaching of elements as weathering progresses. Weathering indices have been used as a tool to evaluate soil fertility and development (Delvaux *et al*., 1989) and also used to stress the impact of climate on rock weathering (Neall, 1977), and provides an understanding of mobile elements during weathering.

The isovolumetric technique is based on the assumption that during sub aerial weathering, addition or removal of elements to the profile does not affect bulk density; thus bulk density and not depth of sample serves as a measure of the extent of weathering (Grant, 1964). The relegation of sample depth as a measure of the extent of weathering, is because

deeper samples may be highly weathered. However, weathering indices for weathering profile samples have conventionally been calculated using the molecular proportions of major element oxides; this is done through the percent of the oxide based on weight. This method is based on the assumption that aluminum is immobile (Fedo *et al*., 1995), while weathering indices from volumetric concentrations plotted against bulk density allows for the mobility of all elements.

Indices that include Fe in their formulae are unsuitable for use in weathered regolith. The mobility of Fe depends on its oxidation state; whether 'Fe^{3+}' which is immobile or 'Fe^{2+}' which is highly mobile. Oxidation during weathering increases Fe^{3+} and reduces Fe^{2+}; this variance has not been captured in indices that include Fe in its formulation and may skew a weathering index to reflect a less advanced stage of weathering. Also, the abundance of Fe is influenced by redox reactions which are erratic in a weathering profile (Harnois, 1988). According to Price and Velbel (2003), a good chemical index of weathering should permit comparison between studies in different localities, on different parent materials and on profiles of different ages. According to them, a useful chemical weathering index should:

i. be easy to use and should involve elements that are easy to obtain in soil analysis (Harnois, 1988),

ii. incorporate elements with a range of mobility in the weathering environment,

iii. have chemically appropriate trends with increased weathering and vary greatly with increasing weathering,

iv. be applicable in a wide range of profiles over varying rock types,

iv. not assume that any element is immobile as this may not be the situation in nature.

Studies that are related to indices of alteration are more interested in the changes of the index with depth in a profile and quantify the chemical changes as weathering proceeds through the indices. This is on the assumption that the soil, represented by its horizons had the same initial composition (Deepest soil is the freshest) and deeper soils are less weathered than surface samples (Price and Velbel, 2003). The findings of Grant (1964) who emphasized bulk density as a measure of the extent of weathering are however contradicted in the former study.

As a function of their formula, some indices may increase in value as weathering progresses while others decrease with increased weathering. Ideally, the predicted trend should be consistently observed regardless of the bedrock on which the weathering profile is developed. As the intensity of chemical weathering increases, the content of highly mobile elements relatively decreases while the immobile elements (Al, Ti) increase.

Chemical weathering indices are used for characterizing weathering profiles; a term used in geochemical studies that refers to soil profiles in pedology, however, studies on mineral composition assesses the intensity of weathering (Bautista *et al*., 2011).

Weathering alters the volume of soil compared to fresh rocks, it is imperative therefore to normalize the element content in the lithological groups to an immobile element before comparing variation of element content with chemical index of alteration values (Dhannoun *et al*., 2010). Al_2O_3 and TiO_2 are often used as normalizing elements as they are considered immobile. The following are examples of weathering indices;

i. Ruxton Ratio (R)

Ruxton index is also referred to as silica/alumina ratio (Che *et al*., 2012). It relates silica loss to total element loss and considers alumina (Fiantis *et al*., 2010) and other sesquioxides to be immobile during weathering (Ruxton, 1968; Chittleborough, 1991). Ruxton index was proposed by Ruxton in 1968 and described as Ruxton ratio in 1991 by Chittleborough. The ratio is most suitable for soil profiles with uniform acid and intermediate rocks. Ruxton's results were based on the conclusion that R correlates with total element loss. The index does not, however, have definite weathering trends when plotted against bulk density and soil depth (Price and Velbel, 2001). It is represented by:

$$R = \frac{SiO_2}{Al_2O_3}$$

ii. Weathering Index (WI)

Weathering index is directly related to the age of soil and provides support knowledge for soil profile development from parent materials (Evans and Cameron, 1979). The formula below is used in calculations:

$$\frac{CaO+MgO+K_2O+Na_2O}{CaO+MgO+K_2O+Na_2O+Si_2O_3+Al_2O_3+Fe_2O_3} \times \frac{100}{1}$$

Variation in weathering intensity depends on the nature and composition of parent rock as well as its source if transported (Walia and Rao, 1999). Weathering index increases with soil depth and indicates a fall in weathering intensity with increasing soil depth (Evans and Cameron, 1979).

iii. Molar Ratio (MR)

Evaluating soil genesis using elemental analysis of oxides includes the use of the ratio SiO_2:R_2O_3 (R is a trivalent cation; Al or Fe), to represent soil weathering index (Jackson, 1973). The MR evaluates the extent of chemical changes in soils (Bera *et al*., 2015). Molar ratio of SiO_2:R_2O_3 reflects the distribution of SiO_2 and sesquioxides in the soil profile, and

reveals the relative loss or accumulation of aluminum and Fe oxides (Kyrylchuk and Haskevych, 2018).

Comparative movement of SiO_2 and R_2O_3 and their calculated ratios from total composition of soils provide a basis for understanding profile development from parent materials (Bera *et al*., 2015). The ratio is calculated on the basis that the resistant oxides of Fe and Al are immobile in the weathering environment (Colman, 1982). Generally, weathering advances when molar ratio decreases and more mobile oxides are removed from the soil weathering environment (Birkeland, 1984). The lower the MR, the more advanced weathering has progressed over time (Burt *et al*., 2003).

2.9 Soil Survey and Land Use Planning, Remote Sensing versus GIS in Soil Survey Studies

2.9.1 Soil Survey and Land Use Planning

Soil survey is the systematic examination, description, classification and mapping of soils in an area (Esu, 2010; Odoemena and Uchua, 2014). It characterizes, classifies and plots the soils in an area on a map, and also stores information about the soils in an organized data base and makes predictions based on its suitability and limitations to diverse uses (SSDS, 2017). Furthermore, soil survey provides a more systematic and accurate inventory of the kinds and distribution of soils in an area for the prediction of its potential for agriculture, forestry, wildlife, recreation, water shed protection etc. and is seen as a prerequisite for developing rational land use plan for development (Manchanda *et al*., 2002). Soil survey also investigates the distribution of soils in an area, characterizes, delineates map units and ultimately evaluates the mapping units for specific types of land use as well as predicting the effects of the land use on the environment (Odoemena and Uchua, 2014; Deckers *et al*. 2001; SSDS, 2017).

Besides the generation of soil data, information on climate, water resources, vegetation, land forms, land use and the socioeconomic setting of the people is required for proper land evaluation, particularly for agricultural purposes (Deckers *et al*., 2001). Thus, similar environments found in different places may give rise to similar soils. This regularity of soils based on similarity in environment permits prediction of the location of many different kinds of soil (SSSD, 2017). This principle is fundamental and makes soil surveys practical (Hudson, 1992).

Decision making on land use is thus, aided during land use planning so that limited resources are put to the most beneficial use and conserving same for the future.

Soil survey started in the United States in the late 1890s (Simson, 1990) and then in Europe in the early 1900s and reached its full scale practice during the 1950s, 1960s and 1970s when soil surveys were carried out in Africa, Asia and Latin America for the development of projects (Deckers *et al*., 2001). It applies the principles of Soil Science and involves geomorphology, theories of soil formation, physical geography, analysis of vegetation and land use patterns (Odoemena and Uchua, 2014). Most of the soil surveys published before 1910 were strongly influenced by geologic concepts which regarded soils as the products of geologic formations defined by land forms and litho-compositions. Furthermore, early soil surveys were made to help farmers identify soils that responded to different crops and management of various soil types (SSDS, 2017). With the advancement in soil surveys, soil properties were later found not necessarily as products of geology only, but drainage, slope and topography also had chunks of influences on them. However, geology provides the most reliable and robust context for understanding natural earth systems, including soils and defines its composition and arrangements (SSDS, 2017).

The identification of land type, vegetation, land use, slope and relief are of major importance in soil survey. Soils are surveyed and mapped following the interpretation of remote sensing imagery (Mulder, 1987), field survey and cartography (Sehgal *et al*. 1989).

Land use is the manipulation of land resources in order to derive benefits while land use planning is the process of allocating uses to tracts of land and resources to those uses and puts the resources of the environment to the most beneficial use whilst conserving those resources for the future. Proper land use planning maximizes economic benefits and helps in conflict resolution (Odoemena and Uchua, 2014). The systematic assessment of land resources for the purpose of selecting and adopting land use options without degrading the resource as well as the adoption of measures that encourage such land uses is called land use planning (FAO, 1999). Sustainable land use planning may therefore put limited environmental resources to new kinds of productive use if the previous does not seem to yield desired results; especially when there are competing uses for the same tract of land. The fitness of soils for land use can however not be assessed in isolation from other aspects of the environment and hence, it is land which is often employed as the basis for suitability evaluation. It is however more appropriate to assess existing land use options using the integrated natural resources approach before introducing more beneficial forms of land uses (Chopra, 2012).

Land capability classification is used to evaluate the capability of land to support a range of land uses, on a long term sustainable basis and considers the physical nature of the

land including; geology, soils and slope as well as climate and erosion hazard which may influence the long term sustainable use of the land for agriculture (Odoemena and Uchua, 2014). It takes into account limitations due to salinity, stoniness, drainage and flooding but does not take into account the economics of agriculture and sociopolitical factors. In the case of permanent limitation or where it is technically not feasible for an individual farmer, the land is classified according to the nature of its present limitation.

Twelve limitations and hazards have been grouped under four categories; for instance, unspecified erosion (e) limitation could be; aeolian (a), water (h) or mass movement (m) while unspecified wetness limitation (w) could be due to; flooding (f) or drainage (d). Similarly, unspecified soil limitations (s) could be linked to either; coarse fragments (g), rockiness (r), electrical conductivity (k) or limiting layer (l) while unspecified limitation due to climate may be presented in form of; precipitation (p), temperature (t) and complex topography (x).

Land suitability classification is the actual or potential fitness of a given tract of land for a defined use and takes into account the economics and sociopolitical factors during land evaluation. Land capability classification grades land for broad scale agricultural uses, while land suitability classification is applied to clearly defined land uses (Idoga *et al*. 1995). Suitability classification may be done based on the present status of the tract of land or after minor improvements. Topographic characteristics, climatic conditions as well as the quality of soil in an area are the most important determinant parameters of land suitability evaluation (Al-Mashreki *et al*. 2010). According to Drohan *et al*. (2003), the purity of a mapping unit is less than 50 % and may lead to erroneous conclusions especially when such maps are used to produce suitability maps (Riezebos, 1989; Ziadat *et al.*, 2003); moreover, the accuracy of site specific suitability using highly detailed soil map is 60-70 % and is questionable when providing reliable information for land use planning (Ziadat, 2000).

The aim of land suitability evaluation is to find out parcels of land that best support commonly grown crops based on the physical and chemical properties of the soils in the area and make recommendations for improved yield. Variations are often recorded due to varying limitations in terms of slope, drainage, $CaCO_3$ content, texture and gravel which influence root penetration, water and nutrient retention (Al-Mashreki *et al*., 2010). When evaluating land from soil survey data, soil variability is always an important factor to look out for; especially as it concerns soil mapping units. Thus, land suitability is assigned to the mapping unit by calculating the average values of the soil parameters at each of the several observation points (Khalil *et al*. 1995).

2.9.2 Remote Sensing and GIS in Soil Survey Studies

Soil surveys using the conventional approach is subjective, time consuming and labour intensive while soil surveys with satellite imageries is faster, less laborious and more precise in delineating soil boundaries (Ojanuga, 1986., Manchanda *et al*., 2002 and Esu, 2010).

According to Bautista *et al* (2011) and Rees (1990), the use of satellite imageries to identify land forms is more accurate than soil maps in aerial photos; while the SSDS (2017) emphasized the relevance of integrating digital mapping techniques such as digital elevation models and geology, global positioning systems and multispectral bands to enhance the chances of locating map unit boundaries. Ultimately, soil maps are made to supply soil related information to its end users for proper management and decisions.

Remote sensing is the art and science of obtaining information about a phenomenon through the analysis of data acquired by a device that is not in physical contact with the phenomenon under investigation (Odoemena and Uchua, 2014). It is also described as a means of sensing and measuring the earth's resources with the aid of instrument(s) on a plat form above the earth surface (Odoemena, 2011). The use of imaging sensor technologies including instruments found in air and space crafts, as well as those found in electrophysiology are examples. The remote sensing system is however, an integration of disciplines such as optics, photography, computer, electronics, telecommunication, satellite lunching etc. It offers researchers the opportunity to investigate inaccessible areas and aids faster acquisition of data from a large area. Other advantages of the technology are; no security restriction, high level of automation, repetitive observations, and vast wealth of information, synoptic view of an area, reliability, flexibility and ease of measurements (Odoemena and Uchua, 2014). On the other hand, remote sensing is faced with limitations such as image distortion, scale, coverage and resolution.

Digital elevation model (DEM) is an electronic model of the earth's surface that can store and manipulate data in a computer (Brough, 1986). It is a relatively new tool especially in the third world countries introduced by advances in computer technology *via* developments in remote sensing and geographic information system which provides approaches to meet the demands of resource related modelling (Mermut and Eswaran, 2001; Salehi *et al*., 2003).

Integration of remote sensing technique such as DEM within a GIS database can decrease cost and time, and increase details during soil surveys (Green, 1992). Digital elevation model had been used by Bautista *et al*. (2011) in combination with Landsat and Spot satellite imageries as well as digital maps of vegetation from the National Forest

Inventory to study the spatial distribution and development of soils in tropical karst regions in Mexico. However, geology and chemical weathering indices as well as micromorphology were not fairly considered as recommended by Alexandrovskiy (2001). As important as they are in soil surveys, topography and vegetation are a unique combination of factors that are used to predict soil boundaries and properties within the bordered areas (SSDS, 2017). The present study therefore integrates these parameters for a better research outcome.

Geographic information system (GIS) is a computer-based system that is capable of integrating, storing, editing, analyzing, sharing and displaying geographically referenced information which may later help in decision making systems (Odoemena and Uchua, 2014). The technology of GIS can be used in diverse ways including urban planning, environmental impact assessment and in accessing soil information, soil mapping and map interpretation as well as in criminal investigations (Yekola *et al*. 2012). Information processed in the GIS software is often generated using a Global Positioning System (GPS), especially during soil surveys.

Global positioning system is the most modern means of surveying and mapping, and has immense contribution in remote sensing and aerial photography, and has found applications in GIS data collection (Odoemena and Uchua, 2014). The integration of GIS and remote sensing methods has mutual benefit since both technologies have similar applications by similar professionals, and require identical hard and software (Uboldi and Chuvieco, 1997). This has resulted in improved management and use. While remote sensing (raster information) collects spatial data, GIS (vector information) is a recognized powerful tool for turning such large volumes of spatial data into useful information and so, both systems attain optimal potentials when integrated (Ndukwe, 1997).

Digital elevation model is used during landform characterization to derive landscape attributes (Dobos *et al*. 2000). It is often manipulated during mapping to give a quantitative description of landforms and variability of soils. The manipulation gives rise to a network of drainages as well as maps of elevation, slope and aspect (Brabyn, 1997). These attributes, together with the images can be used to improve their capabilities for soil mapping (Lee *et al*., 1988); however, Moore *et al.* (1992) emphasized the integration of geologic information for the prediction of soil types by DEM. Similarly, Bayramin (2001) developed a pre-model for soil mapping by the integral use of DEM, satellite data and digital geological data to improve soil map quality. As an alternative and improved soil mapping technique, Aksoy *et al.* (2009) successfully used DEM and Landsat TM imagery to survey the soils in a hill terrain, in order to develop a land capability as well as irrigation suitability maps of

Northwest province, Turkey. Digital elevation model is therefore well established in solving soil survey related issues globally.

CHAPTER THREE
MATERIALS AND METHODS

3.1 Location of the Study

The research was located in Cross River State, Nigeria. Cross River State (CRS) is within latitudes 5º32′ and 4º27′N and longitudes 7º50′ and 9º28′E. The state is bordered by Benue State in the Northeast, Ebonyi and Abia States in the West, Akwa Ibom State in the Southwest, and the Atlantic Ocean and the Cameroons in the South and East, respectively (Fig. 1). Three locations, namely: Ishibori, Agoi Ibami and Mfamosing were selected. Each location uniquely belongs to a local government area (LGA) and agricultural zone. For instance, Ishibori, Agoi Ibami and Mfamosing belong to Ogoja, Yakurr and Akamkpa LGAs; northern, central and southern agricultural zones, respectively. The selection of these communities, namely: Ishibori, Agoi Ibami and Mfamosing, was based mainly on the formations of limestone in the various agricultural zones.

3.2 Geology of the Study Area

Limestone occurs throughout Nigeria in both Basement Complex rocks and the Sedimentary Basins and can be grouped into Cretaceous and Tertiary limestone on the basis of geological age (Fatoye and Gideon, 2013). The Sedimentary limestone of the Cretaceous and Tertiary ages is associated with shale, siltstone and fine-grained sandstone. They are often hard, grey and shelly (Abatan *et al*., 1993), while the Cretaceous varieties extend from Calabar in the south-east, through Agila, Igumale and Yandev in central Nigeria to Ashaka and Gombe in the northeast of the country.

The Oban-Obudu hills form the Basement Complex of CRS and are made up of Precambrian Schist and Gneiss with intrusives of igneous rocks (Ekwueme, 1987). The sedimentary limestone of Cretaceous and Tertiary ages in Cross River State is common in the Ikom depression (Mamfe Rift) and Calabar flank. Limestone in the state is intercalated with shale, siltstone, and fine-grained sandstone (Figs. 2 and 3) (Okpiliya *et al.,* 2011; Ogungbesan and Akaegbobi, 2011, Fatoye and Gideon, 2013). This has resulted in shale, limestone and sandstone (SLS) (Ishibori), sandstone and limestone (SL) (Agoi Ibami), and shale and limestone with sandstone intercalation (SLSi) (Mfamosing) (Table 1).

3.3 Climate of the Study Area

Cross River State is characterized by a humid tropical climate and has distinct wet and dry seasons which vary slightly in duration from the southern guinea savannah in the Ishibori

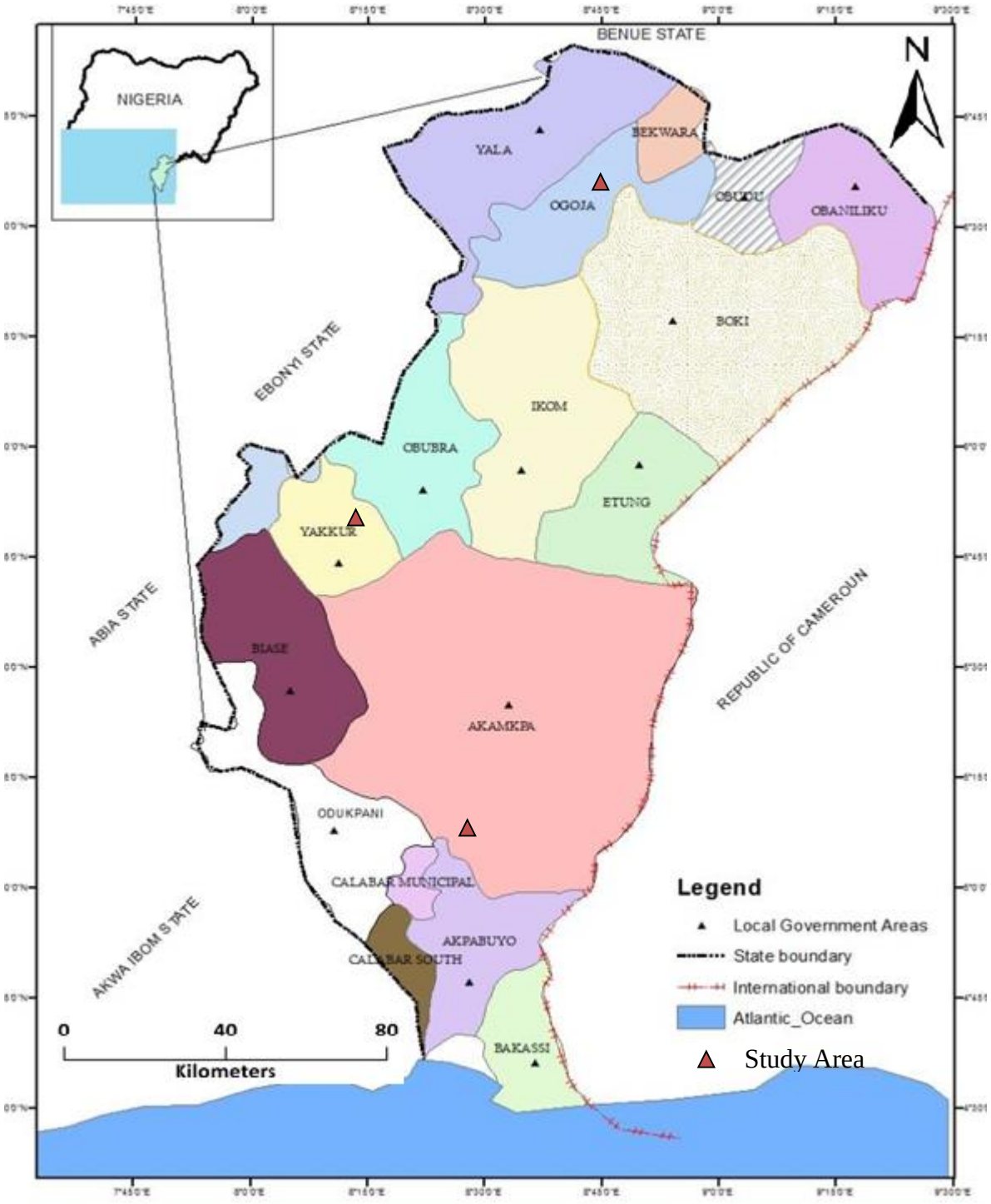

Fig. 1: Political Map of Cross River State showing Local Government Areas and the study locations
Source: Department of Geography and Regional Planning, UNN (2017)

Table 1: Location of sampling areas in the three geological formations

Northern agricultural zone {shale, limestone and sandstone (SLS)}			
Ogoja LGA			
Ishibori	06°39'17"N	08°47'51"E	70.1 m (asl)
Central agricultural zone {sandstone and limestone (SL)}			
Yakurr LGA			
Agoi Ibami	05°43'27"N	08°10'37.2"E	82.0 m (asl)
Southern agricultural zone {shale and limestone with sandstone intercalation (SLSi)}			
Akamkpa LGA			
Mfamosing	05° 04'41.8"N	08° 27'49.8" E	33.5 m (asl)

Source: Google Map (2018); Nigerian Geological Survey Agency (NGSA) (2006)

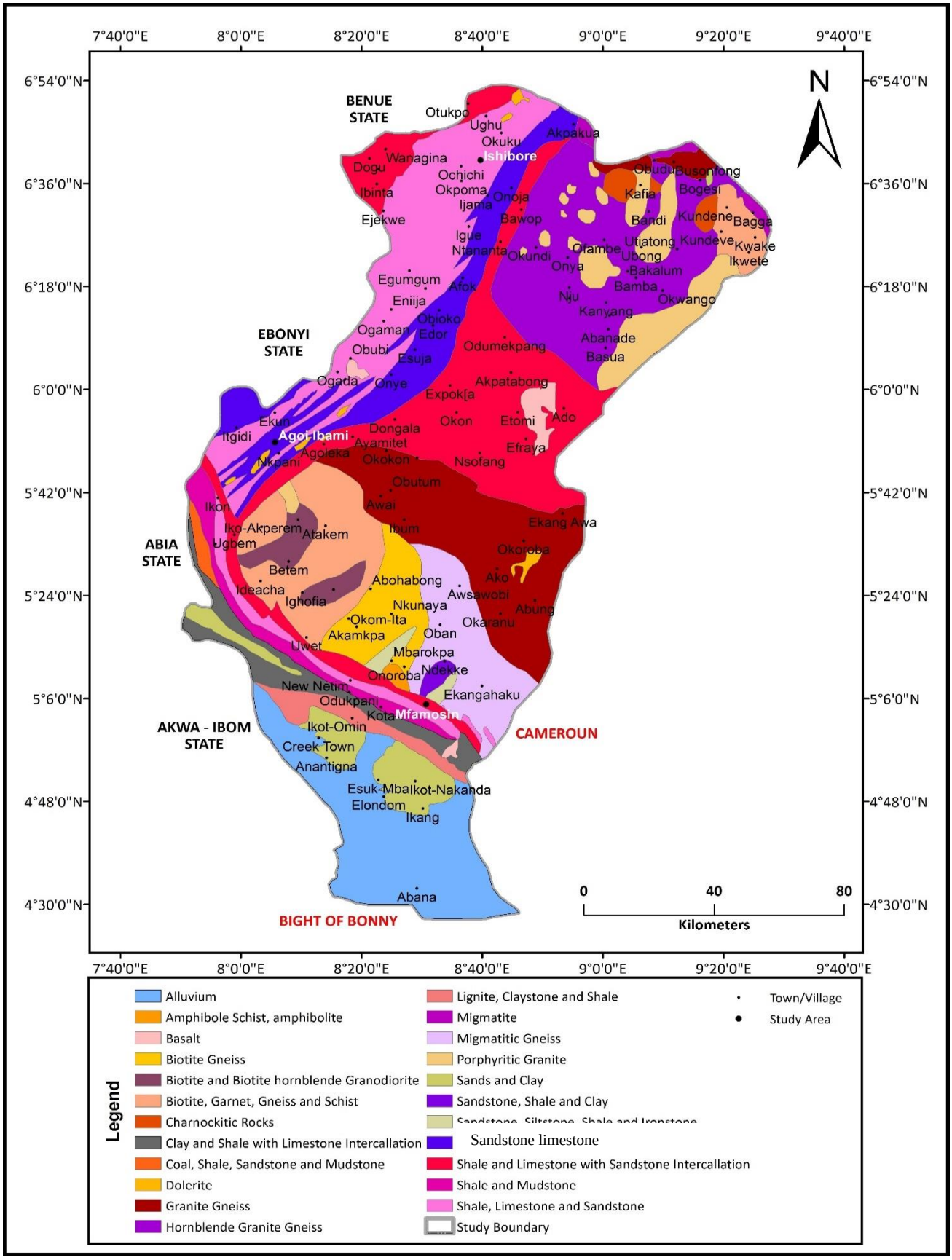

Fig. 2: Geological map of Cross River State
Source: Department of Geography and Regional Planning, UNN (2017)

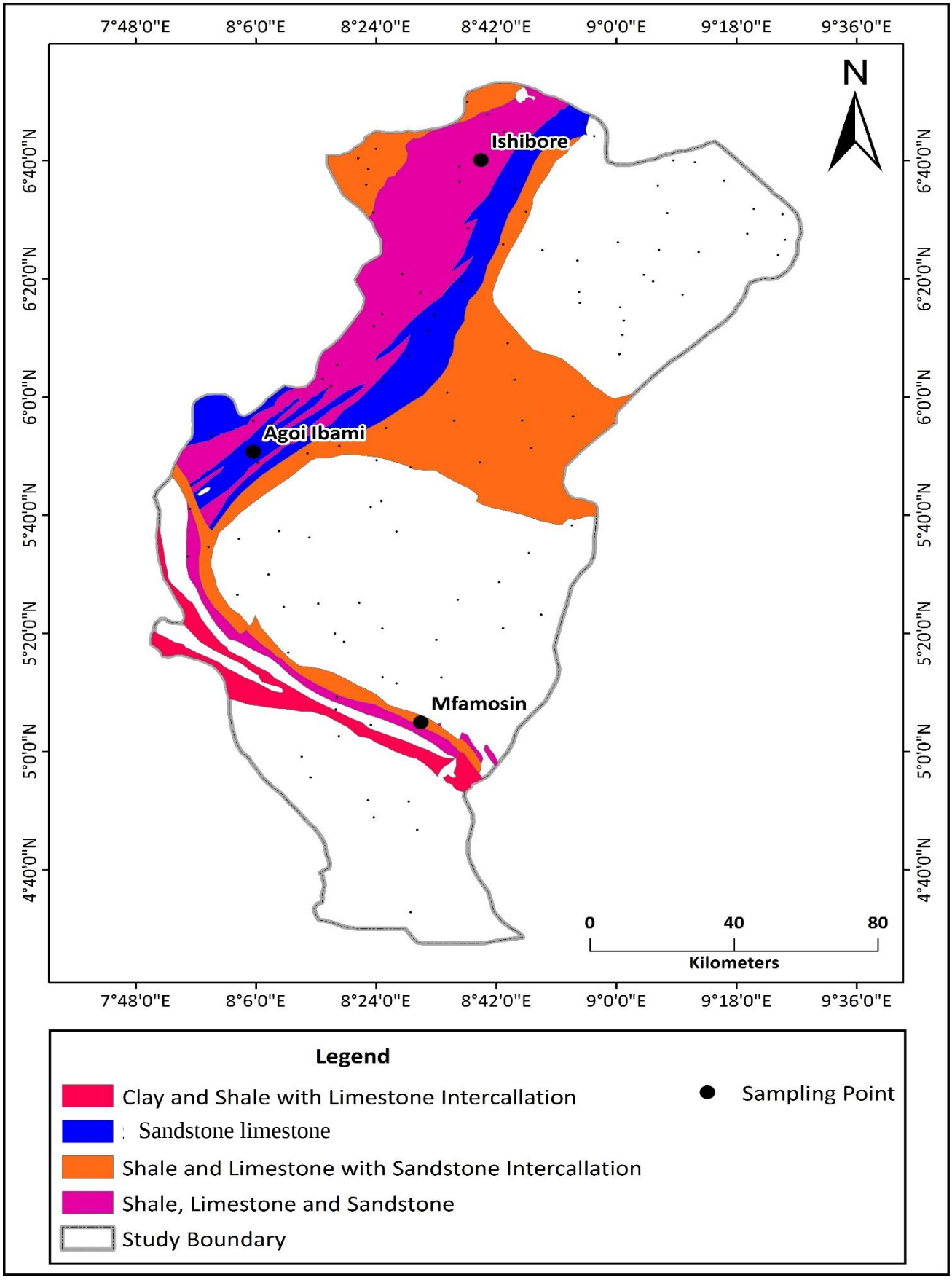

Fig. 3: Distribution of limestone from the Northern to Central and Southern Cross River State
Source: Department of Geography and Regional Planning, UNN (2017)

area to the tropical rainforests of Agoi Ibami and Mfamosing. The Ishibori area is characterized as a moist subhumid, while Agoi Ibami and Mfamosing are moist humid to per humid. Rainfall varies from 1251.4 and 3347.8 mm/annum with a mean of 1983.02 mm/annum in the Ishibori area, while a range of 1760.3-2683.8 mm/annum and mean of 2258.0 mm/annum are typical of the Agoi Ibami area. Consequently, Mfamosing has a range of 2109.0-3770.8 mm/annum and a mean of 2894.1 mm/annum of rainfall (Sambo *et al*. 2016b).

Temperature varies from 22.96 and 33.75 ^{o}C with a mean of 27.85 ^{o}C in the Ishibori area and a range of 22.56-31.89 ^{o}C with a mean of 26.96 ^{o}C in the Agoi Ibami area. The Mfamosing area has a range of 23.53-31.95 ^{o}C and a mean of 27.23 ^{o}C (Sambo *et al*. 2016a). Mean relative humidity in the Ishibori area is reported as 72.14 %, while evaporation rate has a mean value of 2.24 mm/day and a range of 1.8-2.8 mm/day. In the Agoi Ibami area, a mean value of 81.71 % had been obtained for relative humidity, while the evaporation rate has a mean of 3.61 mm/day with a range of 2.9-5.0 mm/day. In the Mfamosing area, relative humidity has values of 9.2-80.75% with a mean of 84.99 %. The evaporation rate in the Mfamosing area has a range of 0.9-3.0 mm/day and mean of 2.62 mm/day (Sambo *et al*. 2016b). The climatic condition of the state varies in such a way that atmospheric moisture decreases and sunshine intensity increases as the southern guinea savanna in the Ishibori area is being approached from the southern tropical rainforests.

3.4 Vegetation and Land Use

The vegetation maps of Nigeria and Cross River State are presented in Figs. 4 and 5, respectively. Cross River State has a multi-vegetational system. Its vegetation ranges from the mangrove swamps in the Southern coastal areas (Bakassi, Calabar South and parts of Akpabuyo Local Government Areas) through the tropical rainforest in the southern upland (Calabar Municipal, Akpabuyo, Odukpani, Akamkpa and Biase Local Government Areas) (Eni *et al*. 2011) and central areas of the state (Yakurr, Abi, Obubra, Ikom and Boki Local Government Areas) to the southern guinea savannahs in the northern parts of the state (Ogoja, Yala, Bekwarra and parts of Obudu Local Government Areas) (Fon *et al*., 2014). Montane parkland dominates the Obudu-Obanliku plateaux.

Cross River State is agrarian and agriculture is practiced at commercial scale by the Government and companies, while subsistence farming is practiced by individuals and households as means of livelihood. Cocoa (*Theobroma cacao*), oil palm (*Elaeis guineensis*), gmelina (*Gmelina arborea*), rubber (*Hevea brasiliensis*), sesame (*Sesamum indicum*) and tree fruits as well as neem (*Azadirachta indica*) are tree crops commonly grown at commercial

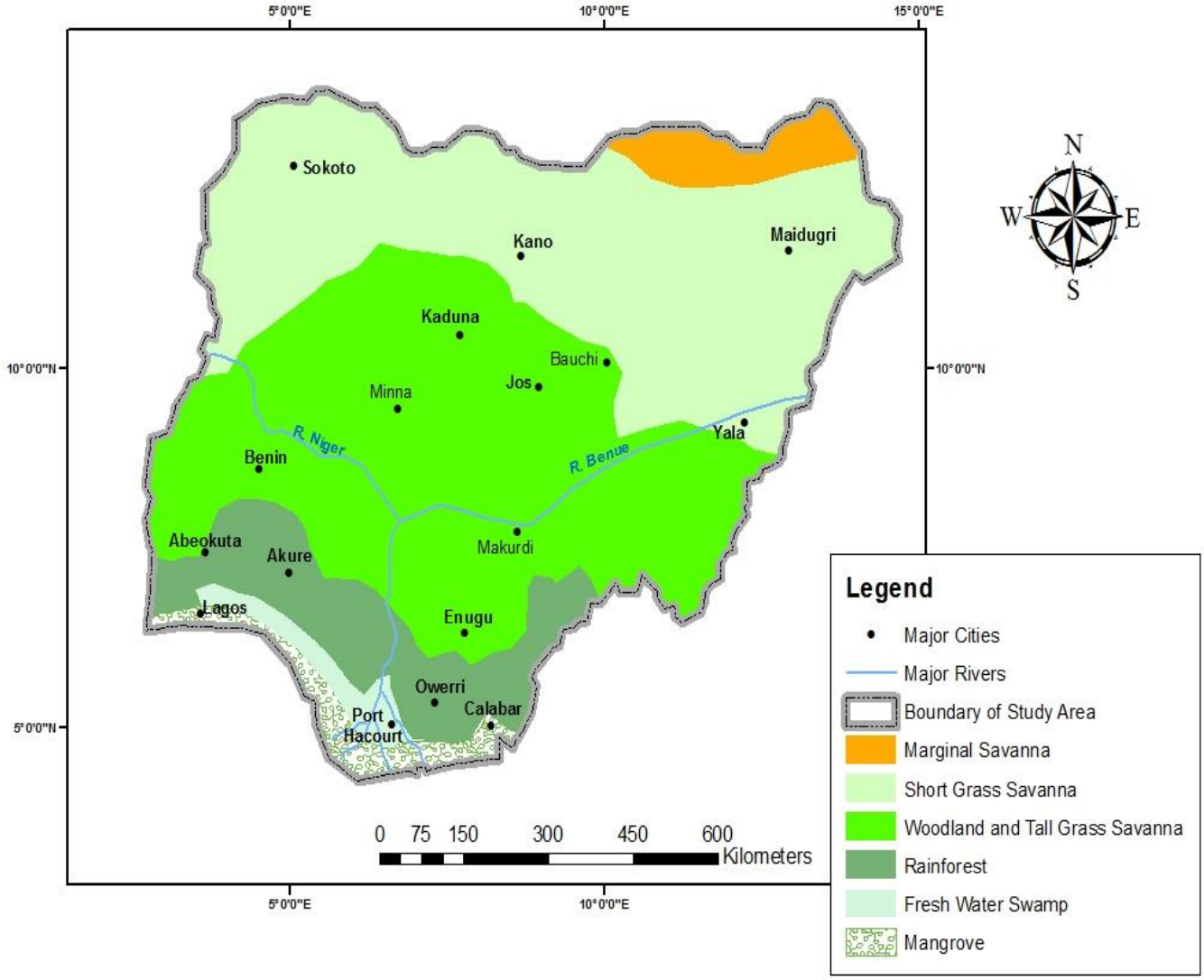

Fig. 4: Vegetation belts of Nigeria.
Source: Department of Geography and Regional Planning, UNN (2017)

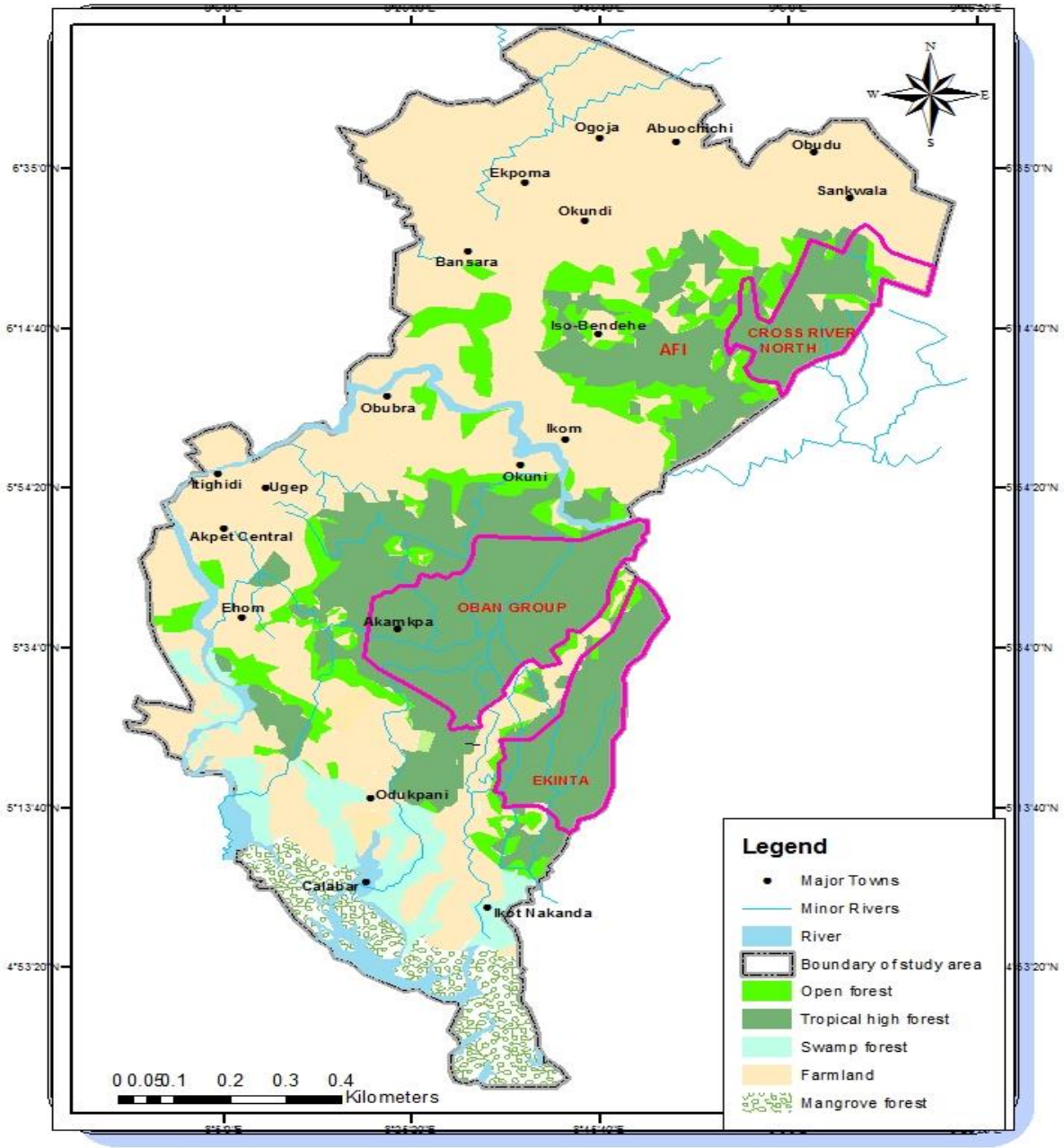

Fig. 5: Vegetation and land use map of Cross River State
Source: Department of Geography and Regional Planning (2017)

scale. Commonly produced food crops include maize (*Zea mays*), cassava (*Manihot esculenta*), yam (*Dioscorea spp.*) and rice (*Oryza sativa*), as well as vegetables like cucumber (*Cucumis sativus*), okra (*Abelmoschus esculentus*) and watermelon (*Citrullus lanatus*); however, many other foods and cash crops have been successfully cultivated with economic benefits. Besides its agricultural land uses, CRS is home to many minerals including; baryte ore, quartz, mica, tourmaline, limestone and titanium in the Basement Complex, while the Sedimentary Basin has limestone, pyrite, rock salts and uranium prospects (Njar, 2018). Exploitation of these minerals has positively affected the state's economy leaving an adverse impression on the underlying soils.

3.5 Soils of Cross River State

According to Nsor and Ibanga (2008), soils derived from sandstone-shale-siltstone are characterized by sandy loam and loamy sand surface textures, while the Basement Complex soils were mainly loamy sand and sandy loam to clayey textures in the basaltic soils of Central Cross State. On the otherhand, Aki (2012) obtained sandy loam to clay and clay loam in the wetland soils developed from limestone in Southern Cross River State. In a recent study in Yakurr, Ofem *et al*. (2017) reported loamy sand and sandy loam textures, while Afu *et al*. (2017) obtained sandy clay loam and loam in limestone derived soils in Odukpani. Furthermore, Aki *et al*. (2014) obtained loamy sand and sandy loam in the surface soils developed from biotite-hornblende-gneiss in Akamkpa.

Soil pH in the coastal plain soils of Bakassi had a range of 2.0-4.6 (Abua, 2012), while Aki (2012) obtained 7.81-8.20 in the limestone soils in the wetlands of Southern CRS. Organic carbon content in the soils of CRS varied from 0.8-17.6 g/kg in the soils developed from biotite-hornblende-gneiss in Akamkpa (Aki *et al*., 2014) to 2.8-13.1 g/kg in the wetland soils developed from limestone in Southern CRS (Aki, 2012). Ofem *et al.* (2017) obtained a range of 1.1-11.8 g/kg in the soils of Yakurr, as relatively higher values of 1.4-32.7 g/kg were obtained in the schist derived soils of Biase (Ofem and Esu, 2015). Total N ranged from 4.8 to 11.1 g/kg in Bakassi and higher in Akpabuyo (Abua, 2012), while 0.1-1.0 g/kg was obtained in the wetland soils developed from limestone in southern Cross River State (Aki, 2012), and mean surface value of 1.4 g/kg was stated for the soils derived from limestone (Afu *et al*., 2017). However, in the schistose soils of Biase, Ofem and Esu (2015) obtained 0.42-1.82 g/kg, while Ofem *et al*. (2017) obtained 0.3-1.0 g/kg in the Yakurr area for total nitrogen.

Abua (2012) recorded high exchangeable Ca and Mg in the soils of Bakassi and low values in the coastal plain soils of Akpabuyo, while Aki (2012) obtained moderate levels of exchangeable Ca and Mg but low K and Na in limestone soils of wetlands in southern CRS.

According to Afu *et al.* (2017), exchangeable Mg was low (1.8-3.2 cmol/kg) in limestone, sandstone and shale derived soils in Odukpani. Nevertheless, Ofem and Esu (2015) had rated the schistose soils in Biase low in exchangeable bases with ranges of 1.4-4.0, 0.3-1.1, 0.08-0.38, and 0.08-0.37 cmol/kg for exchangeable Ca, Mg, K, and Na, respectively.

Mineralogy of the clay fraction indicates the presence of quartz, illite, nacrite and mixed layer structure of kaolinite and montmorillonite in the flood plain soils of Onwu River (Akpan-Idiok and Ogbaji, 2013), while Ofem and Esu (2015) obtained quartz, kaolinite and muscovite as dominant minerals as well as plagioclase, microcline and sepiolite in the schistose soils of Biase. In the Basement Complex soils of Akamkpa, Aki *et al.* (2014) obtained quartz, microcline and kaolinite.

The wetland soils developed from limestone in Southern Cross River State were classified as Aeric Endoaqualfs (Eutric Gleysols), Typic Dystrudepts (Plinthic Aerisols, Dystric Cambisols) (Aki, 2012)., while Akpan-Idiok and Ogbaji (2013) classified the flood plain soils of the Onwu River as Typic Kandiudults (Haplic Acrisols) and Fluvaquentic Humaquepts (Gleyic/Vertic Cambisols). In Biase, the schistose soils were classified as Typic Paleudults and Typic Kanhapludults (Clayic and Skeletic Acrisols) (Ofem and Esu, 2015), while soils on coastal sediments in Southern Calabar were classified as Typic and Haplic Sulfaquents (Gleysols) (Ofem and Esu, 2016).

3.6 Field and Laboratory Studies

Google satellite imageries of the areas underlain by limestone were obtained as found in Table 1 and used as base maps and later, for the selection of sampling areas. Sample areas underlain by limestone were selected in each of Ishibori, Agoi Ibami and, Mfamosing to represent soils underlain by shale-limestone-sandstone (SLS), sandstone-limestone (SL) and, shale and limestone with sandstone intercalation (SLSi), respectively. This was done according to the criteria set by Esu (2010) and Ezeaku (2011) for selecting sampling areas during soil surveys. Such areas must be representative and include all soil units within the survey area, accessible and allow for very intensive observation. Soil boundaries were delineated on the imageries by visual interpretation.

Field reconnaissance visits were then carried out to the selected areas underlain by SLS, SL and, SLSi in Ishibori, Agoi-Ibami and Mfamosing, respectively. These areas have

previously been mapped and shown to be underlain by these formations by the Nigerian Geological Survey Agency (NGSA, 2008). As a means of confirming the presence of limestone, 10 % HCl was used on rock outcrops. The foamy reaction was a confirmation of the presence of carbonate-rich rocks.

Digital elevation models (DEMs) of the study locations were obtained from USGS Explorer SRTM 1 arc-second Global (http://earthexplorer.usgs.gov/) at a resolution of 30 m (Figs. 6, 8 and 10). Using ArcGIS (ESRI, US) software, the DEMs were used to generate the slope maps of the study areas (Aksoy *et al*., 2009) (Figs. 7, 9 and 11). Data generated by this means provides a comparatively higher accuracy compared to that of the satellite imageries. The elevation ranges created in the slope maps were used to delineate slope transition from high to low elevations. This was in turn used to identify the soil mapping units, which include IH1, IH2 (SLS), AI1, AI2 and AI3 (SL), and MF1, MF2 and MF3 (SLSi).

Guided by the elevation ranges, two profile pits were randomly sited and dug in each of the eight (8) mapping units to represent the soils. Auger soil samples were obtained from depths of 0-20 and 20-40 cm in each mapping unit transiting to the next to check transition zones between elevation ranges. However, the auger soil samples were not analyzed in the laboratory. The soil profiles were described according to the criteria of Schoeneberger *et al*. (2012). Thirteen (13) soil samples were obtained from soils overlying SLS, while twenty (20) soil samples were collected each from soils overlying SL, and SLSi, all from the pedogenic horizons. Fifty-three (53) soil samples were collected from 16 soil profile pits and used for physical and chemical analyses. Sampling was done from bottom to top to determine physical, chemical and mineralogical properties and transported in labelled polythene bags to the laboratory**.** On the other hand, nine (9), fourteen (14) and twelve (12) soil samples were obtained from depth intervals of 30 cm in soils overlying SLS, SL, and SLSi, respectively. Thirty-five (35) soil samples were therefore obtained and subjected to elemental oxide analysis using XRF.

Similarly, six (6) soil samples were obtained from ABC horizons in SLS, and eleven (11) each from SL and SLSi formations and used for dithionite and oxalate forms of Fe and Al, as well as clay mineralogy using x-ray diffractometer (XRD). Only soil samples in the highest and medium elevation ranges were used for these analyses (forms of Fe and Al, and clay mineralogy). This was done with a focus on the soils developed from limestone formations.

The soil samples were air-dried under laboratory conditions, ground and passed through a 2 mm mesh of sieves. Air-dry soil samples meant for elemental oxide analysis were

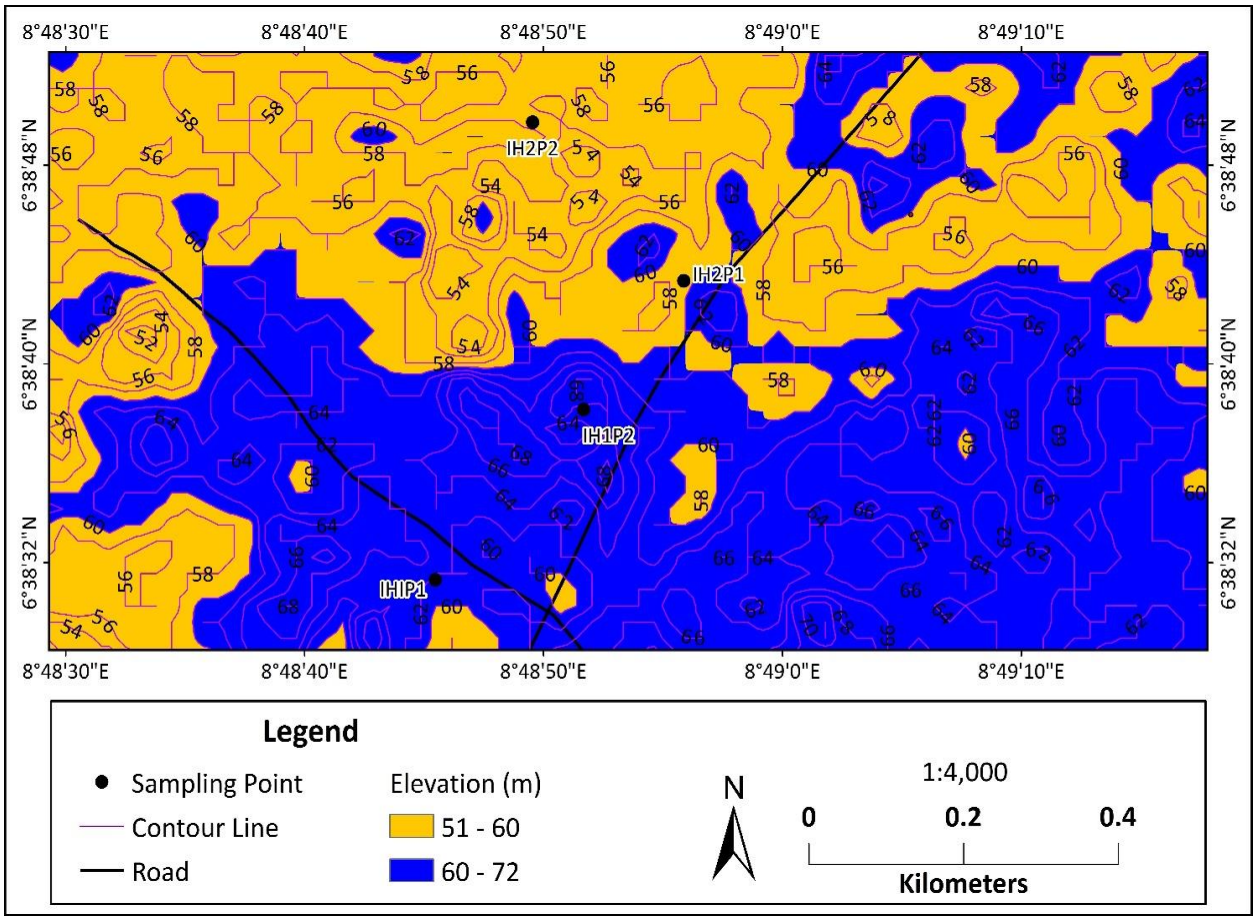

Fig. 6: DEM map of Ishibori, underlain by shale, limestone and sandstone

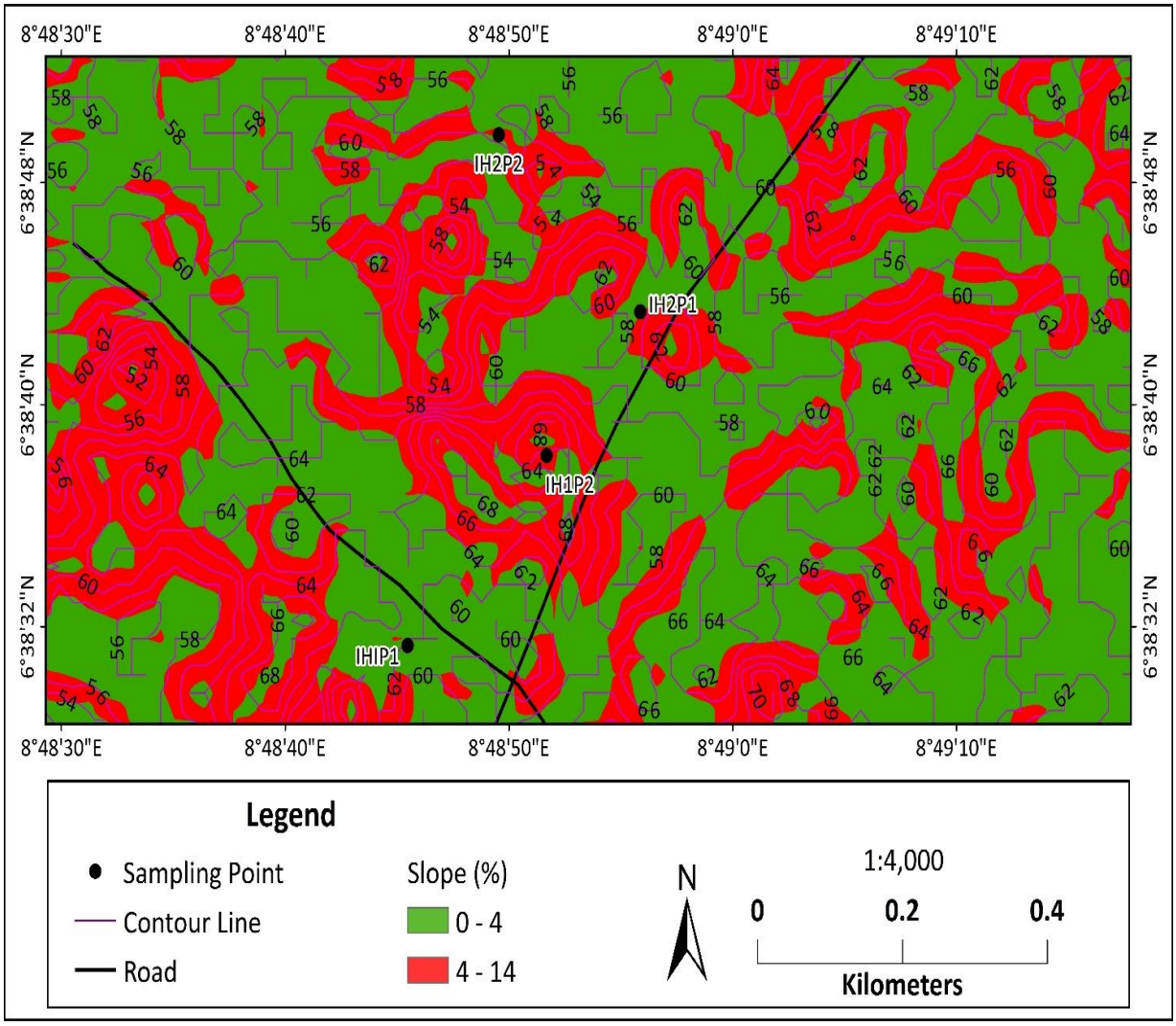

Fig. 7: Slope map of Ishibori, underlain by shale, limestone and sandstone

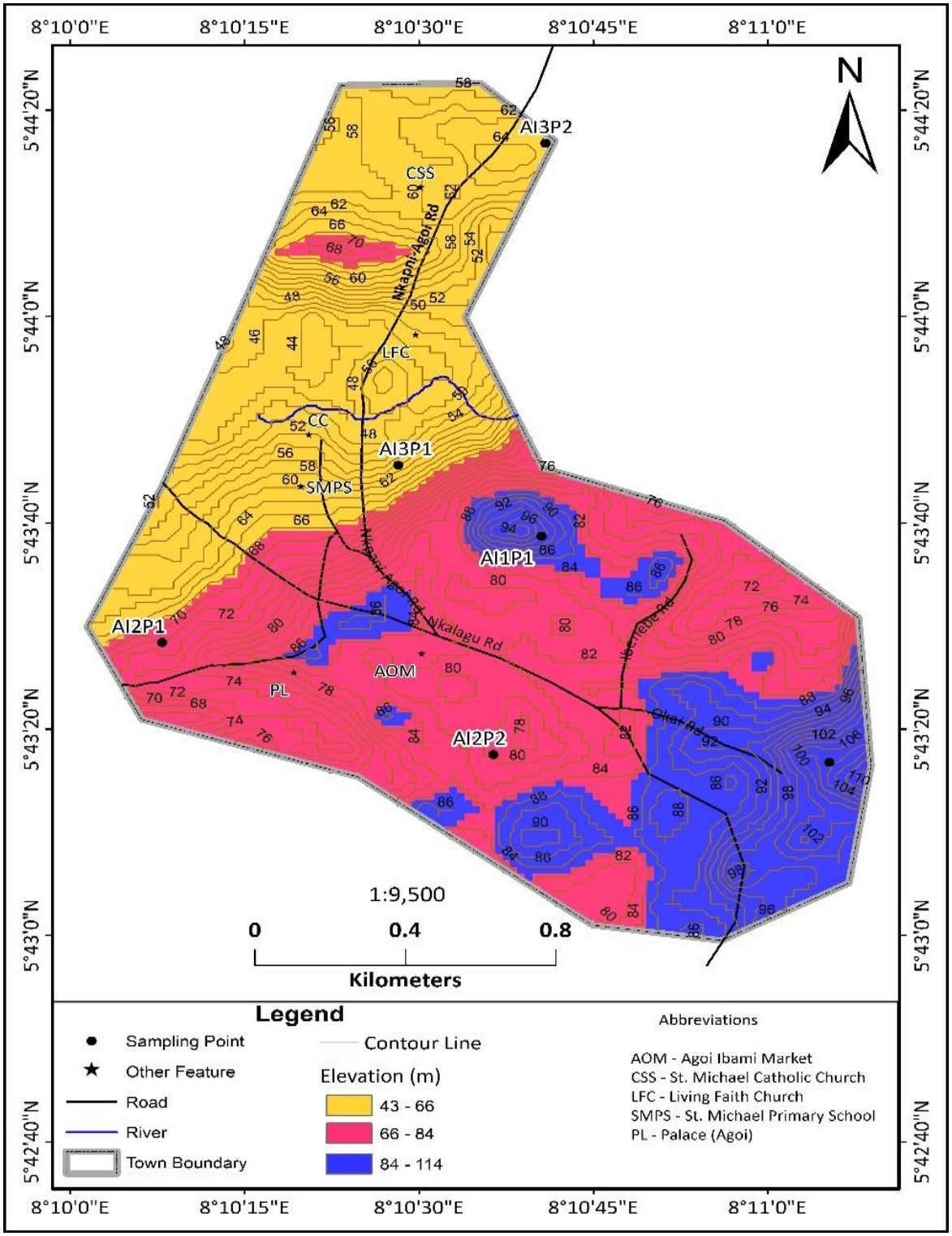

Fig. 8: DEM map of Agoi Ibami, underlain by sandstone and limestone

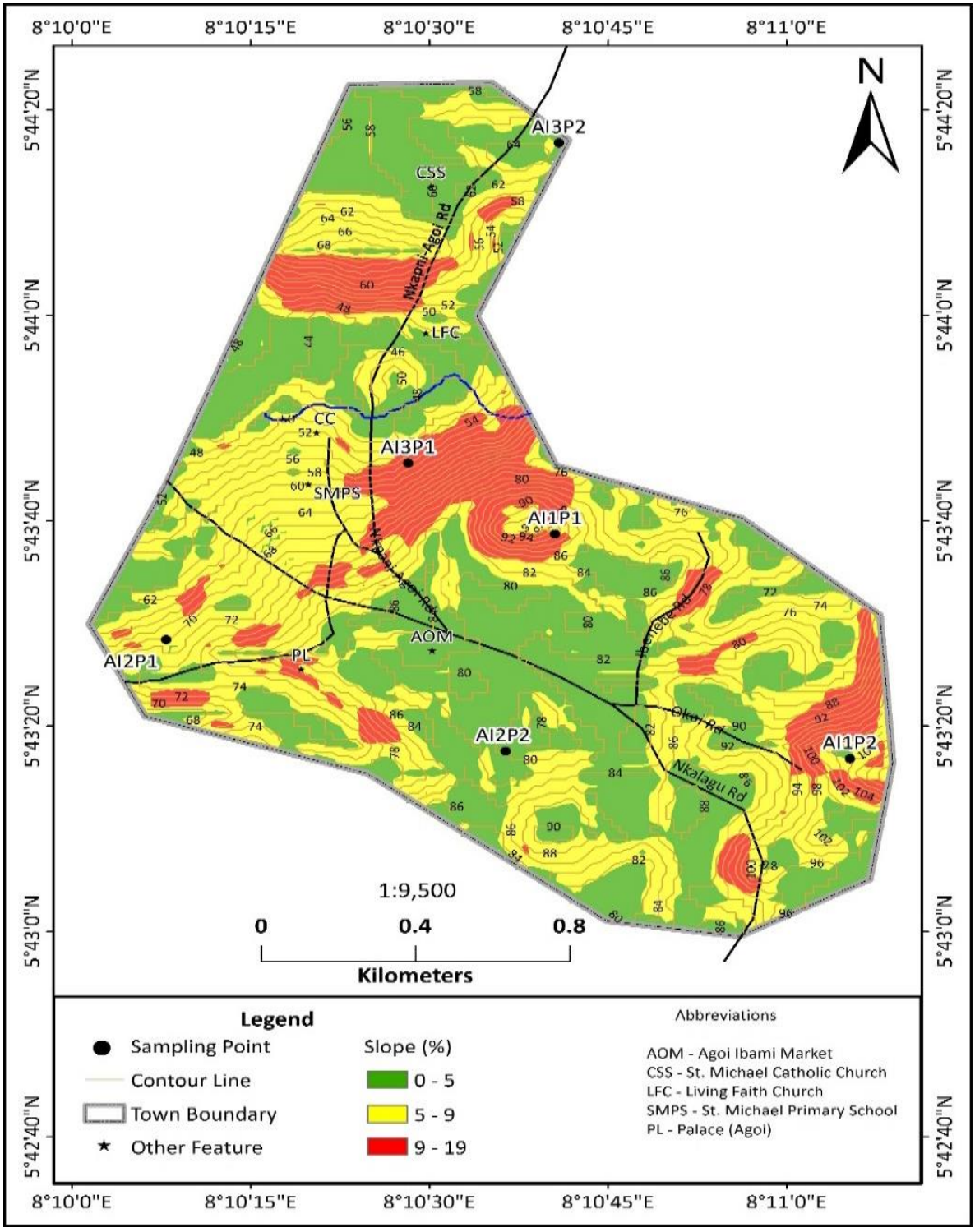

Fig. 9: Slope map of Agoi Ibami, underlain by sandstone and limestone

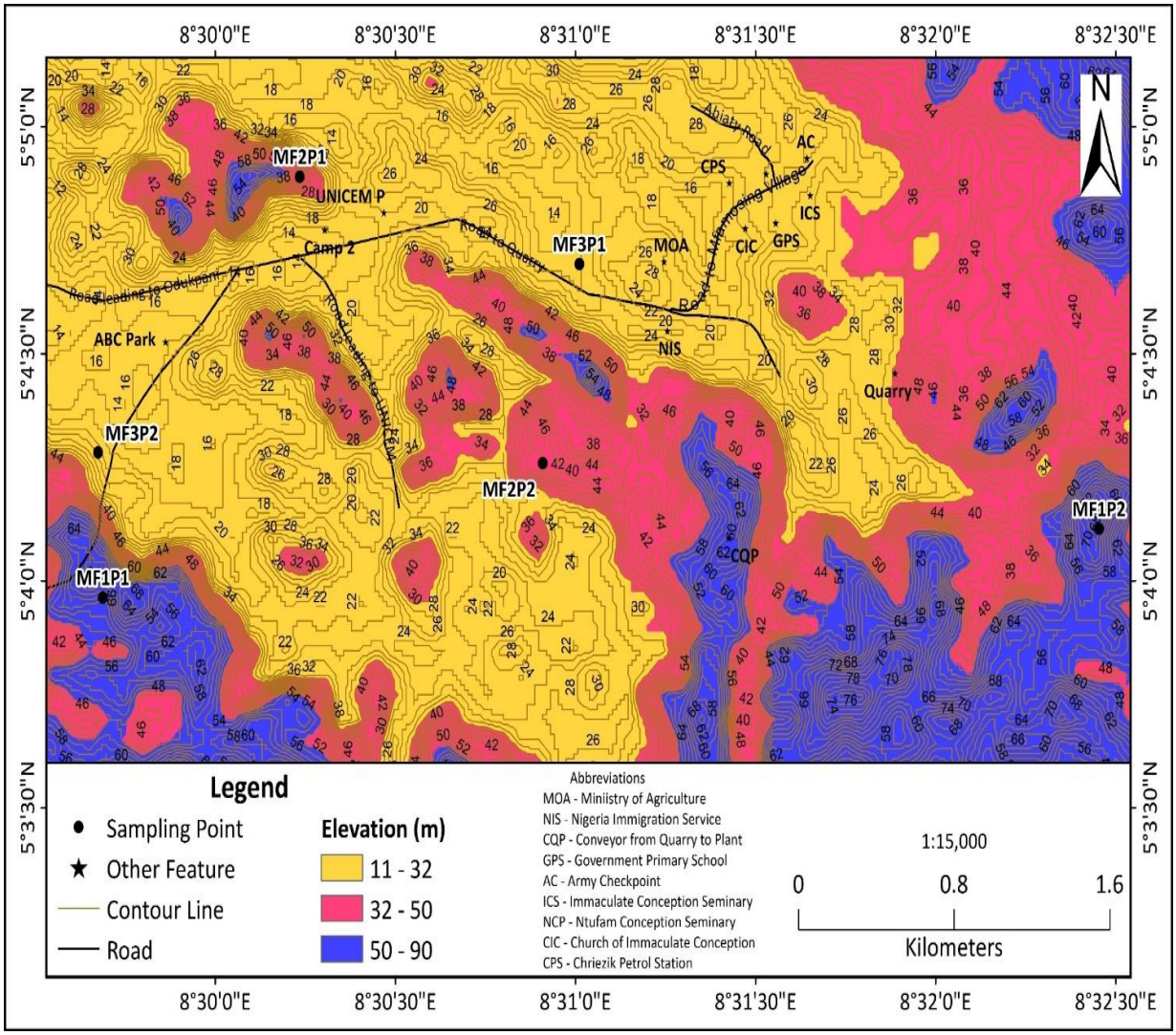

Fig. 10: DEM map of Mfamosing underlain by shale and limestone with sandstone intercalation

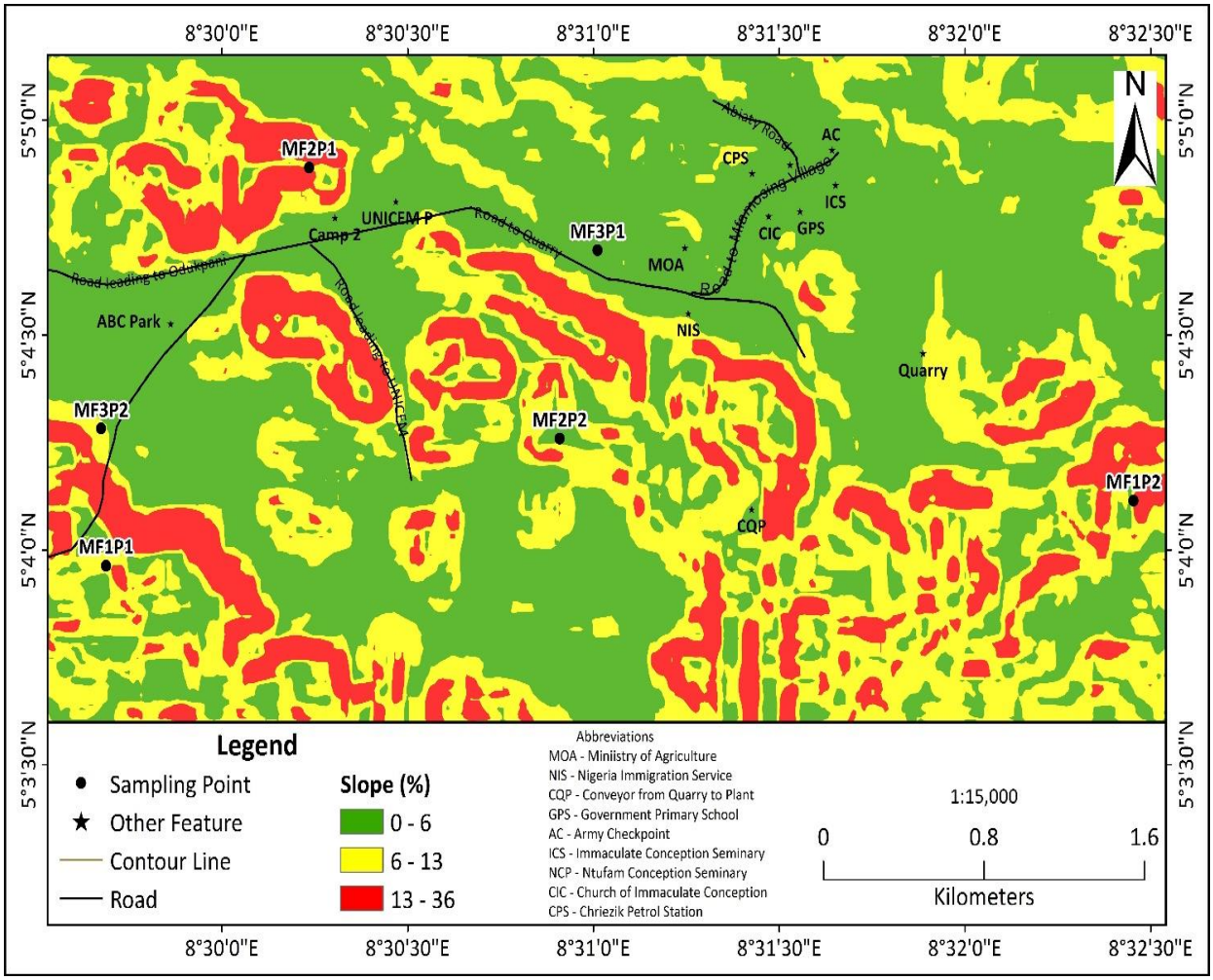

Fig. 11: Slope map of Mfamosing underlain by shale and limestone with sandstone intercalation

further pulverized with an automatic mill and homogenize into a fine powder (3-4 µm) before any further analysis.

Two standard core cylinders with volumes of 137.98 cm^3 and 98.21 cm^3 were used vertically for the collection of undisturbed soil samples meant for saturated hydraulic conductivity (Ksat.), total porosity, air space porosity, capillary porosity and soil moisture content at saturation and after being drained at 30, 60 and 90 cm tensions. In any case, soil samples meant for sesquioxide forms and mineralogical analysis were obtained from the A, B and C horizons of soil profile pits per mapping unit except from IH2, AI3 and MF3.

3.7 Laboratory Studies

Soil samples meant for physical, chemical and mineralogical studies were processed by air-drying under laboratory conditions for about 48 hours to constant weight, ground with a wooden pestle and sieved with a 2 mm sieve. The fine earth fractions were then used for laboratory analysis.

3.7.1 Soil Physical Properties

3.7.1.1 Saturated Hydraulic Conductivity (Ksat)

Hydraulic conductivity of saturated soils (Ksat) was determined by the constant head method of Klute and Dirksen (1986). This method is based on the direct application of Darcy's equation to a saturated soil column of uniform cross-sectional area. A hydraulic head difference as imposed on the soil column, and the resulting flux of water measured using the relation:

$Ks = VL/At\ (H_2 - H_1)$

Where, V = Volume of water that flows through the sample of cross sectional area (A) in time (t).

$(H_2 - H_1)$ = Hydraulic head difference

L = Length of sample

The following properties were determined by the method of Obi (2000a):

3.7.1.2 Total Porosity (St)

Soil total porosity was calculated from:

$$St = \frac{\text{Volume of water in soil at saturation}}{\text{Volume of cylinder}} \times 100$$

3.7.1.3 Air space porosity (non-capillary porosity) (NCP)

Soil air space porosity was calculated as:

$$NCP = \frac{\text{Vol. of water drained at 60 cm of tension}}{\text{Volume of cylinder}} \times 100$$

3.7.1.4 Capillary Porosity (CP)

Soil capillary porosity was calculated as the

$$CP = \frac{\text{Vol. of water retained at 60 cm of tension}}{\text{Volume of cylinder}} \times 100$$

3.7.1.5 Soil Moisture Content (Sm)

Soil moisture content was calculated as:

$$Sm = \frac{\text{Mass of water at saturation}}{\text{Oven dry mass of soil}} \times 100$$

3.7.1.6 Soil moisture content at 60 cm of tension

Soil moisture content at 60 cm of tension was calculated as a ratio:

$$= \frac{\text{Mass of water in soil}}{\text{Oven dry mass of soil}} \times 100$$

3.7.1.7 Soil Bulk density (Bd)

Soil bulk density was determined by calculation after oven drying soil at 105 ^{o}C for 48 hours or to constant weight, using the relation:

$$Bd = \frac{\text{Oven dry mass of soil}}{\text{Volume of soil}} \times 100$$

Volume of soil in cylinder was determined by using the formula; $\pi r^2 h$.

Where, π is a constant with value of 22/7

r^2 = square of the internal radius of the core

h = height of core.

3.7.1.8 Pore Size Distribution

Water desorption method was used for pore size distribution by saturating soil samples in standard core cylinders with water, and draining with successively increased tension as outlined by Danielson and Sutherland (1986). Three tension levels of 30, 60 and 90 cm were used. According to Obi (2000b), equivalent radii of pores were obtained from:

$r = 2T/\rho gh$

where, r = equivalent radius of pore

T = surface tension

ρ = density of water

g = acceleration due to gravity

h = suction (height of water column).

The fraction of total pore space filled with water (Θv/St) may then be plotted as a function of pore diameter to express the pore size distribution (Danielson and Sutherland, 1986).

3.7.1.9 Soil Particle Size Analysis

Soil particle size distribution was determined on the fine-earth fraction by the Bouyoucos hydrometer method using sodium hexametaphosphate (Calgon) as the dispersant (Soil Survey Staff, 2014a). Percent sand, silt and clay were then traced on a soil textural triangle to determine the soil textural class. Particle fractionation into coarse sand (CS) and fine sand (FS) was carried out.

3.7.2 Chemical Soil Properties

3.7.2.1 Soil pH

Soil pH was determined in a soil to solution ratio of 1:2.5 using a glass electrode pH meter in water and 1 *M* $CaCl_2$ as reported by SSS (2014a). This was done after standardizing the pH meter with a buffer of pH 7. Delta pH was also obtained as the difference between pH in water and 1 *M* $CaCl_2$.

3.7.2.2 Organic Carbon

The Walkley-Black modified acid-dichromate procedure was used to determine organic C using diphenol amine as an indicator and potassium dichromate as the oxidizing agent as outlined by SSS (2014a). This procedure measures only active or decomposable organic matter in the soil.

3.7.2.3 Total nitrogen

Total N was determined by the macro-Kjeldahl digestion method through the careful use of a macro- Kjeldahl digestion-distillation apparatus (SSS, 2014a).

3.7.2.4 Available phosphorus

Available P was determined by the Bray P1 extraction procedure (Soil Survey Staff, 2014a).

3.7.2.5 Exchangeable Bases

The soils were leached with 1 *N* NH_4OA_C (pH 7) in a 1:1 soil-solution ratio. Exchangeable K and Na in the extract were determined with the aid of a flame photometer, while Ca and Mg were determined by the versenate EDTA titration procedure (SSS, 2014a).

3.7.2.6 Exchangeable Acidity ($H^+ + Al^{3+}$)

Exchangeable acidity was obtained by leaching the soil samples with $BaCl_2$-triethanolamine (TEA) solution buffered at pH 8.2 by using the back titration procedures of Holmgren *et al.* (1977) as modified in SSS (2014a).

3.7.2.7 Cation Exchange Capacity (CEC) by NH_4OAc at pH 7.0

Cation exchange capacity (CEC) by NH_4OAc at pH 7.0 was obtained by saturating the soil exchange sites with an index cation (NH_4^+) at pH of 7.0 by treating the soil with 1 *M* NH_4OA_C at pH 7.0 and the concentration of NH_4^+ in the extract was determined by distillation and titration (SSS, 2014a).

3.7.2.8 Base Saturation

The base saturation was calculated by expressing the sum of exchangeable bases as a percentage of CEC by ammonium acetate at pH 7.0 (Soil Survey Staff, 2014a).

3.7.2.9 Cation Exchange Capacity – clay (CEC-clay)

The cation exchange capacity of the mineral clay colloids was estimated by dividing the CEC by percent clay (SSS, 2014b).

3.7.2.10 Calcium Carbonate Equivalent (CCE)

Calcium carbonate equivalent was measured by treating the soil samples with 1 *M* HCl and titrating with 1 *M* NaOH in the presence of I ml of bromothymol blue indicator (Udo *et al*., 2009). The % $CaCO_3$ was calculated from:

$$\% \ CaCO_3 \text{ equivalent} = \frac{M \times a - b}{s} \times 50 \times mcf$$

Where a = ml NaOH is used for blank

b = ml NaOH used for sample

s = air-dry sample weight (g)

M = molarity of NaOH solution (g)

50 = equivalent weight of $CaCO_3$

mcf = moisture correction factor

3.7.2.11 Dithionite and oxalate extractable Fe and Al

Dithionite-extractable Fe and A1 were measured by the method of Mehra and Jackson (1960) and as modified in SSS (2014a) using the dithionite-citrate system buffered with sodium bicarbonate. Fe and A1 in the extracting solution were determined colorimetrically using inductively coupled plasma mass spectrometry (ICP-MS).

Oxalate forms of Fe and Al were estimated by extraction with 0.2 *M* acid ammonium oxalate at pH 3.0 (McKeague and Day, 1966). Dithionite and oxalate forms of Fe and Al

were determined using the Perkin Elmer NexION 450D ICP-MS model. The equipment has a detection limit of 0.01 g/kg. Other derived forms of Fe and Al were obtained, as shown in Table 2.

3.7.3 Elemental oxide content using X-Ray fluorescence

An automatic mill was used to pulverize and homogenize air-dry soil samples into fine powder (3-4 µm) and then analyzed with XRF Delta Premium Spectrometer 2019 using the procedures of Tejnecky *et al*. (2015). X-ray emission spectrography is elemental analysis obtained by measuring the X-ray fluorescence produced by bombarding a sample with high-energy X-rays. Each element yields fluorescent radiation of unique wavelengths, one of which is selected for measurement by using an analyzing crystal that diffracts according to Bragg's law. The intensity of the fluorescent radiation is proportional to the amount of the element present. However, the process is affected by sample homogeneity, particle size and the absorption and enhancement of radiation by any other elements present in the sample (Soil Survey Staff, 2014a).

3.7.4 Mineralogical Analysis

3.7.4.1 Pretreatment of Soils for Clay Mineralogy

Pretreatment of soils is intended to have minimum effect on soil constituents (Kunze and Dixon, 1986). The removal of free iron oxides, organic matter, soluble salts and calcium carbonates cleans coarser soil fractions and frees soil aggregates of the cementing action of Fe oxides. The removal of these oxides results in well-defined x-ray diffraction patterns. However, the efficiency of H_2O_2 is low in an alkaline medium and requires carbonates to be removed with acid sodium acetate if organic matter is to be oxidized.

Iron-containing samples, such as oxisols and some ultisols, are generally to disperse unless the free Fe oxides are removed. The sodium dithionite citrate method, has been successfully used with minimum destruction of the clay minerals (Mehra and Jackson, 1960). Sodium bicarbonate acts as a buffer, while sodium dithionite reduces the Fe.

Soil samples were dispersed and separated into particle size fractions with the use of Calgon. Fifty grams (50g) of MgOAc was added to the clay suspension and allowed to flocculate overnight (Kunze and Dixon, 1986). The supernatant solution was decanted and transferred into a 100mls centrifuge tube. Centrifugation was done for 5 minutes. A spatula was used to scoop the clay into weighed crucibles and oven-dried at 105 ^{o}C overnight.

The mineralogy of the clay fraction was determined with the use of an X-ray diffractometer with Ni-filtered Cu-Kα radiation at 40 kV and 30 mA at a wavelength of 1.54 Å. The type of clay mineral in the sample was reflected on the diffraction peaks (Soil Survey Staff, 2014a).

Table 2: Forms of Fe and Al and their derivatives

Horizon	Depth	Dith Forms		Oxal. Forms		Cryst. Forms		Deg. of act.		Clay	Co-migration of	
		Fed	Ald	Feo	Alo	Fed-Feo	Ald-Alo	Feo/Fed	Alo/Ald		Fed/clay	Ald/clay
	Cm	← g/kg →								g/kg		

Footnote: Fed: dithionite Fe, Ald: dithionite Al, Feo: oxalate Fe, Alo: oxalate Al, Fed-Feo: crystalline Fe, Ald-Alo: crystalline Al, Feo/Fed: degree of activation of Fe, Alo/Ald: degree of activation of Al, clay/Fed: co-migration of Fed, clay/Ald: co-migration of Ald

3.8 Soil Classification

Soil classification was based on the climatic information as well as the morphological, physical and chemical as well as clay mineralogy of the soils. The soils were classified according to the requirements of the United States Department of Agriculture (USDA) System of soil classification (Soil Survey Staff, 2014b) and correlated with the World Reference Base (WRB) for Soil Resources System (FAO/WRB, 2014). The USDA system of soil classification was done at the family level of classification, while the WRB for Soil Resources System was carried out at the second level of classification.

3.9 Soil Chemical Indices of Weathering

Results obtained from X-ray fluorescence (XRF) were used in the total elemental composition of soil samples. Considering that the major chemical effects of the weathering processes are the leaching of the mobile components (Ca, Na, Mg and K) and the concentrations of the residual Al and Fe. Indices of soil chemical weathering used in the study are presented in Table 3.

3.10 Land Use Recommendation using Land Capability Classification System

According to Ezeaku (2011), land capability classification is assessed by comparing the characteristics of soil mapping units with critical limits set for each capability class. Critical limits as reported by Sinclair and Dobos (2006) and Sys *et al.* (1991), Dobos (2006) and Lynch *et al*. (2009) were used in the land capability class category and presented in Table 4.

According to Lynch *et al*. (2009), Subclass allocation priority when more than one kind of limitation is considered indicates that erodibility is given a preferential consideration as compared to wetness, soil and climate limitations:

e → w → s → c

erodibility excessive wetness rooting zone limitations climate

Land suitability evaluation was thereafter done for oil palm cultivation. Oil palm is cultivated across the state, hence its choice. Land suitability evaluation was done for tracts of land that were classified in the land capability classes I to IV. This was done to further strengthen land use recommendations.

Table 3: Summary of weathering indices with weathering limits

Indices	Optimum fresh Value	Optimum weathered value	Ideal trend of index up profile (increase in weathering)	Formula	References
R	>10	0	Negative	$R = SiO_2/Al_2O_3$	Ruxton, (1968)
WI	10	0	Negative	$\{(CaO+MgO+K_2O+Na_2O/ CaO+MgO+K_2O+Na_2O+Si_2 O_3+Al_2O_3+Fe_2O_3) * 100\}$	Evans and Cameron, (1979)
MR	10	0	Negative	$SiO_2:R_2O_3$	Bera *et al*., (2015)

R: Ruxton index, WI: Weathering index, MR: Molar ratio

Table 4: Guide for Classifying Soils into Land Capability Classes

	Land capability class – Degree of limitations, restrictions or hazards							
Soil properties	I	II	III	IV	V	VI	VII	VIII
Physical soil conditions (s)								
**Soil depth (cm)	≥120	80-120	50-80	20-50	<20	-	-	-
Surface texture	Sl,fsl,vfsl,l, sil,scl,cl,sicl,(non arenic–ls,lfs, fs)	Ls,lfs,fs,sic,sc,c(<60%clay)muck,mucky peat	c (≥60% clay)	Cs	Same criteria as class I	As in class II	As in class III	Not class determining
Rock fragments (surface)	<15%	≥15-<35%	≥35-<60%	≥35-<60%	≥15%	Not class determining		
Wetness (w)								
**Drainage	Well or moderately well(5YR and redder)	Moderately well or somewhat poor(mottles between 80-100 cm)	Imperfectly drained(mottles between 40-80 cm)	Poorly drained (mottling between 0-40 cm)	Not class determining			
Flooding	None during growing season. Crop selection not restricted	Rare-Occassional. Slight crop damage. 0 to < 20% yield reduction or crop selection slightly affected	Occasional. Moderate crop damage. ≥20-<35% yield reduction or crop selection moderately affected	Frequent. Severe crop damage. ≥35-50% yield reduction or crop selection severely affected	Class I if overcome and protected from flooding	Class II and III if overcome	Class IV if overcome	Tidal flats
Fertility (f)								
Reaction (pH)	Favourable: easy to modify		Unfavourable: high lime or difficult to modify	Unfavourable: very difficult to modify	Not generally class – determining			Cat clays; unfavourable reaction; impractical to modify
***BS (%)	>80	60-80	40-60	20-40	<20	-	-	-
Climate								
Precipitation effectiveness (mm/annum)	≥1100	≥780-<1100	≥630-<780	≥480-<630	Not class determining	≥250-<480	<250	Not class determining
Cumulative days dry in moisture control section	<135 – Udic or Udic Ustic	≥135-<180 – Typic Ustic	≥180-<220 Aridic Ustic	≥180-<220 Aridic Ustic	Not class determining	≥220-<270 – Ustic Aridic	≥270 Typic Aridic	≥270
Erosion (e)								
*slope(degrees)	0-<7	0-≤7	0-≤15	0-≤20	0-≤25	0-≤35	0->35	0->35
**Erosion	Low	Susceptible to erosion	Susceptible to erosion	Susceptible to erosion	-	Erosion	erosion	Erosion hazard

Foot note: Ls: loamy sand, lfs: loamy fine sand, fs: fine sand, Sic: silty clay, Sc: sandy clay, c: clay, Sl: sandy loam, fsl: fine sandy loam, vfsl: very fine sandy loam, l: loam, sil: silty loam, scl: sandy clay loam, cl: clay loam, sicl: silty clay loam

Source: Sinclair and Dobos (2006). *Lynch *et al*., 2009, **Sys *et al*., 1991***Holland *et al*. (1989)

3.11 Procedure for Land Suitability Evaluation for Oil palm Production

Among the various cash crops grown in CRS, oil palm uniquely stands out. It is grown across the state, from the northern to southern agricultural zones. In addition, oil palm has non of its part as waste before or after harvesting and processing.

3.11.1 Land Evaluation Procedures

The eight mapping units were assessed for their suitability for oil palm cultivation according to Sys *et al*. (1993) (Table 5). The pedons were placed in suitability classes by comparing the primary and secondary data to oil palm requirement information (Table 5). The most limiting factor was assumed to determine the overall suitability ratings in accordance with Liebig's law of minimum. For the parametric (square root method) method, each limiting characteristic was rated as shown in Table 5. The index of productivity (IP) for each pedon was computed using the equation:

IP = A √B/100 * C/100 …. E/100

Where;

A = Overall lowest characteristic rating;

B, C … E = The lowest characteristic rating for each land quality group except the land quality group of which A characteristic is a member..

For the purpose of this research, the land quality groups; climate (c), topography (t), wetness (w), physical characteristics (s), soil fertility (f) and alkalinity (n) were used. Since there are often strong correlations within a land quality group, only one member with the least rating in each of c, t, w, s, f and n was used to calculate the index of productivity.

Index of productivity was calculated for current or actual and potential productivity. In potential productivity, properties that are easily altered by soil management procedures were masked. In this case; soil pH, BS and organic carbon were masked. For instance, the index of productivities of MF1P2 were calculated thus;

IPc = f √c/100 * t/100 * w/100* s/100* n/100
IPc = 58 √100/100 * 95/100 *100/100*96/100* 100/100 = 55.4

IPp = f √c/100 * t/100 * w/100* s/100* n/100
IPp = 95 √100/100 * 100/100 *96/100*100/100* 100/100 = 93.1

Where;

IPp = Potential index of productivity. Where BS, pH (H_2O) and organic carbon are not part of the "f" group since they are easily altered.

IPc = Current or actual index of productivity. In this case, BS, pH (H_2O) and organic carbon are part of the "f" group.

Table 5: Land use requirements for suitability classes for oil palm cultivation (limitation-parametric method of evaluation)

	Class, degree of limitation and rating scale					
Land	S1		S2	S3	N1	N2
Characteristics	0	1	2	3	4	
	100-95	95-85	85-60	60-40	40-25	25-0
Suitability to land use:	High		moderate	marginal	currently	permanently
Climate (c)						
MAR(mm/yr)	>2000	2000-1700	1700-1450	1450-1250	-	<1250
MAT (°C)	>25	25-22	22-20	20-18	-	<18
Topography (t)						
Slope (%)	0-4	4-8	8-16	16-30	>30	-
Wetness (w)						
Flooding	Fo (No flooding limitation)	Fo (No flooding limitation)	F1 Slight limitation; no longer than 1-2 mths	F2 Moderate limitation; 2-3mths of flood in every 5-10 yrs	-	F3 Severe limitation; Every year, 2-4 mths of flood.
Drainage	imperfect	Moderate well	Moderate well	Poor; aeric	Poor; drainable	V. Poor; not drainable
Physical characteristics (s)						
Texture	L,SCL,CL	CL, SCL, L	C, SCL	SCL-LfS	-	C, CS
CF (Vol. %)	0-15	15-35	35-55	55-75	-	>75
Soil depth (cm)	>150	150-120	120-100	100-80	-	<80
Soil fertility (f)						
CEC (cmol/kg)	>10	8-10	6-8	<6	-	-
BS (%)	>35	20-35	>20	-	-	-
pH (H_2O)	7.2-7.0, 7.2-7.5	7.0-6.2, 7.5-8.0	6.2-5.8, 8.0-8.2	5.8-5.5, 8.2-8.5	<5.5	>8.5
Mg:K	>3.5	>3.5	2.0-3.5	1.0-2.0	-	-
Org. C (%)	>1.5	1.5-0.8	0.8-0.4	<0.4	-	-
Alkalinity (n)						
ESP (%)	0-15	15-25	25-35	35-45	-	>45

MAR = mean annual rainfall, MAT = mean annual temperature, CF = coarse fragment, CEC = Cation exchange capacity, BS = Base saturation, ESP = exchangeable sodium percent

Source: Sys *et al.* (1993);

3.12 Data analysis

1. Data was subjected to statistical analysis using STATISTICA version 13.3.

2. Descriptive statistics of mean, range and coefficient of variability were used to compare data between elevation ranges and limestone formations.

3. Correlation analysis was used to establish the relationship between soil properties.

4. The most influential properties causing soil variation between limestone formations were identified by principal component analysis

CHAPTER FOUR
RESULTS AND DISCUSSION

4.1 Morphological Properties of the Soils

Morphological properties of the soils overlying SLS in the Northern agricultural zone (IH1 and IH2), SL in the central agricultural zone (AI1, AI2 and AI3) and SLSi in the southern agricultural zone (MF1, MF2 and MF3) are presented in Tables 6-13.

4.1.1 Soil Horizon and Horizonation

In the soils of SLS in the Northern agricultural zone (Table 6), the dominant epipedon in soil mapping unit IH1 was the plough layer which was represented by Ap horizon, while the transition AB and CB horizons were typical in the endopedons. However, pedon IH1P2 had well-developed B and C horizons with pronounced illuviation of clay as cutans, resulting in Bt and Ckt horizons with the marked occurrence of clay and Fe cutans. Also, a thin layer of a somewhat whitish material representing carbonate accumulation occurred concurrently with illuviated clay. Hence, the occurrence of Ckt horizon designation (Plate 2). The sequence of horizons in IH1P1 and IH1P2 seemed to be different, as IH1P1 was represented by Ap, AB, CBk and CB, while IH1P2 was represented by Ap, Bt and Ckt horizon sequences. It is evident from the sequences that IH1P2 may have had more periods for pedogenesis that may have resulted in Bt and Ckt horizons, while IH1P1 was relatively younger with the dominance of transition horizons. Similar results were obtained by Girao *et al.* (2014) and Irmak *et al.* (2007), who reported that Ck horizon indicates the accumulation of carbonates, while Buringh (1979), Buol *et al.* (1980) and Dinc *et al*. (1987) emphasized that cambic B horizons are developed alongside calcic C horizons in limestone soils of semi-arid regions. Calcic horizons are often traced to arid and semi-arid regions but develop as an illuvial horizon in carbonate-rich soils (Rabenhorst *et al*., 1991; Khormali *et al*., 2003).

The poorly drained soils in mapping unit IH2 (Table 7) were also dominated by the plough layer (Ap horizon) in the epipedons, while the endopedons were dominated by transition BC horizons and Cg horizons in both pedons. The presence of Cg horizon is an indication of gleying, due mainly to prolong soil wetting. However, transition BC horizon in both pedons was an indication of the possible presence of an Inceptisol. Gleyed horizons as indicated by colour are clear signs of gleization which may have occurred as a result of the continuos inundation of the soils with water, filling the macro and micro pores of the soils. Apparently, the soils were used for the cultivation of paddy rice. Horizon sequence in this

Table 6: Morphological Properties of Soils in Mapping Unit IH1

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
			IH1P1 (64 m ASL)					
Ap	0 – 37	Dark reddish brown (5YR 3/3)	-	scl	2m sbk	wss; mfr	cs	Many medium pores; common medium and coarse roots; common medium Ants and Worms; many Fe and Mn concretions
AB	37 – 64	Dark red (10R 3/6)	c,f-m, d (Very pale brown) (10YR 7/3)	scl	2m sbk	wss; mfi	cs	Few moderate clay and Fe cutans on ped faces; common medium pores; common medium roots; few Ants; few Fe and Mn concretions
CBk	64 – 130	Brown (7.5YR 5/3)	m, c, d (gray) (5Y 6/1)	scl	2m sbk	ws; mfi	cw	Very few thin clay cutans on ped faces; few very fine pores; few to common fine roots; few Mn nodules;
CB	130 – 195	Very dark Grayish brown (10YR 3/2)	f, f-m, f (gray) (5Y 6/1)	sl	2m gr	wss; mfr	-	Few and common fine pores; evidences of $Fe^{3+}$and Mn concretions
			IH1P2 (65 m ASL)					
Ap	0 – 42	Black (5YR 2.5/1)	-	sl	1m gr	wss; mfr	cw	many medium pores; many coarse roots; many Ants; few Fe nodules.
Bt	42 – 70	Light brown (7.5YR 6/4)	-	scl	2m sbk	wss; mfr	gw	Few thin clay and Fe cutans on ped faces; common fine pores; very fine and medium roots; very few Ants; many Fe nodules.
Ckt	70 – 150	Gray (7.5YR 5/1)	c, m, d (reddish yellow) (7.5YR 6/8)	scl	2c abk	ws; mfi		Very few thin clay cutans on ped faces; few very fine pores; thin layer of weathered $CaCO_3$ crossing the layer at mid-way.

Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

Table 7: Morphological properties of soils in Mapping Unit IH2

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
			IH2P1 (54 m ASL)					
Ap	0 – 26	Dark brown (7.5YR 3/2)	-	sl	2m cr	wss; mfr	cs	Common fine and medium pores; many fine roots; many Worms and Ants.
BC	37 – 84	Gray (2.5Y 6/1)	m, m/c, p (brown) (7.5YR 4/4)	scl	2m sbk	ws; mfi	gi	Few fine pores; few very fine roots; few Ants; few fine charcoal; very few thin clay cutans on ped faces
Cg	84 – 145	Gray (2.5YR 6/1)	c, m, d (brown) (7.5YR 4/4)	scl	massive	wvs; mvfi		Few Fe nodules; common fine charcoal
			IH2P2 (51 m ASL)					
Ap	0 – 13	Dark brown (7.5YR 3/2)	-	sl	1m cr	wss; mvfr	cs	Many medium pores; many medium and coarse roots; many Ants and few earth Worms.
BC	13 – 46	Gray (2.5Y 5/1)	c, f- m, d (red) (2.5YR 4/8)	scl	2m sbk	ws; mfi	cw	Very few thin clay cutans in pores; few fine to medium pores; common fine and medium roots; few Earth worms
Cg	46 – 87	Very dark greenish gray (Gley1 3/1)	F, f- m, d (red) (2.5YR 4/8)	sl	Massive	ws; mfi		

Foot note: Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

Table 8: Morphological properties of soils in Mapping Unit AI1

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
			AI1P1 (93 m ASL)					
Ap	0 – 9	Dark brown (7.5YR 3/2)	-	ls	1f gr	wns; ml	cs	Many medium and coarse pores; many fine and medium roots; few fine and medium Ants
BA	9 – 54	Dark yellowish brown (10YR 4/6)	-	sl	1m sbk	wss; mfr	gs	Common fine and medium pores; many coarse roots; few Ants; cracks extending to the Bt horizon
Bt	54 – 122	Yellowish red (5YR 4/6)	F, m, f, light yellowish brown (2.5Y 6/4)	scl	2m sbk	wss; mfr	cw	Few thin clay cutans on ped faces; common fine pores; few very fine roots; many medium Worm casts
Ckt	122 – 200	Yellowish red (5YR 4/6)	M, m/c, p light yellowish brown (2.5Y 6/4)	scl	2m sbk	ws; mfr	-	Few thin clay cutans on ped faces; few very fine pores; few fine roots; many Mn-Fe concretions with quartz.
			AI1P2 (106 m ASL)					
Ap	0 – 15	Very dark gray (10YR 3/1)		ls	1f gr	wns; ml	cs	common to many medium and coarse pores; many fine to coarse roots; many Ants. Many fine pores; many medium and coarse roots; few Ants and Worm casts.
AB	15 – 67	Brown (7.5YR 5/4)	-	sl	1m gr/sbk	wss; mvfr	cs	Few thin clay cutans on ped faces; few very fine pores; few fine roots; common quartz; few Ants; common cracks.
Btv	67 – 146	Very pale brown (10YR 7/3)	M, m/c d, dark red (7.5R 3/8)	scl	2m abk	ws; mfi	cw	Few thin clay cutans on ped faces; few very fine pores; many Fe nodules and quartz; common to many charcoal.
Ckt	146 - 180	Light brown (7.5YR 6/4)	C, f, f, gray (2.5Y 6/1)	scl	2m sbk	ws; mfi		

Foot note: Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

Table 9: Morphological properties of soils in Mapping Unit AI2

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
			AI2P1 (70 m ASL)					
Ap	0 – 12	Brown (7.5YR 4/2)	-	ls	2m gr	wns; ml	cs	Many fine and medium pores; many fine and medium roots.
Bt	12 – 59	Reddish brown (5YR 4/4)	-	sl	1m sbk	wss; mvfr	gw	Common fine and coarse pores; few fine and coarse roots; few Ants and Worm casts; common cracks in the horizon.
BC	59 – 120	Yellowish brown (10YR 5/4)	F, f, d, red (10R 4/8)	sl	2m sbk	wss; mfi		Common moderate clay cutans on ped faces; few fine to medium pores; few very fine roots; common Fe-Mn concretions.
			AI2P2 (80 m ASL)					
Ap	0 – 13	Brown (7.5YR 4/2)		sl	2m gr	wss; m(l-vfr)	cs	Many medium and coarse pores; many medium and coarse roots; many Ants.
AB	13 – 60	Yellowish red (5YR 4/6)	C, m, d, red (2.5YR 4/6)	sl	2/3m sbk	wss; mfr	cs	Common moderate clay and Fe cutans on ped faces; common fine to medium pores; common fine roots; few Ants; few quartz and Fe nodules.
BC	60 – 82	Brown (7.5YR 5/4)	F, f, f, brownish yellow (10YR 6/6)	sl	2m sbk	wss; mfr	cw	Few thin clay cutans on ped faces; common fine pores; few fine roots; few Fe nodules.
Ckt	82 – 132	Grayish brown (2.5Y 5/2)	F, f/m, d, yellowish red (5YR 5/6)	scl	3c sbk	wvs; mfi	-	Few thin clay cutans on ped faces; few medium pores; few Fe-Mn concretions; many cracks between 1.0 and 1.7 cm wide.

Foot note: Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

Table 10: Morphological properties of soils in Mapping Unit AI3

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
AI3P1 (54 m ASL)								
Ap	0 – 21	Dark reddish brown (5YR 3/2)	-	sl	1f gr	wns; ml	gs	Many medium and coarse pores; many medium and coarse roots; many Ants and few beetles.
AB	21 – 80	Dark reddish gray (5YR 4/2)	F, f- m, distinct red (10R 4/8)	sl	1m sbk	wss; mvfr	-	Few fine pores; common fine and medium roots; very few Ants and Worm casts; few Fe nodules.
AI3P2 (62 m ASL)								
Ap	0 – 50	Dark gray (5YR 4/1)	-	ls	1f gr	wns; ml	cs	Many medium and coarse pores; many fine to medium roots; few Ants and Termites; many charcoal.
AB	50 – 94	Gray (5YR 5/1)	-	ls	1m gr	wns; ml	cw	Common medium pores; common medium roots; few Fe nodules.
Cg	94 – 124	Gray (7.5YR 5/1)	M, m, d, strong brown (7.5YR 5/8)	ls	1m sbk	wss; mvfr	-	Few fine pores; few very fine and medium roots; few Fe nodules.

Foot note: Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

Table 11: Morphological properties of soils in Mapping Unit MF1

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
			MF1P1 (58 m ASL)					
Ap	0 – 17	Very dark gray (7.5YR 3/1)	-	sl	2m sbk	wss; mfr	cs	Few to common medium and coarse pores; few fine roots; Termite cast and few Ants.
Bt	17 – 56	Light yellowish brown (10YR 6/4)	-	sl	2m sbk	wvs; mvfi	cw	Very few thin clay cutans on ped faces; many coarse pores; few fine roots; wide cracks of about 3-7 cm.
BC	56 – 121	Pale brown (10YR 6/3)	C, c, d, gray (2.5Y 6/1)	sl	3m/c abk	wvs; mvfi		Common thin clay cutans on ped faces; common to many medium and coarse pores; few Ants; many coarse boulders of limestone covering about 75 % of pit.
			MF1P2 (70 m ASL)					
Ap	0 – 14	Brown (7.5YR 4/2)		ls	1f gr	wns; ml	cs	Many medium and coarse pores; many fine and medium roots; few charcoal; thick and closely interwoven fine to medium roots.
AB	14 – 43	Reddish brown (5YR 4/3)		ls	1f-m sbk	wss; mvfr	gs	Many medium pores; many medium and coarse roots; few Earthworms and Termites.
Bt	43 – 81	Reddish brown (5YR 4/4)		sl	1-2f/m sbk	wss; mvfr	cw	Few thin clay cutans on ped faces; few medium pores; few fine and coarse roots; few charcoal; many boulders; few plinthites.
Crv	81 – 153	Red (10R 4/6)	Few fine and medium distinct brownish yellow (10YR 6/6)	sl	2f/m sbk	ws; mfi		Very few thin clay cutans on ped faces; few fine pores; rock covering over 50 % of the surface with plinthites and hard pan at this depth.

Foot note: Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

Table 12: Morphological properties of soils in Mapping Unit MF2

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
MF2P1 (44 m ASL)								
Ap	0 – 16	Very dark gray (7.5YR 3/1)	-	Sl	1f sbk	wss; mfr	cs	Many medium and coarse pores; many fine and medium roots; many Ants; few Termites; few charcoal.
BC	16 – 44	Brown (7.5YR 5/3)	-	sl	1m sbk	wss; mfr	gw	Common fine and medium pores; common medium roots; few Fe nodules; common Termite cast and few Ants; few charcoal.
CB	44 – 97	Gray (7.5YR 5/1)	C, m, d, red (2.5YR 5/8)	sl	1,2m pl	ws; mfi	gs	Few fine and medium pores; few very fine roots.
Cr	97 - 161	Gray (7.5YR 5/1)	C, m, d, red (2.5YR 5/8) and reddish black (2.5YR 2.5/1)	sl	2m pl	ws; mfi		Common fine and medium pores, few Fe concretions; duripan.
MF2P2 (43 m ASL)								
Ap	0 – 16	Very dark gray (7.5YR 3/1)		ls	1f gr	wns; ml	cs	Common fine and medium pores; many fine, medium and coarse roots; many large Ants; few fine quartz grains.
AB	16 – 45	Brown (7.5YR 4/3)		ls	1f/m sbk	wss; mvfr	cw	Common fine pores; common coarse roots; few large Ants; few charcoal.
Bt	45 – 107	Brown (7.5YR 4/3)		sl	2m sbk	wss; mfr	gw	Few thin clay cutans on ped faces; few very fine pores; few very fine roots; many fine to medium quartz grains; many medium charcoal with boulders of sandstone outcrops.
BC	107 – 200	Strong brown (7.5YR 5/8)		sl	2m sbk	wss; mfr		Few thin clay cutans on ped faces; few very fine pores; few fine and coarse roots; very few charcoal with cobbles of sandstone.

Foot note: Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

Table 13: Morphological properties of soils in Mapping Unit MF3

Horizon	Depth (cm)	Munsell Colour (moist)	Mottles	Tex.	Struc.	Consist.	Bound.	Other characteristics
			MF3P1 (18 m ASL)					
Ap	0 – 15	Dark gray (5YR 4/1)	C, m, d, red (2.5YR 4/8)	sl	2m sbk	ws; mfi	cs	Common fine and medium pores; common fine roots; few charcoal.
Cg	15 – 48	Yellowish brown (10YR 5/4)	C, m, d, strong brown (7.5YR 5/8)	sl	Massive	ws; mfi	-	Few very fine pores; few very fine roots; few charcoal.
			MF3P2 (23 m ASL)					
Ap	0 – 16	Black (5YR 2.5/1)	-	sl	1m gr	wss; mvfr	cs	Common medium pores; many fine and medium roots; many charcoal; highly ramified by fine and medium roots.
BC	16 – 28	Gray (5YR 5/1)	F, f, f, red (2.5YR 5/8)	sl	2m sbk	ws; mfi	cw	Common medium pores; common fine roots; few charcoal and many crabs.
Cg	28 – 49	Dark gray (5YR 4/1)	-	ls	Single grained	wns; ml	-	Few very fine pores; few fine and medium roots; few charcoal and many crabs.

Foot note: Text. = texture, gr = gravel, co = cobbles, l = loam, s = sand, c = clay; struc.= structure: 1,2,3 = weak, moderate, m, c = medium, coarse; gr, cr, sbk = granular, crumby and sub-angular blocky structure; consist.= consistence: w = wet, m = moist, ss = slightly sticky, fr = friable, fi = firm, v = very; bound.= boundary: cs = clear smooth, cw = clear wavy, gw = gradual wavy; gi = gradual irregular, ASL: above sea level

mapping unit was somewhat similar to those of Folmer (1998) who obtained A/Bw/C/R in a karst depression.

In the soils overlying SL in Central agricultural zone (Table 8), horizons and horizonation in soil mapping unit AI1 shows the dominance of Ap horizons in the epipedons, while the endopedons were mainly B horizons with modifications into Bt, Btv and transition BA horizons which all occurred at depths of 9 – 146 cm. Illuviation of clay however extended to the Cr horizon in AI1P2 giving rise to Ckt horizon at depth of 146-180 cm. Though modified, horizon sequence was somewhat regular in the soils and followed the order A, B and C horizons, with a resultant BA and AB horizons at 9-54 cm and 15-67 cm in AI1P1 and AI1P2, respectively.

Though with the presence of plough layer occurring at similar horizon thicknesses of 0-12 and 0-13 cm, in AI2P1 and AI2P2, respectively, mapping unit AI2 had dissimilar horizon sequences. In mapping unit AI2, AI2P1 had a sequence of Ap, Bt and BC, while AI2P2 had Ap, AB, BC and Ckt (Table 9) which shows that argillation had extended to depths of 82-132 cm in AI2P2 as against 12-59 cm in AI2P1, indicating that pedogenesis may have occurred over greater periods in AI2P2. Nwaka (1990) and Ayolagha (2001) in earlier studies reported that soil weathering state can be deduced from the kinds and sequences of horizons in soil profiles. Both pedons had pronounce B horizons with clear illuviation of clay, however AI2P2 had deposition of carbonates at 82-132 cm which gave rise to Ckt horizon. Furthermore, AI2P1 was impeded by large limestone cobbles which occupied a significant proportion of the digging surface. This further emphasizes that pedogenesis may have taken place for a longer period in AI2P2 compared to AI2P1. Similar results were obtained by Ferreira *et al*. (2016) in the well-drained profiles over karst in Apodi plateau, Brazil. According to Bautista *et al*. (2011), Luvisols occur in areas with strong evidence of lessivation and high rainfall, and show the accumulation of secondary Ca carbonates. Luvisol (Table 27) was obtained in AI2P2, while other soil units failed to meet its criteria in the WRB classification System.

Mapping unit AI3 was designated by transition B horizons in the endopedons; however, AI3P2 had Cg horizon at depth of 94-124 cm (Table 10). Such transition B horizons are clear indications of cambic B horizons and the soils are likely to be Inceptisols. Similar horizon sequence of Ap, AB and Cg seemed to have occurred in mapping unit AI3, except that AI3P1 had not been inundated for too long as to result in the Cg endopedon. Ferreira *et al*. (2016) obtained A/C horizon sequence in a poorly drained karst soil, while

Sedov *et al*. (2008) obtained A/Bg/C in soils in karst depressions. These reports appeared similar to results obtained in the present study.

In the soils overlying SLSi in Southern agricultural zone (Table 11), the horizon designations in mapping unit MF1 shows the dominant presence of Bt horizons in the endopedons and Ap in the epipedons. Transition AB horizon in MF1P2 occurred within a thin depth of 14-43 cm with plinthic horizon represented by Crv at 81-153 cm. The presence of Bt horizon designates maturity especially as regards the illuviation of clay materials. Also, MF1P2 seems to have been exposed to more periods of pedogenesis, which has resulted in the formation of Bt at a greater depth with a resultant Crv horizon which may have been formed with time in MF1P1 just as the argillic B horizon formed at a depth closer to the soil surface. Again, this may be due to the influence of surficial erosion as thin Ap horizons (0-16 cm) dominated the surface soils, while thick B horizons comprising BC and Bt horizons occurred in the subsurface with the C horizon occurring at depths of 44-161 cm in MF1P1. MF1P2 had quite an extensive Bt horizon occurring at 45-107 cm. Again, this is indicative of relative soil maturity especially in MF2P2 (Table 12) as there was no marked Bt horizon in MF1P1. Similar results were reported by Carroll and Hathaway (1953) in the Lenoir limestone soils and Afu *et al.* (2017) in the Odukpani limestone soils, who obtained A, transition B and C horizons.

A regular trend of horizon sequence occurred in all the pedons in MF3 with Ap and Cg horizons dominating the epipedons and endopedons, respectively (Table 13). However, transition BC horizon occurred at 16-28 cm in MF3P2, and indicates a likely occurrence of entisols in the mapping unit.

Except IH1P1 and MF2P1, and pedons in the poorly drained soils, all other studied soils had argillic B or C horizons with the presence of clay or Fe cutans. This is an indication that silicate clays may have been illuviated in the underlying horizons as a result of soil maturity. Consequently, soils in the poorly drained mapping units when compared to those in the well drained areas were adjudged young. Such soils are present in new areas where sediments are being deposited (Esu, 2010). Also, the absence of calcic horizons (Bk or Ck) in the soils may be as a result of dissolution and leaching of carbonates, however, calcic horizons are common in arid or semi arid environments but develops as an illuvial horizon in carbonate rich soils (Rabenhorst *et al*., 1991; Khormali *et al*., 2003).

4.1.2 Soil Depth

In the soils overlying SLS in Northern agricultural zone, soil mapping unit IH1 had thickness ranging from 37 to 42 cm in the epipedons with an overall depth of over 150 cm

(Table 6). Soils in mapping unit IH1 were rated as very deep (Osman, 2018; SSSD, 2017) as soil depth exceeded 150 cm. IH1P1 was not impeded by either high water table or bed rock, while IH1P2 was impeded by bed rock. Such deep soils are most likely to support the growth of most arable and cash crops. It will encourage the anchorage of even plants with deep tap root system. These depths exceeded those obtained by VandenBygaart and Protz (1994), Carroll and Hathaway (1953), Emadi *et al*. (2008) and Irmak *et al*. (2007) for similar soils elsewhere. Soils in mapping unit IH1 had thicker epipedons than those in IH2. Thick epipedons in the area may be due to very little surficial erosion occasioned by low rainfall in the southern guinea savannah area and nearly flat topography that does not support speedy removal of soil materials. VandenBygaart and Protz (1994), and Evans and Cameron (1979) reported increasing soil depth or thickness as a mark of progressive soil development, while Bautista *et al*. (2011) attributed increase in soil depth to increasing limestone rock dissolution. Also, Bt horizons seem to be absent in IH1P1, while in IH1P2 Bt horizon was quite thin and occurred at 42-70 cm of depth. This suggests that the soils are relatively young and are at their active stage of soil development. Also, the thick Ap horizons, especially in IH1P2 is a reflection of the land use type. The area has a long standing Teak plantation which has been left to fallow for a long period. Leaf fall and subsequent decay may have also contributed to the thickness of the Ap horizon. Moreso, a bedrock was encountered at 150 cm. Furthermore, the rate of formation of the A horizon may be faster than the rate at which sheet or surficial erosion occurs. According to Krasilnikov *et al*. (2007), soils on steep backslopes that have epipedons that are 30–50 cm thick, experience slower sheet erosion than A horizon formation.

Soil mapping unit IH2 had a range of thickness from 13 to 26 cm in the epipedon while water table was encountered at depths of 87 and 145 cm in IH2P2 and IH2P1, respectively (Table 7). Water impeded soils may cause limitation to crop growth, especially when wetting occurs at a prolong period. Such soils may most likely encourage the growth of water tolerant crops like paddy rice and sugar cane. Findings by Emadi *et al.* (2008) show that water table was encountered at depths of 100 and 130 cm for two soil profiles in the semi arid region of southern Iran.

In the soils overlying SL in Central agricultural zone, the thickness of epipedons ranged from 9 to 15 cm in mapping unit AI1 (Table 8), from 12 to 13 cm in mapping unit AI2 (Table 9) and, from 21 to 50 cm in mapping unit AI3 (Table 10). Soil depth varied generally from 120 in AI2P1 and 200 cm in AI1P1. Soil depth was generally rated as deep to very deep (SSSD, 2017; Osman, 2018) in the well-drained soils of the area. Soil depths in mapping unit

AI2 were impeded by limestone bedrock at 120 and 132 cm for AI2P1 and AI2P2, respectively. Such limitation may pose a challenge to the proliferation of tap rooted cash crops and may obstruct the permeability of water. Hence, resulting in poorly aerated soils. In their study, Ferreira *et al.* (2016) observed consolidated limestone deposit at 27 cm. There seem to be a variation in soil depth between the mapping units of soils overlying SL within the Central agricultural zone; such variations within a short distance may constitute variation in limestone rock (Bautista *et al*., 2011) which Girao *et al*. (2014) attributed to variation in soil depth from 1.9 m to over 4.3 m within a horizontal distance of 70 m.

Mapping unit AI3 was characterized by the occurrence of water table at 80 and 124 cm for AI3P1 and AI3P2, respectively (Table 10) and were rated as moderately deep to deep (SSSD, 2017). Soils in mapping unit AI3 obviously had thicker Ap horizons which reflect the steep nature of slope in the upland area with the occurrence of surficial erosion which may have eroded surface soils down the upland and deposited same in the low lying and poorly drained AI3. This process may have resulted in thicker Ap horizons in AI3. This contradicts findings in the soils over SLS in northern agricultural zone which were poorly affected by surficial erosion. Deeper depths of 100 and 130 cm to water table were however obtained by Emadi *et al*. (2008) in the semi arid region of Iran.

In the soils overlying SLSi in Southern agricultural zone, the thickness of epipedons ranged from 14 and 17 cm for MF1 and MF2 (Tables 11 and 12), while poorly drained MF3 had thickness ranging from 15 and 16 cm (Table13) which happens to be within the range of soils in MF1 and MF2. Soil depth ranged from 121 cm in MF1P1 impeded by limestone bedrock through 153 cm in MF1P2 impeded by hard pan to 200 cm in MF2P2, for the well-drained soils. The soils were rated deep to very deep on the scale of SSSD (2017). Such roots and water limiting layers observed in MF1P1 and MF1P2 are likely to act as a bareer to the penetration of roots and permeability of water and may result in soil mottling. The poorly drained MF3 soils were less than 50 cm deep and described as shallow on the scale of SSSD (2017). Sedov *et al.* (2008) obtained a comparatively deeper soil in the poorly drained soils than those in the uplands. Such soils tend to limit the kind of crops they support to a few.

The rainfall amount and temperature of the soils over SL in Central and that over SLSi in Southern agricultural zone are similar and the areas are together found within the tropical rainforest vegetation. For instance, in the Central and Southern agricultural zones, rainfall and temperature have ranges of 1760.3-2683.8 mm/annum and 22.56-31.89 ^{o}C, while the South has 2109.0-3770.8 mm/annum and 23.53-31.95 ^{o}C, respectively. Soils overlying SLS in the Northern agricultural zone experience relatively lower amount of rainfall (1251.4-

3347.8 mm/annum) but higher temperature (22.96-33.75 ^{o}C) and described as Southern guinea savannah. From the foregoing, the thicknesses of epipedons of soils in the high elevations of SLS in Northern agricultural zone appear to be thicker (37-42 cm) than those in other limestone formations (9-17 cm). This variation in the thickness of epipedons may also be linked to surficial erosion occasioned by high rainfall in the Central and Southern agricultural zones. It is also not surprising to have the thickness of the epipedons in the poorly drained mapping units AI3 and MF3 of the Central and Southern zones, respectively greater (15 – 50 cm) than those in the Northern zone (13-26 cm). The surficially eroded soils in the high elevations may have been deposited in the low lying elevation ranges in the poorly drained areas (Figs. 6-11). This deposition process is more in the Central and Southern zones, and has resulted to greater depositions (Tables 10 and 13) compared to those in the Northern zone (Table 7). The deposition of such sediments may be ascribed to colluviation and subsequent deposition of eroded materials from areas of higher elevation ranges *via* sheet erosion (Krasilnikov *et al*, 2007). Singleton and Lavkulich (1987) reported that soil horizons along a chronosequence became progressively deeper and more differentiated with soil age, while VandenBygaart and Protz (1994) stated that, Fe and soil organic matter were likely transported to the B horizons where they were immobilized as coatings or cutans on sand grains in the limestone soils.

4.1.3 Soil Colour

Matrix colour in the soils overlying SLS in the Northern agricultural zone ranged from black (5YR 2.5/1) and dark reddish brown (5YR 3/3) in the surface soils of mapping unit IH1 (Table 6). The subsurface soils were mainly modified from brown and gray with dominant hue of 7.5YR and value varying from 3 to 6, while chroma ranged from 2 to 6. Black soils indicate a significant presence of organic matter, while dark reddish brown signifies a combination of humified organic matter and oxidized Fe (Nsor and Ibanga, 2006) in the surface soils. Gray subsurface soil colours indicate impeded drainage or that the soils are frequently saturated with water (Nsor and Ibanga, 2006) by high clay content or bedrock. Very pale brown, gray and reddish yellow mottles occurred at depths of 37-64, 64-195 and 70-150 cm, respectively. Reddish yellow may indicate a transition from unhydrated Fe oxide in a well-drained soil to hydrated Fe oxides in a somewhat moister condition than it is in red soils. Similarly, iron released from the weathering of the parent material in the dry season may have precipitated and given rise to pedogenic hematite (Yaalon, 1997). Reports by Irmak *et al*. (2007), and Carroll and Hathaway (1953) seem to be similar to those obtained in the present study, except that gray and light gray colours were obtained in the subsurface soils.

In mapping unit IH2, soil surface colour was dark brown (7.5YR 3/2) (Table 7); an indication that humified organic matter may be high in the soils. In the subsurface soils, matrix soil colour ranged from gray (2.5YR 6,5/1) to very greenish (Gley 1 3/1). Very greenish soils indicate that the soils have remained inundated for a long period. Mottles dominated the subsurface soils of this unit and ranged from few common to many fine to medium and coarse distinct brown (7.5YR 4/4) and red (2.5YR 4/8) mottles. Red mottles appeared in streaks and indicated the presence of Fe minerals; perhaps, Jarosites and pyritic minerals.

In the Central agricultural zone, soil colour ranged from very dark gray (10YR 3/1) to dark brown (7.5YR 3/2) in the surface soils of mapping unit AI1 (Table 8) and brown (7.5YR 4/2) in mapping unit AI2 (Table 9). In the subsurface soils of mapping unit AI1, matrix soil colour ranged from dark yellowish brown (10YR 4/6) through very pale brown (10YR 7/3) to yellowish red (5YR 4/6). Reddish soil colour may be associated with the presence of Fe oxides under well-drained soil conditions. In the subsurface soils of AI2, reddish brown (5YR 4/4) and yellowish brown to brown were obtained. Brown or dark surface soil colours indicate that the soils are likely to be high in organic matter, while ferromagnesian or manganiferrous minerals impart dark colours on subsurface soils. This accords earlier reports by SilvaNeto (2008), and Tardi and Nahon (1985) who attributed reddish soil colours in carbonate rich materials to rapid precipitation of oxides of iron occasioned by high pH which favours the precipitation of ferrihydrites; a precursor of hematite rather than goethite. Soils containing goethite are yellow or brown, and ferrihydritic soils are dark reddish-brown (Jaworska *et al*., 2016). Results obtained in this study contradicts that of Sedov *et al*. (2008) who obtained dark reddish brown to very dark grayish brown.

In the poorly drained soils of AI3, dark reddish brown (5YR 3/2) to dark gray (5YR 4/1) in the surface soils and dark reddish gray (5YR 4/2) to gray (5YR 5/1) in the subsurface soils were obtained (Table 10). Gray soil colours are indicators of unfavourable wetness conditions, while reddish colours suggest that the soils may contain iron coatings in the form of Jarosites and pyritic minerals (especially in such poorly drained environment), and concretions inherited from earlier pedogenesis (Yakubu and Ojanuga, 2013). Bigham *et al*. (1978) attributed redder soil colours to increase in the proportion of hematite to goethite; this probably was the situation especially in the well- drained soils of northern and central agricultural zones.

Mottles ranged from few to common and many, medium and coarse faint to distinct light yellowish brown (2.5Y 6/4) to dark red (7.5Y 3/8) and gray (2.5Y 6/1) mottles at depths

of 54-200 cm in mapping unit AI1 (Table 8). In mapping unit AI2, mottles occurred between depths of 13 and 132 cm and ranged from few to common, fine to medium faint and distinct red (10R, 2.5YR 4/8,6) to yellowish red (5YR 5/6) (Table 9). Mottles were obtained between 21 and 124 cm depths in mapping unit AI3 and ranged from few to many, fine to medium, distinct red (10R 4/8) and strong brown (7.5YR 5/8) (Table 10).

In the soils over SLSi in Southern agricultural zone, soil colour ranged from very dark gray (7.5YR 3/1) to brown (7.5YR 4/2) and black (5YR 2.5/1) in the surface soils of mapping unit MF1 (Table 11), while the subsurface soils varied from light yellowish brown (10YR 6/4) and red (10R 4/6). The surface soils of mapping unit MF2 were dominated by very dark gray (7.5YR 3/1) with subsurface matrix colour ranging from gray (7.5YR 5/1) through brown (7.5YR 4/3) to strong brown (7.5YR 5/8) (Table 12). These results are similar to those obtained elsewhere (Carroll and Hathaway, 1953; Irmak *et al*., 2007) for similar soils. Greyish soil colours indicate that the soils were frequently saturated with water or poorly drained due to bed rock impediment (Nsor and Ibanga, 2006).

Soil colour in the poorly drained mapping unit MF3 varied from dark gray (5YR 4/1) and black (5YR 2.5/1) in the surface soils and ranged from gray (5YR 5/1) and dark gray (5YR 4/1) to yellowish brown (10YR 5/4) in the subsurface soils (Table 13). Variation in soil colour may be as a result of a difference in the soils mineral origin or change in soil development.

Mottles in the soils overlying SLSi in Southern agricultural zone varied from few to common, fine to coarse distinct gray (2.5Y 6/1) and brownish yellow (10YR 6/6) mottles at depths of 56-153 cm for mapping unit MF1 (Table 12). In mapping unit MF2, mottles were present as common, medium, distinct red (2.5YR 5/8) and reddish black (2.5YR 2.5/1) (Table 12). The poorly drained mapping unit MF3 had common medium distinct red (2.5YR 4/8) mottles in the surface soils and few to common, fine to medium, faint and distinct strong brown (7.5YR 5/8) to red (2.5YR 5/8) mottles in the subsurface soils (Table 13). The presence of mottles in the soils could be due to prolonged periods of inadequate aeration occasioned by continuous inundation or reduction during periods of the year, leading to cycles of oxidation – reduction reactions (Esu, 2010). Moreover, the gray soil mottles may be attributed to prolong water logging.

4.1.4 Soil Texture

In the soils overlying shale, limestone and sandstone in northern agricultural zone, soil texture in mapping unit IH1 ranged from sandy loam to sandy clay loam in the surface soils and from sandy loam in the CB horizon of IH1P1 to sandy clay loam in other subsurface

horizons (Table 6). In the poorly drained soils of IH2, the surface soils had sandy loam textures, while sandy clay loam textural class dominated the subsurface soils. However, sandy loam was obtained in the Cg horizon of IH2P2 (Table 7). The entire soils seem to be relatively young as the translocation of clay, lessivation or obvious presence of clay cutans were not recorded except in IH1P2 where illuviation of clay occurred at 42-70 cm depth. The presence of transition horizons signifies cambic horizons, while Cg horizons that are manifestations of gleization occurred in the subsurface horizons. These results align with those of Afu *et al*. (2017) and Ferreira *et al*. (2016), and similar to those of Aksoy *et al*. (2009) but contradicts findings by Girao *et al*. (2014) and Emadi *et al*. (2008) who obtained clay loam, sandy clay, clay and loamy textures in soils of similar parent materials. An obvious situation in the study area is that, particle sizes do not seem to have moved significantly to poorly drained low lying areas (Tables 6 and 7) as textural classes remained as sandy loam and sandy clay loam. This was due to the nearly level landscape which does not permit sufficient surficial erosion.

In the soils of SL in Central agricultural zone, loamy sand textural class dominated the surface soils of mapping unit AI1 (Table 8), while the subsurface soils had textural classes that ranged from sandy loam in the transition AB and BA horizons to sandy clay loam in the B and C horizons. In soil mapping unit AI2, textural class ranged from loamy sand to sandy loam in the surface soils. In the subsurface soils, textural class was mainly sandy loam except in the Ckt horizon of AI2P2 where sandy clay loam was obtained (Table 9). Illuviation of clay seemed to have occurred in the Bt and Ckt horizons; however, the dominance of transition horizons possibly resulting from the sandy loam texture of the underlying horizons is an indication that the soils are young and may not have been exposed to prolonged pedogenesis or weathering. The dominance of sand in the soils may be linked to the contribution of sandstone to limestone parent material. Parent material is the main factor determining soil texture in southeast, Nigeria (Akamigbo and Asadu, 1983).

The poorly drained soils of mapping unit AI3 had soil texture that varied from sandy loam to loamy sand in the surface and subsurface soils (Table 10). These soils are young and do not show evidences of lessivation; hence the development of transition horizons. The result obtained in the well-drained soils of AI1 and AI2 are in tandem with those of Afu *et al*. (2017), Carroll and Hathaway (1953) and Ferreira *et al*. (2016) but in contrast to those of Girao *et al*. (2014), Aksoy *et al*. (2009) and Emadi *et al*. (2008). The later authors obtained clay loam, sandy clay, clay, and loamy textures.

In the soils of SLSi in Southern agricultural zone, soil textural class ranged from sandy loam to loamy sand in the surface and subsurface soils of MF1 (Table 11). This shows clear dominance of sand and indicates that sand may not have had sufficient time to weather into finer particle sizes like silt and clay and may be adjudged as being relatively young. In mapping unit MF2, sandy loam and loamy sand textural classes dominated the surface and subsurface soils with transition horizons such as BC, CB, AB and BC dominating the subsurface soils except at depth of 45-107 cm where Bt horizon was obtained in MF2P2 (Table 12). All these suggest that the soils are relatively young as only few thin clay cutans were obtained at depths of 45-200 cm in MF2P2. The poorly drained soils in MF3 had sandy loam textural class dominating the surface and subsurface soils except in the Cg horizon of MF3P2 where sand was obtained at 28-49 cm (Table 13). These soils are young as they are formed by seasonal deposition of eroded materials from areas of high elevation ranges (Figs. 10 and 11) through colluviation. Finer textural classes were obtained by Afu *et al*. (2017), Carroll and Hathaway (1953), Girao *et al*. (2017) and Emadi *et al*. (2008) from soils of similar materials. The dominance of finer particle sizes may infer advanced weathering.

4.1.5 Soil Structure

In the soils overlying SLS in Northern agricultural zone, soil structure was weak medium granular and moderate medium subangular blocky in the Ap horizons. In the endopedons, moderate medium subangular blocky and granular to moderate coarse subangular blocky structures were obtained in mapping unit IH1 (Table 6). This aligns with findings by Girao *et al*. (2014) and Sedov *et al*. (2008) and contradicts those of Emadi *et al*. (2008) and Ferreira *et al*. (2016) who rather obtained angular blocky structures. In IH2, soil structure was mainly weak medium and moderate medium crumby in the surface soils. The subsurface soils were dominantly represented by moderate medium subangular blocky and structureless massive soils (Table 7). Structureless massive was also reported by Irmak *et al.* (2007) and Emadi *et al*. (2008) for similar soils elsewhere; such soils are coherent and found in medium to fine textured soils (Esu, 2010). Moderate medium subangular blocky structures were obtained in horizons with increased clay which may have been responsible for soil aggregation, especially in the Ckt horizon of IH1P2. Structureless massive soil structures were obtained in the Cg horizons where clay seemed to be high in the poorly drained soils of IH2 as the soil particles were clumped in massive chunks (Plate 5).

In the soils of SL in Central agricultural zone, the structure of soils in mapping unit AI1 was mainly weak fine granular. This may be attributed to the relatively sandy textures, while that of the subsurface soils ranged from weak medium granular and subangular blocky

structures in the transition AB and BA horizons to moderate medium subangular blocky structure (Table 8). Blocky structures seem to have been obtained in the Bt and Ckt horizons mainly as a result of higher clay which may have been responsible for soil aggregation. In mapping unit AI2, the surface soils were dominated by moderate medium granular structures, while soil structures in the subsurface soils ranged from weak medium subangular blocky, moderate to strong, medium, coarse subangular to angular blocky structures (Table 9). These findings are similar to reports by Emadi *et al*. (2008), Ferreira *et al*. (2016), Girao *et al*. (2014) and Sedov *et al*. (2008), however, there seem to be a deviation in the grades and classes of soil structures. Granular structures in the surface soils are likely due to high organic matter content, while gradually increasing structural class and grade especially in the Ckt horizon were strong coarse subangular blocky. These may be as a result of increasing clay amount.

In the poorly drained soils of mapping unit AI3, weak fine granular structures dominated the surface soils, while subsurface soil structures ranged from weak medium granular to weak medium subangular blocky structures (Table 10). Soils in the poorly drained and low elevation areas may have been enriched by organic matter, perhaps due to the anaerobic environment which favours the accumulation of organic matter. This may have led to the formation of granular structures. Strong structural grade in the subsurface soils may be due to the increase in clay amount within the pedon (Yakubu and Ojanuga, 2013).

In the soils of SLSi in Southern agricultural zone, soil structures varied from weak fine granular to moderate medium subangular blocky in the surface soils of MF1, while the subsurface soils ranged from weak to moderate and strong, fine and medium to coarse subangular blocky and angular blocky structural types (Table 11). Mapping unit MF2 had weak fine granular and subangular blocky structures in the surface soils, while the subsurface structures had weak and moderate, fine and medium subangular blocky structure and weak to moderate medium platy structures (Table 12). Low clay amount may have been responsible for the weak and fine grades, and classes while the platy structures were likely due to the underlying shaley parent material, or due to leaching or compaction by grazing animals. Granular, subangular and angular blocky structures had been reported by Girao *et al*. (2014), Ferreira *et al*. (2016) and Sedov *et al.* (2008) for limestone soils in various parts of the world. Weak and fine structural grades and classes seem to have been obtained in the transition AB, BC and CB horizons; however, moderate medium subangular blocky and platy structures were obtained in the Bt and Cr horizons.

4.1.6 Soil Consistence

In the soils of SLS in Northern agricultural zone, soil consistence was slightly sticky (wet) and friable (moist) in the Ap horizons of IH1. The endopedons were dominated by slightly sticky to sticky (wet), and firm to friable (moist) consistences (Table 6). In mapping unit IH2, the top soils were slightly sticky (wet) and friable to very friable (moist), while the subsoils were sticky to very sticky (wet) and firm to very firm (moist) (Table 7). The consistences of the soils became more sticky with increasing clay amount especially in the subsurface soils. The stickiness observed may be attributed to the parent material. Findings in the present study are corroborated with those of Emadi *et al*. (2008), and Ferreira *et al*. (2016) who obtained friable and firm consistence in the surface soils and friable, firm and slightly hard in the subsurface soils, as well as very firm (moist) and non-plastic through slightly plastic and very plastic consistence.

In the soils of SL in Central agricultural zone, the surface soils of mapping unit AI1 were non sticky (wet) and loss (moist); this was as a result of the sandy texture of the soils. The subsurface soils ranged from slightly sticky to sticky (wet), and friable to very friable and firm (moist) (Table 8). Soil consistence in mapping unit AI2 was also, non-sticky (wet) in the sandy surface soil of AI2P1 and loss (moist), and slightly sticky (wet) and loss to very friable (moist) in the sandy loam surface soil of AI2P2. The subsurface soils were slightly sticky to very sticky (wet), and very friable, friable and firm (moist) (Table 9). These findings were similar to those of Emadi *et al*. (2008) for similar soils in Southern Iran and Ferreira *et al*. (2016) in the Apodi plateau. Poorly drained soils in AI3 were non sticky (wet) and loss (moist) in the surface soils, and non sticky to slightly sticky (wet) and loss to slightly friable (moist) in the subsurface soils (Table 10). Loss consistence was due to the non coherent nature of the soil arising from low clay and relatively low organic matter in the surface soils, while the friable consistence was due to the very weakly coherent nature of the soils; such soils may be ideal for the root growth of plants.

In the soils overlying SLSi in Southern agricultural zone, soils in mapping unit MF1 were non sticky and slightly sticky (wet) and loss to friable (moist) in consistence, while the subsurface soils were slightly sticky in MF1P2 to very sticky (wet) in MF1P1 and firm to very friable (moist) in consistence (Table 11). In mapping unit MF2, the surface soils were non sticky to slightly sticky (wet) and loss to friable (moist), while the subsurface soils were slightly sticky to sticky (wet) and very friable to firm (moist) (Table 12). These results are similar to those of Emadi *et al*. (2008) and Ferreira *et al*. (2016). Non sticky or loose consistences seem to have correlated with high sand content in the surface soils; such soils

exhibit minimal cohesive and adhesive properties (Esu, 2010). Poorly drained soils in MF3 were slightly sticky to sticky (wet) and firm to very friable (moist) in the surface soils, while the subsurface soils were non sticky to sticky (wet) and loss to firm (moist) (Table 13).

4.1.7 Horizon Boundary

In the soils overlying SLS in Northern agricultural zone, clear smooth boundary separated the Ap from AB horizon on the basis of higher gravels and the presence of very pale brown mottles in the AB horizon of IH1P1. Though the presence of mottles in CBk increased, it had less gravels compared to AB and was separated from the overlying AB horizon by clear smooth boundary (Table 6). The underlying CB was separated from CBk by the dominance of dark concretions which gave rise to very dark grayish brown matrix colour as compared to brown in CBk which obviously had gray mottles. In IH1P2, clear wavy boundary separated the Ap from the Bt horizon on the basis of black matrix colour and very coarse roots in Ap and light brown in Bt; however, the dominance of streaks and evidences of deposition of weathered $CaCO_3$ in the Ckt horizon as well as its coarse angular blocky structure separated it from the overlying Bt horizon.

In the poorly drained IH2 mapping unit, the Ap horizons were separated from BC horizon by clear smooth boundary on the basis of their black brown colour as compared to gray matrix colour in BC with an accompanying red mottled colour (Table 7). Consequently, the underlying gleyed Cg horizons were separated from the BC horizon by gradual irregular and clear wavy boundary on the basis of their massive soil structures with gray to very dark greenish gray colour.

In the soils of SL in Central agricultural zone, clear smooth boundary separated the Ap horizon from the BA horizon of AI1P1 on the basis of soil texture, structure and matrix colour (Table 8). The soil texture of Ap horizon was loamy sand with granular structure and dark brown colour against sandy loam, subangular blocky structure and dark yellowish brown colour in the BA horizon. Gradual smooth boundary separated the BA horizon from Bt horizon which had few thin clay cutans and yellowish red colours. Clear wavy boundary separated the Bt horizon from Ckt which had more of Fe concretions and quartz with many reddish mottles.

In AI1P2, the Ap horizon was separated from the AB horizon by clear smooth boundary on the basis of colour and consistence, while AB horizon was separated from the Btv horizon by clear smooth boundary based on the presence of mottles as well as plinthites in Btv horizon. The Btv and Ckt horizons were separated mainly by the presence of Fe nodules and quartz in the Ckt and the presence of plinthites in Btv by clear wavy boundary. In AI2P1, clear smooth

boundary separated the brownish Ap horizon from the reddish brown horizon based on the granular structural type in the Ap horizon and subangular blocky structural type in the Bt horizon (Table 9). However, gradual wavy boundary was used to distinguish between the Bt and BC horizons as the former had coarse pores and worm casts, while the later had Fe-Mn concretions.

In AI2P2, the Ap and AB horizons were separated by clear smooth boundary mainly by the presence of red mottles and yellowish red matrix colour in AB as compared to brown matrix colour in Ap horizon (Table 9). Also, the AB and BC horizons were separated by a clear smooth boundary mainly by brown matrix colour and brownish yellow mottles in BC horizon compared to the yellowish red matrix and red mottled colours in AB horizon. Between the BC and Ckt horizons was clear wavy boundary with the striking presence of Fe-Mn concretions, wide cracks, coarse rock fragments and grayish brown matrix colour in the Ckt horizon. In the poorly drained AI3, gradual smooth boundary separated the Ap and AB horizons in AI3P1 based on dark reddish brown colour in Ap against dark reddish gray in AB which also had distinct red mottles (Table 10). In AI3P2, clear smooth boundary separated dark gray Ap horizon from gray AB horizon; however, the subangular blocky soil structure in Cg as well as strong brown mottles were absent in the AB horizon.

In the soils of SLSi in Southern agricultural zone, clear smooth boundary separated the Ap horizon of MF1P1 from its Bt horizon based on the light yellowish brown matrix colour, clay cutans and many coarse pores and 3-7 cm wide cracks in the Bt horizon against very dark gray colour and termite casts in Ap horizon (Table 11). The BC horizon had pale brown matrix as well as gray mottled colours and coarse boulders which were absent in the Bt horizon and therefore, separated by clear wavy boundary. In MF1P2, clear smooth boundary separated Ap and AB horizons based on matrix colour, soil structure and presence of charcoal in the Ap horizon. However, gradual smooth boundary separated the AB horizon from the Bt horizon based on the sandy loam texture in Bt against loamy sand texture in the AB horizon. Clear wavy boundary separated the reddish brown Bt horizon which also had few charcoal from the reddish Crv horizon which had distinct presence of plinthites with rock occupying over 50 % of the profile pit surface as well as distinct brownish yellow mottles.

In MF2P1, very dark gray matrix soil colour with coarse roots in the Ap horizon separated it from BC horizon with clear smooth boundary, while the BC and CB horizons were separated by gradual wavy boundary mainly as a result of brown colour in the BC horizons against gray matrix colour as well as platy structures in the CB and Cr horizons (Table 12). Gradual smooth boundary distinguished the CB horizon from the Cr horizon

based on the presence of roots; which seemed to be few but very fine in the CB horizon and absent in the Cr horizon. In MF2P2, very dark gray matrix colour in the Ap horizon and brown in the AB horizon were separated by clear smooth boundary, while clear wavy boundary separated the AB horizon from the Bt horizon mainly by the presence of clay cutans and boulders of sandstone. However, gradual wavy boundary separated the Bt horizon from BC horizon mainly due to the presence of many medium charcoal in the Bt horizon.

In the poorly drained MF3P1, clear smooth boundary distinguished the Ap horizon from the Cg horizon based on matrix and munsell colours as well as the soil structural type (Table 13). The Ap horizon had dark gray matrix soil colour, red munsell colour and moderate medium subangular blocky structure, while Cg horizon had yellowish brown matrix colour, strong brown mottled colour and massive soil structure. In MF3P2, black matrix colour was obtained in the Ap horizon, while gray colour was recorded in the BC horizon. Both horizons were separated by clear smooth boundary, while BC and Cg horizons were separated by clear wavy horizons mainly due to red mottled colour in the BC horizon which was absent in the Cg horizon. Furthermore, single grained structure was obtained in the Cg horizon, while subangular blocky structure was obtained in the BC horizon.

4.1.8 Other Soil Characteristics

In the soils overlying SLS in Northern agricultural zone, there were many medium pores in the surface horizons, and common fine and medium to few very fine pores in the subsurface soils of mapping unit IH1 (Table 6). These soils are well aerated and drained and are likely to give way for oxidative processes which will result in a brightly coloured soil (Esu, 2010). In the poorly drained soils of mapping unit IH2, soil pores in the surface soils were many, medium, while common fine to medium pores were obtained in the subsurface soils (Table 7). Such poorly drained soils are poorly aerated and could encourage gleization in the area (Esu, 2010).

Very few thin and few moderate clay and Fe cutans on ped faces were present in the subsurface CBk, Bt and Ckt horizons of the soils in mapping unit IH1 (Table 6); however, only very few thin clay cutans in pores were recorded in the transition BC horizon of mapping unit IH2 (Table 7). The presence of cutans in the B or C horizons is a direct indication that argillic horizons are present and the soils are mature, while Fe and humus cutans may indicate the presence of spodic horizons (Esu, 2010).

In the surface soils of mapping units IH1 and IH2, plant roots were common to many, and medium to coarse, while in the subsurface soils, few to common, very fine to medium

roots were obtained. The presence of roots, especially in the subsurface soils indicate the absence of root restrictive impediments such as hard pans, saline layers, stone line etc.

Animal activities were dominantly represented by ants, worms and termites. In mapping unit IH1, common to many medium ants and worms were present in the surface soils, while very few to few ants occurred in the subsurface soils (Table 6). However, in the poorly drained soils of mapping unit IH2, ants were many, while worms were few in the surface and subsurface soils (Table 7). The presence of animals in their numbers or the occurrence of their activities is an indication of soil mixing; otherwise called faunal pedoturbation, a probable reason for the dominance of transition horizons (Ofem and Esu, 2015).

Many Fe and Mn concretions and few Fe nodules were observed in the surface soils of mapping unit IH1. The concretions and nodules decreased to few Fe and Mn concretions and few Mn or Fe nodules in the subsurface soils (Table 6). Climatic changes and low soil permeability between seasons may lead to redox reactions and the formation of Fe-Mn concretions (Schwertmann and Taylor, 1989). A thin layer of weathered chalky $CaCO_3$ crossing the Cr horizon of IH1P2 was observed. In the poorly drained mapping unit IH2, only few Fe nodules and charcoals were recorded in IH2P1 (Table 7). The occurrence of nodules and concretions in limestone soils is a common phenomenon (Yaalon, 2009) and varies with soil depth and horizontal distance (Alencar, 2002; Mota *et al*., 2007), while Oliveira *et al*. (2000) had earlier obtained Fe-Mn concretions in limestone soils. Jimenez-Millan and Nieto (2008) related the occurrence of Fe-Mn concretions to source material, while Schwertmann and Taylor (1989) attributed it to climate changes. This may not be out of place, as its presence decreased from the soils overlying SLS in the northern agricultural zone (Plate 1) to those overlying SLSi in the southern agricultural zone (Plate 16).

In the soils overlying SL in central agricultural zone, soil pores were common to many, medium to coarse pores in the surface soils of mapping unit AI1, while the subsurface soils had common to many, very fine to medium pores (Table 8). In mapping unit AI2, many fine to medium and coarse pores were recorded in the surface soils, while the subsurface soils were dominated by few to common, fine to medium pores (Table 9). In the poorly drained soils of AI3, many medium and coarse pores were obtained in the surface soils, while the subsurface soils had few to common, fine to medium pores (Table 10). Soil pores are of great significance in pedogenesis; aerated soils are brighter and better drained (Esu, 2010). Emadi *et al*. (2008) obtained many, fine to very fine pores in the surface soils and few to many and fine to very fine pores in the subsurface of carbonate rich soils in Southern Iran.

Few thin clay cutans on ped faces were obtained in the subsurface B horizons which were accompanied by cracks that mainly extended to the Bt horizons in mapping unit AI1 (Table 8). In mapping unit AI2, few to common thin to moderate clay cutans were obtained with accompanying cracks which extended increasingly to the Bt horizons (Table 9). However, cutans were generally not identified in soils of mapping unit AI3 (Table 10). In the Calcareous soils of Southern Iran, Emadi *et al*. (2000) obtained few thin clay cutans on ped faces of the Bt horizons, while Sedov *et al*. (2008) obtained thin clay cutans over soil aggregates in the tropical karst in Yucatan. Ballagh and Rung (1970) attributed continuous clay cutans on ped faces to clay accumulation, while Esu (2010) identified argillans in B horizon as good indicators of argillic horizons.

Plant roots were mainly many, fine to medium and coarse in the surface soils of mapping unit AI1, while the subsurface soils had few to many, fine to medium and coarse roots which decreased in size and presence with increasing depth of soil (Table 8). In mapping unit AI2, many fine to medium and coarse roots were obtained in the surface soils, while the subsurface soils had few, very fine and coarse roots (Table 9). Plant roots were many, fine to medium in the surface soils of mapping unit AI3, while the subsurface soils were dominated by few to common, fine to medium roots which drastically reduced in number and size as the soil water table was approached (Table 10).

Animal activities in the surface soils of mapping unit AI1 were dominated by few to many, fine to medium ants, while the subsurface soils were few in terms of ants but with common to many worm casts (Table 8). Soils in mapping unit AI2 had many ants in the surface, especially in AI2P2 with few ants and worm casts in the subsurface soils. However, few to many ants and, few termites and beetles were recorded in the surface soils of mapping unit AI3, while in the subsurface soils, very few ants and worm casts were recorded (Table 10). The presence and activities of animals in soils is of pedogenic significance as they aid soil mixing otherwise called faunal pedoturbation, leading to fewer horizon. Consequently, ants, termites and worms are of interest to pedologists during field description of soils.

Many Mn and Fe concretions with many Fe nodules were obtained at depths greater than 122 cm in the subsurface soils of mapping unit AI1 with many quartz and charcoal; however, cracks extended from the transitional horizons to the Bt horizons (Table 8). In mapping unit AI2, Fe-Mn concretions were recorded at depths greater than 59 cm, while Fe nodules with quartz occurred between 13 and 82 cm. Schwertmann and Taylor (1989) identified climatic changes as well as low soil permeability between seasons as facilitators of redox reactions and the formation of Fe-Mn concretions. Oliveira *et al*. (2000) had however

obtained the concretions in similar soils in Minas Gervais, while Hendricks (1991) reported their dominance in calcareous soils. Cracks extended from BC to Ckt horizons which measured about 1.0-1.7 cm in width (Table 9). In mapping unit AI3, concretions were not detected; however, Fe nodules occurred at depths of 21-80 cm and 50-124 cm in AI3P1 and AI3P2, respectively (Table 10). The occurrence of such wide cracks may be an indication that the soils have expanding clay minerals. The presence of smectites and occurrence of smectification were observed in similar soils by Irmak *et al*. (2007) and Yesilsoy (1994), while Petar *et al*. (2016) obtained mixed mineralogy with the presence of expanding 2:1 clay minerals like vermiculites and smectitites.

In the soils overlying SLSi in southern agricultural zone, few to common and many, medium and coarse pores were obtained in the surface soils of mapping unit MF1, while the subsurface soils had few to common and many, fine to medium coarse pores which decreased generally in size and number with an increase in soil depth (Table 11). A similar observation was made by Emadi *et al*. (2008) in the calcareous soils of Southern Iran. Soils in mapping unit MF2 had common to many, fine to coarse pores in the surface soils and very fine to medium, few to common pores in the subsurface soils (Table 12). The surface soils of poorly drained mapping unit MF3 had common, fine to medium pores, while the subsurface soils had few to common, very fine to medium pores (Table 14).

Very few to common, thin clay cutans were present on ped faces of the subsurface B horizons; while very few thin clay cutans were present in the Crv horizon of mapping unit MF1 (Table 11). In mapping unit MF2, few thin clay cutans occurred in the Bt horizons and were not detected in the poorly drained soils of MF3. Clay cutans were also recorded in the calcareous soils of southern Iran (Emadi *et al*. 2000) and in the tropical karst of Yucatan (Sedov *et al*. 2008), while Ballagh and Rung (1970) attributed its continuous occurrence on ped faces to clay accumulation.

Plant roots were few to many, fine to coarse in the surface soils of mapping unit MF1 with thick and closely interwoven fine to medium roots in MF1P2, while the subsurface soils had few to common, medium to coarse roots that extended to depths of about 81 cm in MF1P2 (Table 11). Soils in mapping unit MF2 however, recorded common to many, fine to medium and coarse roots in the surface soils while the subsurface soils had few to common, very fine to medium roots (Table 12). Soils in the low elevation areas were poorly drained and represented by mapping unit MF3. These soils recorded common to many, fine to medium roots (Table 13). Poorly drained soils in mapping unit MF3 recorded common to many, fine to medium roots in the soil surface, while the subsurface soils had few to

common, very fine and medium roots. According to Esu (2010), a large part of the roots of most plants are found in the upper soil horizons and may as well contribute to soil mixing (floral pedoturbation).

Few termite casts and ants were present in the surface soils of MF1 while few ants, worms and termites were present in the subsurface soils (Table 11). Many medium and large ants with few termites were recorded in the surface horizons of mapping unit MF2 and common termite casts with few large ants in the subsurface soils (Table 12). The subsurface soils in the poorly drained mapping unit MF3 however, were dominated by many crabs (Table 13). Few ant and termites in MF1 and MF2 may have been a probable reason for the fewer but distinct soil horizons in the area mainly because of reduced faunal pedoturbation.

In mapping unit MF1, sandstone rock layer covered over 50 % of the digging surface in MF1P2 with plinthites extending from depths of 43 cm through 153 cm and charcoal through the entire depth, while in MF1P1, boulders of limestone covered over 75 % of the digging surface with cracks of about 3-7 cm wide (Table 11). There were however, no concretions and nodules at the studied depths. In soil mapping unit MF2, few Fe nodules occurred at depths of 16-44 cm in MF2P1, while quartz with boulders of sandstone occurred through the depths of MF2P2 (Table 12). However, few to very few charcoal occurred in the entire depths of both pedons. Few to many charcoal fragments were present in the entire depth of the poorly drained soils of mapping unit MF3 (Table 13). According to Yakubu and Ojanuga (2013), soils with ironstone otherwise referred to as plinthites are relatively older and may have low weathering intensity.

4.2 Soil Physical Properties

Soil physical properties are discussed and separated into soils overlying SLS in the Northern agricultural zone, SL in the central agricultural zone and SLSi in the southern agricultural zone, respectively.

4.2.1 Particle Size Distribution

a. Clay: The distribution of clay in soils of SLS in Northern agricultural zone showed mean of 210 g/kg and range of 140-260 g/kg in the A horizons of mapping unit IH1, while the subsoils had mean of 236 g/kg and range of 200-280 g/kg with increase at the Bt horizon of IH1P2 and decrease with depth in IH1P1 (Table 14). The illuviation and subsequent development of argillic horizon in IH1P2 indicates relative maturity in the high elevation areas. In the poorly drained soils of mapping unit IH2 in the low elevations, clay content had a mean of 170 g/kg (160-180 g/kg) in the surface soils, while values in the subsurface soils had mean of 225 g/kg and range of 180-280 g/kg (Table 14). Similar results were obtained by

Table 14: Physical properties of the soils overlying shale, limestone and sandstone formations

Horizon	Depth	Grav	Particle size distribution			BD	K_{sat}	Porosity		
	Cm	%	Clay	Silt	Sand			Total	Air spac	Capillary
				g/kg		g/cm^3	cm/h		%	
IH1P1:										
Ap	0-37	48	280	120	600	1.45	78.18	65	21	43.9
AB	37-64	53	240	100	660	1.48	106.28	54.4	9.4	45
CBk	64-130	58	300	100	600	1.66	0.61	44.4	3.1	41.2
CB	130-195	57	200	120	680	1.61	9.16	49.5	14.5	35
IH1P2:										
Ap	0-42	50	140	140	720	1.54	40.31	46.4	2.3	44.1
Bt	42-70	71	220	200	580	1.66	9.77	45.7	5.8	39.9
Ckt	70-150	0	220	120	660	1.56	0.61	54.5	6.7	47.8
Mean Surface soils		49	210	130	660	1.50	59.25	55.7	11.7	44
Range Surface soils		48-50	140-260	120-140	600-720	1.45-1.54	40.31-78.18	46.4-65	2.3-21	43.9-44.1
Mean Subsurface soils		47.8	236	128	636	1.60	25.29	49.7	7.9	41.8
Range Subsurface soils		0-71	200-280	100-200	580-680	1.48-1.66	0.61-106.28	44.4-54.5	3.1-14.5	35-47.8
IH2P1:										
Ap	0-26	30	180	240	580	1.18	8.55	60.9	5.7	55.2
BC	26-84	0	220	60	720	1.63	0.61	48.1	5.5	42.6
Cg	84-145	20	280	200	520	1.56	70.85	47.3	5.4	41.9
IH2P2:										
Ap	0-13	22	160	110	730	0.71	256.54	69.5	13.1	56.4
BC	13-46	18	220	120	660	0.91	130.71	62.2	4.2	58
Cg	46-87	19	180	140	680	1.3	0.61	53	3.0	50
Mean Surface soils		26	170	175	655	0.95	132.55	65.2	9.4	55.8
Range Surface soils		22-30	160-180	110-240	580-730	0.71-1.18	8.55-256.54	60.9-69.5	5.7-13.1	55.2-56.4
Mean Subsurface soils		14	225	130	645	1.35	50.70	52.65	4.5	48.1
Range Subsurface soils		0-19	180-280	60-200	520-720	0.91-1.63	0.61-130.71	47.3-62.2	3.0-5.5	41.9-58

Foot note: BD: bulk density, Ksat: saturated hydraulic conductivity, Grav = Gravel, Capill = Capillary, Air P= Airfilled porosity

Aksoy *et al*. (2009) for Typic Xerochrepts in Bursa Province of Turkey, and Ferreira *et al*. (2016) in Apodi Plateau, Brazil. Results of particle size distribution by Petar *et al*. (2016) (261-780 g/kg), Carroll and Hathaway (1953) (171-617 g/kg), and Irmak *et al*. (2007) (303.3-536.2 g/kg) indicates higher values for clay, while VandenBygaart and Protz (1994) reported lower values (0.4-25.2 %) in the Pinery dune soils overlying limestone.

In the soils overlying SL in Central agricultural zone, the mean value of clay obtained was 110 g/kg with a range of 100-120 g/kg in the A horizons of mapping unit AI1, while values in the subsoils were 220 g/kg and range of 140-280 g/kg (Table 15). The distribution of clay with depth showed a steady increase in AI1P1, while a clear clay bulge was obtained at 67-146 cm in AI1P2. Eluviation-illuviation process seemed to have been very active in the high elevation soils resulting in clay accumulation in the B horizon, especially in AI1P2 whose depth was limited at 150 cm by a bed rock of limestone. This process may have resulted in the clear formation of argillic horizons with a corresponding presence of clay or Fe cutans in the subsurface soils in a process referred to as lessivation. In mapping unit AI2, mean clay content was 140 g/kg and range was 120-160 g/kg in the top soils, while subsurface soil values had mean of 176 g/kg with a range of 140-220 g/kg. Values increased with soil depth in AI2P1, while AI2P2 had an irregular distribution of clay with depth with clay illuviation at 82-132 cm (Table 15). Ferreira *et al*. (2016) attributed clay increase at the B horizon of such soils to clay flocculation resulting from the accumulation of secondary carbonates in the subsurface soils. The soils seem to be relatively young as they were dominated by transition horizons compared to IH1 in the soils of SLS. This is because the formation of argillic Bt horizons is a slow process and occurs over a long period (Krasilnikov *et al.*, 2007) and was identified as a determining factor in the genesis of soils in the karst region of Spain (Delgado *et al.* 2003). The process is often favoured by alternate wet and dry seasons and the presence of expanding clay minerals which give room for the movement of clay. Similar results were obtained by VandenBygaart and Protz (1994), while Carroll and Hathaway (1953) (17.1-61.7 %), Petar *et al.* (2016) (261-780 g/kg) and Aksoy *et al*. (2009) (214-515 g/kg) recorded relatively higher values. Such high values may results from advanced pedogenesis in the areas relative to the area in the current study.

In the poorly drained soils of mapping unit AI3 of the low elevation areas, a mean of 110 g/kg and range of 100-120 g/kg as was obtained in the surface soils, while in the subsurface soils mean of 140 g/kg and range of 100-180 g/kg were obtained for clay with overall values that seemed to have increased with soil depth especially in AI3P2 (Table 15). Clay content in the low elevations of the poorly drained soils of mapping unit AI3 seems to

Table 15: Physical properties of soils overlying sandstone and limestone formation

Horizon	Depth	%	Particle size distribution			BD	K_{sat}	Porosity		
		Grav	Clay	Silt	Sand			Total	Air P	Capill
	Cm	content		g/kg		g/cm^3	cm/h		%	
AI1P1:										
Ap	0-9	0	100	60	840	1.19	103.23	57.3	2.7	54.6
BA	9-54	1	140	80	780	1.46	29.32	45.8	4.0	41.8
Bt	54-122	1	260	60	680	1.59	1.22	48.8	2.6	46.2
Ckt	122-200	74	260	60	680	1.53	25.65	42.9	7.3	35.5
AI1P2:										
Ap	0-15	0	120	60	820	0.99	106.89	63.4	2.8	60.7
AB	15-67	0	140	60	800	1.40	9.16	48.5	11.3	37.2
Btv	67-146	24	240	80	680	1.40	63.52	51.6	5.5	46.0
Ckt	146-180	8	280	100	620	1.66	5.50	47.4	4.0	43.5
Mean Surface soils		0	110	60	830	1.09	105.06	60.4	2.8	57.7
Range Surface soils		0	100-120	-	820-840	0.99-1.19	103.23-106.89	57.3-63.4	2.7-2.8	54.6-60.7
Mean Subsurface soils		18	220	73	707	1.51	22.40	47.5	5.8	41.7
Range Subsurface soils		1-74	140-280	60-100	620-800	1.4-1.66	1.22-63.52	42.9-51.6	2.6-11.3	35.5-46.2
AI2P1:										
Ap	0-12	6	120	60	820	1.04	29.72	60.6	6.6	54.0
Bt	12-59	4	180	100	720	1.26	28.10	60.0	23.5	36.2
BC	59-120	14	180	120	700	1.60	9.77	46.1	4.2	41.8
AI2P2:										
Ap	0-13	0	160	120	720	1.35	8.28	52.2	3.4	47.8
AB	13-60	63	160	100	740	1.59	6.11	45.7	6.5	39.1
BC	60-82	11	140	100	760	1.45	10.38	42.7	2.9	39.8
Ckt	82-132	55	220	140	640	1.49	36.04	48.8	6.6	42.2
Mean Surface soils		3	140	90	770	1.20	19	56.4	5.0	50.9
Range Surface soils		0-6	120-160	60-120	720-820	1.04-1.35	8.28-29.72	52.2-60.6	3.4-6.6	47.8-54
Mean Subsurface soils		29.4	176	112	712	1.48	18.08	48.7	8.7	39.8
Range Subsurface soils		4-63	140-220	100-140	640-760	1.26-1.60	6.11-36.04	42.7-60.0	4.2-23.5	36.2-42.2
AI3P1:										
Ap	0-21	2	120	100	780	0.98	243.60	59.8	13.0	46.8
AB	21-80	21	140	40	820	1.52	19.49	63.0	23.9	39.1
AI3P2:										
Ap	0-50	0	100	40	860	1.23	28.26	51.8	6.8	44.9
AB	50-94	6	100	40	860	1.41	10.23	45.2	7.1	38.1
Cg	94-124	35	180	80	740	1.57	0.49	45.1	3.1	42.0
Mean Surface soils		1	110	70	820	1.11	135.93	55.8	9.9	45.9
Range Surface soils		0-2	100-120	40-100	760-860	0.98-1.23	28.26-243.60	51.8-59.8	6.8-13	44.9-46.8
Mean Subsurface soils		21	140	53	807	1.5	10.07	51.1	11.4	39.7
Range Subsurface soils		6-35	100-180	40-80	740-860	1.41-1.57	0.49-19.49	45.1-63.0	3.1-23.9	39.1-42.0

Foot note: BD: bulk density, Ksat: saturated hydraulic conductivity, Grav = Gravel, Capill = Capillary, Air P= Airfilled porosity

have the lowest value (100-180 g/kg) compared to those of AI1 (100-280 g/kg) and AI2. This contradicts findings by Malgwi and Abu (2011), who reported that significantly higher clay and silt values are obtained in the valley bottom soils.

In the SLSi of the Southern agricultural zone, soils of mapping unit MF1 had clay values ranging from 100 to 140 g/kg with a mean of 120 g/kg in the surface soils, while values in the subsurface soils ranged from 120 to 180 g/kg with a mean of 160 g/kg (Table 16). Values increased continuously with an increase in soil depth and may have resulted in the development of Bt horizon at depths of 17-56 cm and 43-81 cm in MFIPI and MF1P2, respectively. In mapping unit MF2, clay amount had a range of 120-140 g/kg and a mean of 130 g/kg in the surface soils, while values in the subsurface soils had a range of 100-260 g/kg and a mean of 140 g/kg. The soils were dominated by transition horizons with the development of a thin Bt horizon at 45-107 cm in MF2P2 (Table 16). These soils seemed to be relatively young and were also influenced by animal activities which may have aided faunal pedoturbation. The soils have not been so developed as to allow for the accumulation of clay or carbonates in the B horizons. These ranges of values are within those obtained by VandenBygaart and Protz (1994), and Ferreira *et al*. (2016) but lower than those of Irmak *et al*. (2007), Carroll and Hathaway (1953) and Petar *et al*. (2016). In the poorly drained and low elevation soils of mapping unit MF3, clay content had a mean of 120 g/kg and range of 100-140 g/kg in the surface soils, while in the subsurface soils, the mean was 140 g/kg and the range was 120-160 g/kg (Table 16). These soils may be adjudged young as they were dominated by transition horizons with gleyed C horizons that indicated youthfulness. Contrary to a study by Malgwi and Abu (2011), values of clay in the poorly drained and low elevation soils of MF3 were lower than those of soils in the well-drained and high elevations. Overall low values of clay (100-180 g/kg) in the soils over SLSi may imply youthfulness and may not have been sufficiently weathered to finer particle sizes such as clay, or the clay materials may have been eroded as a result of the high rainfall in the area.

The result of correlation shows that clay positively and significantly ($p<0.05$) correlated with silt. This indicates that increasing the value of clay with soil age will bring about a similar change for silt. Also, a positive and significant (p=5 %) correlation of clay with exchangeable Mg^{2+} and Ca^{2+} as well as with calcium carbonate equivalent indicate increases with soil development. Findings by Olson (1958), Rickert and Tedrow (1967), Cowie (1968) and Thompson (1981) indicate that total silt and clay in soil profiles show straight-line relationships with soil age. Such increase may be due to weathering of coarser fractions *in situ* or influx of finer particles from wind-blown materials. However, VandenBygaart and

Table 16: Physical properties of soils overlying shale and limestone with sandstone intercalation

Horizon	Depth	Grav	Particle size distribution			BD	K_{sat}	Porosity		
			Clay	Silt	Sand			Total	Air S	Capill
	cm	%		g/kg		g/cm^3	cm/h		%	
MF1P1:										
Ap	0-17	21	140	180	680	1.29	0.49	49.1	1.7	47.4
Bt	17-56	13	160	130	710	1.43	6.33	50.5	3.5	47.0
BC	56-121	-	160	130	710	1.39	4.87	46.6	0.9	45.7
MF1P2:										
Ap	0-14	0	100	40	860	1.32	126.67	50.0	2.3	47.7
AB	14-43	13	120	40	840	1.51	41.41	44.4	3.0	41.4
Bt	43-81	42	180	40	780	1.62	33.62	40.8	6.1	34.7
Crv	81-153	0	180	60	760	1.39	4.38	55.6	2.8	52.8
Mean surface soils		11	120	110	770	1.31	63.58	49.6	2.0	47.6
Range surface soils		0-21	100-140	40-180	680-860	1.29-1.32	0.49-126.67	49.1-50	1.7-2.3	47.4-47.7
Mean subsurface soils		14	160	80	760	1.47	18.12	47.6	3.3	44.3
Range subsurface soils		0-42	120-180	40-130	710-840	1.39-1.62	4.38-41.41	40.8-55.6	0.9-6.1	34.7-52.8
MF2P1:										
Ap	0-16	0	140	60	800	0.90	75.52	62.2	3.2	59.0
BC	16-44	84	160	80	760	1.31	4.38	37.1	5.5	31.6
CB	44-97	15	140	100	760	1.40	2.68	48.6	2.5	46.2
Cr	97-161	0	140	60	800	1.41	10.72	43.6	3.2	40.0
MF2P2:										
Ap	0-16	15	120	40	840	1.31	34.10	53.1	3.6	49.4
AB	16-45	0	100	40	860	1.38	65.77	48.4	7.0	57.1
Bt	45-107	0	140	40	820	1.52	37.03	42.3	7.7	34.6
BC	107-200	22	160	120	720	1.54	11.21	40.5	2.6	37.9
Mean surface soils		8	130	50	820	1.10	54.81	57.7	3.4	54.2
Range surface soils		0-15	120-140	40-60	800-840	0.90-1.31	34.10-75.52	53.1-62.2	3.2-3.6	49.4-59.0
Mean subsurface soils		20	140	73	787	1.42	21.97	43.4	4.8	41.2
Range subsurface soils		0-84	100-260	40-120	720-860	1.31-1.54	2.68-65.77	37.1-48.6	2.5-7.7	31.6-57.1
MF3P1:										
Ap	0-15		140	120	740	0.87	0.49	71.7	3.6	68.0
Cg	15-48		160	200	640	1.31	4.87	53.4	2.0	51.4
MF3P2:										
Ap	0-16		100	110	790	1.18	10.72	54.9	2.7	52.3
BC	16-28	19	140	120	740	0.53	1.95	80.2	3.7	76.5
Cg	28-49	24	120	60	820	1.22	4.87	58.0	3.2	54.8
Mean surface soils			120	115	765	1.02	5.60	63.3	3.2	60.2
Range surface soils			100-140	110-120	740-790	0.87-1.18	0.49-10.72	54.9-71.7	2.7-3.6	52.3-68.0
Mean subsurface soils		14	140	127	733	1.02	3.90	63.9	3.0	60.9
Range subsurface soils			120-160	60-200	640-820	1.02-1.31	1.95-4.87	53.4-80.2	2.0-3.7	51.4-76.5

Foot note: BD: bulk density, Ksat: saturated hydraulic conductivity, Grav = Gravel, Capill = Capillary, Air P= Airfilled porosity

Protz (1994) as well as Birkeland (1984), Protz *et al*. (1984), Singleton and Lavkulich (1987) identified soil weathering as the main contributor to the decreasing particle size and support the fact that larger particles weather physically to smaller particle sizes with soil age. Well-drained soils overlying SLS in the northern agricultural zone were dominated by transition horizons, while those developed from SL and SLSi had pronounced Bt or Crt horizons except in MF2P1. Soils over SLS could be adjudged to be the most weathered, followed by those overlying SL in the central agricultural zone and then SLSi in the southern agricultural zone. Reports have been made that impure forms of limestone produce deep soils (Carroll and Hathaway, 1953), while pure limestone produces only thin soils and are poorly developed (Esu, 2010).

b. Silt: In the soils overlying SLS in the northern agricultural zone, silt content ranged from 120 to 140 g/kg with a mean of 130 g/kg in the surface soils of mapping unit IH1, while values in the subsurface soils ranged from 10 to 20 % with a mean of 128 g/kg (Table 14). In the poorly drained soils of mapping unit IH2, the mean value of silt was 175 g/kg with a range of 110-240 g/kg in the surface soils, while the mean of 130 g/kg and a range of 60-200 g/kg was obtained in the subsurface soils (Table 14). From the foregoing, silt may have been more easily eroded from areas with greater elevation ranges down to the poorly drained and low elevation soils of mapping unit IH2. However, silt content in the soils was low compared to clay. This can be explained as either sand may not have undergone weathering to give rise to silt, silt may have weathered to clay, or dispersion of sand size aggregates was not properly done. Previous reports by Petar *et al.* (2016), Afu *et al*. (2017) and Girao *et al*. (2014) on similar soils indicate values within the range of those in the present study, while VandenBygaart and Protz (1994) obtained lower values in similar soils. However, higher values of silt were reported by Carroll and Hathaway (1953), Aksoy *et al*. (2009), Emadi *et al*. (2008) and Ferreira *et al*. (2016).

In the soils overlying SL in the central agricultural zone, silt content had means of 60 and 73 g/kg in the surface and subsurface soils, respectively, with a range of 60-100 g/kg in the subsurface soils for mapping unit AI1 (Table15). In mapping unit AI2, silt ranged from 60 to 120 g/kg with a mean of 90 g/kg in the surface soils, while values in the subsurface soils ranged from 100 to 140 g/kg with a mean of 112 g/kg (Table 15). In the poorly drained soils of low elevations represented by AI3, silt content ranged from 40 to 120 g/kg with a mean of 70 g/kg in the surface soils, while the subsurface soils had values ranging from 40 to 80 g/kg with a mean of 53 g/kg (Table 15). Silt content in the soils of SL in the central agricultural zone was relatively low compared to values in the soils overlying SLS in the

northern zone. However, values appeared to be relatively higher in mapping unit AI2. This contradicts findings by Irmak *et al*. (2007) Aksoy *et al*. (2009) and Ferreira *et al*. (2016) on similar soils elsewhere, while similar results were obtained by Afu *et al*. (2017) and VandenBygaart and Protz (1994).

In the soils developed from SLSi in the southern agricultural zone, silt content in mapping unit MF1 varied from 40 to 180 g/kg with a mean of 110 g/kg in the surface soils, while the subsurface soil values ranged from 40 to 130 g/kg with a mean of 80 g/kg (Table 16). Silt content decreased in mapping unit MF2 relative to MF1 and MF3 to a mean of 50 g/kg and range of 40-60 g/kg in the surface soils, while values in the subsurface soils had a mean of 73 g/kg and range of 40-120 g/kg (Table 16). Silt content seems to have the highest value in the poorly drained and low elevation soils of MF3 with mean and range of 115 g/kg and 110-120 g/kg, respectively in the surface soils, while values in the subsurface soils ranged from 60 to 200 g/kg with a mean of 127 g/kg (Table 16). Soils in MF3 had the highest silt content (60-200 g/kg) compared to those of MF2 (40-120 g/kg) and MF1 (40-180 g/kg). This implies that silt was more readily transported by erosion than other soil particle sizes.

Straight-line relationships have been obtained between total silt and soil age, and this accords to an earlier report that soil age contributes to increasing silt content (VandenBygaart and Protz, 1994). The silt was highest (60-240 g/kg) in soils of SLS in the northern agricultural zone, followed by SLSi in the southern agricultural zone (40-200 g/kg) and then SL in the central zone (40-140 g/kg). Consequently, soils overlying SLS could be adjudged most mature than their counterpart in other geological formations.

Silt content correlated positively and significantly ($p<0.05$) with pH (H_2O), exchangeable Mg^{2+} and Ca^{2+}, as well as base saturation. There is, therefore, a likely increase in the values of these parameters when silt increases. However, silt correlated significantly ($p<0.05$) but negatively and strongly with sand and indicates a good likelihood of decreasing the value of sand with increasing silt content. As the values of sand decreased with weathering, the proportion of silt increases from the weathering of sand fraction.

c. Sand: Sand content in the soils formed from SLS in the northern agricultural zone ranged from 600 to 720 g/kg with a mean of 660 g/kg in the surface soils, while values in the subsurface soils ranged from 580 to 680 g/kg with a mean of 636 g/kg in mapping unit IH1 (Table 14). In the low elevation and poorly drained soils of mapping unit IH2, a mean of 655 g/kg and a range of 580-730 g/kg was obtained in the surface soils, while values in the subsurface soils ranged from 520 to 720 g/kg with a mean of 645 g/kg (Table 14). Sand percentage content in IH2 was similar (520-730 g/kg) to values in the well-drained IH1 (580-

720 g/kg). This indicates that sand particle size in mapping unit IH1 may have been weathered into clay over time and are comparatively more mature than those of IH2, or the sand fraction in the poorly drained soils may have been deposited by receding water. Previous studies by Afu *et al*. (2017) and VandenBygaart and Protz (1994) on similar soils elsewhere indicate similar results, while lower values were reported by Petar *et al*. (2016), Carroll and Hathaway (1953), Girao *et al.* (2014) and Aksoy *et al*. (2009). Low values of sand translate to more intense weathering conditions and more mature soils, while high sand content indicates limited weathering intensity and relative youthfulness.

In the soils overlying SL in the central agricultural zone, sand content ranged from 820 to 840 g/kg with a mean of 830 g/kg in the surface soils, while subsurface soils had values ranging from 620 to 800 g/kg with a mean of 707 g/kg for soils of mapping unit AI1 (Table 15). In mapping unit AI2, sand content ranged from 720 to 820 g/kg with a mean of 770 g/kg in the surface soils, while subsurface soil values had a range of 640-760 g/kg (X=712 g/kg) (Table 15). Sand content seemed to have increased with a decrease in elevation ranges as values in the low lying poorly drained soils had a mean of 820 g/kg and range of 760 to 860 g/kg in the surface soils as well as ranged from 740 to 860 g/kg with a mean of 807 g/kg in the subsurface soil (Table 15). An increase in sand content with decreasing elevation ranges reflected in the horizonation of the soils. For instance, well-developed soils with Bt and Ckt horizons in AI1 correspond with low sand and high clay in the mapping unit, resulting in more mature soils in the unit compared to those of AI3, which may be regarded as recent with the high sand content obtained. Reports by Petar *et al*. (2016), Carroll and Hathaway (1953) and Girao *et al*. (2014) indicate lower values of sand, an indication that either intensity of weathering in the tropical rainforest is slower or the soils may not have had enough time to weather into finer fractions (Birkland, 1984, Protz *et al*. 1984, Singleton and Lavkulich, 1987).

In the soils developed from SLSi in the southern agricultural zone, sand content ranged from 680 to 860 g/kg with a mean of 770 g/kg in the surface soils of MF1, while subsurface soil values ranged from 710 to 840 g/kg with a mean of 760 g/kg with values that increased in MF1P1 and decreased in MF1P2 with an increase in soil depth (Table 16). In mapping unit MF2, the amount of sand ranged from 800 to 840 g/kg with a mean of 820 g/kg in the surface soils, while values in the subsurface soils ranged from 720 to 860 g/kg with a mean of 787 g/kg. The values varied irregularly with increasing soil depth (Table 16) and could probably indicate pedoturbation or youthfulness. These values appear higher than those obtained by Petar *et al*. (2016) (132.4-427.2 g/kg), Girao *et al.* (2014) (428-540 g/kg),

Emadi *et al.* (2008) (60-550 g/kg) and Ferreira *et al.* (2016) (35-426 g/kg) for soils of similar origin. However, VandenBygaart and Protz (1994) recorded very high values, while Afu *et al*. (2017) obtained similar values to the surface soils in the present study. The sand content in soils of the low elevation ranges that were poorly drained in mapping unit MF3 seems to have recorded the least values in the area, with a mean of 765 g/kg and a range of 740-790 g/kg in the surface soils, while values in the subsurface soils ranged from 640 to 820 g/kg with a mean of 733 g/kg (Table 16). Sand content in the poorly drained mapping unit MF3 had the least mean values in the surface and subsurface soils compared to soils in other elevation ranges. These results contradict those of Malgwi and Abu (2011), who obtained significantly higher values of silt and clay in the valley bottom position compared to other landscape positions.

Generally, mean values of sand in the SLS in the northern agricultural zone were lower (520-730 g/kg) than those in the SL (620-860 g/kg), and SLSi (640-860 g/kg). This indicates that some of the coarse fractions in the SLS soils in the northern agricultural zone may have been weathered into finer size fractions, resulting in comparatively older soils as weathering occurs over time. Sand dominated soils are most likely to have acid pH values. For instance, sand correlated poorly and negatively but significantly (p=5 %) with soil pH and indicated increasing acidity with increasing sand content. A similar trend of relationship was obtained with exchangeable Mg^{2+} and Ca^{2+} as well as base saturation. This accords to earlier findings by Ofem and Esu (2015) that exchangeable bases are most likely to be leached in sand dominated soils, especially in the high rainfall areas.

4.2.2 Bulk Density

Values of bulk density in mapping unit IH1 of the soils developed from SLS in the northern agricultural zone ranged from 1.45 to 1.54 g/cm^3 (X= 1.50 g/cm^3) in the surface soils, while in the subsurface soils, the values ranged from 1.48 to 1.66 g/cm^3 with mean of 1.60 g/cm^3 (Table 14). The values generally increased with soil depth. In mapping unit, IH2, bulk density values ranged from 0.71 to 1.18 g/cm^3 with a mean of 0.95 g/cm^3, while values in the subsurface soils ranged from 0.91 to 1.63 g/cm^3 with a mean of 1.35 g/cm^3 (Table 14) in the poorly drained soils of the low elevation areas. Values in the surface soils appeared to be lower than the subsurface values. This may be due to compaction in the subsurface soils or due to organic matter accumulation in the topsoils, which has lower bulk density than mineral particles. These values align with a report by Brady and Weil (2002) as values of bulk density in IH1 were within 1.2 to 1.8 g/cm^3 for the sand dominated soils. Such range of values are likely to encourage aeration and water movement for optimum crop growth (Esu, 2010).

Values obtained by VandenBygaart and Protz (1994) and Aksoy *et al*. (2009) were lower than values obtained in the present study. High bulk density may be a result of soil compaction by animal trampling or traction. Higher bulk densities have been reported for fallow lands (Ezeaku and Eze, 2014).

Soils in mapping unit AI1 of SL in the central agricultural zone had bulk density values ranging from 0.99 to 1.19 g/cm^3 with a mean of 1.09 g/cm^3 in the surface soils, while the subsurface soils had values ranging from 1.4 to 1.66 g/cm^3 with a mean of 1.51 g/cm^3 (Table 15). Bulk density values increased with soil depth; however, surface soils were within the range of 1.1 – 1.4 g/cm^3 suggested for cultivated loams (Donahue *et al.* 1983). In mapping unit AI2, values ranged from 1.04 to 1.35 g/cm^3 with a mean of 1.20 g/cm^3 in the surface soils, while subsurface soil values ranged from 1.26 to 1.60 g/cm^3 with a mean of 1.48 g/cm^3 (Table 15). The trend of values seems to be similar to those obtained in AI1 as values increased with soil depth. In the poorly drained soils of AI3 found in the low elevation areas, ranges of 0.98-1.23 g/cm^3 and 1.41-1.57 g/cm^3 as well as means of 1.11 and 1.50 g/cm^3 were obtained in the surface and subsurface soils, respectively (Table 15). Most of the soils in the present study were within the range of sand-sandy loam or slightly finer. According to Brady and Weil (2002), sand and sandy loam soils have bulk density values of 1.2 and 1.8 g/cm^3. This range of values is not threatened by hardpans and will permit root proliferation as well as air and water movement (Esu, 2010). Somewhat similar values were obtained by VandenBygaart and Protz (1994) and Aksoy *et al*. (2009).

In the soils overlying SLSi in the southern agricultural zone, soils in mapping unit MF1 had bulk density values range of 1.29 to 1.32 g/cm^3 with a mean of 1.31 g/cm^3 in the surface soils, while the subsurface soils had values ranging from 1.39 to 1.62 g/cm^3 with a mean of 1.47 g/cm^3 (Table 16). Bulk density values appeared to be higher in the Bt horizons compared to Cr horizons. In mapping unit MF2, bulk density values ranged from 0.90 to 1.31 g/cm^3 in the surface soils with a mean of 1.10 g/cm^3, while the subsurface soils ranged from 1.31 to 1.54 g/cm^3 with mean of 1.42 g/cm^3 (Table 16). Bulk density values increased consistently with an increase in soil depth, resulting in higher subsurface mean values.

Soils in the low elevations and poorly drained represented by mapping unit MF3 had bulk density values ranging from 0.87 to 1.18 g/cm^3 and a mean of 1.02 g/cm^3, while in the subsurface soil, bulk density values ranged from 1.02 to 1.31 g/cm^3 with mean of 1.02 g/cm^3 (Table 16). Again, subsurface mean values were similar to surface soils (X= 1.02 g/cm^3). Surface and subsurface values are quite similar to those obtained in the soils of SL in the

central agricultural zone but lower than values (1.35 g/cm^3) obtained in the subsurface soils overlying SLSi in the northern agricultural zone.

Soil bulk density values depend on soil mineral make-up and will tend to increase with an increase in sesquioxidic minerals (Fe oxides). The minerals were higher (7.18 %) in the SLS in the northern agricultural zone than in the southern agricultural zone (2.159 %), hence its higher bulk density as previously stated. Fasina *et al*. (2015) linked a significant change in bulk density among and within soils to differences in mineralogy and clay, while Kefas *et al*. (2020) obtained a positive correlation between bulk density and oxalate as well as dithionite forms of Fe and Al. The range of values in the present study aligns with those in the reports by Brady and Weil (2002) and indicates that the soils were not threatened by hardpans and will permit root proliferation as well as air and water movement (Esu, 2010; VandenBygaart and Protz, 1994; Aksoy *et al*. 2009).

Values of correlation coefficient of bulk density with saturated hydraulic conductivity, total and capillary porosities and moisture content indicate an inverse relationship and strong significant correlation ($r^2 > 0.63$) at $p < 0.05$. Similarly, it also correlated negatively and strongly with organic carbon, total N, C/N ratio, exchangeable Na^+ and K^+, as well as available P by Bray 1 method. These parameters are likely to decrease with increased bulk density values.

4.2.3 Saturated Hydraulic Conductivity (Ksat.)

The saturated hydraulic conductivity of surface soils of mapping unit IH1 in the SLS of the northern agricultural zone had values that ranged from 40.31 to 78.18 cm/h with a mean of 59.25 cm/h, while subsurface soil values ranged from 0.61 in the CBk and Ckt horizons to 106.28 cm/h in the AB horizon with mean of 25.29 cm/h (Table 14). Soils in the low elevation and poorly drained mapping unit IH2 had values that varied from 8.55 to 256.54 cm/h with a mean of 132.55 cm/h in the surface soils, while subsurface soils values spanned from 0.61 in the BC horizon to 130.71 cm/h in the transition BC horizon with mean of 50.70 cm/h (Table 14). The soils were rated high to very high in saturated hydraulic conductivity (SSSD, 2017) except in the CBk, Bt and Ckt horizons, where either calcium carbonate or clay accumulation hindered water transmission. This may encourage surficial erosion in an undulating landscape or ponding on a tableland. According to Sarki *et al*. (2014), clay mineralogy and particle size distribution affect saturated hydraulic conductivity. However, in the present study, soil particle sizes were not significantly correlated with Ksat.

In the soils overlying SL in the central agricultural zone, mapping unit AI1 had saturated hydraulic conductivity values of 103.23 to 106.89 cm/h with a mean of 105.06 cm/h

in the surface soils, while the subsurface soils varied from 1.22 in the Bt horizon to 63.52 cm/h in the Btv horizon with a mean of 22.40 cm/h (Table 15). Soils in mapping unit AI2 had values that ranged from 8.28 to 29.72 cm/h with a mean of 19.0 cm/h in the surface soils, while the subsurface soils ranged from 6.11 in the AB horizon to 36.04 cm/h in the Ckt horizon with a mean of 18.08 cm/h (Table 15). In low elevation areas represented by mapping unit AI3, values of Ksat. varied from 28.26 to 243.60 cm/h (X=135.93 cm/h) in the surface soils, while the subsurface soils ranged from 0.49 in the Cg horizon to 19.49 cm/h in the transition AB horizon (X=10.07 cm/h) (Table 15). Surface mean values of Ksat. in soils of mapping unit AI3 seem to be the highest (135.93 cm/h) in the SL, while the subsurface values were the lowest (mean= 10.07 cm/h) compared to AI1 [X=105.06 (surface soil) and 22.40 cm/h (subsurface soil)] and AI2 [X=19.0 (surface soil) and 18.08 cm/h (subsurface soil)] as they exceeded 36.0 cm/h. These values are rated high on the scale of SSSD (2017), while the surface soils of AI1 and AI3 had very high values. However, the Bt, Ckt and Cg horizons seem to have moderately low values. Such low values may restrict the vertical flow of water and subsequently lead to ponding or surficial erosion.

In the soils overlying SLSi in the southern agricultural zone, saturated hydraulic conductivity varied from 0.49 to 126.67 cm/h with a mean of 63.58 cm/h in the surface soils of mapping unit MF1, while the subsurface soils had values ranging from 4.38 in the Crv horizon to 41.41 cm/h in the transition AB horizon with mean of 18.12 cm/h (Table 16). In mapping unit MF2, values of Ksat. ranged from 34.10 to 75.52 cm/h with a mean of 54.81 cm/h in the surface soils, while the subsurface soils ranged from 2.68 in CB to 65.77 cm/h in the transition AB horizon with a mean of 21.97 cm/h (Table 16). The surface soil values were rated very high, while the subsurface soils were rated high on the scale of SSSD (2017). The surface soils of MF3 had comparatively the lowest values in the surface soils with values ranging from 0.49 to 10.72 cm/h with a mean of 5.60 cm/h, relative to the subsurface soil with values ranging from 1.95 to 4.87 cm/h and a mean of 3.90 cm/h (Table 16). The values of Ksat. are rated moderately high on the scale of SSSD (2017) and are not likely to be a threat to farmers' crops. Very high values of Ksat. may encourage leaching or loss of plant nutrients leading to reduced crop growth relative to production cost. Also, values obtained in the subsurface soils of the poorly drained and low elevation soils of MF3 were the least (3.9 cm/h) compared to similar depths in mapping units MF1 (18.12 cm/h) and MF3 (21.97 cm/h).

Saturated hydraulic conductivity correlated positively and significantly ($p< 0.05$); strongly with total porosity ($r^2=0.57$), moisture content ($r^2=0.757$), organic carbon ($r^2=0.639$), C/N ratio ($r^2=0.58$) and exchangeable Na^+ and K^+ ($r^2>0.60$), and weakly with available P and

capillary porosity. Increasing saturated hydraulic conductivity will most likely bring about an increase in the values of the soil parameters.

4.2.4 Soil Porosity

a. Total Porosity

In the soils of SLS in the northern agricultural zone, total porosity ranged from 46.4 to 65 % (X=55.7 %) in the A horizons of mapping unit IHI, while in the subsoils, total porosity ranged from 44.4 in the CBk horizon to 54.4 % in the Ckt horizon with a mean of 49.7 % (Table 14). However, the Ap horizons of the poorly drained and low elevation soils of IH2 had comparatively higher values with a mean of 65.2 % and range of 60.9-69.5 % in the Ap horizons, while the underlying soils had total porosity values ranging from 47.3 in the Cg horizon to 62.2 % in the transition BC horizon with a mean of 52.65 % (Table 14). The soils are high in total porosity as the means were greater than 50 % recommended for a silt-loam surface soil (Brady, 1974). Such values will result in extremely porous soil (Pagliai, 1988). Higher values in the poorly drained soils are a reflection of saturation by water.

In the soils of SL in the central agricultural zone, mapping unit AI1 had values of total porosity ranging from 57.3 to 63.4 %, with a mean of 60.4 % in the surface soils, while in the subsurface soils, the value ranged from 42.9 in the Ckt horizon to 51.6 % in the Btv horizon with a mean of 47.5 % (Table 15). Soils in mapping unit AI2 had values that ranged from 52.2 to 60.6 %, with a mean of 56.4 % in the surface soils, while subsurface soils ranged from 42.7 in the BC horizon to 60.0 % in the Bt horizon with a mean of 48.7 % (Table 15). These values were rated high (Brady, 1974), and the soils were highly porous (Pagliai, 1988). Such porous soils are likely to encourage soil aeration and water movement. Values in the poorly drained and low elevation soils of AI3 contradict those in AI1 and AI2 as the values appear to be relatively lower with a range of 51.8-59.8 % and mean of 55.8 % in the surface soils, while in the subsurface soils, the values ranged from 45.1 % in the Cg horizon to 63.0 % in the transition AB horizon with a mean of 51.1 % (Table 15); the subsurface soil values were relatively higher in mapping unit AI3 than in AI1 and AI2.

In the SLSi soils of the southern agricultural zone, values of total porosity ranged from 49.1 to 50.0 %, with a mean of 49.6 % in the surface soils of mapping unit MF1, while the underlying soils had values ranging from 40.8 in the Bt horizon to 55.6 % in the Crv horizon with a mean of 47.6 % (Table 16). In mapping unit MF2, values ranged from 53.1 to 62.2 %, with a mean of 57.7 % in the Ap horizons, while the subsoils ranged from 37.1 in BC horizon to 48.6 % in transition CB horizon with a mean of 43.4 % (Table 16). These values were rated high (Brady, 1974), and the soils are described as extremely porous by the rating

of Pagliai (1988). Such values are likely to encourage the free movement of air and water. In the poorly drained soils of MF3 in the low elevations, values ranged from 54.9 to 71.7 %, with a mean of 63.3 % in the surface soils, while subsurface soil values ranged from 53.4 in the Cg horizon to 80.2 % in the BC horizon with mean of 63.9 % (Table 16). Values in the surface and subsurface soils of MF3 were higher than those of MF1 and MF2. Fetter (1998) and Rieu and Sposito (1991) reported that soils porosity of greater than 50 % and 45-50 %, respectively, of the total volume of soil, are good agricultural soils. This further affirms the importance of the soils for agricultural purposes.

Total porosity correlated strongly, positively and significantly with capillary porosity and moisture content, indicating a likely increase in values with increasing capillary porosity and moisture content. However, correlation with air-filled porosity, organic carbon, total N, available P and exchangeable bases was weak at a 5 % level of significance. Weathering increases porosity and water holding capacity and reduces bulk density (SSDS, 2017).

b. **Air-Space Porosity (macroporosity or non-capillary porosity)**

In the soils overlying SLS in the northern agricultural zone, mapping unit IH1 had values of air-space porosity ranging from 2.3 to 21 % with a mean of 11.7 % in the surface soils, while the subsurface soils had values varying from 3.1 in the CBk horizon to 14.5 % in the transition CB horizon (Table 14). In mapping unit IH2, air-space porosity ranged from 5.7 to 13.1 %, with a mean of 9.4 % in the surface soils relative to the subsurface soils, which had values ranging from 3.0 to 5.5 % in the Cg and BC horizons, respectively (X=4.5 %) (Table 14). Lower values of air-space porosity were obtained in the poorly drained soils of mapping unit IH2 than those in IH1 in both the surface and subsurface soils; this indicates that the higher values previously recorded in the poorly drained and low elevation soils were mainly as a result of saturation by water. Soil aeration is likely limiting plant growth when air-filled porosity falls below about 10 % (Baver and Farnsworth, 1940; Vomocil and Flocker, 1961). However, Brady (1974) predicted 25 % as the optimal value for a silt loam surface soil. These soils are likely to be threatened by poor soil aeration, especially in the subsurface soils.

In the soils of SL of the central agricultural zone, soils in mapping unit AI1 ranged in values from 2.7 to 2.8 % (X=2.8 %) in the surface soils, while the subsurface soil values ranged from 2.6 in the Bt horizon to 11.3 % in transition AB horizon (X= 5.8 %) (Table 15). Mapping unit AI2 had a range of 3.4-6.6 % and a mean of 5.0 % in the surface soils, while values in the subsurface soils ranged from 4.2 in the BC horizon and 23.5 % in the Bt horizon with a mean of 8.7 % (Table 15). The surface soils of the poorly drained and low elevation

mapping unit AI3 had the highest (3.1-23.9 %) air-space porosity, probably because it recorded the highest values of coarse sand, while its subsurface soil values were somewhat intermediate between values obtained in the mapping unit AI1 and AI2. These values are less than 25 % stated by Brady (1974) for surface silt loam soils. Though the soils have favourable total porosity, as stated earlier in this report, soil aeration may limit plant growth as most values of air-filled porosity were less than 10 % reported by Baver and Farnsworth (1940) and Vomocil and Flocker (1961).

In the soils of SLSi in the southern agricultural zone, air-space porosity in mapping unit MF1 ranged from 1.7 to 2.3 % (X=2.0 %) in the surface soils, while the subsurface soils ranged from 0.9 in BC to 6.1 % in the Bt horizon (X=3.3 %) (Table 16). In mapping unit MF2, a range of 3.2-3.6 % and a mean of 3.4 % were obtained in the surface soils, while subsurface soils had values ranging from 2.5 in CB to 7.7 % in Bt horizon with a mean of 4.8 % (Table 16). The poorly drained soils of mapping unit MF3 found in the lowest elevation range had the lowest values for air space porosity in the subsurface soils with values ranging from 2.0 to 3.7 %, and a mean of 3.0 %, while the surface soil values were somewhat intermediate between their counterpart in MF1 and MF2 with values ranging from 2.7 to 3.6 %, and mean of 3.2 % (Table 16). The values for the surface soils of the upland shows that the soils were poorly drained and low in soil aeration and may not encourage the activities of micro and burrowing organisms. According to Esu (2010), good aeration makes soil-forming processes operate faster. Soil aeration is likely to become limiting to plant growth when the air-filled porosity falls below about 10 % (Baver and Farnsworth, 1940; Vomocil and Flocker, 1961). This porosity helps in the rapid drainage of excess water after heavy rainfall. This allows for air to return to the plant roots for normal functioning. Air-filled porosity did not show any significant relationship with the parameters under study, except with total porosity, where a significant positive correlation was obtained (r^2=0.50) at $p < 0.05$.

c. Capillary Porosity (microporosity)

Capillary porosity of soils developed over SLS in the northern agricultural zone ranged from 43.9 to 44.1 %, with a mean of 44 % in the surface soils of mapping unit IH1, while the subsurface soil values ranged from 35 in transition BC horizon to 47.8 % in the Ckt horizon with mean of 41.8 % (Table 14). In mapping unit IH2, capillary porosity ranged from 55.2 to 56.4 %, with a mean of 55.8 in the surface soils of the low elevation areas, while the subsurface soils ranged from 41.9 in Cg to 58 % in the transition BC horizon of the poorly drained soils (Table 14). The values of capillary porosity in IHI were lower both in the surface and subsurface soils than those in IH2. This indicates that the capillaries of IH2 filled

with water had influenced the soils, especially in the poorly drained soils. Capillary porosity retains the water required for plant growth. The soil capillaries become filled with air as the plant extracts the water needed for normal functioning from these fine pores.

In the soils of SL in the central agricultural zone, capillary porosity ranged from 54.6 to 60.7 %, with a mean of 57.7 % in the surface soils of mapping unit AI1, while the subsurface soils had values ranging from 35.5 in the Ckt horizon to 46.2 % in the Bt horizon (Table 15).

In mapping unit AI2, values ranged from 47.8 to 54 %, with a mean of 39.8 % in the surface soils, while the subsurface soils ranged from 36.2 in the Bt horizon to 42.2 % in the Ckt horizon with a mean of 39.8 % (Table 15). These values are lower than those of AI1. Similarly, soils in the poorly drained mapping unit AI3 ranged from 44.9 to 46.8 %, with a mean of 45.9 %, while the subsurface soils ranged from 39.1 in the transition AB horizon to 42.0 % in the Cg horizon (Table 15). Soils in mapping unit AI3 recorded the lowest values compared to AI1 and AI2 in the surface soils, while the subsurface soils were somewhat high to intermediate between AI1 and AI2.

In the soils of SLSi in the southern agricultural zone, capillary porosity in the surface soils of mapping unit MF1 ranged from 47.4 to 47.7 %, with a mean of 47.6 %, while values in the subsurface soils had values ranging from 34.7 in the Bt horizon to 52.8 % in the Crv horizon with mean of 44.3 % (Table 16). Soils in mapping unit MF2 ranged from 49.4 to 59.0 % (X= 54.2 %) in the surface soils, while values in the subsurface soils ranged from 31.6 in the BC horizon to 57.1 % in the transition AB horizon (Table 16). The poorly drained and low elevation soils in mapping unit MF3 had a range of 52.3-68.0 % (X= 60.2 %) in the surface soils, while the subsurface soils had values ranging from 51.4 to 76.5 % in the Cg and BC horizons, respectively, with a mean of 60.9 % (Table 16). The surface and subsurface of the poorly drained soils in mapping unit MF3 had the highest values (51.4-76.5 %) for capillary porosity when compared to mapping units MF1 (34.1-52.8 %) and MF2 (31.6-59.0 %); with values that appeared to decrease with an increase in elevation ranges. This is not surprising as soil moisture may have dominated capillary porosity, especially in the poorly drained soils of mapping unit MF3.

Capillary porosity showed positive, significant ($p< 0.05$) and strong correlation with moisture content, organic C, available P, exchangeable Na^+ and K^+ and weakly with Ca^{2+}.

4.2.5 Soil Water Content (SWC) (θw)

Soil water content was determined at three tension levels of 30 cm, 60 cm and 90 cm. In the soils formed over SLS in the northern agricultural zone, the soil water content in the

surface soils varied from 28.6 in the 60 and 90 cm tensions to 34.2 % in the 30 cm tension with means of 31.7, 29.45 and 29.75 % for the 30 cm, 60 cm and 90 cm tensions, respectively for soil mapping unit IH1 (Table 17). In the subsurface soils of IH1, values ranged from 20.4 in the transition CB horizon at 90 cm to 34.7 % in the AB horizon at 30 cm tension. Soil water content seems to be highest at 30 cm tension and least at 60 cm tension for the surface and subsurface soils, probably because 30 cm tension exerts the least suction. At low suction pressure, less water is removed from the soil, leaving more moisture in the soils. In the poorly drained mapping unit IH2, values of soil water content in the surface soils ranged from 46.6 at 60 cm tension to 90.1 % at 30 cm tension, while means of 69.65, 62.85 and 64.55 % were obtained at 30 cm, 60 cm and 90 cm tensions, respectively (Table 17). In the subsurface soils, SWC varied from 26.1 in the BC horizon at 60 cm tension to 67 % in the BC horizon at 30 cm tension with means of 40.15, 38.75 and 39.2 % at 30, 60 and 90 cm tensions, respectively (Table 17). Again, soil water content was highest at 30 cm and least at 60 cm in both surface and subsurface soils. Worthy of note is that values of SWC increased in the poorly drained soils of IH2 in both the surface and subsurface relative to the soils in IH1 irrespective of the tension level. Also, the difference in SWC between 30 cm tension and other tension levels seem to be wider compared to 60 and 90 cm tensions in every case.

In the soils of SL in the central agricultural zone, the soil water content in soil mapping unit AI1 ranged from 45.7 at 90 cm to 63.3 % at 30 cm in the surface soils with means of 54.75, 53.55 and 53.45 % at tensions of 30, 60 and 90 cm, respectively (Table 18). In the subsurface soils of AI1, SWC ranged from 23.2 in the Ckt horizon at 60 cm tension to 33.8 % in the Btv horizon at 30 cm tension with means of 30.32, 27.77 and 29.33 % at 30, 60 and 90 cm tensions, respectively (Table 18). Means of SWC decreased with increasing tensions in both surface and subsurface soils (Table 18). In the surface soils of mapping unit AI2, SWC varied from 35.5 at 60 cm tension to 55.9 % at 30 cm tension with means of 46.1, 43.85 and 44.65 % at 30 cm, 60 cm and 90 cm, respectively (Table 18). In the subsurface soils, values varied from 24.4 in the transition AB horizon at 60 cm tension to 32.4 % in the Bt horizon at 30 cm tension with means of 28.68, 27.02 and 27.04 % at 30, 60 and 90 cm tensions, respectively (Table 18). Again, the highest SWC was obtained at 30 cm tension, and the lowest value was obtained at 60 cm tension with the only slight numerical difference between values at 60 and 90 cm tension. Similarly, subsurface means were lower than surface soil means for their corresponding tension levels. In the poorly drained soils of the low elevation range represented by AI3, SWC ranged from 35.4 at 90 cm tension to 49 % at 30

Table 17: More physical properties of the soils overlying SLS

Horizon	Depth	Soil moisture content	(θw) Soil H_2O content			(θv) volume fraction of water			pore size diameters		
			30cm	60cm	90cm	30cm	60cm	90cm	Tens.30cm	Tens.60cm	Tens.90cm
	cm	%		%					96.58µm	48.29µm	32.2µm
IH1P1:										θv/St	
Ap	0-37	44.8	34.2	30.3	30.9	49.59	43.94	44.81	0.76	0.68	0.69
AB	37-64	36.7	34.7	30.4	31.2	51.36	44.99	46.18	0.94	0.83	0.85
CBk	64-130	26.7	25.6	24.8	24.9	42.5	41.17	41.33	0.96	0.93	0.93
CB	130-195	30.7	22.5	21.7	20.4	36.23	34.94	32.84	0.73	0.71	0.66
IH1P2:											
Ap	0-42	30.1	29.2	28.6	28.6	44.97	44.04	44.4	0.97	0.95	0.96
Bt	42-70	27.6	25.2	24.1	24.2	41.83	40.01	40.17	0.93	0.88	0.88
Ckt	70-150	35	31.8	30.7	31.3	49.61	47.89	48.83	0.91	0.88	0.90
Mean surface soils		37.5	31.7	29.45	29.75	47.28	43.99	44.61	0.87	0.82	0.83
Range surface soils		30.1-44.8	29.2-34.2	28.6-30.3	28.6-30.9	44.97-49.59	43.94-44.04	44.4-44.81	0.76-0.97	0.68-0.95	0.69-0.96
Mean subsurface soils		31.3	27.96	26.34	26.4	44.31	41.8	41.87	0.89	0.85	0.84
Range subsurface soils		26.7-36.7	22.5-34.7	21.7-30.7	20.4-31.3	36.23-51.36	34.94-47.89	32.84-48.83	0.73-0.96	0.71-0.93	0.66-0.93
IH2P1:											
Ap	0-26	51.5	49.2	46.6	47.6	58.06	54.99	56.17	0.95	0.90	0.92
BC	26-84	29.5	26.3	26.1	26.2	42.87	42.54	42.71	0.89	0.88	0.89
Cg	84-145	30.3	27.7	26.8	26.9	43.21	41.81	41.96	0.91	0.88	0.89
IH2P2:											
Ap	0-13	97.4	90.1	79.1	81.5	63.97	56.16	57.87	0.92	0.81	0.83
BC	13-46	68.1	67	63.6	64.7	60.97	57.88	58.88	0.98	0.93	0.95
Cg	46-87	40.7	39.6	38.5	39	51.48	50.05	50.7	0.97	0.94	0.96
Mean surface soils		74.5	69.65	62.85	64.55	61.02	55.58	57.02	0.94	0.86	0.88
Range surface soils		51.5-97.4	49.2-90.1	46.6-79.1	47.6-81.5	58.06-63.97	54.99-56.16	56.17-57.87	0.92-0.95	0.81-0.90	0.83-0.92
Mean subsurface soils		42.2	40.15	38.75	39.2	49.63	48.07	48.56	0.94	0.91	0.92
Range subsurface soils		29.5-68.1	26.3-67.0	26.1-63.6	26.2-64.7	42.87-60.97	41.81-57.88	41.96-58.88	0.89-0.98	0.86-0.94	0.89-0.96

Foot note: Tens= tension

Table 18: More Physical properties of soils overlying SL

Horizon	Depth	Soil Moisture content	(θw) Soil H_2O content			(θv) vol. fraction of water			Pore size diameters (θv/St)		
	cm		30cm %	60cm	90cm	30cm	60cm	90cm	Tens.30cm 96.58µm	Tens.60cm 48.29µm	Tens.90cm 32.2µm
AI1P1:											
Ap	0-9	48.2	46.2	46.0	45.7	54.98	54.74	54.38	0.96	0.96	0.95
BA	9-54	31.3	30.4	28.6	29.6	44.38	41.76	43.22	0.97	0.91	0.94
Bt	54-122	30.7	29.9	29.0	29.5	47.54	46.11	46.91	0.97	0.94	0.96
Ckt	122-200	28.0	26.5	23.2	24.2	40.55	35.50	37.03	0.95	0.83	0.86
AI1P2:											
Ap	0-15	63.9	63.3	61.1	61.2	62.67	60.49	60.59	0.99	0.95	0.96
AB	15-67	34.7	33.2	26.7	32.1	46.48	37.38	44.94	0.96	0.77	0.93
Btv	67-146	36.8	33.8	32.9	33.2	47.32	46.06	46.48	0.92	0.89	0.90
Ckt	146-180	28.6	28.1	26.2	27.4	46.65	43.49	45.48	0.98	0.92	0.96
Mean surface soils		56.1	54.75	53.55	53.45	58.83	57.62	57.49	0.98	0.96	0.96
Range surface soils		48.2-63.9	46.2-63.3	46-61.1	45.7-61.2	54.9-62.67	54.74-60.49	54.38-60.59	0.96-0.99	0.95-0.96	0.95-0.96
Mean subsurface soils		31.7	30.32	27.77	29.33	45.49	41.72	44.01	0.96	0.88	0.93
Range subsurface soils		28-36.8	26.5-33.8	23.2-32.9	24.2-33.2	40.55-47.54	35.50-46.11	37.03-46.91	0.92-0.98	0.77-0.94	0.86-0.96
AI2P1:											
Ap	0-12	58.6	55.9	52.2	53.4	58.14	54.29	55.54	0.96	0.90	0.92
Bt	12-59	47.3	32.4	28.7	28.8	40.82	36.16	36.29	0.68	0.60	0.60
BC	59-120	28.9	27.3	26.2	26.3	43.68	41.92	42.08	0.95	0.91	0.91
AI2P2:											
Ap	0-13	38.0	36.3	35.5	35.9	49.01	47.93	48.47	0.94	0.92	0.93
AB	13-60	28.7	25.3	24.5	24.4	40.23	38.96	38.80	0.88	0.85	0.85
BC	60-82	29.4	28.4	27.4	27.0	41.18	39.73	39.15	0.96	0.93	0.92
Ckt	82-132	32.7	30.0	28.3	28.7	44.70	42.17	42.76	0.92	0.86	0.88
Mean surface soils		48.3	46.1	43.85	44.65	53.58	51.11	52.01	0.95	0.91	0.93
Range surface soils		38-58.6	36.3-55.9	35.5-52.2	35.9-53.4	49.0-58.14	47.93-54.29	48.47-55.54	0.94-0.96	0.90-0.92	0.92-0.93
Mean subsurface soils		33.4	28.68	27.02	27.04	42.12	39.79	39.82	0.88	0.83	0.83
Range subsurface soils		28.7-47.3	25.3-32.4	24.5-28.7	24.4-28.8	40.23-44.7	36.16-42.17	36.29-42.76	0.68-0.96	0.60-0.93	0.60-0.92
AI3P1:											
Ap	0-21	60.8	49.0	47.6	46.5	48.02	46.65	45.57	0.80	0.78	0.76
AB	21-80	41.4	27.1	25.7	25.8	41.19	39.06	39.22	0.65	0.62	0.62
AI3P2:											
Ap	0-50	42.0	37.7	36.5	35.4	46.37	44.90	43.54	0.90	0.87	0.84
AB	50-94	32.1	28.4	27.0	27.1	40.04	38.07	38.21	0.89	0.84	0.85
Cg	94-124	28.8	28.2	58.0	27.0	44.27	91.06	42.39	0.98	2.02	0.94
Mean surface soils		51.4	43.35	42.05	40.95	47.2	45.78	44.56	0.85	0.83	0.8
Range surface soils		42-60.8	37.7-49	36.5-47.6	35.4-46.5	46.3-48.02	44.9-46.65	43.54-45.57	0.80-0.90	0.78-0.87	0.76-0.84
Mean subsurface soils		34.1	27.9	36.9	26.63	41.83	56.06	39.94	0.84	1.16	0.80
Range subsurface soils		28.8-41.4	27.1-28.4	25.7-58.0	25.8-27.1	40.0-44.27	38.07-91.06	38.21-42.39	0.65-0.98	0.62-2.02	0.62-0.94

Tens = tension

cm tension with means of 43.35, 42.05 and 40.95 % at 30, 60 and 90 cm tensions, respectively for the surface soils (Table 18). In the subsurface soils, values of SWC ranged from 25.7 in the transition AB horizon at 60 cm tension to 58 % in the Cg horizon at 60 cm tension with means of 27.9, 36.9 and 26.63 % at 30, 60 and 90 cm tensions, respectively.

Soil water content decreased with a decrease in elevation ranges for the surface soils and some subsurface soils for 30, 60 and 90 cm tensions with the 30 cm tension recording the highest values of SWC, while 60 cm tension recorded the lowest value for SWC irrespective of the elevation range. Irrespective of limestone formations, elevation range or soil depth, SWC was higher in the surface relative to the subsurface soils. The high percentage of SWC in the surface soils will favour the availability of nutrients to roots as well as their growth but affect soil aeration status and oxygen availability. Plants are likely to proliferate shallow roots in response to the high SWC in surface soils in order to take advantage of water and nutrient availability in the short run.

In the soils over SLSi in the southern agricultural zone, soils in mapping unit MF1 had SWC of 25.7 at 90 cm tension and 36.8 % at 60 cm tension with means of 34.1, 36.5 and 29.95 % at 30, 60 and 90 cm tensions, respectively for the surface soils (Table 19). In the subsurface soils, SWC was 19.4 in the Bt horizon at 30 cm tension and 38 % at 60 cm tension in Crv horizon with means of 29.2, 30.5 and 30.34 % at 30, 60 and 90 cm tensions, respectively (Table 19). Soil surface values appeared to be higher at 30 and 60 cm tensions and not at 90 cm. , The water content of soils in mapping unit MF2 ranged from 32.6 at 90 cm tension to 65.3 % at 60 cm tension with means of 48, 51.6 and 43.3 % at 30, 60 and 90 cm tensions, respectively for the surface soils (Table 19). Subsurface soils had water content values as 18.6 in the Bt horizon at 90 cm and 33.4 % in the CB horizon at 60 cm tension with means of 25.7, 27.3 and 22.82 % at 30, 60 and 90 cm tensions, respectively (Table 19). SWC at 60 cm were comparatively higher than those obtained at tensions of 30 and 90 cm. The poorly drained soils of mapping unit MF3 had values of SWC as 41.1 at 90 cm tension and 77.8 % at 60 cm tension with means of 59.5, 61.0 and 57.8 % at 30, 60 and 90 cm tensions, respectively (Table 19). In the subsurface soils, values varied from 37.6 in Cg horizon at 90 cm tension to 144 % at 60 cm tension in the BC horizon with means of 74.7, 76.1 and 71.4 % at 30, 60 and 90 cm tensions, respectively (Table 19).

In the entire soils overlying SLSi in the southern agricultural zone, SWC was lower in the subsurface soils at 30 cm and 60 cm tensions and higher in the subsurface soils at 90 cm tension. Contrary to the above, values in the poorly drained and low elevation soils of mapping unit MF3 were higher in the subsurface soils than the surface soils, although without

Table 19: More Physical Properties of Soils Overlying SLSi

Horizon	Depth	Soil	(θw) Soil H_2O content			(θv) vol. fraction of water			pore size diameters (θv/St)		
	cm	mois.cont	30cm	60cm	90cm	30cm	60cm	90cm	Tens.30cm	Tens.60cm	Tens.90cm
			%						96.58µm	48.29µm	32.2µm
MF1P1:											
Ap	0-17	38.2	35.8	36.8	34.2	46.18	47.47	44.12	0.94	0.97	0.90
Bt	17-56	35.4	31.7	32.9	30.4	45.33	47.05	43.47	0.90	0.93	0.86
BC	56-121	33.6	32.0	32.9	30.8	44.48	45.73	42.81	0.95	0.98	0.92
MF1P2:											
Ap	0-14	37.8	32.3	36.1	25.7	42.64	47.65	33.92	0.85	0.95	0.68
AB	14-43	29.5	25.8	27.5	22.3	38.96	41.53	33.67	0.88	0.94	0.76
Bt	43-81	25.2	19.4	21.4	34.1	31.43	34.67	55.24	0.77	0.85	1.35
Crv	81-153	40.1	37.3	38.0	34.1	51.85	52.82	47.40	0.93	0.95	0.85
Mean surface soils		38.0	34.1	36.5	29.95	44.41	47.56	39.02	0.90	0.96	0.79
Range surface soils		37.8-38.2	32.3-35.8	36.1-36.8	25.7-34.2	42.64-46.18	47.47-47.65	33.92-44.12	0.85-0.94	0.95-0.97	0.68-0.90
Mean subsurface soils		32.8	29.2	30.5	30.34	42.41	44.36	44.52	0.89	0.93	0.95
Range subsurface soils		25.2-40.1	19.4-37.3	21.4-38.0	22.3-34.1	31.43-51.85	34.67-52.82	33.67-55.24	0.77-0.95	0.85-0.98	0.76-1.35
MF2P1:											
Ap	0-16	68.9	59.8	65.3	54.0	53.82	58.77	48.60	0.87	0.94	0.78
BC	16-44	28.3	22.3	24.1	20.3	29.21	31.57	26.59	0.79	0.85	0.72
CB	44-97	35.2	31.5	33.4	30.2	43.47	46.09	41.68	0.89	0.95	0.86
Cr	97-161	30.8	30.0	28.6	25.6	42.30	40.33	36.10	0.97	0.93	0.83
MF2P2:											
Ap	0-16	40.6	36.2	37.8	32.6	47.42	49.52	42.71	0.89	0.93	0.80
AB	16-45	35.2	26.6	30.1	20.4	36.71	41.54	28.15	0.76	0.86	0.58
Bt	45-107	27.9	20.9	22.8	18.6	31.77	34.66	28.27	0.75	0.82	0.67
BC	107-200	26.3	23.0	24.7	21.8	35.42	38.04	33.57	0.87	0.94	0.83
Mean surface soils		54.8	48.0	51.6	43.3	50.62	54.15	45.66	0.88	0.94	0.79
Range surface soils		40.6-68.9	36.2-59.8	37.8-65.3	32.6-54.0	47.42-53.82	49.52-58.77	42.71-48.60	0.87-0.89	0.93-0.94	0.78-0.80
Mean subsurface soils		30.6	25.7	27.3	22.82	36.48	38.71	32.39	0.84	0.89	0.75
Range subsurface soils		26.3-35.2	20.9-31.5	22.8-33.4	18.6-30.2	29.21-43.47	51.57-46.09	26.59-41.68	0.75-0.97	0.82-0.95	0.58-0.86
MF3P1:											
Ap	0-15	82.0	76.2	77.8	74.5	66.29	67.69	64.82	0.92	0.94	0.90
Cg	15-48	40.8	38.0	39.3	37.6	49.78	51.48	49.26	0.93	0.96	0.92
MF3P2:											
Ap	0-16	46.4	42.7	44.2	41.1	50.39	52.16	48.50	0.92	0.95	0.88
BC	16-28	150.8	142.3	144.0	134	75.42	76.32	71.02	0.94	0.95	0.89
Cg	28-49	47.7	43.8	45.1	42.7	53.44	55.02	52.09	0.92	0.95	0.90
Mean surface soils		64.2	59.5	61.0	57.8	58.34	59.925	56.66	0.92	0.95	0.89
Range surface soils		46.4-82.0	42.7-76.2	44.2-77.8	41.1-74.5	50.39-66.29	52.16-67.69	48.5-64.82	-	0.94-0.95	0.88-0.90
Mean subsurface soils		79.8	74.7	76.1	71.4	59.55	60.94	57.45	0.93	0.95	0.90
Range subsurface soils		40.8-150.8	38-142.3	39.3-144.0	37.6-134	49.78-75.42	51.48-76.32	49.26-71.02	0.92-0.94	0.95-0.96	0.89-0.92

Foot note: Tens= tension

any regular trend in values with increasing soil depth. Also, values of SWC were highest at 60 cm tension in all the soils overlying SLSi.

4.2.6 Pore Size Distribution

Pore sizes with diameters of 96.58, 48.29 and 32.2 µm were obtained at 30, 60, and 90 cm tensions, respectively (Tables 17-19). Soil pores larger than 50 µm allow for the circulation of water and air, while those less than 50 µm retain soil water (Florian *et al.*, 2011). Soil pore diameters obtained at 60 and 90 cm tensions will therefore retain water. The pore size distribution for soils overlying SLS represented by mapping units IH1 and IH2 suggest that the fraction of total pore space filled with water (θv/St) decreased at field capacity (tension of 60 cm) where pore diameter of 48.29 µm was obtained. However, the CB horizon in IH1P1 instead experienced an increase in the fraction. This trend in values implies that soil moisture could only be drained at a tension of 90 cm, representing micropores, as tensions of 30 and 60 cm still had water in-pores.

The trend in curves for soils overlying SL in the central agricultural zone was similar to those of SLS except for the loamy sand and sandy loams obtained in the Ap and Cg horizons of AI1P1, AI3P1 and AI3P2, which increased in θv/St at 60 cm tension. The increase for sandy loams or loamy sands align with similar graphs plotted by Danielson and Sutherland (1986) and shows a transition in soils between the convex facing curves obtained in the soils of SLS and concave shaped curves in the soils of SLSi. The entire graphs of pore size distribution showed an opposite trend to those of the SLSi and had convex shapes for most of its lines as trends of θv/St indicated increases at 60 cm tension with very steep slopes. This suggests an abrupt loss in moisture from 30 to 60 cm tension compared to 60 and 90 cm tension.

4.2.7 Factor Analysis for Variable Loadings: Physical Properties of the Soils

The 19 physical soil variables were compressed into 3 principal components (PC): PC1, PC2 and PC3, and presented in Table 20.

The 3 PCs contributed 67.7 % of the total variation in soils overlying SLS, SL, SLSi and ALV. Out of this variation, 39.7 % was accounted for by PC1, the most influential PC, while 15.0 % of the variability was contributed by PC2. The least influential PC was PC3 which explained only 13.0 % of the total variation between the soils.

Total P, capillary P, soil moisture content, θ_{30}, θ_{60}, θ_{90}, θ_{v30}, θ_{v90} and BD explained the dominance of PC1 over PCs 2 and 3 to the variability in soils over SLS, SL, SLSi and ALV, with over 85 % contribution by each soil property to the variability. All the properties had negative correlations with PC1 except BD. PC2 was explained by sand (85.8 %) and clay (-73.1 %), while the contribution of air porosity (-83.8 %) manifested in PC3. Soil properties

Table 20: Factor Analysis for Variable Loadings: Physical Properties of the Soils

Variables	PC1	PC2	PC3
Grav	0.331	-0.435	-0.353
Clay	0.255	-0.731	-0.346
Silt	-0.130	-0.671	-0.225
Sand	-0.090	0.858	0.353
BD	0.894	-0.291	0.093
Ksat	-0.320	0.256	-0.343
Total_P	-0.865	0.090	-0.402
Air_P	0.132	0.250	-0.838
Capillary_P	-0.911	0.020	0.092
SMC	-0.932	0.130	-0.228
θ_{30}	-0.957	0.037	-0.123
θ_{60}	-0.953	0.043	-0.001
θ_{90}	-0.957	-0.009	-0.094
θ_{v30}	-0.939	-0.189	-0.045
θ_{v60}	-0.790	-0.144	0.290
θ_{v90}	-0.859	-0.390	0.042
$\theta_{v/St30}$	-0.328	-0.569	0.591
$\theta_{v/St60}$	-0.163	-0.246	0.660
$\theta_{v/St90}$	-0.175	-0.562	0.500
Formations {SLS}	0.231	-0.433	-0.474
Formations {AL}	-0.522	-0.027	-0.114
Formations {SL}	0.095	-0.147	0.133
Formations {SLSi}	0.264	0.499	0.339
Eigenvalues	9.12	3.46	2.99
% total variance	39.67	15.05	12.99
Cummulative Eigenvalues	9.12	12.59	15.57
Cumulative %	39.67	54.72	67.71

that manifested their influence on soil variability in PC3 were of less importance in soil variation.

Properties manifested in PC1 were the most significant contributors to soil variation between the lithologies. The high % contribution of the soil properties to PC1 explains their importance in soil variation. Therefore, $\theta_{30}=\theta_{90}>\theta_{60}>\theta_{v30}$>SMC>Capillary porosity>BD>Total porosity>θ_{v90} in that order showed good correlation with the most significant roles in explaining the variations due to PC1. Variable influence to soil variation reduced from PC1 to PCs 2 and 3.

Soils over ALV were the most influential in PC1, while those over SLSi contributed to the greatest variation in PC2 and the SLS in PC3. This (ALV>SLSi>SLS) indicates the order in which variability in lithology affected the physical soil properties. The high Eigenvalues for the 3 PCs exceeding 2 indicates that the 3 PCs can be adopted.

4.3 Soil Chemical Properties

Soil chemical properties are presented in Tables 21-23 and discussed according to soils overlying SLS in the northern agricultural zone, SL in the Central agricultural zone and SLSi in the Southern agricultural zone.

4.3.1 Soil pH

In the soils on SLS in the northern agricultural zone, soil pH (H_2O) values ranged from 5.5 to 7.3 (X=6.4) in the Ap horizons of mapping unit IH1, while soil pH ($CaCl_2$) had a range of 4.4-6.2 with a mean of 5.3 (Table 21). In the subsoils, pH (H_2O) values ranged from 6.2 in the CBk horizon to 6.8 in the Ckt horizon with a mean of 6.5, while pH ($CaCl_2$) ranged from 4.1 in the Ckt horizon to 5.0 in the transition CB horizon (X= 4.4) (Table 21). In all cases, pH ($CaCl_2$) was lower than pH (H_2O) and showed a steady decrease with increased soil depth except in the transition CB horizon, while soil pH (H_2O) increased irregularly with soil depth. However, the highest values of pH in both solutions occurred in the Ap horizon of IH1P2 and the transition CB horizon of IH1P1.

In mapping unit IH2, soil pH(H_2O) in the topsoils had a range of 6.1-6.7 with a mean of 6.4, while soil pH ($CaCl_2$) had a mean of 5.0 (Table 21). In the underlying soils, pH (H_2O) ranged from 5.2 in the transition BC horizon to 5.9 in the Cg horizon with a mean of 5.6, while soil pH ($CaCl_2$) ranged from 4.2 in the BC horizon to 4.8 in the Cg horizon with a mean of 4.5 (Table 21). Corresponding lowest and highest values of pH in both solutions were obtained in the same horizon. A similar trend was noticed for mapping unit IH1, and this shows a strong correlation between the two solutions used in determining soil pH. The

Table 21: Chemical properties of the soils overlying SLS

Horizon	Depth	soil pH		Org. C	Total N	C/N	Exch. Bases				CEC	Base	Exch. Acidity		Avail. P	CaCO3
		H_2O	$CaCl_2$				Na^+	K^+	Mg^{2+}	Ca^{2+}	NH_4OAc	Sat.	Al^{3+}	H^+	Bray1	Equiv
	cm	1:2.5	1:2.5	← g/kg →			← $cmol_c/kg$ →					%	cmolc/kg		mg/kg	g/kg
IH1P1:																
Ap	0-37	5.5	4.4	7.55	1.54	5	0.04	0.07	3.4	3.4	28.0	24.7	1.4	1.4	7.94	28.6
AB	37-64	6.4	4.4	4.66	1.12	4	0.02	0.05	4.0	5.8	35.6	27.7	1.8	1.0	6.62	28.3
CBk	64-130	6.2	4.2	2.75	1.12	3	0.02	0.05	1.4	7.2	35.2	24.6	2.8	3.2	Tr	29.2
CB	130-195	6.7	5.0	6.18	0.98	6	0.03	0.07	1.8	3.8	23.6	24.2	0.6	0.6	3.97	22.1
IH1P2:																
Ap	0-42	7.3	6.2	40.3	1.54	26	0.10	0.14	3.4	5.4	27.6	32.8	0.2	0.6	15.89	38.8
Bt	42-70	6.4	4.4	6.18	1.12	6	0.03	0.07	1.8	4.2	27.6	22.1	2.0	1.0	7.94	25.8
Ckt	70-150	6.8	4.1	1.37	0.98	1	0.02	0.05	3.4	6.8	42.8	23.1	3.0	2.8	3.97	36.8
Mean surface soils		6.4	5.3	23.93	1.54	16	0.07	0.11	3.4	4.4	27.8	28.7	0.8	1.0	11.92	33.7
		5.5-	4.4-				0.04-	0.07-		3.4-		24.7-	0.2-	0.6-	7.94-	28.6-
Range surface soils		7.3	6.2	7.55-40.3	-	5-26	0.10	0.14	-	5.4	27.6-28	32.8	1.4	1.4	15.89	38.8
Mean subsurface soils		6.5	4.4	4.23	1.06	4	0.024	0.06	2.5	5.6	33.0	24.3	2.0	1.7	4.50	28.4
		6.2-	4.1-		0.98-		0.02-	0.05-	1.4-	3.8-		22.1-	0.6-	0.6-		22.1-
Range subsurface soils		6.8	5.0	1.37-6.18	1.12	1-6	0.03	0.07	4.0	7.2	23.6-42.8	27.7	3.0	3.2	0-7.94	36.8
IH2P1:																
Ap	0-26	6.7	5.0	32.24	2.66	12	0.09	0.12	5.0	7.2	28.4	43.7	0.6	0.4	7.94	17.8
BC	26-84	5.8	4.4	3.77	0.98	4	0.02	0.05	2.2	4.4	22.8	29.3	1.2	0.8	3.97	29.4
Cg	84-145	5.9	4.8	2.06	0.56	4	0.02	0.05	1.4	7.6	29.2	31.1	0.6	0.8	10.59	28.2
IH2P2:																
Ap	0-13	6.1	5.0	86.64	2.80	31	0.28	0.29	4.2	8.2	54.4	23.9	0.6	0.6	19.86	27.0
BC	13-46	5.2	4.2	52.39	2.80	19	0.11	0.15	2.4	5.2	40.0	19.7	3.0	3.4	19.86	30.0
Cg	46-87	5.6	4.7	28.21	1.68	17	0.05	0.09	4.2	5.0	15.2	61.5	0.4	0.4	15.89	17.7
Mean surface soils		6.4	5.0	59.44	2.73	22	0.19	0.21	4.6	7.7	41.4	33.8	0.6	0.5	13.9	22.4
		6.1-		32.24-	2.66-		0.09-	0.12-	4.2-	7.2-		23.9-		0.4-	7.94-	17.8-
Range surface soils		6.7	-	86.64	2.80	12-31	0.28	0.29	5.0	8.2	28.4-54.4	43.7	-	0.6	19.86	27.0
Mean subsurface soils		5.6	4.5	21.61	1.51	14	0.05	0.09	2.6	5.6	26.8	35.4	1.3	1.4	12.58	26.3
		5.2-	4.2-	2.06-	0.56-		0.02-	0.05-	1.4-	4.4-		19.7-	0.4-	0.4-	3.97-	17.7-
Range subsurface soils		5.9	4.8	52.39	2.80	4-19	0.11	0.15	4.2	7.6	15.2-40.0	61.5	3.0	3.4	19.86	30.0

Table 22: Chemical properties of the soils overlying SL

Horizon	Depth	soil pH		Org. C	Total N	C/N	Exch. Bases				CEC	Base Sat.	Exch. Acidity		Avail. P	$CaCO_3$
		H_2O	$CaCl_2$				Na^+	K^+	Mg^{2+}	Ca^{2+}	NH_4OAc		Al^{3+}	H^+	Bray 1	equiv.
AI1P1:	cm	1:2.5	1:2.5	← g/kg →			←		cmol$_c$/kg	→		%	cmol$_c$/kg		mg/kg	g/kg
Ap	0-9	7.2	6.5	8.92	1.12	8	0.05	0.09	1.0	3.6	12.0	21.5	1.4	1.6	18.53	21.5
BA	9-54	6.4	5.1	3.43	0.98	4	0.02	0.05	1.2	2.0	17.6	16.0	0.4	0.6	6.62	16.0
Bt	54-122	5.3	4.3	1.72	0.42	4	0.02	0.05	0.2	1.8	17.6	21.3	4.4	3.6	Tr	21.3
Ckt	122-200	5.4	4.2	2.40	0.98	2	0.02	0.05	0.8	0.8	19.6	24.0	4.4	4.8	Tr	24.0
AI1P2:																
Ap	0-15	6.8	6.1	10.29	1.68	6	0.06	0.09	0.6	3.8	12.4	36.7	TR	2.0	10.59	27.0
AB	15-67	6.1	5.3	4.12	1.26	3	0.02	0.05	0.4	1.8	17.6	12.9	TR	1.8	6.62	16.7
Btv	67-146	6.4	5.1	2.06	0.28	7	0.02	0.05	0.6	1.8	17.6	14.0	TR	2.2	3.97	22.0
Ckt	146-180	7.7	6.5	2.06	0.84	2	0.02	0.05	2.2	6.0	19.2	43.1	0.4	2.0	2.65	30.0
Mean surface soils		7.0	6.3	9.61	1.40	7	0.055	0.09	0.8	3.7	12.2	38.1	0.7	1.8	14.56	24.3
Range surface soils		6.8-7.2	6.1-6.5	8.92-10.29	1.12-1.68	6-8	0.05-0.06	-	0.6-1.0	3.6-3.8	12-12.4	36.7-39.5	0-1.4	1.6-2.0	10.59-18.53	21.5-27.0
Mean subsurf soils		6.2	5.1	2.63	0.79	3	0.02	0.05	0.9	2.37	18.2	18.1	1.6	2.5	3.31	21.7
Range subsurface soils		5.3-7.7	4.2-6.5	1.72-4.12	0.28-1.26	2-7	-	-	0.2-2.2	0.8-6.0	17.6-19.6	8.5-43.1	0-4.4	0.6-4.8	0-6.62	16-30
AI2P1:																
Ap	0-12	6.4	5.5	9.95	1.40	7	0.05	0.09	0.8	2.4	36.4	9.2	TR	2.0	13.24	21.5
Bt	12-59	6.4	5.0	3.77	0.84	4	0.02	0.05	0.4	3.6	26.8	15.2	TR	2.2	9.28	25.5
BC	59-120	6.3	5.5	2.40	2.52	1	0.01	0.04	0.2	4.8	15.6	32.4	TR	2.0	1.32	23.5
AI2P2:																
Ap	0-13	6.7	6.0	7.55	1.40	5	0.04	0.07	1.4	1.6	4.4	70.7	TR	2.0	7.94	28.2
AB	13-60	5.7	4.7	4.12	0.84	5	0.02	0.06	0.8	1.6	5.2	47.7	0.8	1.8	5.3	15.8
BC	60-82	5.7	4.6	1.72	0.70	2	0.02	0.04	0.4	2.6	6.0	51.0	0.6	1.8	5.3	30.0
Ckt	82-132	6.5	6.0	3.43	0.84	4	0.02	0.05	2.2	5.6	15.2	51.8	TR	1.6	2.65	30.0
Mean surface soils		6.6	5.8	8.75	1.40	6	0.045	0.08	1.1	2.0	20.4	39.9	0	2.0	10.59	24.9
Range surface soils		6.4-6.7	5.5-6.0	7.55-9.95	-	5-7	0.04-0.05	0.07-0.09	0.8-1.4	1.6-2.4	4.4-36.4	9.2-70.7	-	-	7.94-13.24	21.5-28.2
Mean subsurface soils		6.1	5.2	3.09	1.15	3	0.018	0.048	0.8	3.64	13.76	39.6	0.28	1.88	4.77	25.0
Range subsurface soils		5.7-6.5	4.6-6.0	1.72-4.12	0.70-2.52	1-4	0.01-0.02	0.04-0.06	0.2-2.2	1.6-5.6	5.2-26.8	15.2-51.8	0-0.8	1.6-2.2	1.32-9.29	15.8-30.0
AI3P1:																
Ap	0-21	5.3	4.8	22.16	2.10	11	0.06	0.10	0.4	1.8	10.8	21.9	1.4	2.0	7.94	27.5
AB	21-80	5.2	4.2	2.78	0.56	5	0.01	0.04	0.6	0.6	6.0	20.8	2.0	2.6	3.97	20.7
AI3P2:																
Ap	0-50	5.7	5.1	5.15	1.54	3	0.03	0.06	0.2	1.4	10.0	16.9	0.2	2.2	15.89	28.2
AB	50-94	6.7	6.1	1.72	0.84	2	0.01	0.03	1.2	2.0	20.8	15.6	TR	1.8	5.3	28.1
Cg	94-124	7.5	6.7	1.03	1.26	1	0.01	0.03	0.8	3.6	30.8	14.4	TR	1.8	9.28	25.2
Mean surface soils		5.5	5.0	13.66	1.82	8	0.045	0.08	0.3	1.6	10.4	19.4	0.8	2.1	11.92	27.9
Range surface soils		5.3-5.7	4.8-5.1	5.15-22.16	1.54-2.10	3-11	0.03-0.06	0.06-0.10	0.2-0.4	1.4-1.8	10-10.8	16.9-21.9	0.2-1.4	2-2.2	7.94-15.89	27.5-28.2
Mean subsurface soils		6.5	5.7	1.84	0.89	2	0.01	0.033	0.86	2.07	19.2	16.9	0.67	2.07	6.18	24.7
Range subsurface soils		5.2-7.5	4.2-6.7	1.03-2.78	0.56-1.26	1-5	-	0.03-0.04	0.6-1.2	0.6-3.6	6.0-30.8	14.4-20.8	0-2	1.8-2.6	3.97-9.28	20.7-28.1

Table 23: Chemical properties of soils overlying SLS

Horizon	Depth	soil pH		Org. C	Total N	C/N	Exch. Bases				CEC pH7	Base Sat.	Exch. Acidity		Avail. P	CaCO3
		H_2O	$CaCl_2$				Na^+	K^+	Mg^{2+}	Ca^{2+}	NH_4OAc		Al^{3+}	H^+	Bray 1	equiv.
MF1P1	Cm			g/kg			cmol$_c$/kg					%	cmol$_c$/kg		mg/kg	g/kg
Ap	0-17	6.6	5.9	13.73	3.36	4	0.06	0.11	0.6	6.0	19.2	18.5	TR	2.0	3.97	19.2
Bt	17-56	6.1	5.2	2.06	0.84	2	0.01	0.03	1.4	5.0	23.8	48.8	1.0	1.6	11.91	23.8
BC	56-121	5.6	5.1	0.69	0.56	1	Tr	0	0.4	5.6	19.6	31.9	1.0	1.8	90.02	19.6
MF1P2																
Ap	0-14	5.7	5.0	21.96	0.98	22	0.07	0.12	0.8	1.4	10.8	22.1	0.4	1.8	7.94	26.0
AB	14-43	5.4	4.3	6.86	0.98	7	0.04	0.08	1.0	0.8	10.8	17.8	2.4	2.4	1.32	18.3
Bt	43-81	5.8	4.7	7.21	1.26	6	0.04	0.07	0.4	1.2	12.8	13.4	1.8	3.4	Tr	24.9
Crv	81-153	5.4	4.4	3.43	0.28	12	0.02	0.05	1.2	0.4	13.2	12.7	3.2	2.0	6.62	23.6
Mean surface soils		6.2	5.5	17.85	2.17	8	0.065	0.12	0.7	3.7	25.2	20.3	1.9	0.2	1.9	22.6
Range surface soils		5.7-6.6	5-5.9	13.7-31.96	0.98-3.36	4-22	0.06-0.07	0.11-0.12	0.6-0.8	1.4-6.0	10.8-39.6	18.5-22.1	0-0.4	1.8-2.0	3.97-7.94	19.2-26
Mean subsurface soils		5.7	4.7	4.05	0.78	5	0.022	0.05	0.88	2.6	13.8	24.9	1.9	2.2	21.97	22.0
Range subsurf soils		5.4-6.1	4.3-5.2	0.69-7.2	0.28-1.26	1-12	0-0.04	0-0.08	0.4-1.4	0.4-5.6	10.8-18.8	12.7-48.8	1-3.2	1.6-3.4	0-90.02	18.3-24.9
MF2P1																
Ap	0-16	5.6	5.0	46.34	2.52	18	0.11	0.15	3.2	7.2	41.2	25.9	1.8	2.2	6.62	20.8
BC	16-44	5.1	4.3	7.89	4.62	2	0.04	0.08	2.0	2.0	62.8	6.6	18.2	15.0	5.3	21.1
CB	44-97	5.3	4.2	3.09	0.84	4	0.02	0.05	1.4	0.8	57.2	4.0	20.2	13.8	1.32	20.2
Cr	97-161	5.4	4.4	1.37	0.84	2	0.02	0.05	4.8	3.0	16.8	46.9	6.4	13.0	1.32	26.1
MF2P2																
Ap	0-16	6.4	5.9	20.59	0.42	49	0.07	0.12	0.6	2.2	18.0	16.6	Tr	1.6	10.59	20.3
AB	16-45	5.4	4.4	7.55	0.84	9	0.04	0.09	0.4	1.2	21.6	8.0	1.6	2.0	9.28	15.2
Bt	45-107	5.3	4.4	6.52	0.70	9	0.04	0.08	1.0	0.4	24.4	6.2	3.4	1.8	3.97	20.0
BC	107-200	5.4	4.3	3.09	0.56	6	0.02	0.05	0.8	0.4	50.0	2.5	3.0	3.4	Tr	18.8
Mean surface soils		6.0	5.5	33.47	1.47	23	0.09	0.14	1.9	4.7	29.6	21.2	1.9	0.9	1.9	20.6
Range surface soils		5.6-6.4	5-5.9	20.59-46.3	0.42-2.52	18-49	0.07-0.11	0.12-0.15	0.6-3.2	2.2-7.2	18-41.2	16.6-25.9	0-1.8	1.6-2.2	6.6-10.59	20.3-20.8
Mean subsurface soils		5.3	4.3	4.91	1.40	4	0.03	0.067	1.73	1.3	38.8	12.4	8.8	8.2	3.53	20.2
Range subsurface soils		5.1-5.4	4.2-4.4	1.37-7.9	0.56-4.62	2-9	0.02-0.04	0.05-0.09	0.4-4.8	0.8-3.0	16.8-62.8	2.5-46.9	3-20.2	1.8-3.4	0-9.28	15.2-26.1
MF3P1																
Ap	0-15	6.7	6.7	15.08	4.20	4	0.06	0.10	2.8	9.0	40.0	29.9	Tr	2.0	5.3	6.1
Cg	15-48	6.5	6.3	2.51	1.12	2	0.01	0.03	0.8	4.0	40.0	12.1	Tr	1.8	13.24	22.8
MF3P2																
Ap	0-16	6.1	5.9	19.39	5.58	3	0.07	0.13	0.8	7.4	25.6	32.8	Tr	2.2	31.77	14.5
BC	16-28	6.1	5.7	7.89	2.10	4	0.04	0.09	0.4	5.0	11.2	49.4	Tr	2.2	30.45	23.7
Cg	28-49	6.3	6.1	6.82	1.12	6	0.04	0.08	0.2	2.2	12.6	20.0	Tr	2.2	18.53	45.7
Mean surface soils		6.4	6.3	17.24	4.89	4	0.065	0.115	1.8	8.2	32.8	31.4	2.1	Tr	2.1	10.3
Range surface soils		6.1-6.7	5.9-6.7	15.08-19.4	4.20-5.58	3-4	0.06-0.07	0.10-0.13	0.8-2.8	7.4-9.0	25.6-40	29.9-32.8	-	2.0-2.2	5.3-31.77	6.1-14.5
Mean subsurface soils		6.3	6.0	5.74	1.45	4	0.03	0.067	0.47	3.7	21.27	27.2	0	2.1	20.74	30.7
Range subsurface soils		6.1-6.5	5.7-6.3	2.5-7.89	1.12-2.10	2-6	0.01-0.04	0.03-0.09	0.2-0.8	2.2-5.0	11.2-40	12.1-49.4	-	1.8-2.2	13.2-30.45	22.8-45.7

irregular decrease in values of soil pH with soil depth is not surprising, as the deposition of various soils and soil materials at various periods in the poorly drained low elevation soils may have resulted in such irregularity. Soil pH (H_2O) was greater than soil pH ($CaCl_2$), with over 1 pH unit in all the studied soils. Mean soil pH in both solutions decreased from the surface to subsurface soils in all the mapping units. These values were rated as slightly acid to near neutral on the scale of Holland *et al.* (1989) and will discourage the weathering of primary minerals, hence the dominance of the clay mineralogy by 2:1 minerals. Alumina becomes insoluble at a pH range of 5-9, while silica becomes more soluble within the pH (Tan, 1998) and results in the formation of kaolinite, gibbsite and bauxite (Ollier, 1975). Similar ranges of values for similar soils have been reported (VandenBygaart and Protz, 1994, Afu *et al*. 2017, Sedov *et al*., 2008, Samonil, 2007 and Aksoy *et al*., 2009), and even higher values in Turkey and Brazil (Irmak *et al*., 2007, Girao *et al*., 2014 and Ferreira *et al*. (2016). The high values obtained by the above researchers may be as a result of low rainfall amount in their study areas. However, Petar *et al*. (2016) and Carroll and Hathaway (1953) recorded lower values for similar soils. Such low pH values may be a result of the loss of organic matter and soil minerals and the effects of acid fertilizers. Values of pH($CaCl_2$) were by more than 1 pH unit lower than pH (H_2O) in all horizons. The highest and lowest values of pH (H_2O) and pH ($CaCl_2$) were obtained in the same or adjoining horizons. Regular decrease in soil pH values down to the Ckt horizons where pH values increased was obtained.

In the soils overlying SL in the Central agricultural zone, soil pH (H_2O) had a range of 6.8-7.2 with a mean of 7.0 in the topsoils of mapping unit AI1, while soil pH ($CaCl_2$) ranged from 6.1 to 6.5 with a mean of 6.3 (Table 22). In the underlying soils, pH (H_2O) ranged from 5.3 in the Bt horizon to 7.7 in the Ckt horizon with a mean of 6.2; similarly, soil pH ($CaCl_2$) ranged from 4.2 in the Ckt to 6.5 in the Crt horizons with a mean of 5.1 in the subsoils (Table 22). Values of pH ($CaCl_2$) were lower than pH (H_2O) by more than 1 pH unit. The highest and lowest values of pH (H_2O) and pH ($CaCl_2$) occurring in the same or adjoining horizons with a regular decrease in soil pH values down to the Ckt horizons where pH values increased.

In soil mapping unit AI2, soil pH (H_2O) ranged from 6.4 to 6.7 (X=6.6) while pH ($CaCl_2$) ranged from 5.5 to 6.0 (X=5.8) in the surface soils (Table 22). In the subsurface soils of AI2, pH (H_2O) (H_2O) was 5.7 in the transition AB and BC horizons and 6.5 in the Ckt horizon with a mean of 6.1, while pH ($CaCl_2$) had values that ranged from 4.6 in the BC horizon to 6.0 in the Ckt horizon (X=5.2) (Table 22). Soil pH values decreased with soil depth steadily down to the Ckt horizon where values increased, an indication that free

carbonates may have been deposited or illuviated in such horizons. This is, however, not out of place as the horizon had already been identified as the Ckt horizon, which designates an accumulation of carbonates (k) and illuvial clay (t). Soil pH (H_2O) and pH ($CaCl_2$) had their lowest and highest values in the same or adjoining horizons, pointing out a clear correlation between the two solutions used in their determination. Values of soil pH ($CaCl_2$) were lower than pH (H_2O) by more than 1 pH unit.

In the poorly drained and low elevation soils of mapping unit AI3, soil pH (H_2O) ranged from 5.3 to 5.7 with a mean of 5.5 in the Ap horizons, while soil pH ($CaCl_2$) had a range of 4.8-5.1 with a mean of 5.0. in the subsoils (Table 22). Values of pH (H_2O) ranged from 5.2 in the AB horizon to 7.5 in the Cg horizon with a mean of 6.5, while pH ($CaCl_2$) ranged from 4.2 in the transition AB horizon to 6.7 in the Cg horizon with a mean of 5.7 in the subsurface soils (Table 22). Soil pH values in salt were less than pH in water. Mean soil pH in both solutions decreased in the subsurface soils of all the mapping units except in the poorly drained AI3 soils which increased in the subsurface soils. The above range of values indicate that the soils in mapping unit AI1 and AI2 were moderately acid to near neutral in pH, while those in AI3 were moderately acid on the scale of Holland *et al*. (1989). A similar range of values was obtained by Petar *et al*. (2016), Afu *et al*. (2017), VandenBygaart and Protz (1994) and Aksoy *et al*. (2009), while Girao *et al*. (2014), Ferreira *et al.* (2016) and Khormali *et al*. (2003) obtained higher values for similar soils elswhere. However, lower values were reported by Carroll and Hathaway (1953).

In the soils of SLSi in the Southern agricultural zone, soil pH (H_2O) in the Ap horizons of mapping unit MF1 varied from 5.7 to 6.6 (X= 6.2), while pH ($CaCl_2$) ranged from 5.0 to 5.9 with a mean of 5.5 (Table 21). In the underlying soils, values of pH (H_2O) were 5.4 in the Crv horizon and 6.1 in the Bt horizon with a mean of 5.7, while values of pH ($CaCl_2$) varied from 4.3 in the transition AB horizon to 5.2 in the Bt horizon (Table 23). The highest and lowest values of soil pH in H_2O were obtained in the same horizons as in pH ($CaCl_2$) with a regular decrease in pH values with increased soil depth.

In soil mapping unit MF2, soil pH (H_2O) ranged from 5.6 to 6.4 (X= 6.0) in the Ap horizons, while pH ($CaCl_2$) ranged from 5.0 to 5.9 with a mean of 5.5 (Table 23). Soil pH (H_2O) in the subsoils ranged from 5.1 in the BC horizon to 5.4 in the transition AB and CB horizons with mean of 5.3, while soil pH ($CaCl_2$) ranged from 4.2 in the CB horizon to 4.4 in the AB and Bt horizons, with mean of 4.3 (Table 23). Soil pH values decreased irregularly with increased depth. However, soil pH ($CaCl_2$) values were about 1 pH unit lower than soil pH (H_2O). In soil mapping unit MF3, soil pH (H_2O) ranged from 6.1 to 6.7 (X=6.4) in the Ap

horizons, while pH ($CaCl_2$) ranged from 5.9 to 6.7 (X= 6.3) (Table 23). In the subsoils, values of pH (H_2O) were 6.1 in the BC horizon and 6.5 in the Cg horizon with mean of 6.3, while soil pH ($CaCl_2$) values were 5.7 in the BC and 6.3 in the Cg horizon (X=6.0) (Table 23). Soil pH in both solutions decreased irregularly with increase in soil depth, however, at every point, soil pH values in $CaCl_2$ were lower than those in H_2O. Also, mean soil pH values decreased from the surface to the subsurface soils presenting lower values in the subsurface soils. Mean soil pH was highest in the surface and subsurface soils of mapping unit MF3 compared to those of other mapping units while the least values were obtained in soils of mapping MF2.

The entire soils, irrespective of mapping units were moderately acid by the scale of Holland *et al*. (1989) and do not indicate that significant amounts of Al^{3+} and H^+ are present to affect plant growth (Esu, 2010). Similar results have been reported by some researchers (Sedov *et al*., 2008, Aksoy *et al*., 2009, Petar *et al*., 2016, and VandenBygaart and Protz, 1994), while Carroll and Hathaway (1953) reported lower values, and Afu *et al*. (2017) reported slightly higher values. However, Irmak *et al.* (2007), Girao *et al*. (2014) and Ferreira *et al*. (2016) reported higher values in Turkey and Brazil than those reported in the present study for similar soils. When carbonate minerals weather, the released Ca^{2+} becomes leached and replaced by H^+, resulting in lower pH values which encourage the solubility of silica, especially at 5 – 9 pH and the insolubility of alumina (Tan, 1998) that later result in the formation of kaolinite (Ollier, 1975). Soils overlying SLS in the northern agricultural zone had comparatively lower 'profile mean' values and appear to be most weathered as soil pH tends to decrease with soil age (VandenBygaart and Protz, 1994).

Soil pH in water correlated positively, strongly and significantly ($p< 5$ %) with pH ($CaCl_2$) as well as exchangeable Ca^{2+} but negatively and weakly with exchangeable H^+ and Al^{3+}. This trend is anticipated as increased soil pH is most likely to increase soil Ca^{2+} content or vis versa, while increasing soil pH will result in a decrease in Al^{3+} and H^+ contents. On the other hand, pH in $CaCl_2$ correlated weakly but positively and significantly ($p< 0.05$) with organic C and available P.

4.3.2 Soil Organic Carbon

In the soils overlying SLS in the northern agricultural zone, soil mapping unit IH1 had organic carbon values that ranged from 7.55 to 40.3 g/kg with a mean of 23.93 g/kg in the A horizons, while values in the underlying soils ranged from 1.37 in the Ckt horizon to 6.18 g/kg in the Bt horizon with mean of 4.23 g/kg (Table 21). In mapping unit IH2, soil organic carbon values varied from 32.24 to 86.64 g/kg with a mean of 59.44 g/kg in the A horizons,

while values in the subsoils were 2.06 in the Cg horizon and 52.39 g/kg in the BC horizon (X= 21.61 g/kg) (Table 23).

Organic carbon regularly decreased with an increase in soil depth, giving rise to higher values in the surface soils. Organic carbon in IH1 was rated medium, while IH2 was rated high based on the scale of Holland *et al*. (1989). However, the poorly drained and low elevation soils of mapping unit IH2 recorded higher values than those in IH1, perhaps due to the slow decomposition of organic matter occasioned by the poorly drained soils in the low lying area. The relatively fallow area with teak trees and leaf droppings may have been responsible for this high value, especially in IH1P2. The accumulation of organic matter in karstic planes encourages the development of mollic epipedons, leading to phaeozems in areas with more moisture (Bautista *et al*., 2011).

The poorly drained and low elevation soils of IH2 experienced deposition and a slow rate of decomposition. The well-drained condition in IH1 encouraged the mineralization of organic matter to simpler organic compounds or mineralized nutrients. The process may have been facilitated by micro-organisms in the well-drained soils. Sedov *et al*. (2008) attributed the accumulation and stabilization of humus in limestone soils to near neutral pH, which results from the dissolution of $CaCO_3$ in the soils. Ferreira *et al*. (2016) reported a similar range of values, while VandenBygaart and Protz (1994) obtained relatively higher values. However, lower values were reported for similar soils in Cross River, Turkey, Brazil, and Iran (Afu *et al*., 2017, Irmak *et al*., 2007, Girao *et al*., 2014, Aksoy *et al*., 2009, Emadi *et al*., 2008).

In the soils overlying SL in the central agricultural zone, soil mapping unit AI1 had organic carbon values that ranged from 8.92 to 10.29 g/kg with a mean of 9.61 g/kg in the topsoils, while the subsoils ranged from 1.72 in the Bt horizon to 4.12 g/kg in the transition AB horizon (Table 22). In soil mapping unit AI2, it had a range of 7.55-9.95 g/kg with a mean of 8.75 g/kg in the A horizons, while values in the subsoils ranged from 1.72 in the BC horizon to 4.12 g/kg in the AB horizon with mean of 3.09 g/kg (Table 22). Values of organic carbon decreased with an increase in soil depth. Organic carbon content in the poorly drained and low elevation soils of mapping unit AI3 ranged from 5.15 to 22.16 g/kg with a mean of 13.66 g/kg in the A horizons, while values in the subsoils ranged from 1.03 in the Cg horizon to 2.78 g/kg in the transition AB horizon with mean of 1.84 g/kg (Table 22). Organic carbon content decreased with increased soil depth in the entire soils, with soils in AI3 recording the highest mean values in the surface soils but the least subsurface soil values compared to AI1 and AI2. The steep slope coupled with high elevation ranges of AI1 and AI2 with somewhat

sandy texture may have encouraged the leaching of organic matter down to the low lying and poorly drained areas through sheet erosion, as earlier reported by Krasilnikov *et al*. (2007). Also, the poorly drained condition may not have given room for mineralization of organic matter compared to AI1 and AI2. Low soil organic matter may be linked to rapid mineralization against humification (Haynes, 2005) or high temperatures that favour mineralization (Ibia, 2001). Similarly, Ferreira *et al*. (2016) identified rapid mineralization as a reason for low organic carbon in limestone soils. Previous studies by Irmak *et al*. (2007) and Emadi *et al*. (2008) in other areas indicated lower values, while higher values were reported by VandenBygaart and Protz (1994) and Girao *et al*. (2014) for similar soils in Ontario and Brazil. These higher values were obtained in temperate regions, which are reported by Esu (2010) to be areas with a low level of organic matter decomposition, thereby encouraging its accumulation.

In the soils of SLSi in the Southern agricultural zone, organic carbon content ranged from 13.73 to 31.96 g/kg with a mean of 17.85 g/kg in the A horizons of mapping unit MF1, while values in the subsoils were 0.69 and 7.21 g/kg in the BC and Bt horizons, respectively with mean of 4.05 g/kg (Table 23). In mapping unit MF2, organic carbon ranged from 20.59 to 46.34 g/kg (X= 33.47 g/kg) in the A horizons, while values in the subsoils were 1.37 in the Cr horizon and 7.89 g/kg in the in the BC horizon (X= 4.91 g/kg) (Table 23). In the poorly drained soils of mapping unit MF3, organic carbon content ranged from 15.08 to 19.39 g/kg (X=17.24 g/kg) in the topsoils, while values in the subsoils ranged from 2.51 to 7.89 g/kg in the Cg and BC horizons, respectively (X= 5.74 g/kg) (Table 23). Contrary to results recorded in other limestone formations, mean surface soil values of organic carbon in the low elevation range of MF3 were the least (17.24 g/kg) compared to MF1 (17.85 g/kg) and MF2 (33.49 g/kg), especially in the surface soils; however, organic carbon decreased regularly with an increase in soil depth in all the pedons studied. Organic carbon was rated high in MF2 and medium in MF1 and MF3 based on the scale of Holland *et al.* (1989); this may be due to either leaf fall or slow organic matter mineralization occasioned by high rainfall in the rainforest area. Relatively higher values were obtained by VandenBygaart and Protz (1994) in Ontario, while lower values were reported by Afu *et al.* (2017), Irmak *et al.* (2007) and Emadi *et al*. (2008) in Cross River, Turkey and Iran, respectively. However, results by Girao *et al*. (2014) and Ferreira *et al*. (2016) in Brazil were similar to those obtained in the present study. The soils contain either umbric or mollic epipedon as values of organic C in the upper 18 cm were higher than 6 g/kg (Esu, 2010).

Organic C in the soils correlated positively, significantly (p= 0.05) and strongly with total N (r^2=0.591), C/N (r^2=0.892), exchangeable Na (r^2=0.959) and K (r^2=0.962), and available P (r^2=0.626) but weakly with CEC, exchangeable Ca and Mg. Though relative, these parameters will increase with increasing soil organic matter. Decomposed organic matter (humus) acts as a source of N and P and provides exchange sites for the exchangeable bases.

4.3.3 Total nitrogen

In the soils over SLS in the northern agricultural zone, total N in the A horizons of mapping unit IH1 had a mean of 1.54 g/kg, while subsoils had a mean of 1.06 g/kg with values ranging from 0.98 in the CB and 1.12 g/kg in the B horizons (Table 21). In the poorly drained soils of mapping unit IH2, total N ranged from 2.66 to 2.8 g/kg with a mean of 2.73 g/kg in the topsoils, while in the underlying soils, the values ranged from 0.56 in the Cg horizon to 2.8 g/kg in the BC horizon with a mean of 1.51 g/kg (Table 21). Total N decreased with increasing soil depth, while soils in mapping unit IH2 showed higher mean values at both depths than those in IH1. The poorly drained soils in IH2 had higher organic carbon content than IH1, and correlated positively and significantly with total nitrogen (r^2=0.591). This suggests that higher organic carbon may be responsible for higher nitrogen in the IH2 compared to IH1. These values are rated medium based on the scale by Holland *et al*. (1989) and are within the range of values obtained by Afu *et al*. (2017) and Samonil (2007) for similar soils in Cross River and the Bohemian Karst region, respectively. Such values indicate that the processes of organic matter mineralization and nutrient immobilization are both active as total nitrogen is neither low nor high. Again, organic matter mediates nitrogen; therefore, both increase or decreases linearly in soil landscape.

In SL soils of the central agricultural zone, total N ranged from 1.12 to 1.68 g/kg with a mean of 1.40 g/kg in the A horizons, while values in the subsoils ranged from 0.28 in the Btv horizon to 1.26 g/kg in the transition AB horizon with mean of 0.79 g/kg for mapping unit AI1 (Table 22). In mapping unit AI2, topsoils had a mean of 1.40 g/kg, while subsoils had a mean of 1.15 g/kg with values ranging from 0.70 to 2.52 g/kg in the BC horizons (Table 22). The poorly drained soils of mapping unit AI3 had a range of 1.54-2.10 g/kg and a mean of 1.82 g/kg in the A horizons, while the subsoils had 0.56 in the transition AB horizon and 1.26 g/kg in the Cg horizons with mean of 0.89 g/kg (Table 22). Total N regularly decreased in most of the pedons studied except in AI3P2. The irregular decrease in the value of total N in AI3 may be traced to the seasonal or periodic deposition of organic materials over time. These soils appear to have the highest values among the soils in this formation, an

indication that organic matter in high elevations (AI1 and AI2) may have been deposited in AI3 or the rate of mineralization was lower here than in AI1 and AI2, a condition that was probably favoured by pH values less than 6.0 due to few or inactive microbes. Bacteria and protozoa, which are responsible for the decomposition of organic matter, are most active at a pH range of 6-7 (Agbede, 2009). These values are also rated medium based on the scale of Holland *et al.* (1989) and indicate a balance between mineralisation and immobilisation processes. These values are within the range of values obtained by Afu *et al.* (2017) and Samonil (2007) for similar soils elsewhere.

In the soils overlying SLSi in the southern agricultural zone, total N ranged from 0.98 to 3.36 g/kg (X= 2.17 g/kg) in the topsoils of mapping unit MF1, while the subsoils had 0.28 in the Crv horizon and 1.26 g/kg in the Bt horizon (X= 0.78 g/kg) (Table 23). In mapping unit MF2, total N ranged from 0.42 to 2.52 g/kg (X= 1.47 g/kg) in the A horizons, while the subsoils had a range of 0.56-4.62 g/kg (X= 1.40 g/kg) (Table 23). In the poorly drained soils of MF3 in the low elevations, total N ranged from 4.2 to 5.58 g/kg with a mean of 4.89 g/kg in the A horizons, while values in the underlying soils ranged from 1.12 in the Cg horizon to 2.10 g/kg in the BC horizon (X= 1.45 g/kg) (Table 23). Total N was highest in the A horizons of the poorly drained mapping unit MF3 (X= 4.89 g/kg) and least in MF2 (X=1.47 g/kg), an indication of the process of deposition or reduced mineralization of organic matter in the poorly drained soils. Values of total N in MF1 and MF2 were rated medium, while MF3 was rated high based on the scale of Holland *et al*. (1989). The topsoils values exceeded 4.5 g/kg specified for soils in the tropics. Decomposition of soil organic matter for the release of N was therefore facilitated by the moderately acid to near neutral pH, which encourages the build-up of soil bacteria and protozoa (Agbede, 2009). The range of values obtained in MF1 and MF2 were within those reported by Afu *et al*. (2017) and Samonil (2007) for similar soils in the Cross River and Bohemian karst region, while values in MF3 appeared higher than those recorded in the previous studies. Furthermore, soils overlying SLSi in the southern agricultural zone had the highest total N values compared to those of the SLS and SL. This indicates that the rate of mineralization was lower in the southern rainforest area than in the southern guinea area.

Total N correlated positively and significantly (p= 0.05) with CEC (r^2=0.596) and exchangeable bases (r2>0.35) except Ca^{2+}. This finding is in agreement with those of Elamin *et al.* (2007). According to them, $CaCO_3$ treatment on soil significantly reduced total N and available P but significantly increased exchangeable Ca. Thus, the presence of $CaCO_3$ in the study areas dominantly decreased total N and available P.

4.3.4 C/N Ratio

In the soils overlying SLS in the northern agricultural zone, values of C/N ratio had ranges of 5-26 and 1-6 in the A horizons and underlying soils with means of 16 and 4, respectively, for IH1. In the poorly drained soils and low elevation soils of IH2, a range of 12-31 with a mean of 22 in the A horizons, while a range of 4-19 and a mean of 14 was obtained in the subsoils (Table 21).

In the soils overlying SL in the Central agricultural zone, soils in AI1 had a range of 6-8 with a mean of 7 in the topsoils, while the underlying soils had a range of 2-7 and a mean of 3. In AI2, values were generally low, with a range of 1-7 in the entire soils (Table 22). In the poorly drained and low elevation soils of AI3, 3-11 and 1-5 were obtained in the top and subsoils, respectively (Table 22).

In the SLSi soils in the southern agricultural zone, values with ranges of 4-22 and 1-12 in the A horizons and underlying soils, respectively, while means of 8 and 5 were obtained in MF1 (Table 23). In MF2, ranges of 18-49 and 2-9 were obtained in the top and subsoils, respectively. However, the poorly drained and low elevation soils of MF3 had very low values with ranges of 3-4 and 2-6 in the top and subsoils, respectively (Table 23).

Irrespective of limestone formations and elevation ranges, C/N ratios decreased in values with increasing soil depth. Young and Aldag (1982) attributed such a decrease to decreasing organic carbon content. The narrow C: N ratios with soil depth reflect high levels of microbial activities and rapid decomposition of organic matter in the A horizons with a concomitant release of nutrient elements into soil solution for crop plant uptake (Akpan-Idiok, 2012) or leaching. Excess carbon compared to nitrogen, especially in IH1P2, IH2P2 and MF2P2, which was indicated by C/N ratios higher than the separating index of 25 (Paul and Clark, 1989), is not very healthy for the soil environment. In such a situation, soil micro-organisms tend to draw any available soil nitrogen in the proper proportion to make use of available carbon to complete the nitrogen cycle and continue decomposition. This is known as “robbing” the soil of nitrogen and delays the availability of nitrogen as a plant nutrient. High values of the C/N ratio typify uncultivated soils (SSS, 2014b). However, in the majority of the soils, values of the C/N ratio were less than 25. This indicates the mineralization of organic matter to release inorganic N and NH_4^+. This process adds to the already available N and is a healthy process.

The ratio of C to N correlated positively, strongly and significantly ($p< 0.05$) with available P ($r^2=0.596$) and the monovalent exchangeable bases ($r^2>0.80$). This implies a

mutual relationship as the values of the C/N ratio are most likely to increase with an increase in P, Na^+ and K^+.

4.3.5 Available phosphorus

In the soils overlying SLS in the northern agricultural zone, available P ranged from 7.94 to 15.89 mg/kg with a mean of 11.92 mg/kg in the A horizons, while values in the underlying soils ranged from trace to 7.94 mg/kg in the Bt horizon with a mean of 4.5 mg/kg (Table 21). The distribution of available P values showed a decrease with soil depth in mapping unit IH1.

In the poorly drained and low elevation soils of mapping unit IH2, available P had a range of 7.94-19.86 mg/kg (X= 13.9 mg/kg), while values in the underlying soils ranged from 3.97 in the BC horizon to 19.86 mg/kg in the BC horizon with mean of 12.58 mg/kg (Table 21). Available P values were slightly lower in the subsoils, with values that regularly decreased in IH2P2 and irregularly in IH2P1. Soils in the poorly drained IH2 had comparatively higher values (3.97-19.86 mg/kg) than those of IH1 (0-15.89 mg/kg) in both top and subsoils. This is probably because the pH values (6.1-6.7) of the soils in IH2 encouraged organic matter decomposition *via* the support of soil bacteria (Agbede, 2009). These values were rated low based on the scale of Holland *et al*. (1989) and indicate that available P may either have been leached or present in unavailable forms. Carbonatic soils have also been associated with P-fixation (USDA, 1995). Higher values were obtained by Afu *et al*. (2017) for similar soils in Cross River, while Ferreira *et al*. (2016) reported lower values for similar soils in Brazil.

In the soils overlying SL in the central agricultural zone, available P had a range of 10.59-18.53 mg/kg (X= 14.56 mg/kg) in the A horizons of mapping unit AI1, while values in the underlying soils ranged from trace values to 6.62 mg/kg (X= 3.31 mg/kg) (Table 22). Available P values decreased continuously from the A horizons downwards to the last horizon. In mapping unit AI2, values of available P ranged from 7.94 to 13.24 mg/kg with a mean of 10.59 mg/kg in the A horizons, while values in the subsoils varied from 1.32 in the BC horizon to 9.28 mg/kg in the Bt horizon with a mean of 4.77 mg/kg (Table 22). Available P values decreased regularly with soil depth resulting in overall lower values in the subsoils. In the poorly drained and low elevation soils of mapping unit AI3, available P had a range of 7.94-15.89 mg/kg (X= 11.92 mg/kg) in the topsoils, relative to values in the subsoils, which varied from 3.97 to 9.28 mg/kg (X= 6.18 mg/kg) (Table 22). Higher values were therefore obtained in the surface soils, though decreasing irregularly with soil depth. The topsoils of AI3 (11.92 mg/kg) had values that were intermediate between the topsoils of AI1 (14.56

mg/kg) and AI2 (10.59 mg/kg), while the subsoils were lower relative to the surface soils of all the mapping units in soils over the SL. The soils were rated low in available P based on the scale of Holland *et al*. (1989), a condition which may have been facilitated by pH values greater than 6.0, especially in the high elevation soils of AI1 and AI2. According to Agbede (2009), pH values greater than 6.0 encourage the activities of decomposing bacteria. Higher values were obtained by Afu *et al*. (2017) for similar soils in Cross River, while Ferreira *et al*. (2016) reported lower values in Brazil for similar soils.

The soils formed on SLSi in the Southern agricultural zone had values of available P ranging from 3.97 to 7.98 mg/kg with a mean of 5.96 mg/kg in the topsoils, while values in the underlying soils ranged from trace values to 90.02 mg/kg in the BC horizons with mean of 21.97 mg/kg in mapping unit MF1 (Table 23). Available P increased irregularly with soil depth. In mapping unit MF2, available P ranged from 6.62 to 10.59 mg/kg with a mean of 8.61 mg/kg in the A horizons, while values in the subsoils had values ranging from trace quantities in the BC horizon to 9.28 mg/kg in the transition AB horizon with a mean of 3.53 mg/kg (Table 23). The results show that available P decreased continuously with soil depth, leaving the topsoils higher in their values than the underlying soils.

In mapping unit MF3, the low elevation soils had available P with a range of 5.3-31.77 mg/kg and a mean of 18.54 mg/kg in the A horizons, while the endopedons had values ranging from 13.24 mg/kg in the Cg horizon to 30.45 mg/kg in the BC horizons with mean of 20.74 mg/kg (Table 23). Available P increased in MF3P1 and decreased regularly in MF3P2, while mean surface values showed a general increase in available P with a decrease in elevation ranges, resulting in the highest value in mapping unit MF3. A trend that may be otherwise attributed to the enrichment of the low elevation range by the nutrient or favourable pH range (6.0-7.0) for decomposing bacterial activities. Available P was rated low in the higher elevation soils of SLSi represented by MF1 and MF2, and medium in the low elevation soils of MF3 based on the scale of Holland *et al*. (1989). Afu *et al*. (2017) reported values in Cross River higher than those obtained in MF1 and MF2 but similar to those of MF3, while a report by Ferreira *et al*. (2016) on similar soils in Brazil indicate values within the range obtained in the present study; however, lower values were reported by Sedov *et al*. (2008) for similar soils in Yukatan.

Available P correlated positively and significantly ($p< 0.05$) with saturated hydraulic conductivity ($r^2=0.482$), total porosity ($r^2=448$), soil pH ($r^2=0.159$), organic C ($r^2=0.626$) and the exchangeable bases ($r^2>0.35$) except Mg^{2+}. Available P should therefore increase as the values of the above parameters increase. According to Asadu *et al.* (2020), the availability of

P relies on soil pH. Available P is most likely to increase with an increase in soil pH. Nevertheless, a strong negative correlation was found between available P and bulk density and suggest a decrease in values when the values of bulk density are high.

4.3.6 Exchangeable Bases

a. Exchangeable Ca

In the soils overlying SLS in northern agricultural zone, exchangeable Ca had values ranging from 3.4 to 5.4 cmol/kg with a mean of 4.4 cmol/kg in the A horizons, while values in the underlying horizons ranged from 3.8 in the CB horizon to 7.2 cmol/kg in the CBk horizon with a mean of 5.6 cmol/kg (Table 21). Mean values increased in the subsoils especially in the carbonate and clay enriched horizons of CBk and Ckt for soils in the high elevation range represented by mapping unit IH1 relative to the topsoils with mean of 4.4 cmol/kg. In the poorly drained and low elevation range soils represented by mapping unit IH2, exchangeable Ca ranged from 7.2 to 8.2 cmol/kg (X= 7.7 cmol/kg) in the A horizons relative to the subsoils with values ranging from 4.4 in the BC horizon to 7.6 cmol/kg in the Cg horizons and mean of 5.6 cmol/kg (Table 21). The soils were rated medium in exchangeable Ca based on the scale of Holland *et al.* (1989). Higher values were obtained in the subsoils of IH1, and indicate that the underlying parent rock may have contributed to the chemical make-up of the soils through dissolution of carbonates, weathering of feldspars and ferromagnesian minerals (Irmak *et al.*, 2007). Previous studies indicate lower values by Girao *et al*. (2014) in Brazil, while higher values for similar soils were reported by Ferreira *et al*. (2016) in Brazil, Aksoy *et al*. (2009) in Turkey, Markoski *et al*. (2018) in Macedonia and Irmak *et al*. (2007) in Turkey.

The poorly drained surface soils of IH2 had values which seemed to be higher in exchangeable Ca than those in the well-drained IH1. This may perhaps be because the young soils in IH2 may not have had sufficient time to leach the nutrients through weathering or clay illuviation which does not seem to be occurring reasonably has not offered exchangeable Ca the opportunity to be co-migrated to the subsurface soils.

In the soils formed on SL in Central agricultural zone, exchangeable Ca ranged from 3.6 to 3.8 cmol/kg with mean of 3.7 cmol/kg in the A horizons, while values in the subsoils ranged from 0.8 to 6.0 cmol/kg in the Ckt horizons with mean of 2.37 cmol/kg (Table 22) in AI1. Though mean values decreased in the subsoils, exchangeable Ca seems to have been accumulated in the clay-illuviated parts of the underlying soils of BA and Ckt. In mapping unit AI2, exchangeable Ca ranged from 1.6 to 2.4 cmol/kg with mean of 2.0 cmol/kg in the A horizons, while values in the subsoils ranged from 1.6 in the AB horizon and 5.6 cmol/kg in

the Ckt horizon with mean of 3.64 cmol/kg (Table 22). Exchangeable Ca increased in the subsoils with high values (3.6 and 5.6 cmol/kg) in the clay illuviated horizons of Bt and Ckt horizons, respectively. This points to the co-migration of clay with Ca down to the subsoils. In the poorly drained soils of AI3, exchangeable Ca ranged from 1.4 to 1.8 cmol/kg with a mean of 1.6 cmol/kg in the A horizons, while values in the subsoils ranged from 0.6 in the transition AB horizon to 3.6 cmol/kg in the Cg horizon with a mean of 2.07 cmol/kg (Table 22). These soils are young, with weak indications of clay movement to the subsurface soils and low evidences of exchangeable Ca accumulation in the subsurface soils. Values of exchangeable Ca were rated low in AI1, AI2 and AI3 based on the scale of Holland *et al*. (1989) and may have been leached sequel to the high rainfall amount, and relatively steep landscape, especially in the high elevation areas. Comparatively higher values were obtained by Irmak *et al*. (2007) in Turkey, Girao *et al*. (2014) in Brazil, Markoski *et al*. (2018) in Macedonia and Ferreira *et al*. (2016) in Brazil for similar soils, than those obtained in the present study. Such high values may be due to low and delayed rainfall regime, as well as juvenile nature of the soils.

In the soils overlying SLSi in southern agricultural zone, exchangeable Ca in the area ranged from 1.4 to 6.0 cmol/kg with mean of 3.7 cmol/kg in the A horizons of mapping unit MF1, while values in the subsoils were 0.4 in the plinthic horizon of Crv and 5.6 cmol/kg in the BC horizon and a mean of 2.6 cmol/kg (Table 23). The distribution of exchangeable Ca within the profiles shows that clay may have co-migrated with Ca as values seemed higher in the Bt horizon. In mapping unit MF2, exchangeable Ca ranged from 2.2 to 7.2 cmol/kg (X=4.7 cmol/kg) in the topsoils, while values in the subsoils were 0.8 in the CB horizon and 3.0 cmol/kg in the Cr horizon with a mean of 1.3 cmol/kg (Table 23). Mean values decreased in the underlying soils and shows that there have not been significant movement of exchangeable Ca to the subsoils or the cation may have been lost to regions beyond plant root zone. The poorly drained soils of mapping unit MF3 had values of exchangeable Ca ranging from 7.4 to 9.0 cmol/kg with mean of 8.2 cmol/kg in the topsoils, while values in the subsoils ranged from 2.2 in the Cg horizon to 5.0 cmol/kg in the BC horizons with mean of 3.7 cmol/kg (Table 23). According to the rating scale of Holland *et al*. (1989), values in the high elevation ranges of MF1 and MF2 were low, while those in the low elevation areas of MF3 were rated medium. This indicates that exchangeable Ca in the higher elevation ranges may have been leached to enrich the low lying and poorly drained soils of MF3. Such medium to low values imply that, carbonates in MF1 and MF2 in this limestone formations have been influenced by weathering, dissolution and leaching. The mean values show decrease in

exchangeable Ca in the underlying soils with values that decreased regularly with increase in profile depth; this affirms that the soils in the poorly drained areas are still young and may not have had sufficient time for the leaching of exchangeable Ca away from the surface soils or for it to be moved in company of clay to the subsurface soils.

In the soils overlying SLS in northern agricultural zone, mean surface values increased with decrease in elevation ranges with the highest values occurring in soils of the lowest elevation range (mapping unit MF3); an indication that leached exchangeable Ca in elevated areas had enriched the low lying areas. Higher values were reported by Irmak *et al.* (2007) in Turkey and Markoski *et al*. (2018) in Macedonia, while Girao *et al*. (2014) and Ferreira *et al*. (2016) reported similar results for similar soils in Brazil.

Exchangeable Ca dominated the soils' exchange complex irrespective of the limestone formations. Many authors (Spalevic *et al*., 2017., Mukaetov *et al*., 2000 and Djordjevic, 1993) had pointed out that exchangeable Ca dominated the adsorption complex of soils developed from limestone. It is therefore, not surprising to have it dominate the soils under study. The cation is involved in the creation of stable granular structures, especially in the surface soils of the present study and may create insoluble Ca-humates from neutralized minerals and humic acids (Markoski *et al*., 2018). Low exchangeable bases on the exchange complex in all the soils, is an indication that leaching may have occurred (Yakubu and Ojanuga, 2013) due to high rainfall or sandy soil textures associated with the area.

Exchangeable Ca correlated positively and significantly ($p<0.05$) with CEC (r2=0.429), BS ($r^2= 0.436$), CCE ($r^2= 0.413$) and available P ($r^2= 0.369$). Increasing the value of Ca *via* the addition of soil amendment may bring about an increase in the above parameters, though to a weak extent.

b. Exchangeable Mg

In the soils over SLS in Northern agricultural zone, exchangeable Mg had means of 3.4 and 2.5 cmol/kg in the top and subsoils, respectively with values ranging from 1.4 in the CBk horizon to 4.0 cmol/kg in the AB horizon of the subsoils for mapping unit IH1 (Table 21). In the poorly drained soils of mapping unit IH2, exchangeable Mg ranged from 4.2 to 5.0 cmol/kg (X= 4.6 cmol/kg) in the A horizons, while values in the underlying soils were 1.4 and 4.2 cmol/kg in the Cg horizons with mean of 2.6 cmol/kg (Table 21). Such values are rated high on the scale of Holland *et al*. (1989), especially in the A horizons. Exchangeable Mg decreased irregularly with soil depth in all the soils studied as the AB and Ckt horizons recorded relatively higher values. Consequently, values in the poorly drained soils of IH2 were higher in exchangeable Mg than those in IH1. Possibly because of enrichment of Mg

from leached areas in higher elevation ranges. Values obtained in the present study are higher than those obtained by Afu *et al.* (2017) in Cross River, Girao *et al*. (2014) and Ferreira *et al*. (2016) in Brazil for soils on similar parent materials, while those obtained by Samonil (2007) in the Bohemian karst region were higher than those obtained in the present study. However, similar values were obtained by Markoski *et al*. (2018) in Macedonia.

In the soils overlying SL in Central agricultural zone, exchangeable Mg values ranged from 0.6 to 1.0 cmol/kg with a mean of 0.8 cmol/kg in the A horizons, while values in the underlying horizons were 0.2 in the Bt horizon and 2.2 cmol/kg in the Ckt horizon (X= 0.9 cmol/kg) for soils in mapping unit AI1 (Table 22). Mean values showed a slight increase in exchangeable Mg in the subsoils. In mapping unit AI2, exchangeable Mg ranged from 0.8 to 1.4 cmol/kg with a mean of 1.1 cmol/kg in the A horizons, relative to values in the subsoils which had 0.2 in the BC horizon and 2.2 cmol/kg in the Ckt horizon (X= 0.8 cmol/kg) (Table 22). Unlike values in AI1, exchangeable Mg in AI2 decreased slightly in the subsoils, while the highest value of 2.2 cmol/kg occurred in the Ckt horizon. In the poorly drained and low elevation soils of mapping unit AI3, exchangeable Mg ranged from 0.2 to 0.4 cmol/kg with mean of 0.3 cmol/kg in the A horizons, while values in the underlying horizons ranged from 0.6 to 1.2 cmol/kg in the transition AB horizons with mean of 0.86 cmol/kg (Table 22). The values of exchangeable Mg increased in the subsoils of mapping unit AI3 with clear indications that mean of 0.3 cmol/kg obtained in the A horizons of AI3 was the least value compared to soils in higher elevation ranges with the highest value occurring in the Ap horizon of AI2. The soils are rated low in exchangeable Mg using the scale of Holland *et al*. (1989) and contradicts higher values obtained in the SLS. Comparatively, higher values were obtained by Afu *et al.* (2017) (1.8-3.2 cmol/kg) in Cross River, Girao *et al*. (2014) (0.9-10.2 cmol/kg), Ferreira *et al*. (2016) (0.9-7.5 cmol/kg) in Brazil, Samonil (2007) (28-245.4 cmol/kg) in the Bohemian karst region, and Markoski *et al*. (2018) (X= 4.58 cmol/kg) in Macedonia for similar soils.

In the soils of SLSi in southern agricultural zone, exchangeable Mg ranged from 0.6 to 0.8 cmol/kg with mean of 0.7 cmol/kg in the surface soils, while values in the subsurface soils were 0.4 and 1.4 cmol/kg in the Bt horizons with a mean of 0.88 cmol/kg in mapping unit MF1 (Table 23). Mean values showed a slight increase in the endopedons, though irregularly distributed with soil depth. In mapping unit MF2, exchangeable Mg ranged from 0.6 to 3.2 cmol/kg with a mean of 1.9 cmol/kg in the A horizons, while the underlying soils had values ranging from 0.4 in the AB horizon to 4.8 cmol/kg in the Cr horizon (X= 1.73 cmol/kg) (Table 23). Mean value decreased slightly in the subsoils though irregularly

distributed in the soil profiles. In the poorly drained mapping unit MF3, exchangeable Mg varied from 0.8 to 2.8 cmol/kg with a mean of 1.8 cmol/kg in the A horizons, while values in the underlying soils were 0.2 and 0.8 cmol/kg in the Cg horizons with a mean value of 0.47 cmol/kg (Table 23). From the foregoing, the highest mean value occurred in mapping unit MF2 and was rated medium by the scale of Holland *et al.* (1989), while soils in MF1 and MF3 were rated low in exchangeable Mg. These values were relatively lower than those obtained by Afu *et al.* (2017) in Cross River, Samonil (2007) in the Bohemian karst region and Markoski *et al*. (2018) in the Macedonia. However, soils of MF2 were similar in values to reports by Girao *et al*. (2014) and Ferreira *et al*. (2016) on similar soils in Brazil. Higher rainfall amount in the tropical rainforest areas of SL and SLSi may have led to the dissolution and leaching of exchangeable Mg from the soils, hence the comparatively lower rating. Rainfall amount and relative humidity in the SLS were however lower than obtained in the SL and SLSi areas.

Exchangeable Mg^{2+} correlated positively and significantly with exchangeable Ca^{2+} (r^2= 0.686), CEC (r^2= 0.445) and base saturation (r^2= 0.396). An increase in Mg^{2+} containing compounds may bring about an increase in CEC, base saturation and exchangeable Ca^{2+}.

c. Exchangeable potassium

Soils overlying SLS in northern agricultural zone, had values of exchangeable K ranging from 0.07 to 0.14 cmol/kg with a mean of 0.11 cmol/kg in the A horizons of mapping unit IH1, while values in the subsoils were 0.05 in the B and C horizons, and 0.07 cmol/kg in the Bt and transition CB horizons (X= 0.06 cmol/kg) (Table 21). In the A horizons, values were higher in all pedons than the endopedons and decreased with increase in soil depth. In the poorly drained soils of mapping unit IH2, exchangeable K ranged from 0.12 to 0.29 cmol/kg with a mean of 0.21 cmol/kg in the A horizons, while values in the subsoils were 0.05 in the Cg horizons and 0.15 cmol/kg in the transition BC horizon with a mean of 0.09 cmol/kg (Table 21). Values of exchangeable K in the topsoils of mapping unit IH2 almost doubled that of IH1 with higher values also recorded in the subsoils of IH2 than IH1. The soils are low in exchangeable K based on the scale of Holland *et al*. (1989), and indicates that the soils may either be low in mica (Khresat and Taimeh, 1998) or K is leached away from plant root zone (Markoski *et al*. 2018). Exchangeable K is low in limestone soils and has small capacity of adsorption compared to Ca and Mg and often easily leached from plant root zone and soil exchange complex (Markoski *et al*., 2018). A review of available literatures reported by Afu *et al*. (2017) in Cross River, Irmak *et al*. (2017) in Turkey, Girao *et al*.

(2014), Ferreira *et al*. (2017) in Brazil, and Markoski *et al*. (2018) in Macedonia indicates similar ranges of values for similar soils in different locations.

In the soils on SL in central agricultural zone, means of 0.09 and 0.05 cmol/kg were obtained in the A horizons and the underlying horizons, respectively with values decreasing with increasing soil depth in mapping unit AI1 (Table 22). In mapping unit AI2, exchangeable K ranged from 0.07 to 0.09 cmol/kg with a mean of 0.08 cmol/kg in the topsoils, while subsoil values were 0.04 in the BC horizons and 0.06 cmol/kg in the AB horizon (X= 0.048 cmol/kg) (Table 22). Exchangeable K decreased with soil depth in all the studied soils. In mapping unit AI3, exchangeable K ranged from 0.06 to 0.10 cmol/kg with mean of 0.08 cmol/kg in the A horizons, while values in the underlying soils ranged from 0.03 in the Cg horizon to 0.04 cmol/kg in the AB horizon with a mean of 0.033 cmol/kg (Table 22). Exchangeable K seemed to have decreased with a decrease in elevation ranges; with soils in the lowest elevation range (AI3) having the least values. These values were rated very low on the scale of Holland *et al*. (1989), and may be due to active leaching process in the high rainfall region. The values of exchangeable K are lower than those obtained by Afu *et al*. (2017) in Cross River, Ferreira *et al*. (2017) in Brazil, and Markoski *et al*. (2018) in Macedonia, for similar soils.

In the soils on SLSi in southern agricultural zone, exchangeable K ranged from 0.11 to 0.12 cmol/kg with a mean of 0.12 cmol/kg in the A horizon, while values in the underlying soils ranged from trace values to 0.08 cmol/kg with a mean of 0.05 cmol/kg in mapping unit MF1. Values of exchangeable K decreased with an increase in soil depth (Table 23). In soil mapping unit MF2, exchangeable K ranged from 0.12 to 0.15 cmol/kg (X= 0.14 cmol/kg) in the A horizons, while values in the subsoils ranged from 0.05 to 0.09 cmol/kg with a mean of 0.067 cmol/kg and values that decreased regularly with soil depth (Table 23). Similarly, values of exchangeable K in the low lying poorly drained soils of MF3 decreased with soil depth. In the A horizons, values ranged from 0.10 to 0.13 cmol/kg with a mean of 0.115 cmol/kg, while subsoil values varied from 0.03 and 0.09 cmol/kg with a mean of 0.067 cmol/kg (Table 23).

Values of exchangeable K were comparatively higher in soil mapping unit MF2 and generally higher in the Ap horizons of the soils irrespective of the soil mapping unit. The soils were rated low in exchangeable K on the scale of Holland *et al*. (1989). Such low values may be attributed to leaching occasioned by high rainfall in the region. Similar ranges of values were obtained by Afu *et al*. (2017) in Cross River, Ferreira *et al*. (2017) in Brazil, and Markoski *et al*. (2018) in Macedonia for similar soils.

Soils with low total K are regarded as highly weathered and have low mineral reserve, while those with high values are not highly weathered with high mineral reserve in their exchange complex (Yakubu and Ojanuga, 2013).

Exchangeable K correlated significantly and in a like manner as exchangeable Na. From the results of correlation, an increasing value of exchangeable K is likely to bring about an increase in Mg^{2+}, Ca^{2+}, CEC and available P; though weakly so.

d. Exchangeable Na

In the soils overlying SLS in the northern agricultural zone, exchangeable Na in soils of mapping unit IH1 ranged from 0.04 to 0.10 cmol/kg with mean of 0.07 cmol/kg in the A horizons, while values in the underlying soils varied from 0.02 in the AB, CBk and Ckt horizons to 0.03 cmol/kg in the CB and Bt horizons with a mean of 0.024 cmol/kg (Table 21). Values of exchangeable Na decreased with soil depth in this mapping unit. In the poorly drained soils of mapping unit IH2, exchangeable Na ranged from 0.09 to 0.28 cmol/kg with a mean of 0.19 cmol/kg in the surface soils, while values in the subsoils ranged from 0.02 in the BC and Cg horizons to 0.11 cmol/kg in the BC horizon (X= 0.05 cmol/kg) (Table 19). Such values are rated low on the scale of Holland *et al.* (1989), and the soils are not likely to be sodic. Values of exchangeable Na decreased regularly with increased soil depth resulting in higher values in the A horizons. However, the values were comparatively higher in the poorly drained soils of IH2 and indicates that the cation may have been leached from soils of higher elevation ranges and may be potentially sodic if the enrichment process is sustained. Higher values were obtained by Aksoy *et al.* (2009) in Turkey, Samonil (2007) in the Bohemian karst region, and Markoski *et al*. (2018) in Macedonia, while Afu *et al.* (2017) in Cross River, and Ferreira *et al*. (2016) in Brazil reported similar results for similar soils.

In the soils overlying SL in Central agricultural zone, exchangeable Na ranged from 0.05 to 0.06 cmol/kg with a mean of 0.055 cmol/kg in the A horizons, while values in the endopedons had a mean of 0.02 cmol/kg. Subsoils had mean value that was comparatively lower than those of the A horizons of mapping unit AI1 (Table 22). In mapping unit AI2, values of exchangeable Na varied from 0.04 to 0.05 cmol/kg (X= 0.045 cmol/kg) in the A horizon, while in the subsoils, values were 0.01 in the BC horizon and 0.02 cmol/kg in the underlying horizons of AI2P2 with an average of 0.018 cmol/kg. Exchangeable Na in the A horizon appeared slightly higher than values in the subsoils (Table 22). In the poorly drained and low elevation soils of mapping unit AI3, exchangeable Na ranged from 0.03 to 0.06 cmol/kg with mean of 0.045 cmol/kg in the A horizon, while values in the underlying soils had mean of 0.01 cmol/kg (Table 22). The values are over three times lower than values in

the A horizons. Exchangeable Na decreased in the order AI1>AI2>AI3, which implies a decrease in values with decrease in elevation ranges. This is rather surprising as one would expect such values to be higher in the poorly drained soils of mapping unit AI3. The soils are rated very low in exchangeable Na by the scale of Holland *et al*. (1989), and are not likely to be sodic. Higher values had been reported by Aksoy *et al*. (2009) in Turkey, Samonil (2007) in the Bohemian karst region, and Girao *et al*. (2014) in Brazil, while Ferreira *et al*. (2016) in Brazil reported similar values for similar soils.

In the soils over SLSi in the southern agricultural zone, values of exchangeable Na in mapping unit MF1 ranged from 0.06 to 0.07 cmol/kg with a mean of 0.065 cmol/kg in the A horizons, while the underlying soils had values ranging from trace quantities to 0.04 cmol/kg (X= 0.022 cmol/kg) (Table 23). Exchangeable Na values decreased with soil depth leaving the A horizons with higher values. In mapping unit MF2, values of exchangeable Na varied from 0.07 to 0.11 cmol/kg with a mean of 0.09 cmol/kg in the A horizon, while values in the subsoils had 0.02 in the BC horizons and, 0.04 cmol/kg in the Bt and BC horizons (X= 0.03 cmol/kg) (Table 23). Mean values in the A horizons were higher than values in the endopedons.

In the poorly drained soils of MF3, exchangeable Na varied from 0.06 to 0.07 cmol/kg with a mean of 0.065 cmol/kg in the A horizons, while the underlying horizons had values varying from 0.01 to 0.04 cmol/kg in the Cg horizon with mean of 0.03 cmol/kg (Table 23). Again, higher values in the A horizon relative to the endopedons were obtained. Soils in mapping unit MF2 seem to have recorded the highest values compared to those in MF1 and MF3; however, comparatively higher values were obtained in the endopedons of MF2. The soils are rated low in exchangeable Na on the scale of Holland *et al*. (1989). Previous studies by Ferreira *et al*. (2016) in Brazil and Afu *et al*. (2017) in Cross River reported similar results for similar soils, while higher values were reported by Samonil (2007) in the Bohemian karst region, Aksoy *et al.* (2009) in Turkey and Markoski *et al*. (2018) in Macedonia.

Exchangeable Na correlated positively and significantly ($p<0.05$) with other exchangeable bases, CEC and available P. All the same, correlation with exchangeable K was very high (r= 0.987) just like exchangeable Ca^{2+} correlated highly with exchangeable Mg^{2+}. This indicates that bases of the same group in the periodic table of elements correlate better than those of other groups.

4.3.7 Cation Exchange Capacity

In the soils underlain by SLS in Northern agricultural zone, cation exchange capacity by NH_4OAc varied from 27.6 to 28.0 cmol/kg with a mean of 27.8 cmol/kg in the A horizons of mapping unit IH1, while values in the endopedons were 23.6 in the transition CB horizon and 42.8 cmol/kg in the Ckt horizon (X= 33.0 cmol/kg) (Table 21). Mean cation exchange capacity increased in the endopedons, especially in the Ckt horizon where lessivation had occurred. This was the case in the soils found in high elevation areas. In the poorly drained and low elevation soils of mapping unit IH2, CEC had a range of 28.4-54.4 cmol/kg and mean of 44.4 cmol/kg in the A horizons, while values in the endopedons ranged from 15.2 to 40.0 cmol/kg (X=26.8 cmol/kg) (Table 21). Unlike soils in the high elevation ranges (mapping unit IH1), those in the low elevation range (mapping unit IH2) experienced a decrease in mean values in the endopedons with an irregular distribution of values with soil depth. The soils were rated high in CEC by the scale of Holland *et al*. (1989). High values of CEC in IH1 seem to have been contributed by high clay content in the endopedons. This aligns with findings by Irmak *et al*. (2007) on similar soils in the Harran region of Turkey. Emadi *et al.* (2008) however, attributed increasing CEC with soil depth in Iran to transformation of illite to smectite. Higher values in the A horizons of the poorly drained soils may have been contributed by high organic matter content in the A horizons relative to the underlying horizons. Similar values were reported by Aksoy *et al.* (2009) in Turkey and Emadi *et al*. (2008) in Iran. Studies by Afu *et al.* (2017) in Cross River, and Ferreira *et al*. (2016) in Brazil, however contradicted findings in the present study as lower values were reported by them.

In the soils underlain by SL in central agricultural zone, CEC of soil mapping unit AI1 varied from 12 to 12.4 cmol/kg with mean of 12.2 cmol/kg in the A horizons, while values in the subsoils ranged from 17.6 in the B horizons to 19.6 cmol/kg in the Ckt horizon with mean of 18.2 cmol/kg (Table 21). The soils experienced an increase in CEC with soil depth, especially at Ckt horizons where clay seems to have been illuviated, an indication that clay may have been responsible for the increase in CEC with soil depth.

In soil mapping unit AI2, CEC had a range of 4.4-36.4 cmol/kg and mean of 20.4 cmol/kg in the A horizons, while values in the endopedons ranged from 5.2 in the AB horizon to 26.8 cmol/kg in the Bt horizon (X= 13.76 cmol/kg) (Table 22). In AI2P1, values decreased with increase in soil depth, while in AI2P2, values increased with soil depth. However, the distribution of CEC with soil depth was regular. In the poorly drained soils of mapping unit AI3, CEC had a range of 10.0-10.8 cmol/kg and mean of 10.4 cmol/kg in the A horizons,

while values in the subsoils were 6.0 in the transition AB horizon and 30.8 cmol/kg in the Cg horizon with mean of 19.2 cmol/kg (Table 22). Cation exchange capacity values in the subsoils are higher than values in the A horizons; this contradicts results obtained in the soils underlain by SLS in northern agricultural zone for mapping unit IH2. However, soils in mapping unit AI2 had the highest values in the A horizons, while soils in mapping unit AI3 had the highest overall value. The soils were low to moderate in CEC by the scale of Holland *et al*. (1989). The values indicate low soil reactivity resulting from comparatively coarser soils in the region. Higher values of CEC were reported by Emadi *et al*. (2008) in Iran, Aksoy *et al*. (2009) in Turkey, and Irmak *et al*. (2007) Turkey for similar soils, while Afu *et al*. (2017) in Cross River and Girao *et al*. (2014) in Brazil presented lower values for the soils.

Soils over SLSi in the Southern agricultural zone had CEC range of 10.8-39.6 cmol/kg and mean of 25.2 cmol/kg in the A horizons of mapping unit MF1, while values in the underlying soils were 10.8 in the transition AB horizon and 18.8 cmol/kg in the BC horizon and mean of 13.8 cmol/kg (Table 23). Mean values decreased in the endopedons and relatively higher in the BC and Crv horizons. In mapping unit MF2, CEC varied from 18.0 to 41.2 cmol/kg and mean of 29.6 cmol/kg in the A horizons, while values in the endopedons ranged from 16.8 in the Cr horizon to 62.8 cmol/kg in the BC horizon with mean of 38.8 cmol/kg (Table 23). Cation exchange capacity in the endopedons was higher than values in the A horizons, especially in the BC and CB horizons where the highest values were recorded. In the poorly drained and low elevation soils of mapping unit MF3, CEC values had a range of 25.6-40 cmol/kg and mean of 32.8 cmol/kg in the A horizons, while values in the endopedons ranged from 11.2 in the BC to 40.0 cmol/kg in the Cg horizons and a mean of 21.27 cmol/kg (Table 23).

Mean value of CEC in the endopedons was lower as values in the soil profiles decreased irregularly with soil depth. However, CEC was higher in the A horizons and increased with decrease in elevation ranges. This trend is similar to that observed in the soils for organic carbon and suggests that it may have contributed more to the decreasing values with soil depth. Similar values were reported by Irmak *et al*. (2007) and Aksoy *et al*. (2009) in Turkey, while Afu *et al.* (2017) in Cross River and Girao *et al*. (2014) in Brazil recorded lower values. However, higher values were reported by Samonil (2007) in the Bohemian karst region for similar soils elsewhere.

Cation exchange capacity significantly ($p< 0.05$) correlated positively with organic C ($r^2=0.387$), total N ($r^2=0.596$) and the exchangeable cations ($r^2>0.35$). Correlation of CEC

with organic C (r^2=0.387) and not with clay indicates that organic matter contributed more to the soils' CEC.

4.3.8 Exchangeable Acidity

In the soils underlain by shale, limestone and sandstone in northern agricultural zone, exchangeable acidity had a range of 0.8-2.8 cmol/kg and mean of 1.8 cmol/kg in the A horizons of mapping unit IH1 (high elevation), while values in the endopedons ranged from 1.2 in the transition CB horizon to 5.8 cmol/kg in the Ckt horizon (X= 3.76 cmol/kg) (Table 21). Higher values occurred in the endopedons, this is rather surprising especially with the occurrence of calcic Ckt horizon which would have reduced soil acidity. Exchangeable acidity seems to have been jointly contributed by exchangeable Al and H as nearly equal values were obtained with irregular distribution of these ions with soil depth. In the poorly drained and low elevation soils of mapping unit IH2, exchangeable acidity varied from 1.0 to 1.8 cmol/kg and mean of 1.4 cmol/kg in the A horizons, while values in the subsurface soils ranged from 0.8 in the Cg horizon to 6.4 cmol/kg in the transition BC horizon (X= 2.65 cmol/kg) (Table 21).

Exchangeable acidity was also jointly contributed by exchangeable Al and H in the poorly drained soils. Mean values were also higher in the endopedons. Higher values were reported by Markoski *et al*. (2018) in Macedonia, Samonil (2007) in the Bohemian karst region, and Petar *et al*. (2016) in Pivska Planina, while Ferreira *et al*. (2016) in Brazil and Afu *et al.* (2017) in Cross River recorded lower values for similar soils; however, similar values were reported by Girao *et al*. (2014) in Brazil.

Soils overlying SL in Central agricultural zone, had a range of 2.0-3.0 cmol/kg and a mean of 2.5 cmol/kg in the A horizons for exchangeable acidity in AI1. Values in the underlying soils of AI1 ranged from 1.0 in the transition horizon of BA to 9.2 cmol/kg in the Ckt horizon and a mean of 4.1 cmol/kg (Table 22). The trend of exchangeable Al and H ions down the soil profiles showed that exchangeable H ions contributed over 60 % to the exchangeable acidity of the soils. A probable indication that the hydrolytic reaction of Al^{3+} had proceeded at a faster rate leading to the formation of H^+. Consequently, higher values were obtained in the subsoils for exchangeable Al and H ions. In mapping unit AI2, exchangeable acidity ranged from 1.6 in the Ckt horizon to 2.6 cmol/kg in the AB horizon with a mean of 2.16 cmol/kg in the subsoils, while the A horizons had mean of 2.0 cmol/kg (Table 22). A striking feature of soils in this mapping unit is the near sole contribution of exchangeable H^+ to the exchangeable acidity of the soils as most horizons had trace values for exchangeable Al^{3+}. In the poorly drained and low elevation soils of mapping unit AI3,

exchangeable acidity had a range of 2.4-3.4 cmol/kg and a mean of 2.9 cmol/kg in the A horizons, while values in the subsoils ranged from 1.8 in the Cg horizon to 4.6 cmol/kg in transition AB horizon (X= 2.7 cmol/kg) (Table 22). Though the mean exchangeable acidity was higher in the A horizons, the contribution by exchangeable H to the exchangeable acidity complex of the soils was far greater than that of exchangeable Al^{3+}. Soils in mapping unit AI2 had the least values of exchangeable acidity in Agoi Ibami (soils over SL). Higher values were reported by Markoski *et al*. (2018) in Macedonia, Samonil (2007) in the Bohemian karst region, and Petar *et al*. (2016) in Pivska Planina for similar soils, while Ferreira *et al*. (2016) in Brazil and Afu *et al.* (2017) in Cross River recorded lower values; however, similar values were reported by Girao *et al*. (2014) in Brazil.

General observation of soils overlying SL in central agricultural zone is that the exchange complex occupied by exchangeable acidity was dominated by exchangeable H^+ with values of exchangeable Al^{3+} far less than 20 %, an indication that Al toxicity is not a limiting problem to the growth of crops. Furthermore, it may be deduced that the trace to low amounts of Al in the soils was due to relatively low weathering rate or pedogenesis in the area.

In the soils underlain by SLSi in Southern agricultural zone, exchangeable acidity had a range of 2.0-2.2 cmol/kg and mean of 2.1 cmol/kg in the A horizons, while values in the subsoils had 2.6 in the Bt horizon and 5.2 cmol/kg in the plinthic Crv horizon and a mean of 4.12 cmol/kg (Table 23). Mean values increased in the subsoils with exchangeable H^+ contributing to most of the acidity of the soils, and values increasing regularly with increasing soil depth. In mapping unit MF2, exchangeable acidity ranged from 1.6 to 4.0 cmol/kg (X= 2.8 cmol/kg) in the A horizons, while values in the underlying soils were 3.6 in the AB horizon and 34 cmol/kg in the CB horizon with mean of 17 cmol/kg (Table 23). Though exchangeable acidity values seem to have been contributed by Al and H, values in the subsoils of MF2P1 were very high and may be injurious to plant roots and productivity. Soils in mapping unit MF3 had values that ranged from 2.0 to 2.2 cmol/kg in the BC and Cg horizons with mean of 2.1 cmol/kg (Table 23). Trace values were obtained for exchangeable Al in all the horizons, while exchangeable H increased gradually with increasing soil depth. Markoski *et al*. (2018) in Macedonia, Samonil (2007) in the Bohemian karst region and Petar *et al*. (2016) in Pivska Planina reported higher values, while Ferreira *et al*. (2016) in Brazil and Afu *et al.* (2017) in Cross River State obtained lower values for similar soils; however, similar values were reported by Girao *et al*. (2014) in Brazil.

Compared to mapping units MF1 and MF3, exchangeable acidity values were highest in the top and subsoils of mapping unit MF2. Exchangeable Al^{3+} also seems to be trace in most horizons or correspondingly lower than H^+, an indication that hydrolytic process may have proceeded at a faster rate leading to the formation of more H^+.

Exchangeable Al^{3+} correlated strongly, positively and significantly (p= 0.05) with exchangeable H^+ (r^2= 0.910) an indication that the exchangeable Al^{3+} may have been responsible for high H^+ in the soils. Its weak negative correlation with soil pH (r^2= -0.376), Ca^{2+} (r^2= -0.301) and base saturation (r^2= -0.432) points to acid pH, low exchangeable Ca^{2+} and base saturation when Al^{3+} content is high. Exchangeable Al^{3+} present in the form of aluminosilicates and Al^{3+} bearing oxides, may have been hydrolyzed to give rise to $Al(OH)_3$ which increases soil pH. Al toxicity has been described as the main factor limiting plant growth (Gupta *et al*., 2013; Bose *et al*., 2015).

4.3.9 Base Saturation

In the soils over SLS in Northern agricultural zone, base saturation ranged from 24.7 to 32.8 % and a mean of 28.7 % in the A horizons of mapping unit IH1, while values in the underlying soils were 22.1 in the Bt horizon and 27.7 % in the AB horizon with mean of 24.3 % (Table 21). Mean base saturation decreased in the underlying soils but seemed to be higher in the Bt horizon of IH1P1. In soil mapping unit IH2, base saturation had a range of values of 23.9- 43.7 % (X= 33.8 %) in the A horizons, while values in the subsoils were 19.7 in BC and 61.5 % in the Cg horizon and a mean of 35.4 % in the poorly drained soils (Table 21). Mean values seemed to be higher in the subsoils of IH2 than those of IH1. The soils were rated low in base saturation on the rating scale of Holland *et al*. (1989), and likely to be classified as Ultisols if other criteria for Taxonomic classification are met. This indicates that either basic nutrients may not have occurred in available forms in soil solution for plant uptake or the soils may have been highly weathered and lost its basic cations to leaching. Higher values of base saturation were obtained by Samonil (2007) in the Bohemian karst region, Girao *et al*. (2014) and Ferreira *et al*. (2016) in Brazil for similar soils.

In the soils overlying SL in Central agricultural zone, base saturation in mapping unit AI1 had range of 36.7-39.5 % and a mean of 38.1 % in the A horizons, while values in the underlying soils ranged from 8.5 to 43.1 % in the Ckt horizons with mean of 18.1 % (Table 22). Base saturation values decreased with soil depth but increased in the Btv and Ckt horizons with comparatively higher values in the A horizons than in the underlying soils. In mapping unit AI2, base saturation ranged from 9.2 to 70.9 (X= 39.9 %) in the A horizons, while values in the subsoils had 15.2 in the Bt horizon and 51.8 % in the Ckt horizon with

mean of 39.6 % (Table 22). Though with slight overall decrease in the subsoils, an increase in base saturation was recorded in the Ckt horizon. In mapping unit AI3, base saturation ranged from 16.9 to 21.9 % with a mean of 19.4 % in the A horizons, while values in the subsoils ranged from 14.4 in the Cg horizon to 20.8 % in the transition AB horizon with mean of 16.9 % (Table 22). The soils were rated low in base saturation on the rating scale of Holland *et al*. (1989) and likely to qualify as Ultisols. Such low values may be linked to soils that are highly weathered with leached basic cations or the soil nutrients may not have been present in available forms. Base saturation values generally decreased with soil depth and decrease in elevation ranges; consequently, mean values in mapping unit AI3 were the least compared to those of AI1 and AI2. Reports by Samonil (2007) in the Bohemian karst region, Girao *et al*. (2014) and Ferreira *et al*. (2016) in Brazil indicate higher values for similar soils.

The soils overlying SLSi in Southern agricultural zone presented base saturation values with a range of 18.5-22.1 % and mean of 20.3 % in the A horizons, while values in the subsoils had 12.7 % in the plinthic Crv horizon and 48.8 % in the Bt horizon and a mean of 24.9 % in mapping unit MF1 (Table 23). Mean values increased in the subsoils, while the distribution of base saturation values in the profiles were quite irregular. In mapping unit MF2, base saturation ranged from 29.9 to 32.8 % and a mean of 31.4 % in the A horizons, while values in the underlying soils were 12.1 in the Cg horizon and 49.4 % in the BC horizon and a mean of 27.2 % (Table 23). In mapping unit MF3, the soils had values that ranged from 29.9 to 32.8 % (X= 31.4 %) in the A horizons, while subsoils had range of 12.1-49.4 % and mean of 27.2 % (Table 23). Mean values in the A horizons show that soils in MF3 had the highest values for base saturation which decreased as the elevation ranges increased from MF3 to MF1. Exchangeable bases that may have been leached from the higher elevation ranges (MF1 and MF2) may have enriched the low lying poorly drained soils of MF3 which were inturn rated medium. However, values in the high elevation ranges (MF1 and MF2) were rated low by the scale of Holland *et al*. (1989). The values were similar to report by Samonil (2007) in the Bohemian karst region, while results by Ferreira *et al*. (2016) in Brazil indicated higher values. Base saturation was positively significantly ($p<0.05$) correlated with silt ($r^2=0.415$) as well as exchangeable Ca^{2+} ($r^2=0.436$) and Mg^{2+} ($r^2=0.396$).

4.3.10 Calcium Carbonate Equivalent (CCE)

Calcium carbonate equivalent in mapping unit IH1 for the soils overlying shale, limestone and sandstone in northern agricultural zone had a range of 28.6-38.8 g/kg (X= 33.7 g/kg) in the A horizons, while values in the underlying horizons ranged from 22.1 in the transition CB horizon to 36.8 g/kg in the Ckt horizon and a mean of 28.4 g/kg (Table 21).

Though irregularly distributed with increase in soil depth, CCE seemed to have precipitated in the CBk (29.2 g/kg) and Ckt (36.8 g/kg) horizons, hence the recorded highest values in the mapping unit; however, an average decline in value was recorded in the endopedons (22.1-36.8 g/kg). Higher values of CCE in the A horizons is not pedogenic but likely due to wind-blown calcitic dust from nearby areas like the Gboko cement plant in the Markurdi area.

In the poorly drained soils of mapping unit IH2, CCE ranged from 17.8 to 27.0 g/kg (X= 22.4 g/kg) in the A horizons, while values in the subsoils were 17.7 in the Cg horizon and 30.0 g/kg in the transition BC horizon and a mean of 26.3 g/kg (Table 21). These values are less than those reported by Santos *et al*. (2013) in Brazil and Emadi *et al*. (2008) in Iran for similar soils and within range of those obtained by Aksoy *et al*. (2009) in Turkey and Ferreira *et al*. (2016) in Brazil. The low values of CCE may be attributed to the dissolution and leaching of carbonates as a result of the high rainfall regime in the area. The distribution of $CaCO_3$ in the profiles indicates a $CaCO_3$ bulge between 13 and 80 cm depth of IH1P1 and IH2. This is an indication that $CaCO_3$ had been lost from the topsoils and accumulated deep in the soil (Emadi *et al*., 2008) or it may have been lost by dissolution from underlying rock and subsequently precipitated in the B horizons. In addition, with time, $CaCO_3$ may be more precipitated in the B horizons to give rise to calcic or petrocalcic endopedons.

In the soils overlying SL in Central agricultural zone, CCE had a range of 21.5-27.0 g/kg (X= 24.3 g/kg) in the A horizons of mapping unit AI1, while values in the subsoils ranged from 16.0 in the transition BA horizon to 30 g/kg in the Ckt horizon (X= 21.7 g/kg) (Table 22). Values of CCE seem to decrease in transition BA and AB horizons (indicating weakly formed B horizons) and then increased gradually and regularly up to the Ckt horizons with higher values being recorded in the topsoils. In mapping unit AI2, CCE in the A horizons ranged from 21.5 to 28.2 g/kg with mean of 24.9 g/kg, while values in the subsoils were 15.8 in the AB horizon and 30.0 g/kg in the Ckt horizon (X= 25 g/kg) (Table 22). Calcium carbonate equivalent values in AI2P2 decreased in the AB horizon and then increased gradually with soil depth resulting in slight increase in the mean subsoil values. The decrease in values of CCE at the transition horizons of BA, AB or BC of mapping units AI1 and AI2 suggest active leaching and dissolution of carbonates in the transition horizons. The dissolution and leaching processes may have culminated in the precipitation of the carbonates in the Ckt horizons. Mitkova and Markoski (2019) had recently, identified the processes of dissolution and leaching of carbonates as dominant processes in the limestone soils of Marcedonia.

In the poorly drained soils of mapping unit AI3, CCE varied from 27.5 to 28.2 g/kg and a mean of 27.9 g/kg in the A horizons, while values in the subsoils ranged from 20.7 to 28.1 g/kg in the transition AB horizons (X= 24.7 g/kg) (Table 22). Calcium carbonate equivalent values generally decreased with soil depth in mapping unit AI3; however, mean values in the top and subsoils indicate a steady but gradual increase in CCE with a decrease in elevation ranges. These values are low and indicate that calcium carbonates may have been leached due to high rainfall often experienced in the area. These values are within the range of those obtained by Aksoy *et al*. (2009) in Turkey and Ferreira *et al*. (2016) in Brazil but less than those reported by Santos *et al*. (2013) in Brazil and Emadi *et al*. (2008) in Iran for similar soils.

In the soils overlying SLSi in Southern agricultural zone, CCE in mapping unit MF1 had a range of 19.2-26.0 g/kg (X= 22.6 g/kg) in the A horizons, while values in the subsoils were 18.3 in the transition AB horizon and 24.9 g/kg in the Bt horizon and a mean of 22.0 g/kg (Table 23). Mean values decreased slightly in the subsoils, while soil profile distribution of CCE showed irregular decrease in values. In soil mapping unit MF2, CCE in the A horizons ranged from 20.3 to 20.8 g/kg with mean of 20.6 g/kg, while values in the subsoils ranged from 15.2 in the transition AB horizon to 26.1 g/kg in the Cr horizon (X= 20.2 g/kg) (Table 23).

Calcium carbonate equivalent increased at the Bt horizons but declined slightly using overall mean value of the underlying soils. In the poorly drained soils of mapping unit MF3, CCE had a range of 6.1-14.5 g/kg and a mean of 10.3 g/kg in the A horizons, while values in the subsoils varied from 22.8 to 45.7 g/kg in the Cg horizons with mean of 30.7 g/kg (Table 23). Values of CCE in soil mapping unit MF3 increased regularly with soil depth. However, it showed a decrease with decrease in elevation ranges leaving soils in MF1 with the highest values of CCE in the A horizons, while MF3 had the highest values in the subsoils. Values obtained in the present study are lower than those obtained by Emadi *et al*. (2008) in Iran and Santos *et al*. (2013) in Brazil for soils of similar origin.

On a general note, CCE decreased from the SLS in northern agricultural zone to the SLSi in the southern agricultural zone. For instance, the surface soils of the well-drained soils of the higher elevation ranges were such that; IH1>AI2>AI1>MF1>MF2, while the subsoils presented the trend; IH1>AI2>MF1>AI1>MF2. However, the A horizons of the poorly drained soils of the lowest elevation ranges indicates that; AI3>IH2>MF3, while the trend MF3>IH2>AI3 was obtained for the endopedons of the poorly drained soils. The trends suggest that, for rthe well-drained areas, soils of IH1 had the highest CCE values and were

the least affected by leaching or carbonate dissolution but with the highest calcium carbonate precipitation, perhaps due to the low rainfall amount and relatively higher atmospheric temperature. On the other hand, MF2 had the least value of CCE and experienced more leaching and carbonates dissolution with the least precipitation of calcium carbonates. The climate of MF2 is dominantly tropical rainforest. This is an indication that rainfall which is higher in the southern agricultural zone, may have contributed to the dissolution and leaching of carbonates in the area. Also, literature has identified limestone in the Akamkpa area as being amongst the purest in the country (Fatoye and Gideon, 2013); being a limiting factor to soil development (Esu, 2010), the rate at which it weathered to release carbonates may be very low, hence, the comparatively lower values in the area.

4.3.11 Factor Analysis for Variable Loadings: Chemical Properties of the Soils

Fourteen variables classified as chemical properties were reduced to 3 PCs: PC1, PC2 and PC3. The factor loadings are presented in Table 24.

The three most influential PCs contributed 55.6 % of the total variation with PC1, PC2 and PC3 contributing 25.7, 18.4 and 11.5 %, respectively to soil variation between the lithologies. The first PC was dominated by contributions from soil organic carbon (0.93), exchangeable Na (0.91) and K (0.89), with exchangeable Ca (0.65) and calcium carbonate equivalence (0.06) contributing very low to PC1. This is obviously due to the similarity in soil properties of the various limestone formations.

Unlike the dominance of negative correlations amongst the physical soil properties in PC1, the chemical properties were dominated by positive correlations except for the exchangeable acidity parameters of Al^{3+} and H^{+}. In PC2, variation in soils was mainly due to the influence of exchangeable Al^{3+} (-0.84), H^{+} (-0.77) and pHw (0.63). Like in the contributions to loadings in physical soil properties, soils over ALV contributed the highest loadings to variation in PC1. In PC2, soils over SLSi (-0.70) contributed the highest loading, while in PC3, soils over SLS contributed 0.70 to the loadings. There appears to be a perfect contrast between the best contribution to factor loading in PC2 (SLSi) and the best contribution to factor loading in PC3 (SLS). These formations are both of extreme climates and locations in Cross River State.

4.4 Forms of Fe and Al in the Soils

Forms of Fe and Al are presented in Tables 25 and 26, and discussed with respect to soils overlying, SLS in the Northern agricultural zone, SL in the central agricultural zone and SLSi in the southern agricultural zone, respectively.

Table 24: Factor Analysis for Variable Loadings: Chemical Properties of the Soils

Variables	PC1	PC2	PC3
PHw	0.156	0.633	0.300
SOC	0.929	-0.115	-0.208
TN	0.569	-0.189	0.065
C/N	0.614	-0.127	-0.395
Na	0.906	-0.126	-0.222
K	0.893	-0.150	-0.253
Mg	0.589	-0.058	0.530
Ca	0.648	0.316	0.413
CEC	0.389	-0.560	0.487
BS	0.148	0.534	0.099
Al3+	-0.179	-0.838	0.320
H+	-0.231	-0.773	0.279
av.P	0.194	0.132	-0.210
CCE	0.060	0.284	0.363
SLS	0.122	0.162	0.705
AL	0.422	0.174	-0.113
SL	-0.381	0.404	-0.158
SLSi	-0.141	-0.704	-0.256
Eigenvalus	4.6	3.3	2.1
% total variance	25.7	18.4	11.5
cum eigenvalue	4.6	7.9	10.1
Cum %	25.7	44.2	55.6

Foot note: PC= Principal component

Table 25: Forms of Fe and Al for Soils on SLS and SL

Horizon	Depth Cm	Dith. Forms Fed	Ald	Oxal. Forms Feo	Alo	Cryst. Forms Fed-Feo	Ald-Alo	Deg. of act. Feo/Fed	Alo/Ald	Clay	Co-migration of Fed/clay	Ald/clay
		g/kg								g/kg		
IH1P1		Shale, limestone and sandstone formations										
Ap	0-37	52.5	3.3	0.7	0.7	51.8	2.6	0.013	0.212	280.0	0.187	0.012
CBk	64-130	30.4	3.0	1.2	0.7	29.2	2.3	0.039	0.233	300.0	0.101	0.010
CB	130-195	21.2	2.4	1.0	0.4	20.2	2.0	0.047	0.167	200.0	0.106	0.012
	Mean	34.7	2.9	0.967	0.6	33.73	2.3	0.033	0.204	260.0	0.394	0.011
	CV	46.4	15.8	26.0	28.9	48.3	13.0	53.90	16.50	20.4	36.8	10.2
IH1P2												
Ap	0-42	21.6	2.4	0.4	0.6	21.2	1.8	0.019	0.25	140.0	0.154	0.017
Bt	42-70	22.1	2.7	0.6	0.8	21.5	1.9	0.027	0.296	220.0	0.100	0.012
Ckt	70-150	19.4	1.5	0.5	0.6	18.9	0.9	0.026	0.40	220.0	0.088	0.007
	Mean	21.0	2.2	0.5	0.67	20.5	1.53	0.024	0.315	193.3	0.114	0.012
	CV	6.8	28.4	20.0	17.3	6.9	35.9	18.20	24.30	23.9	30.8	41.7
AI1P1		Sandstone and limestone formations										
Ap	0-9	4.1	0.7	0.1	0.2	4.0	0.5	0.024	0.286	100.0	0.041	0.007
BA	9-54	6.7	1.3	0.2	0.4	6.5	1.9	0.03	0.308	140.0	0.048	0.009
Ckt	122-200	15.5	2.7	0.2	0.9	15.3	1.8	0.013	0.333	260.0	0.060	0.010
	Mean	8.77	1.57	0.17	0.5	8.6	1.4	0.022	0.309	166.67	0.050	0.009
	CV	68.15	65.5	34.64	72.11	69.02	55.79	38.60	7.61	50.0	19.3	17.6
AI1P2												
Ap	0-15	4.7	0.6	0.1	0.2	4.6	0.4	0.021	0.333	120.0	0.039	0.005
Btv	67-146	18.5	2.2	0.3	0.3	18.2	1.9	0.016	0.136	240.0	0.077	0.009
Ckt	146-180	22.1	1.8	0.2	0.3	21.9	1.5	0.009	0.167	280.0	0.079	0.006
	Mean	15.1	1.53	0.2	0.27	14.9	1.27	0.015	0.21	213.3	0.065	0.007
	CV	60.83	54.3	50.0	21.65	61.14	61.32	39.31	50.0	39.03	34.7	31.2
AI2P1												
Ap	0-12	8.3	1.2	0.3	0.3	8.0	0.9	0.036	0.25	120.0	0.069	0.010
BC	59-120+	16.4	2.4	0.4	0.4	16.0	2.0	0.024	0.167	180.0	0.091	0.013
	Mean	12.35	1.8	0.35	0.35	12.0	1.45	0.030	0.21	150.0	0.080	0.012
AI2P2												
Ap	0-13	10.4	1.2	0.3	0.2	10.1	1.0	0.029	0.167	160.0	0.065	0.008
BC	60-82	14.9	1.7	0.3	0.3	14.6	1.4	0.020	0.176	140.0	0.106	0.012
Ckt	82-132	20.8	1.9	0.5	0.3	20.3	1.6	0.024	0.158	220.0	0.095	0.009
	Mean	15.37	1.6	0.37	0.27	15.0	1.33	0.024	0.17	173.3	0.089	0.010
	CV	33.9	22.5	31.5	21.7	34.1	22.9	18.5	5.40	324.0	23.9	21.5

Foot note: Fed: dithionite Fe, Ald: dithionite (dith) Al, Feo: oxalate (oxal) Fe, Alo: oxalate Al, Fed-Feo: crystalline (cryst) Fe, Ald-Alo: crystalline Al, Feo/Fed: degree of activation (deg. of act.) of Fe, Alo/Ald: degree of activation of Al, clay/Fed: co-migration of Fed, clay/Ald: co-migration of Ald

Table 26: Forms of Fe and Al for soils on SLSi

		Dith. Forms		Oxal. Forms		Cryst. Forms		Deg. of act.		Clay	Co-migration of	
Horizon	Depth	Fed	Ald	Feo	Alo	Fed-Feo	Ald-Alo	Feo/Fed	Alo/Ald		Fed/clay	Ald/clay
	Cm	g/kg								g/kg		
MF1P1		Shale and limestone with sandstone intercalation										
Ap	0-17	8.2	1.3	0.8	0.6	7.4	0.7	0.098	0.462	140	0.059	0.009
Bt	17-56	13.1	1.8	0.5	0.6	12.6	1.2	0.038	0.333	160	0.082	0.011
BC	56-121	11.4	1.5	0.3	0.5	11.1	1.0	0.026	0.333	160	0.071	0.009
	Mean	10.9	1.53	0.53	0.57	10.37	0.97	0.054	0.38	153.3	0.071	0.010
	CV	22.8	16.4	47.2	10.2	25.8	26.0	71.4	19.8	7.5	16.3	11.9
MF1P2												
Ap	0-14	7.7	1.2	0.4	0.2	7.3	1.0	0.052	0.167	100	0.077	0.012
Crv	81-153	18.4	3.0	0.2	0.6	18.2	2.4	0.011	0.20	180	0.102	0.017
	Mean	13.05	2.1	0.3	0.4	12.8	1.7	0.032	0.18	140	0.090	0.015
MF2P1												
Ap	0-16	26.0	4.1	0.8	1.3	25.2	2.8	0.031	0.317	140	0.186	0.029
CB	44-97	27.3	5.1	0.7	2.1	26.6	3.0	0.026	0.418	140	0.195	0.039
Cr	97-161	32.9	3.9	1.5	1.3	31.4	2.6	0.046	0.333	140	0.235	0.028
	Mean	28.73	4.37	1.0	1.57	27.73	2.8	0.034	0.36	140	0.205	0.032
	CV	12.8	14.7	43.6	29.5	11.7	7.1	30.3	15.2	0.0	12.7	19.0
MF2P2												
Ap	0-16	7.6	1.4	0.2	0.2	7.4	1.2	0.026	0.143	120	0.063	0.012
Bt	45-107	18.3	4.1	0.3	0.9	18.0	3.2	0.016	0.22	140	0.131	0.029
BC	107-182	18.2	3.7	0.1	0.8	18.1	2.9	0.005	0.216	160	0.114	0.023
	Mean	14.7	3.07	0.2	0.63	14.5	2.4	0.016	0.19	140	0.103	0.021
	CV	41.5	47.5	50.0	59.8	42.4	44.3	67.0	22.5	14.3	34.5	40.4

Foot note: Fed: dithionite Fe, Ald: dithionite (dith) Al, Feo: oxalate (oxal) Fe, Alo: oxalate Al, Fed-Feo: crystalline (cryst) Fe, Ald-Alo: crystalline Al, Feo/Fed: degree of activation (deg. of act.) of Fe, Alo/Ald: degree of activation of Al, clay/Fed: co-migration of Fed, clay/Ald: co-migration of Ald

4.4.1 Forms of Fe

i. Dithionite Extracted Fe

In the soils underlain by SLS in northern agricultural zone, dithionite-citrate extractable Fe (Fed) had means of 34.7 and 21.0 g/kg in IH1P1 and IH2P2, respectively with a regular decrease in IH1P1, while values in IH1P2 increased slightly in the Bt horizon (Table 25). The slight increase in the Bt horizon indicates relative maturity compared to IH1P1 which had a cambic horizon influenced by carbonate accumulation. The comparatively higher mean value in IH1P1 does not depict more intense weathering condition or soil development. It is mainly as a result of higher values in the epipedon arising from low weathering of minerals in the epipedons. According to Juo *et al*. (1974), Fed in most Nigerian soils increases with soil depth. Coefficient of variability (CV) of 46.4 % in IH1P1 and 6.8 % in IH2P2 were obtained. According to Wilding (1985), the CV of IH1P1 is high, while that of IH2P2 is low.

In the soils overlying SL in central agricultural zone, Fed had means of 8.77, 15.1, 16.4 and 15.37 g/kg for AI1P1, AI1P2, AI2P1 and AI2P2, respectively (Table 25). The distribution of Fed with soil depth indicates a regular increase in values with increasing soil depth in all the pedons overlying SL. Felspatic minerals in the Ap horizons may have been weathered and free Fe moved to the endopedons by eluviation. The variability of Fed as shown by CV values of 12.35 – 68.15 % indicates low to high variability (Wilding, 1985) in the soils. By the mean values of each pedon, the trend AI1P1>AI2P2≈AI1P2>AI2P1 may be implied for increasing soil development as the concentration of Fed increases with increasing weathering. However, this may not be totally true for the soils as more Fed seem to have been moved to the Btv horizon in AI1P2 than in any other pedon. This somewhat negates findings by Esu (2010), that soils in low geomorphic areas are younger compared to those in the uplands.

In the SLSi soils of southern agricultural zone, Fed had means of 10.9, 18.4, 28.73 and 14.7 g/kg for MF1P1, MF1P2, MF2P1 and MF2P2, respectively (Table 26). The values of Fed increased in the B horizons with a subsequent decrease in the C horizons, resulting in the formation of outward bulges in MF1P1 and MF2P2. In MF1P2 and MF2P1, the values increased continuously with soil depth. These soils have undergone more advanced weathering and weathered minerals in the epipedons release free Fe which is then moved to the endopedons. The eluviation-illuviation process occurs over time and results in the depletion of the epipedons of Fed. The values of CV for these soils indicate low to high variations with values ranging from 12.8 to 41.5 %. Rating of CV was according to the scale of Wilding (1985).

The Fed in soils over SLSi was such that; MF2P1>MF2P2>MF1P2>MF1P1. According to Schwertmann (1993), increasing concentration of Fe is synonymous with increasing weathering, therefore the trend above indicates that MF2P1 was the most weathered and developed soil. The distribution of Fed reflects variable intensity of weathering at various depths. Higher values of Fed were reported for the Ap horizons of soils over SLS, and the endopedons of soils over SL and SLSi. Higher values of Fed in the endopedons may be interpreted as more developed soils since the movement of Fed occurs over time.

Irrespective of limestone formations and elevation ranges, considering the profile mean values of Fed, decreasing soil development can be stated thus; IH1P1>MF2P1>IH1P2>AI2P2≈AI1P2>MF2P2>MF1P2>AI2P1>MF1P1>AI1P1 for the soils studied. This indicates that soils over SLS in northern agricultural zone were the most weathered and developed, while those over sandstone and limestone, and shale and limestone with sandstone intecalation were similar. These soils both are found in the tropical rainforest. This accords earlier findings that the percentage of free iron increases with increasing weathering and soil age, and decreases with increasing poor drainage (Udo *et al*., 2009, and Dolui and Bera, 2001). Values obtained in IH1P1, IH1P2 and MF2P1 were higher than those obtained by Girao *et al.* (2014) and Irmak *et al*. (2007), while those of other mapping units were within the range of values obtained. The trend in weathering may not be true in all sense as even the adjudged most developed IH1P1, though having the highest concentration of Fed, had no indications of weathered minerals in the epipedons and movement to the endopedons. However, there were possibilities that Fed had moved to subsurface horizons in soils over SLSi. On this basis, soils over SLSi may be regarded as the most developed. Dolui and Bera (2001), observed that the concentration of Fe increased with soil depth and age in some soils. Again, the values of Fed exceeded Feo in all horizons in the soils studied. This indicates that a considerable fraction of Fe is present in crystalline form (Dolui and Mondal, 2007). In the Apodi plateau karst region, Girao *et al*. (2014) reported a range of 2.57-43.91 g/kg for Fed.

Dithionite-citrate extractable Fe correlated positively and highly ($p< 0.05$) with dithionite-citrate extractable Al ($r^2 = 0.712$) as well as the crystalline forms of Fe and Al, that is, Fe_d-Fe_o (crystalline Fe) ($r^2 = 0.987$) and Al_d-Al_o (crystalline Al) ($r^2 = 0.728$). Values of r^2 for correlation of Fe_d with TiO_2 ($r^2= 0.865$), MnO_2 ($r^2= 0.496$) and Fe_2O_3 ($r^2= 0.816$) indicate high probability of the values of these oxides increasing with increasing Fed. Similarly, Fe_d had a weak inverse relationship with Alo/Ald ($r^2= -0.354$), SiO_2 ($r^2= -0.334$) and MR ($r^2= -$

0.472) and strong inverse relationship with K_2O (r^2= -0.685) and WI (r^2= -0.70). These relationships are anticipated as increased weathering brings about an increase in Fe and Al as well as their crystalline forms. In environments of high weathering intensity, SiO_2 and K_2O are often leached, hence the relationship obtained.

ii. Oxalate extracted Fe

Oxalate extracted Fe (Feo) had mean values of 0.967 and 0.50 g/kg in IH1P1 and IH1P2, respectively (Table 25) for the soils over SLS in the northern agricultural zone. Values of Feo increased in the Bt horizons of the soils whereas, low values were obtained in the Ap horizons. Relatively high values of Feo obtained in the B horizons may be due to the presence of amorphous forms of Fe oxides, or the Fe may have been moved in company of organic matter. Comparatively high concentrations of Feo in the endopedons is due to the expressive occurrence of low crystalline hydroxides in the illuvial horizons resulting from the dissolution of minerals in the epipedons by ferrolysis. Such minerals are dissolved in the epipedons and accumulated in the endopedons (da Silva *et al*., 2019). According to Osayande *et al*. (2013) and Obi *et al*. (2009), non-crystalline or amorphous forms of Fe are associated with soil organic matter. According to Juo *et al*. (1974), Feo either decreases or increases with soil depth in Nigerian soils. In temperate soils, McKeague and Day (1966) and McKeague *et al*. (1971), opined that oxalate Fe constitutes 30-60 % of the dithionite – Fe. The irregular distribution of Feo with soil depth is indicated by coefficient of variability greater than 20 %. According to Wilding (1985), such CV is moderate.

For soils developed from SL in the central agricultural zone, mean values of 0.17, 0.20, 0.35 and 0.37 g/kg were reported for AI1P1, AI1P2, AI2P1 and AI2P2, respectively. Values of CVs were greater than 30 % for the soils (Table 25) indicating high variability of values (Wilding, 1985) with soil depth. The distribution of values with depth in all the pits indicate either a bulge in the B horizons (AI1P2) or a regular increase (AI1P1, AI2P1 and AI2P2) in the values with soil depth. Profile mean values indicate that Feo in AI2P2>AI2P1>AI1P2>AI1P1.

For soils overlying SLSi in the southern agricultural zone, profile mean values of 0.53, 0.30, 1.00 and 0.20 g/kg were obtained for MF1P1, MF1P2, MF2P1 and MF2P2, respectively. Values decreased regularly with soil depth in MF1P1 and MF1P2 (Table 26), while an inward bulge was obtained in MF2P1. However, values of Feo increased at the Bt horizon of MF2P2 indicating a possible movement of weathered amorphous minerals to the B horizon. Soil profile values decreased in the order MF2P1>MF1P1>MF1P2>MF2P2. Wilson *et al*. (2017) reported values of Feo less than 2.0 g/kg as low. Similar values were obtained

for the studied soils. Such low values indicate the presence of allophane and imogolite in the soils and rather dominated by crystalline minerals. The low values may also be due to high temperatures which are responsible for low content of amorphous Fe (Juo *et al*. 1974). Raised temperatures enhance dehydration of Fe oxides and causes a shift to a system of lower crystallinity (Sherman *et al*., 1964). In the karst region of Apodi plateau, Girao *et al*. (2014) reported a range of 0.39-4.35 g/kg for Feo in soils over limestone.

Irrespective of limestone formations, the soils in the high elevation areas had values of oxalate Fe occurring in the decreasing order, such that;
MF2P1>IH1P1>MF1P1>IH1P2>AI2P2>AI2P1>MF1P2>MF2P2>AI1P2>AI1P1. This may be regarded as a decreasing order of soil development. According to Seal *et al*. (2006), Feo is more in illuvial horizons of soils. This finding was in-line with the results of MF2P2, while other pedons deviated from their observations.

Oxalate Fe correlated positively and highly (p= 5 %) with P_2O_5 (r= 0.716) and MnO_2 (r^2= 0.624) and weakly so with Fe_2O_3 (r^2= 0.442). It correlated negatively with WI (r^2= -0.459), K_2O (r^2= -0.445) and SiO_2 (r^2= -0.361). Positive relation of Fe_o with P_2O_5 is most likely to bring about an increase in Fe_o as P_2O_5 increases.

iii. Crystalline form of Fe

In the soils overlying SLS in the northern agricultural zone, soil profile means of 33.73 and 20.5 g/kg were obtained in IH1P1 and IH2P2, respectively (Table 25). Values decreased with increasing soil depth with high (48.3 %) and low (6.9 %) CVs in IH1P1 and IH2P2, respectively according to the scale of Wilding (1985). Mean values indicate that IH1P1 is more mature or weathered relative to IH2P2 as higher values were obtained in IH1P1. However, it is observed that weathering had occurred in the epipedon of IH1P2 and some materials had moved to the endopedons. This may warrant one to admit that IH1P2 had more weathering time and was inturn more developed.

In the soils overlying SL in central agricultural zone, 8.6, 14.9, 12.0 and 15.0 g/kg mean values were obtained in AI1P1, AI1P2, AI2P1 and AI2P2, respectively. Crystalline Fe increased continuously with increase in soil depth. Consequently, AI2P2>AI1P2>AI2P1>AI1P1 may be regarded as the order of soil development, with AI2P2 zbeing the most developed. The variability in values with soil depth indicates CV values greater than 30 % for all the profile pits studied, except in IH1P2 where 6.9 % was obtained. This reflects high variability of crystalline Fe in the soils ecept in IH1P2 (Wilding, 1985).

In the soils overlying SLSi in southern agricultural zone, values of crystalline Fe either increased gradually and continuously (MF2P1, MF2P2) or bulged at the B horizon

(MF1P1, MF1P2). Mean values of 10.37, 12.8, 27.73 and 14.5 g/kg were obtained in MF1P1, MF1P2, MF2P1 and MF2P2, respectively (Table 26). Mean values indicate that MF2P1 was more developed and weathered than MF1P1, MF1P2 and MF2P2. Order of soil development can be summarised as MF2P1> MF2P2> MF1P2>MF1P1. Clear outward bulges were obtained in MF1P1 and MF1P2, while values increased continuously with depths in MF1P2 and MF2P2. This negates earlier findings by Esu (2010), that soils in higher geomorphic surfaces are more mature.

Irrespective of limestone formations, profile mean values were of the decreasing order of development, such that;
IH1P1>MF2P1>IH1P2>AI2P2>AI1P2>MF2P2>MF1P2>AI2P1>MF1P1>AI1P1, from most developed or weathered to least developed. Irrespective of elevation ranges, profile mean values indicate that soils on the SLS in northern agricultural zone were more developed or weathered than their counterparts in other limestone formations. The formation of crystalline Fe (Fe_d-Fe_o) is often facilitated by regions with higher temperature and prolonged dry season (4-5 months) (Seal *et al*., 2006). This was a probable reason for the comparatively higher values in the soils of SLS found in the southern guinea savanna area represented by IH1P1 and IH2P2. With increasing soil age, the crystalline iron oxides increase at the expense of poorly crystalline forms (Seal *et al*., 2006). IH1P1 had the highest values of crystalline Fe and returns the oldest or most weathered amongst others. However, there were obvious movement of crystalline form of Fe down the profiles in the tropical rainforest, ie. SL and SLSi.

Crystalline Fe correlated positively and strongly at $p< 0.05$ with crystalline Al (r= 0.712), TiO_2 (r= 0.848) and Fe_2O_3 (r= 0.762) and negatively so with K_2O (r=-0.627) and WI (r= -0.640). Increased weathering intensity brings about an increase in crystalline Fe and hematite, hence the relationship and also brings about the leaching of K_2O from soils. K_2O is regarded as a mobile oxide and is often leached from soils.

iv. Degree of Activation of Fe

In the soils overlying SLS in northern agricultural zone, IH1P1 and IH2P2 had respective mean values of 0.033 and 0.024 (Table 25) with values increasing regularly with soil depth. Higher values in IH1P1 relative to IH1P2 may be described as more active and coincides with higher clay amount and lower organic matter content. Clay amount may therefore, be a probable reason for a higher activity ratio.

In the SL soils of central agricultural zone, mean values of 0.022, 0.015, 0.030 and 0.024 were obtained in AI1P1, AI1P2, AI2P1 and AI2P2, respectively (Table 26).

Comparatively higher values in AI2P2 and AI2P1 may have been due to the higher clay amount obtained in the soils. The degree of activation of Fe was such that, AI2P1>AI2P2>AI1P1>AI1P2. This may be interpreted as AI2P1 having a more active phase which coincides with a higher clay content. A probable indication that clay amount and type may be responsible for a more active soil.

In the soils of SLSi in southern agricultural zone, 0.054, 0.032, 0.034 and 0.016 were obtained as profile mean values for MF1P1, MF1P2, MF2P1 and MF2P2, respectively (Table 26). From the above, it can be deduced that the degree of activity of MF1P1> MF2P1> MF1P2>MF2P2. Highest values of Fe_o/Fe_d were recorded in the B horizons of soils over SLS, while in soils over SL and SLSi, highest values were obtained in the Ap horizons. Similar findings were reported by Jonczak *et al*. (2015) in humus rich and slightly acidified topsoil layers. Acid soil reactions and humic substances do not favour free Fe oxides crystallization (Cornell and Schwertmann, 1979). Wilson *et al*. (2017) reported the ratio Feo/Fed with values of less than 0.18 as low, while Seal *et al*. (2006) reported low values when Feo/Fed is less than 1. Low values indicate that free Fe oxides in the soils are at an advanced stage of crystallinity or aging (Mahaney *et al*., 1991). The authors further implied a dominance of crystalline Fe oxides in the pedogenic Fe fraction of the soils. The entire studied soils irrespective of limestone formation had values of degree of activation less than 0.18, and reflect high crystallinity. In the karst region of Apodi plateau, Girao *et al*. (2014) obtained a range of 0.04-1.14 for Feo/Fed with lower values recorded in the B horizons.

Irrespective of limestone formations, values of active Fe in the studied soils were such that; MF1P1>MF2P1>IH1P1>MF1P2>AI2P1>AI2P2=IH1P2>AI1P1>MF2P2>AI1P2. This trend would be interpreted as an inverse relationship between soil activity and development, such that MF1P1 with the high value of active Fe was the least developed. Least values of degree of activation in MF2P2 and AI1P2 turned most developed or weathered. The soils that are found in the tropical rainforest may have been weathered and lost most of its mobile components. The ratio Fe_o/Fe_d in the solum is low when values of less than one (< 1) are obtained (Seal *et al*., 2006). This indicates that the soils' degree of activation was generally low. Low values indicate that free iron oxides in the soil are at an advanced stage of crystallinity or aging (Mahaney *et al*., 1991). When the values of Fe_o/Fe_d are within 0.09-0.33, high degree of free Fe oxide crystallization is inferred (Jonczak *et al*., 2015). This assertion affirms that soils of AI1P2, MF2P2, AI1P1 and IH1P2 were more mature or developed as they had the least mean values and further suggest that weathering had progressed to an advanced stage or the soils are older (Seal *et al*., 2006).

v. Co-migration of clay with Fe

The co-migration of clay with Fed presented mean values of 0.394 and 0.114 for IH1P1 and IH1P2, respectively. Mean values of 0.05, 0.065, 0.080 and 0.089 were obtained for AI1P1, AI1P2, AI2P1 and AI2P2, respectively. Furthermore, in MF1P1, MF1P2, MF2P1 and MF2P2, mean values of 0.071, 0.090, 0.205 and 0.103 were obtained, respectively.

Amongst the soils studied, co-migration of clay with Fed increased with increasing soil depth resulting in the formation of outward bulges in AI2P2, MF1P1 and MF2P2, except in IH1P1 and IH1P2, where the trends down the profiles either resulted in an inward bulge or decrease with soil depth. Increase in Fed/clay with soil depth suggests that the soils had sufficient time to warrant the movement of clay in the company of Fed down the depth. In soil profiles with distinct soil horizons, the ratio of Fed/clay increases with soil depth (Dolui and Bera, 2001). Most of the weathered Fe bearing minerals in the Ap horizons may have released Fe which had moved in the company of clay. However, in IH1P1 and IH1P2 mineral weathering was low with a reduced rate of movement of Fe down the profiles, hence, the co-migration of clay with Fed was not sufficient to be identified. When Fed/clay ratio decreases with soil depth, the implication is that Fe movement is partially independent of the movement of clay (Juo *et al*., 1974). According to Juo *et al*. (1974), the trends of Fed and clay are similar in most Nigerian soil profiles. Accordingly, Girao *et al*. (2014) reported a range of 0.03 – 0.10 in the B horizons.

4.4.2 Forms of Al

i. Dithionite extracted Al (Ald)

In the soils overlying SLS in northern agricultural zone, mean values of 2.9 and 2.2 g/kg were obtained for IH1P1 and IH2P2, respectively (Table 25). The values of Ald in IH1P1 resulted in a decrease in values with soil depth, while values in IH1P2 bulged outwards. This indicates that Ald released from weathered minerals in the epipedon of IH1P2 had moved downwards with clay in the process of illuviation. An indication that IH1P1 was comparatively younger than IH1P2. The soils may be arranged in the following decreasing order of maturity; IH1P2>IH1P1. In all the soils studied, Ald exceeded Alo, an indication that a considerable fraction of Al oxide is present in the soil as crystalline form (Dolui and Mondal, 2007). According to the ratings of Wilding (1985), the coefficient of variability of the soil was moderate with a range of 15.8-28.4 %.

In the soils developed from SL in central agricultural zone, 0.50, 0.27, 0.35 and 0.27 g/kg were obtained as mean values in AI1P1, AI1P2, AI2P1 and AI2P2, respectively (Table 25). Though irregularly distributed, values of Ald increased with increasing soil depth and

resulted in a bulge in AI1P2. This indicates that AI1P2 has been exposed to more pedogenetic processes, weathered and released Al compounds which had been illuviated in the B horizon, and have become comparatively more mature. Though, AI2P1 may be adjudged the most mature due to its high Ald, the soils appeared to have very close values and may be within the same stage of development and influenced by similar factors of soil development. Coefficient of variability of the soils was greater than 20 % and rated high (Wilding, 1985).

In the soils developed from SLSi in southern agricultural zone, 1.53, 2.1, 4.37 and 3.07 g/kg were obtained as mean values of Ald in MF1P1, MF1P2, MF2P1 and MF2P2, respectively (Table 26). Values increased with increasing soil depth resulting in the formation of bulges in the B horizons. This foretells maturity of the soils as Ald may have been illuviated in the B horizons from the overlying Ap horizons. The CV ranged from 14.7 in MF2P1 to 47.5 % in MF2P2 and rated low to high (Wilding, 1985).

The order of increasing maturity by virtue of mean Ald concentration in the studied soils was such that; MF2P1> MF2P2> IH1P1> IH1P2> MF1P2> AI2P1> AI1P1> AI1P2=MF1P1>AI2P2. This implies that, irrespective of elevation ranges, soils in the high elevations of SLSi and those over SLS had the highest mean values and by implication more mature than their counterparts in other limestone formations and landscapes. The value reported for Ald are far less than those obtained for Fed and negates findings by Jonczak *et al.* (2015). They observed that the values obtained for Ald was twice as high as Fe concentration. This implies the dominance of Fe bearing weatherable minerals over alluminosilicate minerals. Values of Ald obtained in the limestone region of Apodi (Irmak *et al*. 2007) were higher than those obtained in the present study, while Dolui and Bera (2001) observed that the concentration of Al increased with soil depth and age in some soils.

The highest values of Ald were recorded in the endopedons of the soils except in IH1P1 (SLS formation), where the highest values were obtained in the epipedon. This implies that Ald in soils over SLS had the ability to chelate with organic ligands (Wagai *et al*., 2020). When the oxides of Al are released *via* weathering, they react with organic ligands and are rarely lost to leaching. Tsozue *et al*. (2009) reported highest values of Ald in the humic horizon and lower values in the yellowish red clayey endopedons. In the Harran region of Turkey, Irmak *et al.* (2007) reported a range of 0.70-2.70 g/kg in the limestone soils.

Strong positive correlation was obtained between Ald and Fe_d-Fe_o (r^2= 0.691), Al_d-Al_o (r^2= 0.979), TiO_2 (r^2= 0.587) and Fe_2O_3 (r^2= 0.729) and indicates a likelihood of increase in these parameters with increasing value of Al_d. However, weak positive correlations were obtained with Al_o, CaO and MnO_2. On the other hand, Al_d correlated negatively and strongly

with Al_o/Al_d (r^2= -0.522), SiO_2 (r^2= -0.546), K_2O (r^2= -0.755) and the indices of weathering except the Ruxton index. This trend in correlation appears very similar to that obtained for Fe_d and indicates that Fe_d and Al_d increase in highly weathered soils.

ii. Oxalate extracted Al

In the soils overlying SLS in the northern agricultural zone, 0.60 and 0.67 g/kg were obtained in IHIPI and IH1P2, respectively (Table 25) with indications of the movement of Al to the B horizon in IH1P2. However, values were irregularly distributed with increasing soil depth in IH1P1, with moderate CVs of 28.9 and 17.3 % in IH1P1 and IH2P2, respectively. In IH1P2, minerals seem to have been weathered in the epipedon and moved *via* eluviation to the B horizon. Therefore, IH1P2 has had more time to warrant the movement or it may have been dominated by non crystalline form of Al.

In the soils over SL in central agricultural zone, mean values of 0.50, 0.27, 0.35 and 0.27 g/kg were obtained in AI1P1, AI1P2, AI2P1 and AI2P2, respectively (Table 25). The values of Alo generally increased with increasing soil depth. This however, does not infer maturity but youthfulness, as the soils may not have been exposed for a time sufficient for the Al to be eluviated. Values were unevenly distributed with increasing soil depth and had CV values of 38, 12 and 55 % for AI1P1, AI2P2 and AI3P1, respectively. According to the scale of Wilding (1985), the values are low to high.

In the soils of SLSi in southern agricultural zone, mean values of 0.57, 0.40, 1.57 and 0.63 g/kg were obtained in MF1P1, MF1P2, MF2P1 and MF2P2, respectively (Table 26). Highest oxalate extractable Al value occurred in MF2P1 and corresponds with its high organic carbon content amidst its well-drained condition. Also, Alo had the highest values in the Bt horizons. This aligns with earlier studies by da Silva *et al*. (2019). They attributed the high values of Alo in the Bt horizons to the expressive formation of low crystalline hydroxides in the illuvial horizons. In this case, soil minerals are dissolved by ferrolysis in the epipedons and accumulated in the endopedons. Oxalate Al constitutes greater than 60 % of Al in most Nigerian soils (Juo *et al*., 1974). The values of Alo are low and seem to have been influenced by the high temperatures in the tropics. According to Juo *et al*. (1974), high temperatures may be responsible for low content of amorphous Al. High temperatures cause amorphous Al oxides to dehydrate and shift to a system of greater crystallinity (Sherman *et al*., 1964). Dehydration and crystallization are important factors controlling the presence of amorphous materials in tropical soils (Juo *et al*., 1974).

Values were unevenly distributed with soil depth and had CV values of 10.2, 29.5 and 59.8 % in MF1P1, MF2P1 and MF2P2, and rated low, medium and high, respectively (Wilding, 1985).

Irrespective of limestone formations, the order of occurrence of Al oxalate in the well-drained soils is of the decreasing order such that; MF2P1> MF2P2> IH1P2> MF1P1> AI1P1> MF1P2> AI2P1> AI1P2> AI2P2. Worthy of note is the increasing oxalate Al concentration with increasing organic matter content in the entire soils. One may hypothesize the dependence of Alo on soil organic matter content. Oxalate extractable Al correlated weakly and negatively ($p<0.05$) with K_2O ($r^2= -0.428$) and WI ($r^2= -0.456$).

iii. Crystalline form of Al (Al_d-Al_o)

In the soils over SLS in northern agricultural zone, mean values of 2.3 and 1.53 g/kg were obtained in IH1P1 and IH1P2, respectively (Table 25). Higher values seem to have been obtained in the solum of both pedons, while relatively lower values were recorded in the C horizons. Lower profile mean value in IH1P2 indicates low crystallinity and decreasing youthfulness. This may not be totally so, as an outward bulge obtained in IH1P2 may indicate maturity as the soil minerals, particularly biotite and ferromagnesian minerals may have weathered in the epipedon (Tan, 1998), and had sufficient time to warrant the movement of materials from the epipedon to the endopedon. IH1P1 had low CV of 13 %, while IH1P2 had high CV of 35.9 % on the scale of Wilding (1985).

In the soils of SL in central agricultural zone, mean values of 1.4, 1.27, 1.45 and 1.33 g/kg were obtained in AI1P1, AI1P2, AI2P1 and AI2P2, respectively (Table 25) with a regular increase in values with soil depth, and the formation of bulges in AI1P1 and AI1P2. Increasing crystallinity as obtained in the B horizons except in AI2P2 indicates increasing maturity such AI2P1> AI1P1> AI2P2>AI1P2. However, the trend in values in AI1P1 and AI1P2 indicate that weathered minerals in the epipedons may have been moved through illuviation to the B horizons, hence relatively more mature than AI2P1 and AI2P2. Uneven distribution of Al_d-Al_o with increasing soil depth prompted high CV in AI1P1 and AI1P2 (> 50 %) and moderate CV in AI2P2 (22.9 %).

In the soils of SLSi in southern agricultural zone, mean values of 0.97, 1.70, 2.80 and 2.40 g/kg were obtained in MF1P1, MF1P2, MF2P1 and MF2P2, respectively (Table 26). The least mean value obtained in MF1P1 reflects youthfulness and the presence of poorly crystalline mineral phases. Increasing crystallinity indicates increasing maturity and presents an increasing trend of maturity such as MF1P1< MF1P2< MF2P2< MF2P1; such that MF2P1 is the most crystalline and most mature.

On a general note, irrespective of limestone formations, well-drained soils were rated in terms of decreasing maturity and crystallinity as; MF2P1> MF2P2> IH1P1> MF1P2> IH1P2> AI2P1> AI1P1> AI2P2> AI1P2> MF1P1. From the foregoing, crystalline form of Al was highest in the soils of SLSi, and considered most developed and mature. With increasing soil age, crystalline Al oxides increase in soils (Seal *et al*., 2006).

Crystalline Al correlated positively and strongly with CaO (r^2= 0.533), TiO_2 (r^2= 0.584) and Fe_2O_3 (r^2= 0.703). This indicates that increasing weathering that results in more crystalline Al is most likely to bring about an increase in TiO_2 and Fe_2O_3. On the other hand, it correlated strongly and negatively with Al_o/Al_d (r^2= -0.619), SiO_2 (r^2= -0.526), K_2O (r^2= -0.702) and the indices of weathering except Ruxton index. The indices of weathering correlated similarly with the crystalline forms of Fe and Al.

iv. Degree of activation of Al

In the soils overlying SLS in northern agricultural zone, mean values of 0.204 and 0.315 were obtained in IH1P1 and IH1P2, respectively (Table 25). The IH1P2 was more active as it had higher values of Alo/Ald ratio. Moderate CV was obtained in both soils with values ranging from 16.5 to 24.3 %.

In the soils of SL in central agricultural zone, mean values of 0.309, 0.21, 0.21 and 0.17 g/kg were obtained in AI1P1, AI1P2, AI2P1 and AI2P2, respectively (Table 25). Though values were irregularly distributed with increasing soil depth, comparatively higher values were obtained in AI1P1, and indicates greater activity. When active Fe ratio decreases with increase in soil depth, higher proportions of Fe oxides may be present as crystalline forms in endopedons (Juo *et al*., 1974). In terms of activity, the soils may be arranged in the decreasing order of AI1P1> AI1P2> AI2P1> AI2P2. Coefficient of variability was low in the soils of this formation, but exceeded 35 % AI1P2, indicating high variability in soil activity. This was according to the rating on the scale of Wilding (1985).

In the soils of SLSi in southern agricultural zone, mean values of 0.38, 0.18, 0.36 and 0.19 were obtained in MF1P1, MF1P2, MF2 P1 and MF2P1, respectively (Table 26). The values indicate that MF1P1 and MF2P1 were the most active. Irrespective of limestone formation, the values of Feo/Fed were low and reflect an advanced stage of crystallinity or aging (Mahaney *et al*., 1991). The coefficient of variability is rated low to moderate on the scale of Wilding (1985) with values ranging from 15.2 to 22.5 %.

Irrespective of limestone formations, the order of activity in the high elevation areas was arranged in decreasing order of activity of; MF1P1> MF2P1> IH1P2> AI1P1> AI1P2=AI2P1> IH1P1> MF2P2> MF1P2> AI2P2. When Al_o/Al_d in the solum is low, its

value is often less than one (Seal *et al*., 2006). Low values as obtained in the studied soils indicate that free aluminum oxides in the soil are at an advanced stage of crystallinity or aging (Mahaney *et al*., 1991). Higher degree of crystallinity is associated with lower ratios (Osayande *et al*., 2013). This agrees with findings in the present study as the least values were obtained in MF1P2 and AI2P2, which had earlier been identified as the most developed and mature.

The degree of activation of Al correlated positively and strongly with SiO_2, R and WI with r values of 0.529, 0.571 and 0.661, respectively and weakly negative with CaO and Fe_2O_3. However, it showed strong inverse relationship with Al_d and crystalline Al.

v. Co-migration of clay with Ald

The co-migration of clay with Ald was such that IH1P1 had a mean of 0.011, while IH1P2 had a mean of 0.012 for soils over SLS. For the soils overlying SL, mean values of 0.009, 0.007, 0.012 and 0.010 were obtained in AI1P1, AI1P2, AI2P1 and AI2P2, respectively. Furthermore, MF1P1, MF1P2, MF2P1 and MF2P2 had mean values of 0.010, 0.015, 0.032 and 0.021, respectively. Aubert and Segalen (1966) reported the combined movement of clay and Fe from epipedons with the resultant formation of distinct argillic horizons.

Among the soils studied over the three geologic formations, only soils over SLS had either decreasing Ald/clay ratio (IH1P2) or inward bulging values at the B horizons (IH1P1). The inward bulge indicates that Ald had not been eluviated with clay into the B horizons probably due to low weathering intensity which would have aided the release of weathered minerals from the Ap horizons into the B horizons. This low weathering in the epipedons has reduced the further movement of Ald into the endopedons. These signify youthfulness. Also, decrease in Ald/clay with depth implies that the movement of Al with depth was partially independent of clay movement (Juo *et al*., 1974). Thus, IH1P1 and IH1P2 were relatively less developed. Other soils studied indicated either increasing Ald/clay ratios with soil depth (AI1P1, AI2P1, MF1P2) or an outward bulge at the B horizons, an indication that the soils had sufficient time to co-migrate clay with Ald, hence the formation of bulges in AI1P2, AI2P2, MF1P1, MF2P1, MF2P2. This suggests that the soils had sufficient time to warrant the movement of clay in the company of Ald to the endopedons, resulting in the formation of more developed soils.

4.4.3 FactorAanalysis of Variable Loadings for the Forms of Fe and Al

Ten (10) soil variables were compressed into 3 principal components (PC): PC1, PC2 and PC3, and presented in Table 27.

Table 27: Factor Analysis of Variable Loadings for the Forms of Fe and Al

Variables	PC1	PC2	PC3
Fed	0.828	-0.423	0.224
Ald	0.929	-0.006	-0.280
Feo	0.650	0.178	0.574
Alo	0.806	0.312	-0.115
Fed-Feo	0.822	-0.438	0.209
Ald-Alo	0.864	-0.170	-0.358
Feo/Fed	-0.055	0.683	0.564
Alo/Ald	0.124	0.644	0.329
clay/Fed	-0.900	0.148	0.047
clay/Ald	-0.797	-0.218	0.365
SLS	0.348	-0.404	0.702
SL	-0.668	-0.407	-0.223
SLSi	0.375	0.747	-0.367
Eigenvalues	6.217	2.366	1.882
Total variance %	47.8	18.2	14.5
Cumulative Eigenvalues	6.217	8.582	10.464
Cumulative % variance	47.8	66.0	80.5

Foot note: PC= Principal component

The 3 PCs contributed 80.5 % of the total variation in soils overlying SLS, SL and SLSi. Out of this variation, 47.8 % was accounted for by PC1 which turned the most influential PC, while 18.2 % of the variability was contributed by PC2. The least influential PC was PC3 which explained only 14.5 % of the total variation between the soils.

Principal component (PC) 1 was influenced by Fed, Ald, Alo, Fed-Feo, Ald-Alo, clay/Fed and clay/Ald with over 80 % contribution of each soil variable to the variation of soils overlying the three lithologies. All the soil properties were positively correlated with PC1 except clay/Fed and clay/Ald which correlated negatively with PC1.

Feo/Fed and Alo/Ald exerted their contributions to soil variation in PC2 with 68.3 and 64.4 % contributions, respectively to soil variation. In PC3, the influence of the forms of Fe and Al on soil variation was somewhat insignificant as all the variables had made their contributions to PC1 and PC2. The most influential PC presented SL as the lithology with the most contribution to soil variation. It contributed 66.8 % to soil variation. PC2 and PC3 were most influenced by SLSi (74.7 %) and SLS (70.2 %), respectively. Eigenvalues of PC1 and PC2 exceeded 2 and indicate that the first two PCs can be adapted, while PC3 is discarded.

4.5 Elemental Oxides of the Soils Determined by X-ray Florescence (XRF)

Elemental oxides of the soils determined by X-ray florescence (XRF) are presented on Tables 28 - 30.

a. SiO_2: In the soils overlying SLS in the northern agricultural zone, IHIP1 had values of SiO_2 that ranged from 20.31 at 0-30 cm to 27.72 % at 120-150 cm, while values in the poorly drained IH2P2 ranged from 18.97 at 0-30 cm to 24.85 % at 30-60 cm of depths with the development of bulges at 120-150 cm and 30-60 cm for IH1P1 and IH2P2, respectively (Table 28).

In the soils overlying SL in central agricultural zone, AI1P1 had SiO_2 values that ranged from 21.61 at 150-180 cm to 29.08 % at 0-30 cm depths with values that appeared to have varied irregularly with soil depth (Table 29). In AI2P2, values varied from 22.19 at 90-120 cm to 27.78 % at 0-30 cm depth with values that decreased continuously with increased soil depth, while the trend of values in the poorly drained AI3P1 led to the formation of bulge at 30-60 cm with values ranging from 24.59 to 29.32 % (Table 29).

In the SLSi soils in southern agricultural zone, MF1P1 had values that varied from 24.46 in 30-60 cm to 26.9 % at 0-30 cm depth, while MF2P1 had values ranging from 18.95 at 150-161 cm to 22.29 % in 120-150 cm depth. Values of SiO_2 in the poorly drained MF3P2 ranged from 21.2 to 29.71 % (Table 30).

Table 28: Elemental oxide distribution by X-Ray Florescence in soils over SLS

Horizon	Al_2O_3	SiO_2	P_2O_5	K_2O	CaO	TiO_2	MnO_2	Fe_2O_3	ZrO_2	LE
					%					
IHP1 0-30	8.520	20.313	0.03	0.238		1.104	0.06	9.561	0.045	60.067
IHP1 30-60	10.527	20.960		0.411		0.827	0.08	7.796	0.034	59.290
IHP1 60-90	9.083	20.650	0.02	1.179	0.265	0.703	0.13	6.930	0.021	60.907
IHP1 90-120	8.557	22.187		1.199	0.230	0.743	0.09	6.105	0.032	60.757
IHP1 120-150	6.897	27.717	0.04	0.599	0.407	1.464	0.23	5.629	0.099	56.823
IHP1 150-180	6.220	25.057		0.556	0.203	1.221	0.10	7.047	0.067	59.450
Mean	8.30	22.81	0.03	0.70	0.28	1.01	0.12	7.18	0.050	59.55
IH2P2 0-30	7.880	18.970	0.10	0.426	0.191	0.674	0.24	7.994	0.023	63.420
IH2P2 30-60	7.890	24.847	0.06	0.687		0.967	0.12	5.662	0.055	59.620
IH2P2 60-87	7.300	23.110	0.08	0.593		0.880	0.28	5.546	0.051	62.077
Mean	7.69	22.31	0.08	0.569	0.191	0.840	0.213	6.401	0.043	61.71

LE: Light elemen

Table 29: Elemental oxide distribution by X-Ray Florescence in soils over SL

Horizon	Al_2O_3	SiO_2	P_2O_5	K_2O	CaO	TiO_2	MnO_2	Fe_2O_3	ZrO_2	LE
					%					
AI1P10-30	5.777	29.080	0.027	2.886		0.406	0.081	0.767	0.044	60.887
AI1P1 30-60	8.757	25.057		2.623		0.516	0.065	1.569	0.037	61.320
AI1P160-90	11.820	22.790		2.339		0.471	0.028	2.149	0.025	60.323
AI1P1 90-120	12.067	23.410		2.516		0.509	0.017	2.181	0.023	59.223
AI1P1 120-150	11.947	23.980		2.618		0.529	0.019	2.098	0.024	58.733
AI1P1 150-180	11.553	21.610		2.505		0.452	0.250	3.133	0.021	60.400
AI1P1 180-200	12.093	21.650		2.934		0.452	0.171	4.125	0.022	58.487
Mean	10.57	23.94	0.027	2.63		0.476	0.0901	2.29	0.028	59.910
AI2P2 0-30	6.573	27.783	0.045	4.243		0.485	0.045	1.496	0.096	59.137
AI2P2 30-60	8.443	25.387	0.035	3.984		0.503	0.031	4.158	0.062	57.297
AI2P2 60-90	9.623	23.533		4.481		0.480	0.010	2.707	0.052	59.003
AI2P2 90-120	10.360	22.187		5.040		0.549	0.052	4.156	0.039	57.493
Mean	8.745	24.723	0.04	4.437		0.504	0.035	3.129	0.062	58.23
AI3P1 0-30	6.167	27.553	0.021	3.011		0.351	0.020	0.843	0.031	61.977
AI3P1 30-60	8.573	29.323		3.089		0.335	0.016	1.036	0.029	57.547
AI3P1 60-80	10.060	24.587		3.332		0.346	0.007	1.640	0.023	59.957
Mean	8.267	27.154	0.021	3.144		0.344	0.0143	1.173	0.028	59.827

LE: Light elements

Table 30: Elemental oxide distribution by X-Ray Florescence in soils over SLSi

Horizon	Al_2O_3	SiO_2	P_2O_5	K_2O	CaO	TiO_2	MnO_2	Fe_2O_3	ZrO_2	LE
					%					
MF1P1 0-30	7.903	26.903		2.770	0.133	0.345	0.028	2.070	0.047	59.730
MF1P130-60	8.073	24.463		2.454	0.177	0.370	0.023	2.436	0.035	61.883
MF1P1 60-90	7.993	25.273		2.738	0.284	0.349	0.027	2.006	0.036	61.210
MF1P190-120	7.680	25.270		2.453	0.320	0.319	0.030	2.125	0.029	61.693
Mean	7.912	25.477		2.604	0.229	0.346	0.027	2.159	0.037	61.129
MF2P1 0-30	7.403	22.117	0.019	1.138	0.179	0.484	0.057	4.851	0.034	63.663
MF2P1 30-60	8.383	21.573		1.249		0.522	0.017	5.357	0.019	62.820
MF2P1 60-90	7.927	20.393	0.017	1.234		0.502	0.018	5.996	0.017	63.837
MF2P1 90-120	8.107	22.283	0.026	1.487		0.521	0.080	5.493	0.019	61.900
MF2P1120-150	7.840	22.285	0.035	1.533		0.507	0.118	5.198	0.020	62.375
MF2P1150-161	7.127	18.953	0.044	1.210	0.076	0.446	0.238	9.135	0.015	62.663
Mean	7.798	21.267	0.028	1.309	0.128	0.497	0.088	6.005	0.021	62.876
MF3P2 0-30	8.100	21.197	0.054	1.780	0.370	0.413	0.005	1.380	0.044	66.520
MF3P2 30-49	6.590	29.713	0.045	2.851		0.242	0.005	0.368	0.061	60.063
Mean	7.345	25.455	0.050	2.316	0.370	0.328	0.005	0.874	0.053	63.291

LE: Light elements

In poorly drained IH2P2, AI3P1 and MF3P2, the accumulation of SiO_2 increased with increase in soil depth and indicates losses in the Ap horizons. This trend of values with soil depth negates findings by Akpan and Nkanga (2016) in wetland soils. This indicates that SiO_2 may have been leached from the surface soils as it is considered a 'mobile soil component' and is not a major component of organic matter which often dominates the surface soils (Duzgoren-Aydın *et al*. 2002) of organic rich soils. Bulges of SiO_2 were developed at 120-150 cm in IH1P1 and AI1P1, while poorly drained AI3P1 and IH2P2 had bulges at 30-60 cm. However, soils overlying the SLSi formation as well as AI2P2 on the SL formation had no clearly defined trend with increase in soil depth and indicates alterations from deposition of particles elsewhere. Values in the surface soils of the formations seem to decrease with increasing soil depth and shows that the soils may not have had sufficient weathering time to warrant loss of SiO_2. This aligns with the findings of Esu (2010) and negates earlier claims that soils in high elevations were relatively young compared to those in low elevations. The amount of pedogenic silica increases with increasing duration of weathering, and the depth of maximum silica accumulation coincides with the maximum expression of argillic horizon (Kendrick and Graham, 2004). This assertion was observed in AI1P1, IH1P1, IH2P2 and AI3P1. The dominance of SiO_2 in all the soils is anticipated. Higher values of SiO_2 than Al_2O_3 as it was noticed in all the soils studied (Tables 28-30) foretells poor silica solubilization and subsequent removal from the soils, with the resultant formation of kaolinite. This may have contributed to the dominance of kaolinite in the soils, as a major soil mineral. Silicon is the second most abundant element (Sommer *et al*., 2006) and a major component of silicate minerals (Fraysse *et al*., 2006; Orlov, 1985), and present in virtually all parent materials and soils (Sommer *et al*., 2016).

High accumulation of SiO_2 in the soils may be due to high sand content containing quartz bearing minerals (Gardner, 1992). High soil silica may reduce heavy metal toxicity in plants and increase crop yield in P deficient soils such as this, by decreasing the availability of Fe and Mn in plants (Ma, 2004). According to Yakubu and Ojanuga (2013), when silica content in soils is higher than iron and aluminum oxides, it indicates that the soils are not highly weathered, while Buringh (1979) outlined that with progressive weathering and weathering intensity, more sesquioxides tend to accumulate in soil, while SiO_2 is leached out. Similarly, Bohn *et al*. (2001) and Sharma *et al*. (1996) opined that highly weathered soils contain less silica and more of Fe and Al oxyhydroxides which tend to accumulate. Going by the findings of Bohn *et al*. (2001) and Sharma *et al*. (1996), the soils under study seem to be dominated by SiO_2, hence they were not highly weathered. Also, high accumulation of SiO_2

in the soils could be attributed to the high sand fraction with high quartz bearing minerals (Gardner, 1992).

Silicon oxide (SiO_2) was weakly to strongly correlated with ZrO_2 and Al_o/Al_d as well as the indices of chemical weathering except WI, however, it correlated weakly but positively with K_2O (r^2=446). Inverse relationships were also obtained with Fe_d, Al_d, Fe_o and Al_d-Al_o as well as MnO_2 and Fe_2O_3. Soil aging brings about the accumulation of immobile ZrO_2, Fe, Mn and Al, and encourages the leaching of more mobile components such as; SiO_2 and K_2O. Hence the inverse relationship.

b. Al_2O_3: In the soils overlying SLS in the northern agricultural zone, Al_2O_3 ranged from 6.22 in 15-180 cm to 10.52 % in 30-60 cm soil depths, while values in the poorly drained IH2P2 ranged from 7.3 in 60-87 cm depth to 7.9 % in 30-60 cm depth (Table 28)**.** In both pedons, bulges of Al_2O_3 were formed which reduced gradually to the last depth range.

In the soils overlying SL in central agricultural zone, Al_2O_3 varied from 5.78 in 0-30 cm to 12.09 % in 180-200 cm soil depths with clear and well-formed bulge at 90-120 cm depth. In AI2P2, it ranged from 6.57 in 0-30 cm to 10.36 % in 90-120 cm depth and gradually increased in value up to 90-120 cm depth range (Table 29). In the poorly drained and low elevation AI3P1, values of Al_2O_3 followed a similar trend like in AI2P2 with ranges of values from 6.17 in 0-30 cm to 10.06 % in 60-80 cm depth (Table 29).

In the soils of SLSi in southern agricultural zone, values of 7.68 were obtained in 90-120 cm and 8.07 % in 30-60 cm soil depth in MFIPI, while values in MF2P1 ranged from 7.13 in 150-161 cm to 8.38 % in 30-60 cm depth (Table 30). Though its distribution with increasing soil depth was irregular, there appear to be an overall decrease in values with increasing soil depth resulting in bulges at 30 – 90 cm depths. However, values showed a decrease in MF3P2 from 8.1 at 0-30 to 6.59 % at 30-49 cm (Table 30). Ahmad *et al*. (1966), Carroll and Hathaway (1953) and Girao *et al*. (2014) reported higher values for similar soils elsewhere using similar technique.

As the intensity of weathering and pedogenic processes increase, Al_2O_3 increases (Duzgoren-Aydın *et al*. 2002). Therefore, the subsurface of the studied soils were more weathered than the surface soils in all mapping units. Consequently, soils in the SL were the most highly weathered followed by those of SLS and then SLSi. Old soils are abundant in gibbsite (Wang and Choudhury, 1981; Ogg and Baker, 1999). Gibbsite is the dominant crystalline form of Al_2O_3 because its formation is favoured by organic matter that inhibits crystallization (Hsu, 1989).

Interestingly, the entire soils irrespective of limestone formations and elevation range had Al_2O_3 accumulation in the endopedons and depletion in the epipedons, except for the poorly drained and low elevation soils of IH2P2 and MF3P2 that had marginal variation with increased soil depth and decreased values with depth, respectively. Aluminum depletion is favoured by pH lower than 5 (Chamayou and Legros, 1989) and goes on at a faster rate in the top soil (Egli *et al*., 2011); it however becomes immobile at high pH. At low pH, Al_2O_3/TiO_2 ratio increases with depth; consequently, accumulation of Al_2O_3 in the B horizons indicates its mobilization within the soil due to active soil forming processes (Sidhu *et al*., 2000), while Gardner (1992) attributed high contents of Al_2O_3 to the presence of gibbsite bearing minerals in soils. Highly weathered soils contain more of Al oxyhydroxides which accumulates with age (Bohn *et al*., 2001; Sharma *et al*., 1996).

Strong negative correlation was recorded between Al_2O_3 and the indices of chemical weathering except WI and indicates decreasing values of indices as Al_2O_3 increases with soil development. Weak inverse relationships were also obtained with SiO_2 and P_2O_5 with r^2 values of -0.384 and -0.475, respectively at p= 5 %.

c. Fe_2O_3: In the soils overlying SLS in the northern agricultural zone, Fe_2O_3 ranged from 5.63 in 120-150 cm to 9.56 % in 0-30 cm depth with steady decrease in values with increasing soil depth up to 150 cm in IH1P1, where slight increase in value was noticed (Table 28). In the poorly drained and low elevation IH2P2, Fe_2O_3 decreased gradually with soil depth. The poorly drained and low elevation soil appears young and does not seem to have been developed to the stage of IH1P1 which noticed an increase in value at 150 cm.

For the soils overlying SL in the central agricultural zone, Fe_2O_3 ranged from 0.767 in 0-30 cm to 4.125 % in the 180-200 cm depth with gradual and consistent increase with increasing soil depth in AI1P1. Values in AI2P2 varied from 1.496 in 0-30 cm to 4.158 % in 30-60 cm depth and varied irregularly with increased soil depth. In the low elevation and poorly drained AI3P1, Fe_2O_3 was obtained as 0.843 in 0-30 cm and 1.64 % in the 60-80 cm depth with gradual increase in value with soil depth increase (Table 26). Amongst the soils studied in the SL in central agricultural zone, AI2P2 appeared to have the highest value of Fe_2O_3 in the area and therefore the most developed.

Values of Fe_2O_3 in the soils overlying SLSi in the southern agricultural zone ranged from 2.01 in 60-90 cm to 2.44 % in 30-60 cm depth. Varying irregularly with soil depth, a slight bulge appears to have been formed at 30 cm depth, while in MF2P1 values varied from 4.85 in 0-30 cm to 9.14 % in 150-161 cm depth with the clear formation of bulge at 60-90 cm depth. In the poorly drained and low elevation MF3P2, values varied from 0.368 in 30-49 cm

and 1.38 % in the 0-30 cm depth (Table 26). Values of Fe_2O_3 were higher in MF2P1, especially in the underlying depth ranges compared to MF1P1 and MF3P2 and therefore, comparatively more developed.

The values of Fe_2O_3 in soils of shale, limestone and sandstone decreased, while those of SL increased with soil depth; however, AI2P2 decreased at 60-90 cm. Values of Fe_2O_3 varied marginally and irregularly with soil depth in the SLSi except in MF2P1 where a sharp increase of about 40 % in Fe_2O_3 at 150-161 cm was noticed possibly due to the presence of Fe concretions (Tables 6 -13).

The values of Fe_2O_3 were higher in the subsurface soils especially in the SL and SLS, and indicate that the underlying soils were more highly weathered than those of the topsoils. Based on the limestone formations, soils on SLS were more intensively weathered than those of SL and SLSi; this is because as the intensity of weathering and pedogenic processes increase Fe_2O_3 increases in soils (Duzgoren-Aydın *et al*. 2002). According to Burt *et al.* (2003), the high content of Fe_2O_3 in the B horizons compared to the A horizons can be traced to Fe migration from A to B horizon through cheluviation. As the soil advances in weathering, oxides of Ca and K as well as Si are progressively leached, leaving behind TiO_2, Al_2O_3 and Fe_2O_3 which had comparatively remarkably higher values in all the soils studied. Values of Fe_2O_3 reported by Girao *et al*. (2014) are lower than those of the shale, limestone and sandstone in the northern agricultural zone but higher than those of the SL in central zone, and SLSi in southern agricultural zone in the present study.

Higher Al and Fe oxides and lower silicon oxides in the epipedons of soils over SLS than those over SL and SLSi suggest relatively more active leaching of SiO_2 and abundance of Fe_2O_3 and Al_2O_3 in the soils over SLS.

Negative correlations were obtained between Fe_2O_3 and all the chemical indices of weathering used in the study. However, strong relationships were obtained with WI and MR, and weak relationship was obtained with R. Positive correlation of Fe_2O_3 with TiO_2, Fe_d and Al_d were not surprising as weathering intensity increases the values of these immobile oxides in the same direction.

d. TiO_2: Soils overlying SLS in the northern agricultural zone had values of TiO_2 that ranged from 0.703 in the 60-90 cm to 1.464 % in the 120-150 cm depth with an inward bulge occurring in the 60-90 cm depth of IH1P1 (Table 28). Such inward bulge is an indication of a decrease in value and that TiO_2 has not been illuviated in the endopedons. In IH2P2, values of TiO_2 ranged from 0.674 in 0-30 cm to 0.967 cm in the 30-60 cm with small bulge in 30-60 cm. Values in IH2P2 appeared to be less than those of IH1P1.

In the SL soils in central agricultural zone, values of TiO_2 varied from 0.406 in the 0-30 cm to 0.529 in the 120-150 cm depth with values that varied irregularly with increasing soil depth in AI1P1 and AI2P2. Values in AI2P2 ranged from 0.480 to 0.549 %; however, in the poorly drained and low elevation AI3P1, values were within the range of 0.335-0.351 % (Table 29). Values of TiO_2 do not seem to have changed significantly with decreasing elevation ranges, except in the poorly drained AI3P1 where relatively lower values were obtained.

In the soils overlying SLSi in southern agricultural zone values ranged from 0.319 in 90-120 cm to 0.370 % in 30-60 cm depth with bulge occurring at 30 cm for the MFIP1. In MF2P1, values obtained were from 0.446 in 150-161 cm to 0.522 % in 30-60 cm depth with bulge occurring at 30 and 90 cm and decreasing thereafter. In the poorly drained MF3P2, values decreased from 0.413 in 0-30 cm to 0.242 % in the 30-49 cm depth (Table 30). Values in MF3P2 were lower than those of MF2P1. Values obtained by Girao *et al*. (2014) in Brazil were lower than those obtained in the soils overlying SLS in northern agricultural zone but within range of those in the SL and SLSi. The content of TiO_2 in soils may be less than 4 % in the tropics (SSS, 2014b). in the studied soils, its content was far less than this amount.

Though with not so much of variation with soil depth; probably due to its less mobile property (Curi and Franzmeier, 1987), the subsurface soils had comparatively higher values of TiO_2 than the surface soils, an indication that the subsurface soils were more intensively weathered than the surface soils. Based on the limestone formations, soils formed on SLS and SL seem to have greater values and were more intensively weathered than those on the SLSi. This is because, as the intensity of weathering and pedogenic processes increase in tropical soils, TiO_2 increases or remain unchanged in soils (Duzgoren-Aydın *et al*. 2002). TiO_2 is immobile in soils because it is poorly weathered during pedogenesis (Nesbitt and Markovics, 1997; Egli *et al*., 2011, Brantley and Lebedeva, 2011) hence its low amount in the soils. However, it resides in rutile and other Ti bearing minerals (Zhao and Zheng, 2015). Report by Girao *et al*. (2014) indicates lower values of TiO_2 than those obtained in the SLS, but higher than those in the SL and SLSi.

TiO_2 was strongly positively correlated with dithionite Fe and Al as well as their crystalline forms. However, correlation with MnO_2 and ZrO_2 was somewhat moderate as r values of 0.470 and 0.465 were obtained, respectively. This indicates that every increase in the values of TiO_2 is most likely to bring about an increase in the values of Fe, Al, MnO_2 and ZrO_2, all of which are immobile oxides.

e. MnO_2: In the soils overlying SLS in northern agricultural zone values of MnO_2 ranged from 0.06 in the 0-30 cm to 0.23 % in 120-150 cm depth for IH1P1, while values in the poorly drained mapping unit IH2P2 ranged from 0.12 to 0.28 % (Table 28) and appeared higher than those in IH1P1. In the SL of central agricultural zone, values of MnO_2 ranged from 0.017 in 90-120 cm to 0.25 % in 150-180 cm depth with an inward bulge in the 90-120 cm depth in AI1P1, while those in AI2P2 ranged from 0.01 in the 60-90 cm to 0.052 % in the 90-120 cm depth with an inward bulge at 60-90 cm. this indicates that MnO2 has not been illuviated in the endopedons. Values in the poorly drained AI3P1 decreased continuously from 0.02 to 0.007 % (Table 29); these values appear to be the least in the area.

Soils on SLSi in the southern agricultural zone had values of MnO_2 as 0.023 in the 30-60 cm and 0.030 % in the 90-120 cm with an inward bulge at 30-60 cm in MF1P1. In MF2P1, values of MnO_2 ranged from 0.017 in the 30-60 cm to 0.238 % within the 150-161 cm depth with an inward bulge at 30-60 cm depth (Table 30). Other values increased continuously with increased soil depth; however, in MF3P2, values of MnO_2 were quite low compared to those of MF1P1 and MF2P1 and had a mean of 0.005 %. Values of MnO_2 in the soils overlying shale, limestone and sandstone (SLS) in the northern agricultural zone were higher, while those in other limestone formations were lower than those obtained by Girao *et al*. (2014) in the Karst area of Apodi. The low concentration of this oxide in the soil may have been as a result of its presence in insoluble form; however, when complexed with organic matter or under conditions of low pH, it may be soluble and in available form (Lindsay, 1974). When MnO_2 falls within the range of 0.16–5.0 %, it rather exists as ferroxyhyte instead of forming its own minerals (Vodyanitskii *et al.*, 2004).

Soils in IH1 and IH2 had the highest organic carbon content which indicates that it may have complexed metal ions such Mn^{2+}. Hence, a relatively higher value in the soils of SLS (which had higher organic carbon) compared to those of SL and SLSi. Complexation reduces the incidence of toxicity as the micronutrient is released in only quantities required by the plant (Tan and Binger, 1986).

Correlations between MnO_2, Fe_o, and P_2O_3 were strongly positive. However, negative correlation existed between MnO_2, and K_2O and SiO_2. The inverse relationship indicates reducing values or leaching of K_2O and SiO_2 when weathering progresses.

f. ZrO_2: In the soils underlain by SLS in the northern agricultural zone, ZrO_2 varied from 0.021 in the 60-90 cm of IH1P1 to 0.099 % in the 120-150 cm depth range with an inward bulge at 90-120 cm, while values in IH2P2 ranged from 0.023 in the 0-30 cm to 0.055 % in the 30-60 cm depth range (Table 28). Unlike in IH1P1, values bulged outwards at 30-60

cm depth range. The inward bulge indicates that there was no sufficient time for the illuviation of ZrO_2 in the B horizon.

In the SL soils of central agricultural zone, values of ZrO_2 were found from 0.021 in 150-180 cm to 0.044 % in the 0-30 cm depth range with values decreasing regularly and bulging inwards at 150-180 cm and thereafter increasing slightly in AI1P1. This indicates that zirconium minerals were resistant and unweathered as to deplete the endopedons for subsequent movement to the endopedons. In AI2P2, values decreased continuously from 0.096 in the 0-30 cm to 0.039 % in 90-120 cm depth, while values in the poorly drained AI3P1 ranged from 0.023 in the 60-80 cm to 0.031 % in the 0-30 cm depth with values that decreased regularly as in AI2P2 without any bulges (Table 29). Requlr decrease or decrease in the B horizons before a subsequent increase in values indicate that time had not permitted the movement of ZrO2 as a result of the resistance of the mineral to weathering, to be illuviated in the endopedons. Values of ZrO_2 in AI2P2 seem to be the highest in the area compared to those of other elevation ranges. In the SLS soils in southern agricultural zone, values of ZrO_2 varied from 0.029 in 90-120 cm to 0.047 % in 0-30 cm with values that decreased irregularly with increasing soil depth. In MF2P1, ZrO_2 varied from 0.015 in 150-161 cm to 0.034 % in the 0-30 cm depth range with decrease in values and then bulging at 60-90 cm. Values in the poorly drained soils of MF3P2 ranged from 0.044 in 0-30 cm to 0.061 % in 30-49 cm depth range (Table 30). These values appeared to be the highest in the area, especially in the surface soils. Zircon is resistant to weathering and provides a measure of chemical stability (Schaetzl and Anderson, 2005) and often occurs in highly weathered soils. Its content in the soils under study appear to be low and translates to poorly weathered soils.

Zirconium oxide correlated strongly and positively with SiO_2 and weakly with TiO_2. Negative correlation was however obtained with Al_2O_3 at $p<0.05$ level of significance.

g. CaO: In the soils of SLS in the northern agricultural zone, values of CaO in IH1P1 ranged from trace amounts to 0.407 % with a bulge at 120-150 cm, while 0.191 % was obtained in 0-30 cm of IH2P2 with other depth ranges having trace values (Table 28). However, values of CaO in soils of SL were all trace for the soils studied, while those overlying SLSi ranged from 0.133 in the 0-30 cm to 0.32 % in the 90-120 cm depth with a regular increase in values with increased soil depth. Higher values of CaO in the endopedons is a reflection of the limestone parent material underlying the soils. In MF2P1, CaO varied from trace values to 0.179 % in the 0-30 cm, while 0.37 % was obtained in the 0-30 cm depth of MF3P2 (Table 30). Since the soils under study are developed from limestone geological

formations, one would expect high amounts of CaO in the soils. However, the high rainfall and temperature associated with the area had encouraged chemical weathering, dissolution of the oxide and subsequent loss to leaching. Previous studies by Carroll and Hathaway (1953) and Irmak *et al*. (2007) reported higher values than those obtained in the present study.

Calcium is hosted by plagioclase (Nesbitt *et al*., 1980, Babechuk *et al*., 2014); but the low to trace content in most of the soils could also be as a result of low or absence of plagioclase even in the clay mineralogy, an indication that the soils are developed from neither igneous nor metamorphic rock. Also, Ca is a very mobile element (Zhao and Zheng, 2015) and may have been lost to leaching, especially in the high rainfall region of central and southern agricultural zone where mainly trace amounts were reported. Weathering and translocation have been identified as key sources of losses of CaO in soils (Pavitch, 1989). Correlation of CaO with the indices of weathering were all not significant at 5 % level of significance.

h. K_2O: In the soils overlying SLS in northern agricultural zone, values of K_2O ranged from 0.238 in 0-30 cm and 1.199 % in 90-120 cm depth with bulge of values occurring between 60 and 90 cm of soil depth in IH1P1, while values in the poorly drained IH2P2 ranged from 0.426 in 0-30 cm to 0.687 % in 30-60 cm depth (Table 28). Values in IH2P2 seem to have been consistent with the trend of values for Al_2O_3, SiO_2 and Fe_2O_3 with increasing soil depth. Soil mineral weathering during pedogenesis results in translocation and accumulation of major and trace elements in soils (Kabata-Pendias and Pendias, 1992); in the subsurface soil horizons where accumulation exceeds translocation, a bulge is likely to be formed. This was the situation with the soils in this mapping unit.

In the soils of SL in central agricultural zone values of K_2O ranged from 2.34 in 60-90 cm to 2.93 % in 180-200 cm depth with values that bulged inwards at 90-120 cm as values decreased up to 90-120 cm and increased gradually thereafter (Table 29). In AI2P2, K_2O values ranged from 3.98 in 30-60 cm to 5.04 % in 90-120 cm depth with an inward bulge at 30-60 cm as values decreased down to 30 cm depth and increased thereafter. Decrease in values with increasing soil depth indicates that the weathering of muscovite was inactive or the mineral was absent. Also, there may not have been sufficient time to translocate the weathered product to the endopedons. However, values in the poorly drained AI3P1 increased gradually from 3.01 in 0-30 cm to 3.33 % in 60-80 cm depth (Table 29). Values in AI2P2 were the highest among the elevation ranges studied.

In the soils overlying SLSi in southern agricultural zone, values of K_2O were found from 2.453 in 90-120 cm to 2.738 % in 60-90 cm depth with values decreasing irregularly

with increase in depth in MF1P1. In MF2P1, values of K_2O ranged from 1.138 in 0-30 cm and 1.533 % in the 120-150 cm depth with values that increased irregularly with soil depth leading to the formation of a small bulge at 120-150 cm depth. However, values in MF3P2 varied from 1.78 in 0-30 cm to 2.851 % in 30-49 cm depth with values that appeared to be higher than those of MF2P1 (Table 30). Values of K_2O obtained in the present study are higher than those obtained by Irmak *et al*. (2007) and Carroll and Hathaway (1953) in previous studies elsewhere.

K_2O increased with soil depth in most of the profiles studied in the soils over SLS. Different authors however, present different opinions to this increment with soil depth. Potassium is lixiviated during rock weathering (Egli *et al*., 2011), consequently soil layers that are close to the parent material are higher in K contrary to the highly weathered horizons close to the soil surface that are often depleted in K (Bajard *et al*., 2016). However, Burt *et al*. (2003) attributed increase in K_2O with depth to its loss in the A to B horizon, while Akpan and Nkanga (2016) attributed very low to negligible total base percent to juvenile stage of pedogenesis.

Increased weathering intensity brings about increase in the amount of immobile oxides and decrease in the amount of mobile oxides as they are often leached from the soil once release *via* weathering. According to SSS (2014b), when the value of elemental analysis of K concentration (K_2O) exceeds 4 %, illite is obtained. Illitic clays may not be expected in the soils as the value of K_2O was below 4 % in the studied soils. This has resulted in the strong inverse relationship between K_2O and Fe_d, Al_d, Fe_d-Fe_o, Al_d-Al_o, Ti_2O and Fe_2O_3. This relationship implies that an increase in the amount of the immobile oxides will bring about a decrease in K_2O. this relationship is anticipated.

i. P_2O_5: In the soils overlying SLS in northern agricultural zone, P_2O_5 in IH1P1 ranged from trace values to 0.04 % with values that varied irregularly with soil depth, while values in IH2P2 showed an irregular decrease in values with increased soil depth (Table 28). Values of P_2O_5 in the SL were trace in all underlying depth ranges, while the value in 0-30 cm was 0.027, 0.045 and 0.021 % in AI1P1, AI2P2 and AI3P1, respectively (Table 29).

In the soils of SLSi, values of P_2O_5 were all trace in MF1P1, while in MF2P1, values ranged from trace values to 0.044 % in the 150-161 cm depth range as values increased continuously from 30-60 cm to 150-161 cm. Values in MF3P2 ranged from 0.045 to 0.054 %; which is comparatively higher than the trace values obtained in MF1P1 and MF2P1(0.028) (Table 30). Low phosphates in the soil may be as result of its fixation by interlayer hydroxides of chlorites (Tan, 2011). However, when hydrous oxides of Fe and Al are

involved in ligand exchange reactions with phosphates, its availability may be assured only when needed. Fixation of phosphate may also occur in alkaline or near alkaline soils. Many of such soils contain high amounts of soluble and exchangeable Ca^{2+}, and most often $CaCO_3$. The resultant $Ca_3(PO_4)_2$ is insoluble and often precipitates out of solution (Tan, 2011). This fixation process can however, be arrested by humic acids by chelating the ions taking part in the reaction (Lobartini *et al*., 1997). The process ensures its availability in small quantities only when the need arises. P_2O_5 correlated positively and strongly with Fe_o and MnO and weakly negatively with Al_2O_3 and K_2O.

4.5.2 Factor Analysis of Variable Loadings for Metal Oxides of the Soils

Nine (9) soil variables were compressed into 2 principal components (PC): PC1 and PC2, and presented in Table 31.

The 2 PCs contributed 50.52 % of the total variation in soils overlying SLS, SL, SLSi and ALV. 31.42 % of the total soil variation was accounted for by PC1 which was the most influential principal component, while 19.1 % was accounted for by PC2. In this case, PC3 had eigenvalue less than 2.0 and had insignificant contribution to soil variability and was discarded. Furthermore, the high PC1 was influenced by contributions from K_2O, TiO_2, MnO_2 and Fe_2O_3 with loadings of 0.878, -0.790, -0.619 and -0.836, respectively to the variation of soils overlying the four lithologies. The metal oxides were negatively correlated with PC1 except K_2O which correlated positively with PC1. Al_2O_3, SiO_2 and ZrO_2 exerted their contributions to soil variation in PC2 with loadings of 0.655, -0.771 and -0.764, respectively to soil variation. PC2 is a less influential principal component and suggests that the properties manifested in PC2 were of less influence to soil variation between the lithologies. The principal contributor to the elemental oxide component of soils of limestone formation is CaO. The low loading of CaO to the PCs indicate its low level of variation between the soils, an affirmation that the soils were of a similar origin.

PC1 presented SLS and SL as the lithologies with the most contribution to soil variation with contributions of -0.674 and 0.612, respectively to soil variation. PC2 was the most influenced by ALV (-0.550). Eigenvalues of PC1 and PC2 exceeded 2 and indicate that the first two PCs can be adopted, while PC3 is discarded. However, the influence of SLSi may have manifested at a later principal component.

4.6 Clay Mineralogy of the Soils

The weatherable minerals; montmorillonite, palygorskite, vermiculite, lepidomelane and more resistant minerals like quartz and kaolinite dominated the mineralogy of the soils

Table 31: Factor Analysis of Variable loadings for the Forms of Fe and Al

Variables	PC1	PC2
Al_2O_3	0.350	0.655
SiO_2	0.420	-0.771
P_2O_5	-0.502	-0.448
K_2O	0.878	-0.158
CaO	-0.445	-0.186
TiO_2	-0.790	-0.121
MnO_2	-0.619	0.066
Fe_2O_3	-0.836	0.403
ZrO_2	-0.162	-0.764
SLS	-0.674	0.0376
ALV	-0.0492	-0.550
SL	0.612	0.145
SLSi	-0.0213	0.3310
Eigenvalue	4.08	2.48
% total variance	31.40	19.10
Cumulative Eigenvalue	4.08	6.57
Cumulative % variance	31.42	50.52

Foot note: PC = Principal component

formed on various limestone formations in Cross River State. Quartz, kaolinite and montmorillonite were most well distributed and substantial, occurring in virtually all the limestone formations. Though, not obtained across the limestone formations and not necessarily in small quantities, vermiculites, palygorskite and chlorite-vermiculite-montmorillonite combination were other minerals identified in the clay fraction of the soils.

Mineralogical analysis of the clay fractions of a carbonate containing soil in Bambui, Brazil showed the dominance of quartz and kaolinite (Maranhão, 2016). Studies by Khormali *et al*. (2003) and, Monger and Daugherty (1991) revealed the presence of palygorskite, smectites (montmorillonite) and kaolinite as the dominant minerals in calcareous soils. The minerals did not only vary based on limestone formations, they also varied vertically with soil depth, an indication that soil weathering intensity varies with soil depth. Shankar and Achyuthan (2007) also observed a change in mineralogy with soil depth. Such vertical variability may be due to the sedimentary kind of landscape that the soils are exposed to. The minerals obtained in the present study are summarized in Table 32.

a. Quartz

Quartz was the most abundant mineral in the soils, irrespective of the limestone formations (Table 32). Quartz was abundant in the CBk of IH1P1 and absent in IH1P2 for the soils overlying SLS. However, the synonyms of quartz (quartz, syn) were distributed throughout the profiles in this formation, particularly in the Ap and C horizons. Amongst the A, B, C horizons studied, the amount of quartz in the B horizons was less than its abundance in the A and C horizons. In the soils over SL, quartz was either very abundant in the Ap and C horizons (AI1P1, AI2P1) or abundant in the Ap (AI1P2) and BA (AI1P1) horizons. Quartz was only represented in the transition BC horizon of AI2P2. Quartz, syn was either absent (AI1P1), represented (AI2P2), or abundant in the C or BC horizons (AI1P2 or AI2P1). In the soils overlying SLSi, quartz was very abundant in the Bt (MF1P1) and Ap (MF2P2) horizons as well as represented in the Crv horizon of MF1P2. Quartz, syn were very abundant in MF1P1 and MF2P1, but represented in the Ap horizon of MF1P2 and the BC horizon of MF2P2.

Sandstone is commonly intercalated with limestone across the formations, as well as shale in SLS and SLSi formations. The high quartz content may have been inherited from sandstone; as quartz and its synonyms are the most abundant minerals in sandstone, and by extension widely present in most parent materials and soils. According to Asadu *et al*. (2012), most minerals in parent materials are transferred to the overlying soils upon weathering. This might be a probable reason why the primary mineral (quartz) was higher in the C relative to

Table 32: Mineralogy of the soils formed from limestone formations

Horizon	Depth	Qtz	Qtz,syn	Kao	Mon	Ver	Paly	Mon-21A	Mon-15A	C-V-M	Kao-1Ml	Ce-mica	Hem	Par	Na-Ca-Silicate	kao-mon	Nacrite	Mont-syn	mica	Na-Al-Silicate	Mg-Al-Silicate	Mon-heated
IH1P1	Shale, limestone and sandstone formations																					
Ap	0-37		xxx	xx	xx	xx	xx		xx		xx		xx									
CBk	64-130	xxx			xx			xxx		xxx							xx			xx		
CB	130-195		xxx					xxx		xxx										xxx	xxx	
IH1P2																						
Ap	0-42		xxx				xx			x	xxx									xx		x
Bt	42-70		xx	x	xxx			xx		xxx												
Ckt	70-150		xxx	xxx	xx			xxx			xxx						xx				xx	
AI1P1	Sandstone and limestone formations																					
Ap	0-9	xxxx		xxx	xx	xx																
BA	9-54	xxx		xx	xx		xx	xx		xx	xx	Xx							xx(Ce			
Ckt	122-200	xxxx		xxx									xxxx									
AI1P2																						
Ap	0-15	xxx			xxx						xxx			xx				xxx				
Btv	67-146		xxx	xxx	xxx		x	xx							xxx							
Ckt	146-180		xxx					xxx		xxx	xxx											
AI2P1																						
Ap	0-12	xxxx						xxx			xxx		x		xxx		x					
BC	59-120+		xxx	xxx	xx			xxx					x				xx					
AI2P2																						
Ap	0-13		xxx		xxx		x			xxx	xxx											
BC	60-82	xx			xxxx		xxx						xxx									
Ckt	82-132		xx	xxx	xxx	xxx		xxx											x(Pb)			
MF1P1	Shale and limestone with sandstone intercalation																					
Ap	0-17		xxxx																			
Bt	17-56	xxxx		xxx	xxx																	
BC	56-121		xxxx					xxx		xxx	xxx											
MF1P2																						
Ap	0-14		xx		xx	xxx					xxx		x									
Crv	81-153	xx		xxxx															xx(Pb		xxx		
MF2P1																						
Ap	0-16		xxxx	xxxx		x			xxx							xx						
CB	44-97		xxxx	xxx					xxx							xxx						
Cr	97-161		xxxx	xxx	xxx			xxx													x	
MF2P2																						
Ap	0-16	xxxx			xxx											xxxx						
Bt	45-107	xxx		xxx							xxxx					xx			xx(Mn			
BC	107-182		xx	xxx	xx						xxxx					xxx						

N/B: x: poorly represented (<7 %), xx: represented (7-15 %), xxx: abundant (15-30 %), xxxx: very abundant (> 30 %)

the B horizons. However, it occurred in abundance in the B horizons of AI1P1 and AI1P2. This fraction may have been translocated to the B horizon during soil formation.

Quartz is resistant to weathering (Tuncay *et al*., 2019), inert and chemically inactive (Tan, 2011), as a result dominates most soils and contributes so little to the chemical fertility of the soils as it has low cation exchange capacity and small surface area. The small cation exchanges are attributed to the Si-O broken bonds and Si-OH groups on particle edges. The preponderance of quartz and its relative accumulation may be due to its slow weathering rate or resistance in the soils. The loss of other minerals that are easily weathered or influenced by dissolution may have made the accumulation of quartz more obvious. Such silicated soils may require fertilizers for agronomic crop growth. High values of quartz may imply slow weathering potential of the mineral. This is consistent with results obtained by XRF (Tables 28-30), which indicated less weathering intensity in the soils, particularly in those over SLSi.

Quartz was obtained by Kowalska *et al*. (2017) in South Poland as a main mineral phase as well as small amounts of mica and clay minerals in similar soils. Mota *et al*. (2007), Moreira *et al*. (2000) and Lemos *et al*. (1997) observed quartz in soils formed from calcareous materials in the Apodi plateau, Brazil. However, Ferreira *et al*. (2016), described its occurrence in soils over limestone as 'accessory' and further stated that its abundance depends on the purity of the rock. This further explains its non-uniformity in values across the limestone formations as its abundance varied with purity from the soils over SLS to those over SL and SLSi.

b. Kaolinite

In the soils over SLS, kaolinite increased from the Ap horizon (absent) to the Ckt (abundant) of IH1P2 and was only represented in the Ap horizon of IH1P1. The polymorph of kaolinite (kao-1Md) had a similar distribution as kaolinite in IH1P1 and IH1P2. However, kaolinite-1Md was rather abundant in the Ap horizon of IH1P2, while kaolinite was absent in the same horizon. In the soils overlying SL, kaolinite was abundant in the C horizons of the soils except in AI1P2 where kaolinite was absent. The BA and Btv horizons of AI1P1 and AI1P2 were represented and abundant in kaolinite, respectively. Kaolinite – 1Md was abundant in all the Ap horizons of soils over SLSi and absent in the Ap horizon of AI1P1. In the endopedons of soils over SLSi, kaolinite was in abundance, and absent in the Ap horizon of MF1P2, while the Ap horizon of MF2P1 was very abundant in kaolinite. The BC horizons of MF1P1 and MF2P2 were abundant and very abundant in kaolinite-1Md, respectively.

Higher values of kaolinite in the horizons of clay accumulation (Bt and Ckt) relative to the transition, Cr and Ap horizons in IH1P2, AI1P1, AI1P2, AI2P2, MF1P1 and MF2P2

indicate its abundance in clay illuviated horizons; mainly as a result of soil maturity. The dominance of kaolinite in the studied soils, especially in those formed over SL and SLSi formations indicate that most of the weatherable minerals with significant negative charge have disappeared through the weathering process, leaving very low negatively charged kaolinite as the dominant crystalline mineral. Also, weathering may have proceeded to an advanced stage in the soils over SL and SLSi, giving rise to kaolinite and sesquioxides. A situation that may have warranted the occurrence of more hematite in the soils over SL and SLSi. Its presence may have originated from the transformation of primary (feldspars, micas, quartz, pyroxenes and hornblende) or other secondary minerals or by neoformation through recombination of Si in solution. Also, Illite in the studied soils may have been altered, resulting in the formation of kaolinite (Tan, 2011).

Irrespective of the uniqueness of the soils, tropical soils are known to be dominated by kaolinite (Foth, 1990., Esu, 2010 and Asadu *et al*., 2012). Kaolinite has low cation exchangeable capacity and contributes little to the activity of soils and by extension the chemical fertility of the soils. It is expected that the fertility of such soils be boosted by the application of requisite fertilizers. Boero *et al*. (1992) in Sardinia, and Durn (2003) in the Mediterranean region obtained the dominance of minerals ranging from illite and kaolinite in the clay fraction of calcareous soils.

Miura *et al.* (1988) in southwestern Japan, Irmak *et al*. (2007) in the karst environment of Apordi plateau, Brazil and Ahmed *et al*. (1966) obtained kaolinite in limestone soils at various abundances. Kaolinite was also found to be dominant in the soils formed on massive limestone in Pivska Planina (Petar *et al*., 2016). Carroll and Hathaway (1953) had earlier identified kaolinite and chlorite as the end product of the weathering of montmorillonite and mica in limestone soils. Similarly, Ross and Hendricks (1945) revealed kaolinite as the end product of clay weathering in limestone soils. In the Lenoir limestone soils in Augusta County, the abundance of kaolinite was rated as medium. Also, Girao *et al*. (2014) described kaolinite as a common mineral in soils formed from calcareous parent materials, while Bautista *et al*. (2011) opined that kaolinite is more common in soils with long exposure on karstic landscapes. The dominance of kaolinite in the soils may have been as a result of intense weathering condition in the tropical area. By virtue of relative dominance of kaolinite in the tropical rainforest soils over SL and SLSi, they were therefore adjudged more mature compared to soils over SLS which occurred in the southern guinea savannah. However, small amounts of kaolinite were obtained by Moreira *et al*. (2000) in the

carbonate soils of Apodi, and predominated some soil profiles over limestone (Ferreira *et al*. 2016).

c. Montmorillonite

In the soils formed over SLS, montmorillonite was represented in the Ap and CBk horizons of IH1P1 and abundant in the Bt horizon of IH1P2. In the soils overlying SL, montmorillonite was either represented (AI1P1AI2P1), abundant or very abundant in the soils (AI1P2 and AI2P2). Montmorillonite was present in more abundance in the endopedons or in equal measure as the Ap horizons and was irregularly distributed in the soils overlying SLSi. The amount of montmorillonite in the soils indicate that it was either represented or in abundance. Montmorillonite-21 Å was present in the entire soils irrespective of limestone formation and relatively more abundant in the endopedons. However, montmorillonite-15 Å was represented in the Ap horizons of IH1P1 and IH1P2, and abundant in MF2P1. Montmorillonite-syn was identified only in the Ap horizon of AI1P2, while its heated form was poorly represented in the Ap horizon of IH1P2. In general, montmorillonite was amongst the most well-distributed clay minerals in the soils with the occurrence of its various forms in the studied soils.

Montmorillonite is a high activity and expanding clay mineral that is most likely to impart high-base status and natural fertility to the soils. Montmorillonite had earlier been obtained in similar soils elsewhere. For instance, Irmak *et al*. (2007), Singer (1989) and Yilmaz (1990) had identified montmorillonite at various abundances in carbonate rich parent materials. Ahmed *et al*. (1966) obtained only a trace of it in the bauxite soils overlying pure white limestone and even absent in the Lenoir limestone soils of Augusta county (Carroll and Hathaway, 1953). According to Carroll and Hathaway (1953), montmorillonite and mica are the original minerals in limestone soils which are transformed during soil development under the influence of leaching to kaolinite and chlorite. These transformative properties of montmorillonite and kaolinite may be another reason for the dominance of kaolinite in the soils. This assertion negates the findings of Deepthy and Balakrishnan (2005), that minerals in the smectite group are scarce in soils formed from karst in older landscapes with good drainage.

According to Cady *et al*. (1986), the higher the degree of weathering, the lower the ratio of weatherable to resistant minerals. Considering the distribution of minerals in Table 32, it is observed that the studied soils were more dominated by quartz and kaolinite relative to montmorillonite, hence having a higher ratio of weatherable to resistant minerals. This

implies lower degree of weathering. However, the three minerals were distributed across the limestone formations.

The presence of either cambic or argillic endopedons in the entire soils as well as the dominance of kaolinite, quartz and montmorillonite over illite, palygorskite, paragonite and vermiculites coupled with the classification of the soils as Ultisols, Alfisols or Inceptisols (though not in the oxic subgroup) qualify the soils in the intermediate to ultimate stage of weathering. This is according to the criteria of Sys *et al*. (1991). However, the relatively low abundance of kaolinite in the soils over SLS could qualify them in the intermediate stage of weathering.

d. Palygorskite/attapulgite

In soils over SLS, the abundance of palygorskite indicates that it was only represented in the Ap horizons of IH1P1 and IH1P2. Some peaks were also represented by attapulgite in the Ap horizon of IH1P2 and indicated that it was available in abundance. Both minerals are synonymous terms for the same hydrated magnesium aluminum phyllosilicate and have 2:1 structure. In the soils over SL, it was represented in the BA horizon of AI1P1 and poorly represented in the Btv horizon of AI1P2 but absent in AI2P1. In AI2P2, palygorskite was very poorly represented in the Ap horizon and abundant in the BC horizon. However, palygorskite was absent in the soils studied over SLSi.

Though scanty in some of the studied soils, palygorskite was reported in the clay fraction of some limestone derived soils of southern Iran (Emadi *et al*., 2008) and has been identified as the second most abundant mineral in the soils of a karst environment in Apordi plateau, Brazil (Irmak *et al*., 2007). Emadi *et al*. (2008) traced its origin in limestone soils to inheritance, neoformation and transformation from other minerals in limestone soils. However, Bautista *et al*. (2011) obtained a trace of the mineral in the tropical karst of Yucatan, Mexico while, Krekeler *et al*. (2010) obtained it in good measures in similar soils. Palygorskite in the studied soils seems unstable and easily weathered to more stable minerals like kaolinite and quartz.

e. Mica (Mn, Ce, Pb, paragonite and lepidomelane)

Micas were quite scanty and unevenly distributed in the studied soils and either represented or poorly represented in the B and C horizons of AI1P1, AI2P2, MF1P2 and MF2P2, all found in the tropical rainforest area. Micas in the Ap horizons of the studied soils over SL and SLSi may have been weathered as the high rainfall in the area encourages intense weathering. However, paragonite which is a mineral closely related to muscovite dominated the Ap horizon of AI1P2. Paragonite may have been resistant to weathering in the

studied soils, hence its dominance in the Ap horizon. Such soils may not require the application of potassic fertilizers. Muscovite was reported by Miura *et al*. (1988) in the soils overlying limestone in southwestern Japan, while Egli *et al*. (2008) likened the dominance of muscovite and quartz in the soils developed on carbonate rich lithology to the dissolution of carbonates which have left behind, the insoluble and recalcitrant residue. Micaceous minerals are common in calcareous materials and by implication, soils (Kampf and Curi, 2003). In the Apodi plateau, Brazil Ferreira *et al*. (2016) obtained small amounts of mica as interstratified illite-smectite and illite in the clay fraction of the carbonate rich soils.

f. Vermiculite

Vermiculite in the soils overlying SLS was scanty and only represented in the Ap horizon of IH1P1 and absent in IH1P2. In the soils overlying SL, it was represented in the Ap horizon of AI1P1 and abundant in the Ckt horizon of AI2P2. In the soils over SLSi, vermiculite was abundant and poorly represented in the Ap horizons of MF1P2 and MF2P1, respectively. Except for AI2P2, all other soils had vermiculite identified in the Ap horizons. Such low and scanty amounts of vermiculite in soils of limestone origin have been attributed to high soil pH and low activity of Al in the soils (Emadi *et al*., 2008). Evans and Cameron (1979) observed that at an advanced soil age, illite weathers to vermiculite. This may have been a probable reason for the poor identification of illite in the soils. In the soils formed over massive limestone in Pivska Planina, the clay fraction was dominated by vermiculite (Petar *et al*., 2016). Similarly, Cabadas-Baez *et al*. (2010) recorded vermiculite, kaolinite and quartz in the pedo-sediments of karstic sinkholes of Northeast Yucatan, Mexico; while Sedov *et al*. (2008) reported vermiculite in Phaeozems from tectonic karstic plains.

Due to its poor distribution, depth distribution could not be ascertained for vermiculite in the soils studied. However, in Southwestern Japan, the abundance of vermiculite decreased at the endopedons and ranked few in terms of its relative abundance (Miura *et al*., 1988), while in southern Iran, Emadi *et al*. (2008) reported very low vermiculite within the range of 1-6 % in soils of calcareous origin. Vermiculite is a 2:1 expanding clay mineral with high CEC and potential to adsorb soil nutrients for enhanced plant nutrition. Its availability in the soil, even in small quantities will improve the fertility status of the soils.

g. Interstratified chlorite-vermiculite-montmorillonite (C-V-M)

In the soils formed over SLS, C-V-M was basically abundant in the endopedons and only represented in the epipedon of IH1P2. The admixture (C-V-M) increased in the endopedons of the soils. In the soils overlying SL, C-V-M was abundant in the Ckt and Ap horizons of AI1P2 and AI2P2, respectively and was represented in the BA horizon of AI1P1.

However, the mineral was absent in AI2P1. In the soils formed over SLSi, C-V-M was scanty with abundant occurrence in the BC horizon of MF1P1. The abundance of the mineral decreased as rainfall amount increases from the northern agricultural zone to the southern agricultural zone. The interstratification of the minerals is such that, chlorite, vermiculite and montmorillonite are stacked in various ways to result in C-V-M, which is different from the individual components. The combinations are made possible due to the similarity in structure between the components; they are all made of tetrahedral and octahedral sheets, and are ferromagnesian minerals. Distinctions among smectite, vermiculite and chlorite in thin section are seldom possible (SSS, 2014a). They are usually mixed in the soil, seldom occur in pure form and are mixed with ferrous compounds and humus. Amongst the available literature, no such combination of C-V-M was reported in soils developed from limestone. However, chlorite was reported in limestone soils by Intamo *et al*. (2014) in Western Thailand, Ahmad *et al*. (1966) in India, Petar *et al*. (2016) in Pivska Planina and Emadi *et al*. (2008) in southern Iran. Similarly, montmorillonite and vermiculites have been variously reported.

h. Hematite

Hematite was scantily obtained in the soils studied but recorded in the soils of all the formations. In the soils formed over SLS, hematite was represented in the Ap horizon of IH1P1 and absent in IH1P2. In soils of SL, hematite was very abundant in the Ckt horizon of AI1P1 and very poorly represented in AI2P1. However, it was found in abundant amount in the BC horizon of AI2P2. In the soils of SLSi, hematite was only poorly represented in the Ap horizon of MF1P2.

The presence of hematite has been reported in karstic plains of Yucatan by Bautista *et al.* (2003, 2004 and 2011). Mota *et al*. (2007) and Moreira *et al*. (2000) obtained hematite in soils of karstic landscapes. Similarly, Shankar and Achyuthan (2007) identified hematite in the coarse sand fraction of soils formed over limestone in India. According to Kampf and Curi (2012), high temperatures favour the formation of hematite, while SilvaNeto (2008) and Tardi and Nahon (1985) reported that high pH in carbonate rich soils facilitates the precipitation of hematite. The presence of iron bearing minerals in tropical soils depicts soil maturity.

i. Nacrite

In soils of SLS, nacrite was represented in the CBk and Ckt horizons of IH1P1 and IH1P2, respectively. In the soils over SL, it was poorly represented and represented in the Ap and BC horizons, respectively. Its abundance seems to decrease from the northern

agricultural zone to the southern agricultural zone, which follows a decrease in environmental temperature. At increasing temperature, the transition, kaolinite ⟶ dickite ⟶ nacrite, has been documented (Katsumi, 1989). This negates findings in the soils studied. The presence of nacrite in the Flood plain soils of Onwu River was reported by Akpan-Idiok and Ogbaji (2013). Nacrite, together with kaolinite and dickite have similar chemical composition: 46.5 SiO_2, 39.5 Al_2O_3 and 13.9 % H_2O, leading to the formula $Al_2Si_2O_5(OH)_4$. However, nacrite has a different stacking sequence; it was earlier interpreted as a 6-layer structure, and now seen as a 2-layer structure in which the X and Y axes interchange (Brindley and Brown, 1980).

4.7 Soil Classification

a. Shale, limestone and sandstone formations (SLS) in northern agricultural zone

Soils in IH1 had argillic horizons with base saturation of less than 50 % determined by NH_4OAc (pH 7.0) at 125 cm or more below the soil surface and do not have fragipans. The soils qualified as Ultisols at the soil order category. They were freely drained in the southern guinea savannah with Ustic moisture regime. They met the requirement for Ustults in the suborder category. The epipedon of IH1P1 had colour value of 3 (moist) and subhorizons within 100 cm of the argillic horizon had colours with hue of 2.5YR or redder and value of 3 (moist). The soil qualified as Rhodustult in the great group and as Typic Rhodustult at the subgroup category of the USDA Soil Taxonomy System (SSS, 2014b). It has loamy skeletal particle size class (>35 % gravels; <35 % clay), mixed mineralogy class; superactive cation exchange activity class (> 0.60) and Isohyperthermic soil temperature regime. IH1P1 therefore qualified in the Family category; Loamy Skeletal, Mixed, Superactive, Isohyperthermic, Typic Rhodustults. IH1P1 correlated with Ferric Rhodic Acrisols (Ochric) in the World Reference Base for Soil Resources (FAO, 2014). In the Bohemian karst region, limestone soils were classified as Rendzic, Lithic or Hyperskeletic Leptosols in the World Reference Basae for soil resources system (Samonil, 2007).

On the other hand, IH1P2 did not have a densic, lithic, paralithic and petroferric contact within 150 cm of the mineral soil surface and, with increasing depth did not have a clay decrease of 20 % or more from the horizon of maximum clay accumulation. The soil qualified as Paleustult at the great group and Typic Paleustult at the subgroup category of the USDA Soil Taxonomy System. At the family level, IH1P2 qualified as Loamy Skeletal, Mixed, Superactive, Isohyperthermic, Typic Paleustult. This negates classification of similar soils elsewhere, as Calcic Haploxeralf and Typic Calcixerepts were obtained in the highly

calcareous soils of southern Iran (Emadi *et al*., 2008). IH1P2 correlated with Ferric Acrisols (Cutanic) in the World Reference Base for Soil Resources.

In IH2, the soils had umbric (IH2P1) and ochric (IH2P2) epipedons and cambic endopedons by virtue of sandy clay loam texture, weak indications of argillic horizons (higher clay in the subsurface soils) as well as higher colour value and without inherent rock structure. The soils qualified as Inceptisols in the soil order category. Just beneath the epipedons and within 50 cm of the soil surface, the soils had chroma of 1 with redox concentrations and ferrous Fe in IH2P2. This qualified the soil as Aquept at the suborder category. IH2P1 had an umbric epipedon and qualified as Humaquept, while IH2P2 had high ground water table fluctuating from 45 to 87 cm and endosaturated, and classified as Endoaquept. A colour value of 3 in the Ap horizon as well as base saturation of less than 50 % (by NH_4OAc at pH 7.0) in most part of IH2P2 qualified it as Humic Endoaquept. IH2P2 was correlated with Stagnic Gleyic Cambisol (Humic) in the World Reference Base for Soil Resources. However, IH2P1 had organic carbon content of more than 0.2 % at 125 cm below the mineral soil surface. This qualified it as Fluvaquentic Humaquept in the USDA Soil Taxonomy System which was correlated with Gleyic Cambisol (Humic) in the World Reference Base for Soil Resources. Similar soil orders were obtained by Emadi *et al*. (2008) in the highly calcareous soils of southern Iran. They obtained Typic Calcixerepts and Petrocalcic Calcixerepts.

b. Sandstone and limestone formations (SL) in the central agricultural zone

Soils in AI1 had argillic B horizons, base saturation values of less than 50 % (NH_4OAc at pH 7.0) with ochric epipedons by virtue of their thin Ap horizons (less than 25 cm thick). AI1P1 and AI1P2 were categorized as Ultisols at the soil order category. The soils had less than 0.9 % weighted average of organic carbon in the upper 15 cm above the argillic horizon, udic soil moisture regime, well drained and found in the tropical humid climate. The soils qualified as Udults in the suborder category. With increasing depths, the soils do not have clay decrease of 20 % or more from horizons of maximum clay accumulation. They qualified as Paleudults in the great group category. AI1P1 had more than 50 % colours with hue of 2.5YR or redder and colour value of 3 in the upper 75 cm of the argillic horizon and qualified as Rhodic Paleudults in the subgroup category of the USDA Soil Taxonomy System. It had sandy particle size class (loamy sand textural class), mixed mineralogy, and Isohyperthermic soil temperature class. AI1P1 therefore qualified in the Family class; Sandy, Mixed, Superactive, Isohyperthermic, Rhodic Paleudults. AI1P1 was correlated with Ferric Acrisols (Ochric) in the World Reference Base for Soil Resources.

On the other hand, AI1P2 had more than 5 % plinthites in one horizon within 150 cm of the soil surface and qualified as Plinthic Paleudults at the subgroup category of the USDA Soil Taxonomy System. In the family category, the soil qualified as Fine loamy, Mixed, Superactive, Isohyperthermic, Plinthic Paleudults. In the Harran region of Turkey, similar soils were classified as Xeric Haplocalcid and Lithic Xeric Haplocalcid and correlated with Haplic Calcisols (Irmak *et al*., 2007). AI1P2 therefore, correlated with Plinthic Acrisols (Differentic, Hyperdystric, Ochric) in the WRB for soil resources system.

In AI2; AI2P1 had argillic B horizon at 12-59 cm of the mineral soil surface with base saturation (NH_4OAc at pH 7.0) of less than 50 % and clay content in all horizons of less than 30 %. The soil qualified as Ultisol. The soil had Udic moisture regime, less than 0.9 % weighted average organic carbon in the upper 15 cm of the argillic horizon which had reddish brown colour. It qualified as Udult at the suborder level. With increasing depth, the soil does not have clay decrease of more than 20 % from horizon of maximum content within 150 cm of mineral soil surface. The soil does not have plinthites and kandic horizon and meets the criteria for classification as Paleudult at the great group category. Cracks in the Bt horizon through a thickness of over 40 cm qualified the soil as Vertic Paleudult at the subgroup category of the USDA Soil Taxonomy System. At the family level of classification, AI2P1 qualified as Loamy, Mixed, Superactive, Isohyperthermic, Vertic Paleudult. AI2P1 was correlated with Haplic Acrisols (Differentic, Hyperdistric, Ochric, Profondic) in the World Reference Base for Soil Resources. Similar soils in the Bursa Province of Turkey were classified in a similar way. Vertic Xerochrepts (Haplic Calcisol), Chromic Haploxererts (Eutric Vertisol) and Typic Calcixererts (Calcic Vertisol) were obtained by Aksoy *et al*. (2009).

AI2P2 had an ochric epipedon by virtue of its thin epipedon (13 cm thick), an argillic horizon and high base saturation (> 50 % by NH_4OAc at pH 7.0). The soil qualified as Alfisol at the soil order category. With udic soil moisture regime, the soil qualified as Udalf. It does not have a clay decrease of more than 20 % from horizon of maximum clay accumulation with increasing depth. Redox concentrations with hue of 5YR and 10YR, chroma of 6 in the subhorizons qualified it as Paleudalf. Many wide cracks (1.0-1.7 cm) extending through a thickness of over 50 cm qualified the soil as Vertic Paleudalf at the subgroup category of the USDA Soil Taxonomy System. AI2P2 had coarse loamy (>15 % gravel; <18 % clay) particle size class, smectitic mineralogy class, semi active cation exchange activity class and Isohyperthermic soil temperature class. It qualified in the Family; Coarse loamy, Smectitic, Semi active, Isohyperthermic, Vertic Paleudalf. The classification of similar soils by Emadi

et al. (2008) in southern Iran obtained similar results at the order category as they obtained Calcic Haploxeralf. In the contrary, Aksoy *et al*. (2009) obtained Vertic, Typic and Chromic Xerochrepts for similar soils in the Bursa Province of Turkey. AI2P2 was correlated with Vertic Ferric Luvisols (Differentic, Loamic, Ochric, Profondic) in the World Reference Base for Soil Resources.

The poorly drained soils of AI3 had ochric epipedons over cambic endopedons and qualified as Inceptisols at the soil order category. The soils had aquic conditions within 50 cm of the mineral soil surface with chroma of 2 or less and distinct red mottles (10R 4/8) (redox concentrations) in AI3P1 and strong brown in AI3P2. The soils meet the criteria for Aquepts. They had high water table and were endosaturated and qualified as Endoaquepts in the great group category. The Ap horizon of AI3P1 had colour value (moist) of 3 and base saturation (NH_4OAc at pH 7.0) of less than 50 % in all horizons within 100 cm of the soil surface and qualified as Humic Endoaquept in the subgroup category of the USDA Soil Taxonomy System.

AI3P1 was correlated with Dystric Gleyic Cambisols (Humic, Loamic) in the World Reference Base for Soil Resources. However, AI3P2 had less than 0.2 % organic carbon at about 124 cm depth from the soil surface with values that decreased regularly to a depth of 124 cm. The soil had colour value (moist) of 4 at the Ap horizon and qualified as Typic Endoaquept at the subgroup category of the USDA Soil Taxonomy System. AI3P2 was correlated with Dystric Cambisols (Arenic, Ochric) in the World Reference Base for Soil Resources. Similar soil order was obtained by Aksoy *et al*. (2009). Aksoy *et al*. (2009) classified similar soils in Bursa Province of Turkey as Typic Xerochrepts in the USDA System which correlated with Calcaric Cambisol in the WRB System.

c. Shale and limestone with sandstone intercalation in the Southern agricultural zone

Soils in MF1 had low base saturation with values less than 50 % (NH_4OAc at pH 7.0) in the argillic B horizons. The soils qualified as Ultisols in the soil order category. The soils had 0.9 % and more organic carbon by weighted average in the horizons above the Bt horizon and qualified as Humults in the suborder category. The presence of plinthites between 81 and 153 cm of the soil surface in MF1P2 qualified it as Plinthohumult in the great group and as Typic Plinthohumult in the subgroup category of the USDA Soil Taxonomy System (SSS, 2014b). At the family level of classification, the soil met the criteria for classification as Coarse Loamy, Kaolinitic, Superactive, Isohyperthermic, Typic Plinthohumult. MF1P2 was correlated with Haplic Acrisols (Differentic, Humic, Hyperdystric) in the World Reference Base for Soil Resources. On the other hand, with increasing soil depth MF1P1 did not have a

clay decrease of 20 % or more from horizon of maximum clay content and qualified as Palehumult. The soil did not have plinthites but had udic soil moisture regime and qualified as Typic Palehumult in the subgroup category of the USDA Soil Taxonomy System. MF1P1 had coarse loamy particle size class, siliceous mineralogy class, superactive cation exchange activity class and Isohyperthermic soil temperature class. The soil qualified in the Family class of Coarse loamy, Siliceous, Superactive, Isohyperthermic, Typic Paleuhumult. MF1P1 was correlated with Haplic Acrisols (Humic, Hyperdystric, Loamic, Profondic) in the World Reference Base for Soil Resources. Contrary classifications were obtained by others. For instance, Emadi *et al*. (2008) obtained Calcic Haploxeralf, while Irmak *et al.* (2007) classified the soils as Xeric and Lithic Haplocalcid. Furthermore, Aksoy *et al*. (2009) classified similar soils elsewhere as Vertic, Typic and Chromic Xerochrepts.

In MF2; MF2P1 had an ochric epipedon by virtue of its thin horizon (< 25 cm) and low base saturation (< 50 % by NH_4OAc at pH 7.0) as well as the presence of cambic endopedons within 100 cm depth. The soil qualified as Inceptisol at the soil order level. The udic soil moisture regime as well as Isohyperthemic temperature regime qualified it as Udept in the suborder category. Base saturation of less than 60 % (NH_4OAc at pH 7.0) in all endopedons as well as the absence of fragipan and duripan within 100 cm of soil mineral surface qualified the soil in the great group Dystrudept. The soil qualified as Typic Dystrudept in the subgroup category of the USDA Soil Taxonomy System. Being characterized by coarse loamy particle size class, siliceous mineralogy, superactive cation exchange activity class and isohyperthermic soil temperature class, the soil qualified in the Family class of Coarse loamy, Siliceous, Superactive, Isohyperthermic, Typic Dystrudept. MF2P1 was correlated with Dystric Cambisols (Humic, Loamic) in the World Reference Base for Soil Resources System. On the other hand, the presence of an argillic horizon with a thickness of 62 cm and base saturation less than 50 % (NH_4OAc at pH 7.0) qualified MF2P2 as Ultisol in the soil order category. The soil was freely drained with greater than 0.9 % of organic carbon in the upper 15 cm of the argillic horizon. It qualified as Humult in the suborder category. From horizon of maximum clay accumulation, the soil does not have clay decrease of greater than 20 % and qualified as Palehumult at the great group level. Udic soil moisture regime, absence of plinthites and bulk density values greater than 1.0 Mg/m^3 qualified the soil as Typic Palehumult at the subgroup level of the USDA Soil Taxonomy System. In the family category, MF2P2 qualified as Sandy, Mixed, Superactive, Isohyperthermic, Typic Palehumult. This classification contradicts findings by Silva *et al*. (2013), Irmak *et al*. (2007) and Emadi *et al*. (2008) on similar soils elsewhere. In the World

Reference Base for Soil Resources System, MF2P2 was correlated with Chromic Acrisols (Humic, Hyperdystric).

In MF3 soils; MF3P1 did not have a mollic epipedon and had low base saturation values of less than 50 % (by NH_4OAc at pH 7.0) in all horizons. No cambic but ochric epipedon by virtue of its thin epipedon (<25 cm) and low base saturation. It qualified as an Entisol in the soil order category. The soil was permanently saturated with water and had reducing or redoximorphic properties in all sub-horizons and qualified as Aquent in the suborder category. It had organic carbon values greater than 0.2 % in all its horizons, textural class finer than loamy fine sand and with an Isohyperthemic temperature regime and qualified as Fluvaquent at the great group category. Colour hue of 5YR and redder, value of 5 and less as well as chroma of 1 in the Ap horizon qualified it as Aeric Fluvaquent in the subgroup category of the USDA Soil Taxonomy System (SSS, 2014[b]). MF3P1 was correlated with Dystric Regosols (Loamic, Ochric) in the World Reference Base for Soil Resources.

On the other hand, MF3P2 had an ochric epipedon as well as a cambic endopedon and qualified as Inceptisol in the soil order category. In a horizon (BC) just beneath the epipedon, chroma of 1 was obtained with faint reddish mottled redox concentrations; this qualified the soil as Aquept in the suborder category. The soil was endosaturated with very low SAR and qualified as Endoaquept in the great group category. With organic carbon content of more than 0.2 % in all the horizons of MF3P2, the soil was classified as Fluvaquentic Endoaquept at the subgroup category of the USDA Soil Taxonomy System. MF3P2 was correlated with Dystric Fluvic Cambisols (Humic) in the World Reference Base for Soil Resources. Aksoy *et al*. (2009) correlated Typic Xerochrepts in the USDA System of soil classification with Calcaric Cambisol in the WRB for soil resources system.

The summary of the afore-mentioned soil classifications is shown in Table 33.

4.8 Soil Chemical Indices of Weathering

The results of the soil indices of weathering are presented in Tables 34 and 35.

4.8.1 Ruxton Index (R)

In the soils overlying SLS in northern agricultural zone, Ruxton index (R) had ranges of 1.99-4.03 and 2.41-3.17 in IH1P1 and IH2P2, respectively (Table 34). The values of R increased with soil depth and somewhat so regularly with coefficient of variability of 31.5 and 14.9 % in IH1P1 and IH2P2, respectively. The CV values are moderate to low on the scale of Wilding (1985). In soils overlying SL in central agricultural zone, ranges of 1.79-5.03, 2.14-4.23 and 2.44-4.47 were obtained in each of AI1P1, AI2P2 and AI3P1 (Table 34). Though irregular, values decreased with increasing soil depth with CV greater than 30 %,

Table 33: Summary of Soil Classification

Pedons	USDA System	WRB for Soil Resources
IH1P1	Loamy Skeletal, Mixed, Superactive, Isohyperthermic, Typic Rhodustults	Ferric Rhodic Acrisols (Ochric)
IH1P2	Loamy Skeletal, Mixed, Superactive, Ishyperthermic, Typic Paleustult	Ferric Acrisols (Cutanic)
IH2P1	Fluvaquentic Humaquept	Gleyic Cambisol (Humic)
IH2P2	Humic Endoaquept	Stagnic Gleyic Cambisol (Humic)
AI1P1	Sandy, Mixed, Superactive, Isohyperthermic, Rhodic Paleudults	Ferric Acrisols (Ochric)
AI1P2	Fine loamy, Mixed, Superactive, Isohyperthermic, Plinthic Paleudults	Plinthic Acrisols (Differentic, Hyperdystric, Ochric).
AI2P1	Loamy, Mixed, Superactive, Isohyperthermic, Vertic Paleudult	Haplic Acrisols (Differentic, Hyperdystric, Ochric, Profondic)
AI2P2	Coarse-loamy, Smectitic, Semiactive, Isohyperthermic, Vertic Paleudalf	Vertic Ferric Luvisols (Differentic, Loamic, Ochric, Profondic)
AI3P1	Humic Endoaquept	Dystric Gleyic Cambisols (Humic, Loamic)
AI3P2	Typic Endoaquept	Dystric Cambisols (Arenic, Ochric)
MF1P1	Coarse loamy, Siliceous, Superactive, Isohyperthermic, Typic Paleuhumult	Haplic Acrisols (Humic, Hyperdystric, Loamic, Profondic)
MF1P2	Coarse Loamy, Kaolinitic, Superactive, Isohyperthermic, Typic Plinthohumult	Haplic Acrisols (Differentic, Humic, Hyperdystric)
MF2P1	Coarse-loamy, Siliceous, Superactive, Isohyperthermic, Typic Dystrudept	Dystric Cambisols (Humic, Loamic)
MF2P2	Sandy, Mixed, Superactive, Isohyperthermic, Typic Palehumult	Chromic Acrisols (Humic, Hyperdystric).
MF3P1	Aeric Fluvaquent	Dystric Regosols (Loamic, Ochric)
MF3P2	Fluvaquentic Endoaquept	Dystric Fluvic Cambisols (Humic)

Table 34: Weathering indices of the soils on SLS and SL

Soil depth	Ruxton Index	WI	Molar Ratio
IH1P1			
0-30	2.38	0.62	1.12
30-60	1.99	1.04	1.14
60-90	2.27	3.79	1.29
90-120	2.59	3.73	1.51
120-150	4.02	2.44	2.21
150-180	4.03	1.94	1.89
Mean	2.88	2.26	1.53
Range	1.99-4.03	0.62-3.79	1.12-2.21
CV	31.5	58.7	28.9
IH2P2			
0-30	2.41	1.74	1.20
30-60	3.15	1.76	1.83
60-87	3.17	1.62	1.80
Mean	2.91	1.71	1.61
Range	2.41-3.17	1.62-1.76	1.20-1.83
CV	14.9	4.4	22.1
AI1P1			
0-30	5.03	7.49	4.44
30-60	2.86	6.90	2.43
60-90	1.93	5.98	1.63
90-120	1.94	6.26	1.64
120-150	2.01	6.44	1.71
150-180	1.87	6.46	1.47
180-200	1.79	7.19	1.34
Mean	2.49	6.67	2.09
Range	1.79-5.03	5.98-7.49	1.34-4.44
CV	47.3	8.1	52.1
AI2P2			
0-30	4.23	10.58	3.44
30-60	3.01	9.49	2.02
60-90	2.45	11.11	1.91
90-120	2.14	12.07	1.53
Mean	2.96	10.81	2.23
Range	2.14-4.23	9.49-12.07	1.53-3.44
CV	31.2	10.0	37.6
AI3P1			
0-30	4.47	8.01	3.87
30-60	3.42	7.35	3.08
60-80	2.44	8.41	2.10
Mean	3.44	7.92	3.02
Range	2.44-4.47	7.35-8.41	2.10-3.87
CV	29.5	6.8	29.4

Foot note: CV= Coefficient of variability, WI= weathering index

Table 35: Weathering indices of the soils on shale and limestone with sandstone intercalation

Soil depth	Ruxton Index	WI	Molar Ratio
MF1P1			
0-30	3.40	7.30	2.87
30-60	3.03	7.00	2.33
60-90	3.16	7.89	2.53
90-120	3.29	7.33	2.58
Mean	3.22	7.38	2.58
Range	3.03-3.40	7.00-7.89	2.33-2.87
CV	5.0	5.0	8.6
MF2P1			
0-30	2.99	3.69	1.80
30-60	2.57	3.42	1.50
60-90	2.57	3.47	1.41
90-120	2.75	3.98	1.54
120-150	2.84	4.16	1.69
150-161	2.66	3.52	1.17
Mean	2.93	3.71	1.52
Range	2.57-2.99	3.42-4.16	1.17-1.80
CV	6.0	8.1	14.3
MF3P2			
0-30	2.62	6.55	2.24
30-49	4.51	7.21	4.27
Mean	3.57	6.88	3.26
Range	2.62-4.51	6.55-7.21	2.24-4.27
CV	37.5	6.80	44.1

Foot note: CV= Coefficient of variability, WI= weathering index

such values of CV were rated moderate (Wilding,1985). In the soils overlying SLSi in southern agricultural zone, ranges of 3.03-3.40, 2.57-2.99 and 2.62-4.51 were obtained for MF1P1, MF2P1 and MF3P2, respectively (Table 35). Values varied irregularly with increasing soil depth and very low CV of 5-6 % in MF1P1 and MF2P1, respectively.

Though values of R varied irregularly with increasing soil depth, values in IH1P1 and IH2P2 increased with soil depth, while soils overlying SL and SLSi in central and southern agricultural zones, respectively decreased irregularly with increase in soil depth. Ruxton (1968) had stated however, that an ideal trend for the index is the increase in values up the profile which indicates increase in weathering. Consequently, MF1P1, AI2P2, MF2P1 and AI1P1 seem to be more weathered in the surface soils perhaps due to exposures to increased temperature and precipitation. However, the trend in the soils over shale, limestone and sandstone in northern agricultural zone contradicts those of other formations, probably due to partly contrasting temperature and rainfall regimes as well as disparity in vegetation. A comparison of the well-drained soils in the three limestone formations using mean values of R index presents the following increasing weathering sequence;

AI1P1<IH1P1<MF2P1<AI2P2<MF1P1; an indication that MF1P1 and AI2P2 were the most developed and weathered. For the poorly drained soils, the following increasing trend; IH2P2<AI3P1<MF3P2 was obtained for the R index. This indicates that soils overlying SLSi were the most developed and weathered. with increasing soil depth resulting in a range of CV values of 6.8-10 % in the three pedons. The CV values are rated low on the scale of Wilding (1985).

The trend in values shows that, they appeared to have increased in the B horizons, except for the higher values in the surface soils. In the soils of SLSi in southern agricultural zone, weathering index had ranges of 7.0-7.89, 3.42-4.16 and 6.55-7.21 in MF1P1, MF2P1 and MF3P2, respectively. Values varied irregularly with increasing soil depth with CV values having a range of 5.0-8.1 %. The CV values are rated low (Wilding, 1985).

The values of WI were higher in the soils overlying SL and SLSi in the central and southern agricultural zones, respectively than those overlying SLS in northern agricultural zone. This variation between limestone formations may be due to the contrasting vegetation and climate. The northern agricultural zone is mainly southern guinea savannah, while the central and southern agricultural zones are tropical rainforests. According to Evans and Cameron (1979), WI is directly related to soil age; therefore, higher values of WI indicate more exposure to intense weathering condition. Consequently, the intensity of chemical weathering was higher in the soils over SL and SLSi in central and southern agricultural

zones, respectively which were by far higher in the values of WI compared to those in SLS formations. For the well-drained soils, the order of soil development or maturity as assessed by the mean values of WI in increasing order is:

IH1P1<MF2P1<AI1P1<MF1P1<AI2P2, while the order in the poorly drained soils were arranged in increasing order as:

IH2P2<MF3P2<AI3P1.

The above order further affirms finding on the forms of Fe and Al as well as clay mineralogy of the comparatively more mature status of soils overlying SLSi in the southern agricultural zone. Depth wise, relatively higher subsurface values, especially in the B horizons indicate increased weathering intensity with increasing soil depth. This negates earlier findings by Evans and Cameron (1979), that intensity of weathering is higher in the surface soils. The irregular variation of WI values with increasing soil depth indicate irregular soil weathering with depth. Earlier, Che *et al*. (2012) reported that soil weathering with depth lack uniformity, while Walia and Rao (1999) attributed such irregularity to youthful stage of soil genesis.

4.8.3 Molar Ratio (MR)

In the soils overlying SLS in northern agricultural zone, molar ratio (MR) had ranges of 1.12-2.21 and 1.20-1.83 in IH1P1 and IH2P2, respectively (Table 34). Values increased regularly with increasing soil depth in IH1P1 resulting in CV value of 28.9 %. This CV value is rated moderate on the scale of Wilding (1985). Values of MR in IH2P2 increased in the 30-60 cm depth and decreased slightly thereafter. In the soils overlying SL in central agricultural zone, MR had ranges of 1.34-4.44, 1.53-3.44 and 2.10-3.87 in AI1P1, AI2P2 and AI3P1, respectively (Table 29). Values decreased almost regularly with increasing soil depth resulting in coefficient of variability greater than 29 % in the entire soils. Such CV values are rated moderate on the scale of Wilding (1985). In the soils developed from SLSi in southern agricultural zone, ranges of 2.33-2.87, 1.77-1.80 and 2.24-4.27 were obtained in MF1P1, MF2P1 and MF3P2, respectively (Table 35). Values were somewhat irregularly distributed with increasing soil depth and had CV values of 8-14 % in the well-drained MF1P1 and MF2P1. According to Burt *et al*. (2003) and Msanya *et al.* (2003), weathering is at a comparatively more advanced stage where there are low values than high values. Also, the entire soils had mean values of the ratio less than 3.0 except in MF3P230-49, AI1P10-30, AI2P20-30 and AI3P1. When values of the ratio are less than 3.0, the soils may be said to contain kaolinite and gibbsite (Schaetzl and Anderson, 2005). Soils high in gibbsite and kaolinite are highly weathered. Going by the mean values of studied profiles, in order of

decreasing weathering, the well-drained soils are arranged in the following order: MF2P1> IH1P1>AI1P1>AI2P2>MF1P1, while the order in the poorly drained soils was: IH2P2>AI3P1>MF3P2. Soils overlying SLSi in southern agricultural zone represented by MF2P1 was more weathered or developed than their counterparts in other limestone formations.

Ruxton index correlated strongly and positively with molar ratio (p= 5 %) (r= 0.85), while weathering index correlated weakly but positively with molar ratio (r= 0.477). This indicates that molar ratio was compatible with other indices, while R and WI were not.

4.9 Processes of Soil Formation

An attempt to identify the dominant soil forming processes in the study areas was made following soil temperature and rainfall regimes as well as observed field properties and results from laboratory studies. Identified soil forming processes include:

a. Lessivation or Argilluviation

The presence of fissures in AI2P1, AI2P2, AI1P1, AI1P2, and MF1P1 which act as a path way for the translocation of clay size materials in suspension, cutans on ped faces in the Bt, BC and Ckt horizons of the well-drained upland soils as well as the presence of a significant accumulation of illuviated clay in the Bt or Ckt horizons (ratio of clay in the illuvial:eluvial horizon > 1.2) were possible indicators of mechanical downward movement of clay from Ap, AB or BA horizons. This was seen in the comparatively higher values of clay in the argillic horizons compared to the eluvial horizon or immediate horizon overlying it. Lessivage is inhibited in an environment where clays are coated with humus, as it causes aggregation and subsequent immobilization (Sanborn and Pawluk, 1983).

The argillic B horizons were common in the freely draining soils of IH1P2, AI2, AI2, MF1 and MF2P2. According to Cremeens and Mokma (1986), Bt horizons are common in well-drained upland soils that have depleted nutrients in the humid climates and which have undergone wet-dry cycles. The soils under study were low and barely medium in exchangeable Ca and had the process of lessivage occurring in the soils. Lessivage is weak or inexistent in calcareous soils due to dominance of Ca and Mg ions (Schaetzl and Anderson, 2005). This may not be applicable to the soils under study as Ca was low or barely moderate, and had the process of lessivage occurring in the soils. Decarbonation is often regarded as a prerequisite for the process of lessivage.

b. Calcification-decalcification

The formation or accumulation of chalky calcic horizons indicated by CBk and Ckt in IH1P1, AI1P1, AI1P2, IH1P2 and AI2P2 indicate the accumulation of $CaCO_3$, hence

calcification process. When there is a source of Ca in an area with inadequate moisture to remove it from the profile, secondary carbonates accumulate (Stuart and Dixon, 1973), giving rise to calcic horizons. However, in a tropical humid climate such as the tropical rainforest of Cross River State which is characterized by high rainfall, Ca and Mg are often leached in the process of decalcification.

The absence of clear Bk horizon indicates that the process of calcification was truncated by high rainfall regime associated with the area, hence the occurrence of decalcification at a relatively faster rate than calcification. When CO_2 combines with H_2O, H_2CO_3 is formed. This weak acid renders $CaCO_3$ soluble as Ca^{2+} and HCO^{3-} ions which are more easily lost by decalcification. The process seems to have favoured decalcification mainly because of high moisture and low to medium pH. Calcification occurred alongside lessivage in most calcic horizons. The dissolution and leaching of Ca- carbonates are not alien to soils developed on limestone. According to Petar *et al*. (2016), dissolution and leaching of alkaline earth carbonates are common pedogenic processes in the soils formed on the massive limestone of Pivska Planina.

c. Gleization

The poorly drained and anaerobic soils had gray to dark reddish gray and very dark greenish gray (Gley1 3/1) colours in the subsurface soils. Particularly in mapping unit IH2 (IH2P1 and IH2P2) with the reduction of Fe^{3+} to Fe^{2+} as shown in profile IH2P2 (Plate 4). Hue of 2.5Y and chroma of 1 in the endopedons qualify IH2P2 and IH2P1 for the process of gleization. The presence of reduced Fe and Mn result in the expression of gray colours (FAO, 2014). Over 5 per cent of the exposed surface of the poorly drained soils of MF3 (P1 and P2) had signs of oxidation in the root channels with rusty brown and red colours dominating the BC horizon of MF3P2. These are signs of gleization. The poorly drained soils in Ishibori (IH2) and Mfamosing (MF3) were permanently endo- and sometimes anthro-saturated with high organic carbon content in IH2 and moderate in MF3. However, the somewhat coarse textures in AI3 limited the process of gleization. Fine textured soils facilitate gleying (FAO, 2014).

d. Lateritization and latosolization

The soils are leached of basic cations such as Ca, Mg, K and Na as well as Si especially in the well-drained soils. This is shown in the low content of these cations (Tables 21-23; 28-30). Soil reaction was slightly acid to neutral with mottles occurring in the endopedons of most of the soils. An indication that the soils were moderately drained in the tropical humid environment. Fe_2O_3 determined by XRF was high in all the soils with Fe

cutans occurring in the endopedons of IH1P1, IH1P2, MF2P1 (Tables 6 and 12) and AI2P2 (Table 9). The dominance of Fe cutans in the endopedons of IH1P1, IH1P2, MF2P1 and AI2P2 coincides with the molar ratio rating of the soils as comparatively more mature. The presence of cutans (Ferrans) is indicative of soil maturity. Most importantly, molar ratios within the range of 1.33-2.0 indicate lateritic soils, while values greater than 2.0 are non-lateritic soils (Schaetzl and Anderson, 2005). Except for AI3P1, MF1P1 and MF3P2 that had values greater than this range, other soils particularly those overlying SLS in northern agricultural zone were lateritic as values were within the range of 1.33-2.00. This process seems to have occurred alongside latosolization in which residual sesquioxides accumulate with little assumed in-migration of Fe, as bases and silica are leached from the soil. Some of the bases that are released by weathering are biocycled, leading to slightly higher concentration in the A horizon than at depth (Schaetzl and Anderson, 2005). This phenomenon was observed in virtually all the soils.

e. Faunal pedoturbation

Activities of termites, worms and ants as well as their casts were seen in the surface and subsurface of the entire soils, with number and size decreasing as depth increased. These are further presented on the tables for soil morphological properties. Faunal pedoturbation has been considered a major process that influences soil characteristics (Johnson *et al*., 2003) and pedogenesis. Though these organisms were present in the soils, their effect on soil development was not pronounce as soil properties varied with soil depth. For instance, organic carbon varied from the Ap to underlying horizons. However, the dominance of transition horizons may have been as a result of soil mixing by these organisms. This process may not have reached an advanced stage.

f. Mineralization and immobilization

Except for IH1P2, IH2P2, MF1P2 and MF2P2 which had C/N ratios of 26:1, 31:1, 22:1 and 49:1, respectively (Tables 21 and 23), the soils had C/N ratios less than 20:1 with mean annual temperature greater than 22 ^{o}C (Sambo *et al*. 2016b) in the tropical humid region. According to Deenik (2006) net maximum mineralization occurs at C:N ratio of less than 20:1 in aerobic soil conditions at temperatures between 30 and 35 ^{o}C under moist soil condition for enhanced activities of soil microbes. Having met the conditions, the said soils experienced organic matter mineralization, releasing mineral forms of N for either plant uptake or leaching. However, those with high C/N ratios (listed above) are likely to have experienced immobilization as decomposing organisms will have to mine the soils of available mineral N and NH_4^+ to help in the decomposition of the organic materials. The

'robbing process' acts as an energy source for decomposers. Slow mineralization and accumulation of organic matter were outlined as dominant pedogenic processes in the soils formed on the massive limestone of Pivska Planina (Petar *et al*., 2016).

4.10 Land Evaluation

4.10.1 Land Capability Classification of the Soils

Land characteristics used in classifying the soils into USDA Land Capability Classes are presented in Table 36. Table 37 was obtained by comparing Table 36 to Table 4.

a. In the soils overlying SLS in northern agricultural zone, soils in mapping unit IH1 were deep (> 100 cm) with sandy clay loam and sandy loam textural classes but with gravel content that exceeded 15 %. They were moderately well-drained and were not susceptible to flooding. Rainfall amount in the area exceeded 1500 mm/annum resulting in the low base saturation of less than 40 %. However, the soil reaction of IH1P2 exceeded 6.5 required for crop production in the tropics (Udo e*t al*., 2009). IH1P1 and IH1P2 therefore qualified in the subclass IIsf. The use of cover- and green manure crops as well as animal manure and a recommended dose of mineral fertilizer may increase soil organic matter content and by implication, the saturation of exchangeable bases in the soil exchange complex. The use of such soil amendment/combination is most likely to upgrade the soils to a class I soil.

Soils of mapping unit IH2 were characterized by sandy loam textural class with soil reaction within the range of 5.2 and 6.7 as well as low base saturation. Climate interms of effective precipitation and temperature regime was optimum for crop production. However, the soils were limited by gravel content that exceeded 20 %, poor drainage and flooding condition. Shallow depth in IH2P2 (87 cm) may have further created a limitation for the soils. However, water tolerant crops like paddy rice and sugar cane may not be limited by the high water table. Soils in IH2P1 and IH2P2 therefore met the requirements for classification into land capability subclasses IVwcf and IIIwsf, respectively. The soils were down graded to classes IV and III mainly by wetness limitation, base saturation and gravel content that exceeded 15 % in the surface soils. Such soils with moderate limitations may reduce the choice of crops. According to Sys *et al*. (1991b) the soils require moderate conservation practices as well as careful management. Such conservation practices include; slight drainage to improve air and water relations. The distribution of capability classes of mapping units IH1 and IH2 is shown in Fig. 12, and indicates that the class II soils occupied 252.4 ha (37.2 %), class III occupied 418.9 ha (61.7 %) and class IV soils occupied the least expanse with only 7.8 ha (1.1 %) out of a total of 679.1 ha.

Table 36: Land characteristics for classifying soils into USDA Land Capability Classes

Characteristics	IH1P1	IH1P2	IH2P1	IH2P2	AI1P1	AI1P2	AI2P1	AI2P2	AI3P1	AI3P2	MF1P1	MF1P2	MF2P1	MF2P2	MF3P1	MF3P2
Soil physical conditions. (s)																
Soil depth (cm)	195	150	145	87	200	180	120	132	80	124	121	153	161	200	48	49
Surf. Texture	SCL	SL	SL	SL	LS	LS	LS	SL	LS	SL	LS	SL	SL	LS	SL	SL
Gravels (%)	>30	>30	30	>20	0	0	<10	0	<10	0	>20	0	0	15	0	0
Wetness (w)																
Drainage	Mod.well dr	Mod.well dr	Poorly dr	V.poorly dr.	Well dr.	Mod.well .dr	Mod.well Dr	Well dr	Poorly dr	Poorly dr	Well dr	Well dr	Well dr	Well dr	v.poorly dr	v.poorly dr
Flooding	None	None	Rare-occasional	Frequent	None	None	None	None	occasional	occasional	None	None	none	none	Rare-occasional	Rare-occasional
Fertility (f)																
Reaction(pH)	5.5-6.7	6.4-7.3	5.8-6.7	5.2-6.1	5.3-7.2	6.1-7.7	6.3-6.4	5.7-6.7	5.2-5.3	5.7-7.5	5.6-6.6	5.4-5.8	5.1-5.6	5.3-6.4	6.5-6.7	6.1-6.3
BS %(AandB Horizon)	26.2	27.5	36.7	21.8	19.6	21.2	18.9	56.5	21.4	16.3	33.1	17.8	16.3	8.3	29.9	41.1
Climate (c)																
Eff.Ppt.	1983	1983	1983	1983	2258	2258	2258	2258	2258	2258	2894.1	2894.1	2894.1	2894.1	2894.1	2894.1
Days in MCS	Ustic, <135	Ustic, <135	Not class determining		Udic, <135	Udic, <135	Udic, <135	Udic, <135	Not class determining		Udic, <135	Udic, <135	Udic, <135	Udic, <135	Not class determining	
Erosion (e)																
Slope (%)	6.3	5	4	4	5.3	3.5	5.2	2.0	4.0	4.5	4	4	2.0	5.6	0.88	4
Erosion hazard	Slight sheet erosion	None	None	None	None	Slight sheet erosion	None	Slight sheet erosion	None	None	Slight sheet erosion	Slight sheet erosion	Slight sheet erosion	Slight sheet erosion	Slight sheet erosion	None

Foot note: MCS: moisture control section, SCL: sandy clay loam, SL: sandy loam, LS: loamy sand, dr.: drained, v: very, eff. Ppt.: effective precipitation

Table 37: check list for USDA Land Capability Classification											
Capability	Physical properties (s)			Wetness (w)		Fertility (f)		Climate (c)		Erosion (e)	
Classes	Depth	Texture	Gravels	Drainage	Overflow	Reaction	BS	Eff.Ppt	Days in MCS	Slope	Soil loss to Erosion
Soil unit: IH1P1											
I	*	*			*	*		*	*	*	*
II			*	*							
IV							*				
Classification: IIsf											
Soil unit: IH1P2											
I	*	*			*			*	*	*	*
II			*	*		*					
IV							*				
Classification: IIsf											
Soil unit: IH2P1											
I	*	*				*		*		*	*
II			*		*						
III				*							
IV							*				
V									*		
Classification: IVwcf											
Soil unit: IH2P2											
I		*				*		*	*	*	*
II			*								
III	*			*							
IV					*		*				
Classification: IIIwsf											
Soil unit: AI1P1											
I	*		*	*	*	*		*	*		*
II		*									
III										*	
IV							*				
Classification: IIIesf											
Soil unit: AI1P2											
I	*		*	*	*			*	*	*	*
II		*				*					
IV							*				
Classification: IIsf											
Soil unit: AI2P1											
I	*		*		*	*		*	*	*	*
II		*		*							
IV							*				
Classification: IIsf											
Soil unit: AI2P2											
I	*	*	*	*	*	*		*	*	*	*
III							*				
Classification: IIf											
Soil unit: AI3P1											
I			*					*		*	*
II	*	*				*					
III				*	*						
IV							*				
V									*		
Classification: IVwscf											

Table 37 Contd.

Capability	Physical properties (s)			Wetness (w)		Fertility (f)		Climate (c)		Erosion (e)	
Classes	Depth	Texture	Gravels	Drainage	Overflow	Reaction	BS	Eff.Ppt	Days in MCS	Slope	Soil loss to Erosion
Soil unit: AI3P2											
I	*	*	*			*		*		*	*
II											
III				*	*						
IV							*				
V									*		
Classification: IIIwcf											
Soil unit: MF1P1											
I	*			*	*	*		*	*	*	*
II		*	*								
IV							*				
Classification: IIsf											
Soil unit: MF1P2											
I	*	*	*	*	*	*		*	*	*	*
IV							*				
Classification: IIf											
Soil unit: MF2P1											
I	*	*	*	*	*			*	*	*	*
II						*					
IV							*				
Classification: IIf											
Soil unit: MF2P2											
I	*			*	*	*		*	*	*	*
II		*	*								
IV							*				
Classification: IIsf											
Soil unit: MF3P1											
I		*	*			*		*		*	*
II	*										
III				*	*						
IV							*				
V									*		
Classification: IIIwsf											
Soil unit: MF3P2											
I		*	*			*		*		*	*
II	*										
III				*	*		*				
V									*		
Classification: IIIwsf											

*Individual soil property rating

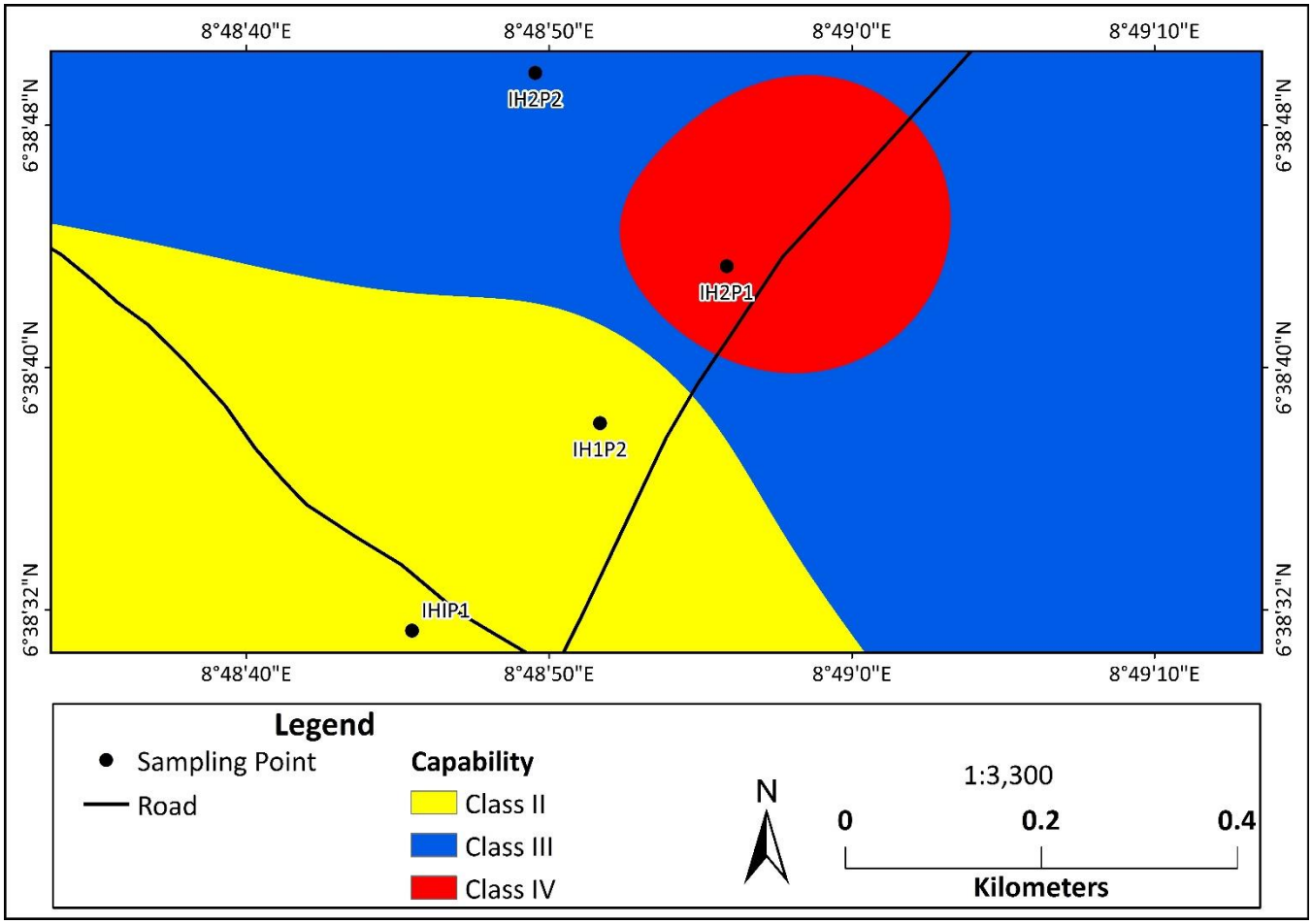

Fig. 12: Distribution of Land capability classes on shale, limestone and sandstone in the Ishibori area

b. In the SL in central agricultural zone, soils in mapping unit AI1 were deep (> 100 cm) with negligible content of gravels in surface soils and favourable wetness and climatic characteristics. AI1P1 and AI1P2 were, however limited by their loamy sand textures and low base saturation, while slope was a limiting property in AI1P1 just as soil pH (H_2O) was high in AI1P2 with values within 6.1-7.7. Soil pH values above 6.5 may be regarded as high for most crops (Sys *et al*., 1991a). AI1P1 and AI1P2 met the requirements for placement into land capability subclass IIIesf and IIsf, respectively (Fig. 13). The class II soil (AI1P2) may require careful to very careful soil management strategies. AI1P1 (class III) may therefore require terracing and surface mulching. These procedures will reduce the speed of running water down slope as well as build up soil organic matter content. It may also require the application of mineral fertilizers so as to boast the saturation of exchangeable bases in the soil exchange complex.

Soils in mapping unit AI2 were deep (> 100 cm) with gravel content less than 10 % as well as favourable wetness condition, climate, erosion and fertility characteristics except base saturation that had values less than 50 % especially in AI2P1. However, AI2P1 was limited by its loamy sand textural class, hence its qualification for placement into land capability subclass IIsf. AI2P1 (IIsf) will therefore require careful management *via* moderate conservation practices like crop rotations with concurrent application of animal manure for increased soil organic matter content. On the other hand, base saturation may be increased by the application of organic fertilizers and lime.

In mapping unit AI3, gravel was less than 10 % with favourable effective precipitation. The soils were limited by base saturation and wetness condition, while AI3P1 was in addition, limited by shallow depth, loamy sand textural class and soil reaction (5.2-5.3). According to the scale of Holland *et al*. (1989), the values of soil pH (H_2O) fall within the range of strongly acid reaction, and may bring about significant amount of exchangeable Al^{3+} on the soil exchange complex (Udo *et al*., 2009). At such pH values, alumina is often soluble (Tan, 1998). AI3P1 and AI3P2 were therefore classified as class IVwscf and IIIwcf, respectively. AI3P1 (class IVwscf) closely suits the requirements of Sys *et al*. (1991b) for class IV soils.

The soil will therefore require very careful management due mainly to its severe limitations of wetness, soil physical properties and base saturation. Special conservation procedures like drainage and flood protection are recommended. The use of soil organic amendments like animal manure as well as the careful use of lime will increase the base

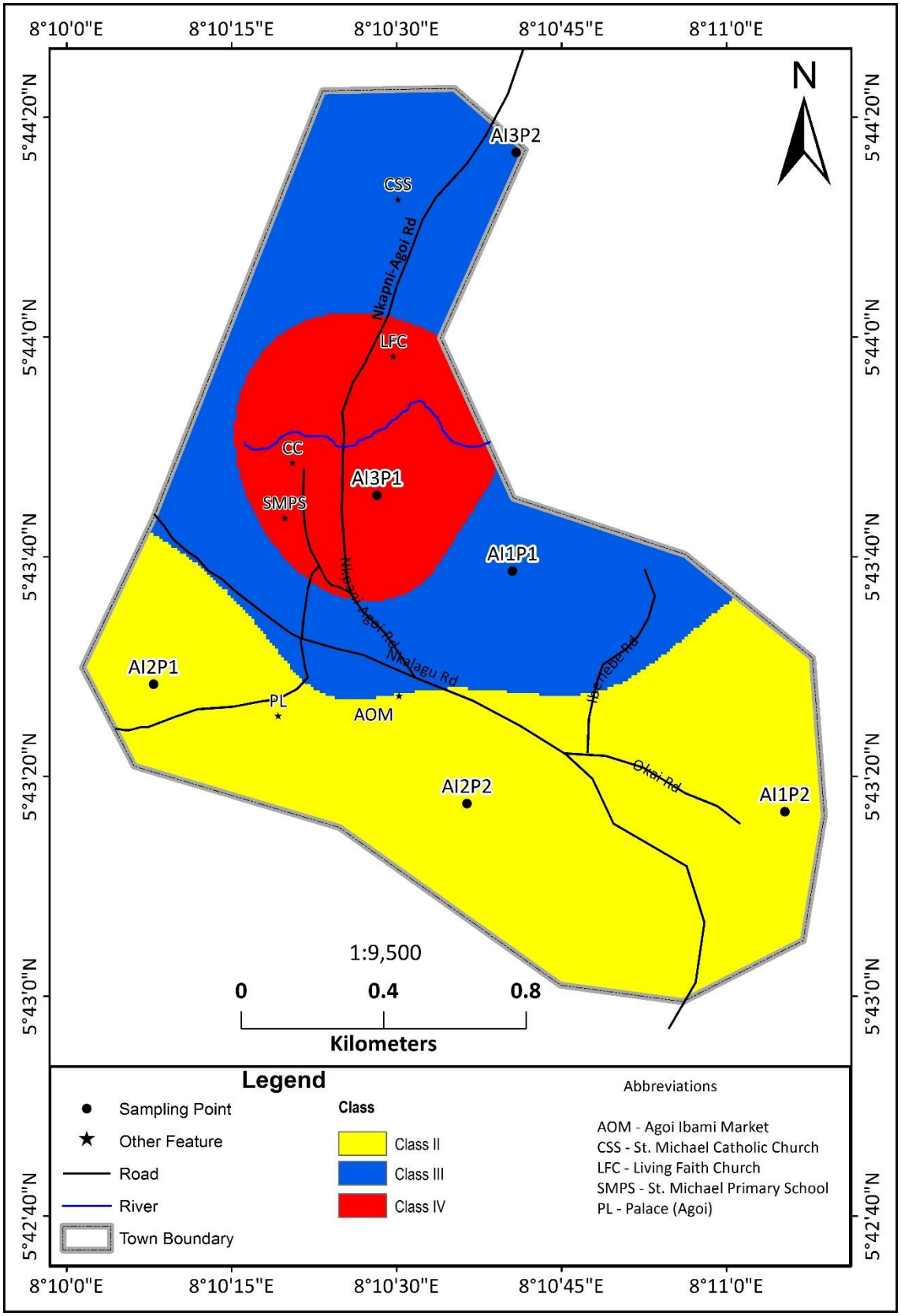

Fig. 13: Distribution of Land capability classes on sandstone and limestone formation in Agoi Ibami

saturation per cent. On the other hand, AI3P2 (class IIIwcf) will require similar treatment however, the use of lime may not be necessary as soil pH was 5.7-7.5.

This range of values is within acceptable limits for arable soils in the tropics (Holland *et al*., 1989). Adequate dosages of calcic and potassic fertilizers can be used to step-up the base saturation of the soils. By this, the soils may be upgraded to class II.

The distribution of capability classes for soils in the central agricultural zone is shown in Fig. 16, and indicates that 137.2 ha (49.0 %) of the land qualified as class II, 98.7 ha (35.3 %) qualified as class III and only 44.0 ha (15.7 %) was classified as class IV land out of an overall hectarage of 279.79 ha. Classes II and III soils are suited for regular cultivation (Esu, 2010) and may require moderately intensive (class II) to intensive (class III) treatment to be safely cultivated. On the other hand, the class IV soils will be best suited for pasture. However, occasional protective cultivation to safeguard it against further deterioration is advocated.

c. In the soils overlying SLSi in the southern agricultural zone, MF1 had favourable wetness conditions, soil reaction (5.4-6.6), as well as erosion and climatic properties (effective precipitation and temperature). Slope per cent was 4 % for soils in MF1 and so offered little or no limitation to agriculture; however, low base saturation was limiting as 12.7-48.8 % was obtained in MF1. Furthermore, loamy sand textural class and greater than 20 % gravels constituted the primary source of limitation for MF1P1. Sys *et al*. (1991b) recommended <15 % of gravel per cent for arable lands (Classes I to IV). MF1P1, therefore, qualified as IIsf, while MF1P2 qualified as IIf in the land capability subclass (Fig. 14). Soils in mapping unit MF1 were good for farming and had very few limitations such as texture, gravels and base saturation. These limitations had down-graded the soils from class I to II. For safe cultivation, the soils need careful management or moderately intensive treatments such as cover cropping and fertilizer application to compensate crop output. The limitations are most likely to reduce the range of crops to be cultivated.

Soils in mapping unit MF2 were deep (> 100 cm) and optimal in terms of climate (precipitation and temperature) as well as slope and wetness condition but sub-optimal in its base saturation which constituted a basic soil fertility limitation. MF2P2 was, however, limited by loamy sand textural class and 15 % gravels, while MF2P1 was limited by low soil pH, which had values that ranged from 5.1 to 5.6. MF2P1 and MF2P2 were classified in the land capability subclass IIf and IIsf, respectively (Fig. 14).

The limitations which have reduced the capability of MF2 from class I to II were moderate, and the soil required moderate conservation practices with careful soil

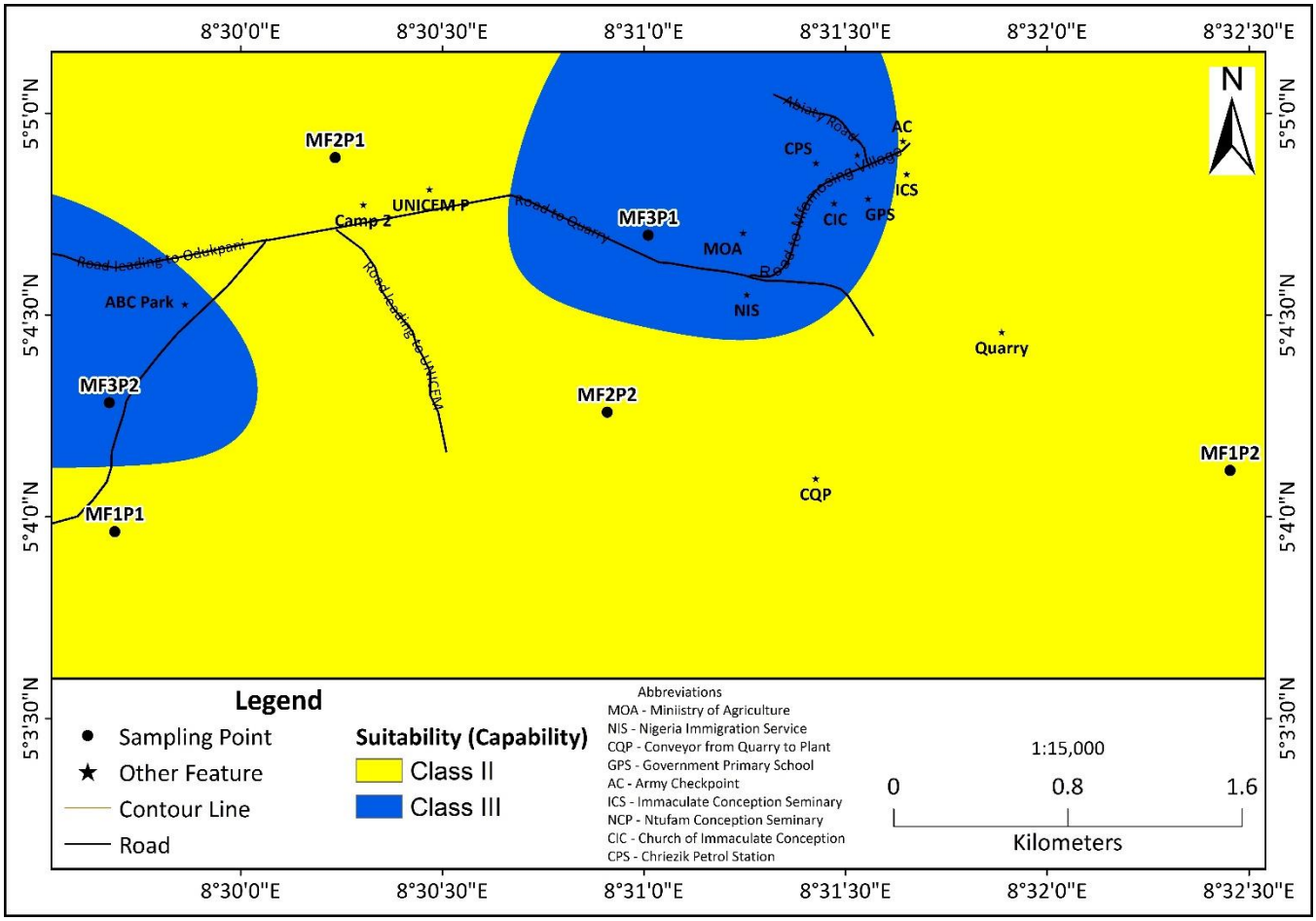

Fig. 14: Distribution of Land capability classes on Shale and limestone with sandstone intercalation in Mfamosing

management procedures. Strip cropping, together with careful application of prescribed doses of mineral fertilizers and lime, will go a long way to address the limitations. This may raise the base saturation and pH to a more favourable level. The soils were suitable for regular cultivation of arable crops.

In mapping unit MF3, soil textural class (sandy loam to loamy sand), gravel percent (negligible in the surface soils), soil reaction (6.1-6.7), precipitation and slope per cent were optimal (Sys *et al*. 1991b and Esu, 2010), while base saturation values (12.1-49.4 %) were sub-optimal. Wetness condition as well as soil depth were however, the main limiting soil properties that qualified the soils for placement into land capability subclass IIIwsf. The limitations were severe and may reduce the choice of crops (Sys *et al*. 1991b). Special conservation practices, as well as very careful management requirements, may increase the land use options. Properties such as frequent overflow accompanied by some crop damage, waterlogging, shallow depth and low fertility outlined by Sys *et al*. (1991b) for class III soils fits the properties of MF3 most appropriately. Surface drainage will most likely improve air and water relations and increase the range of crops that can be cultivated.

The distribution of capability classes is shown in Fig. 14, indicating 1581.8 ha as class II, which represented 71.8 % of the entire area, while 619.85 ha (28.2 %) met the requirements of class III out of the entire area, which measured 2201.65 ha.

4.10.2 Land Suitability Evaluation for Oil palm Cultivation

Land characteristics of the soils are presented in Table 38. These characteristics, in comparison with values in Table 5 gave rise to the suitability classes and scores on Table 39.

(i) Land Requirements for Oil palm

a. Climate: Climate quality was evaluated for temperature and rainfall. In Cross River State, these characteristics were optimum for the cultivation of oil palm (Table 39), with suitability scores ranging from S_1(94) to S_1(100) for mean annual temperature (MAT) and mean annual rainfall (MAR). Climate did not constitute a limitation for the cultivation of oil palm. Near optimum climate was reported for oil palm cultivation in NIFOR by Ogunkunle (1993). According to the study, much of Nigeria is climatically not suitable for the cultivation of oil palm. However, southern Nigeria has been found to have a comparative advantage over other zones in Nigeria for the cultivation of oil palm, probably because of its favourable climate. However, the climate is not a sole determinant of land quality or suitability for oil palm cultivation.

Table 38: Land characteristics data for the soils

Pedons	*MAR mm/yr	*MAT ^{o}C	Slope %	Flooding	Drainage	Texture	CF % Vol.	Depth Cm	CEC cmol/kg	BS %	Mg:K	pH	OC %	ESP %
IH1P1	1983	27.9	6.3	Fo	MWD	SCL	53	195	30.6	25.3	45.6	6.0	0.50	0.09
IH1P2	1983	27.9	5.0	Fo	WD	SL	40.3	150	32.6	26.0	39.3	6.8	1.60	0.17
IH2P1	1983	27.9	4.0	F3	PA	SCL	16.7	145	26.8	34.7	37.9	6.1	1.27	0.16
IH2P2	1983	27.9	4.0	F3	PD	SL	19.7	87	36.5	35.0	25.7	5.6	5.57	0.26
AI1P1	2258	27.0	5.3	Fo	WD	SL	0.7	200	16.7	20.7	13.8	6.3	0.50	0.22
AI1P2	2258	27.0	3.5	Fo	WD	SL	8.0	180	16.7	26.7	17.7	6.4	0.55	0.23
AI2P1	2258	27.0	5.2	Fo	WD	SL	8.0	120	26.3	18.9	7.3	6.4	0.54	0.09
AI2P2	2258	27.0	2.0	Fo	WD	SL	32.3	132	7.7	55.3	21.8	6.2	0.42	0.44
AI3P1	2258	27.0	4.0	F3	PA	LS	11.5	80	8.4	21.4	9.5	5.3	1.25	0.28
AI3P2	2258	27.0	4.5	F3	PD	LS	13.7	124	20.5	15.6	23.3	6.6	0.26	0.12
MF1P1	2894.1	27.2	4.0	Fo	WD	SL	11.3	121	20.9	30.1	17.4	6.1	0.55	0.12
MF1P2	2894.1	27.2	4.0	Fo	WD	SL	13.8	153	11.9	16.5	12.2	5.6	1.00	0.37
MF2P1	2894.1	27.2	2.0	Fo	WD	SL	24.8	161	44.5	20.9	42.6	5.4	1.50	0.12
MF2P2	2894.1	27.2	5.6	Fo	WD	LS	5.0	200	28.5	8.3	9.5	5.7	1.20	0.20
MF3P1	2894.1	27.2	0.88	F3	PA	SL	0	48	40	21	27.4	6.6	0.88	0.08
MF3P2	2894.1	27.2	4.0	F3	PA	LS	14.3	49	16.5	34.1	4.4	6.2	1.71	0.21

*Source: Sambo *et al*. (2016a)

MAR = mean annual rainfall, MAT = mean annual temperature, CF = coarse fragment, ACEC = apparent CEC, SEB = sum of exchangeable bases, ESP = exchangeable sodium percent

Table 39: Suitability classification and scores of the pedons for oil palm cultivation

Pedons	MAR mm/yr	MAT °C	Slope %	Flooding	Drainage	Texture	CF % Vol.	Depth Cm	CEC cmol/kg	*BS %	*pH	*OC %	Mg:K	ESP %
IH1P1	S_1(94)	S_1(100)	S_1(91)	S_1(100)	S_1(97)	S_1(98)	S_2(76)	S_1(100)	S_1(100)	S_1(88)	S_2(82)	S_2(72)	S_1(100)	S_1(100)
IH1P2	S_1(94)	S_1(100)	S_1(92)	S_1(100)	S_1(97)	S_I(96)	S_2(80)	S_1(100)	S_1(100)	S_1(89)	S_1(93)	S_1(100)	S_1(100)	S_1(100)
IH2P1	S_1(94)	S_1(100)	S_1(95)	N_1(35)	S_3(50)	S_1(98)	S_1(92)	S_1(90)	S_1(100)	S_1(98)	S_2(84)	S_1(90)	S_1(100)	S_1(100)
IH2P2	S_1(94)	S_1(100)	S_1(95)	N_1(35)	N_1(38)	S_I(96)	S_1(89)	S_3(48)	S_1(100)	S_1(100)	S_3(58)	S_1(100)	S_1(100)	S_1(100)
AI1P1	S_1(100)	S_1(100)	S_1(94)	S_1(100)	S_1(100)	S_I(96)	S_1(100)	S_1(100)	S_1(100)	S_1(85)	S_1(87)	S_2(72)	S_1(100)	S_1(100)
AI1P2	S_1(100)	S_1(100)	S_1(97)	S_1(100)	S_1(100)	S_I(96)	S_1(98)	S_1(100)	S_1(100)	S_1(89)	S_1(90)	S_2(75)	S_1(100)	S_1(100)
AI2P1	S_1(100)	S_1(100)	S_1(94)	S_1(100)	S_1(100)	S_I(96)	S_1(98)	S_1(88)	S_1(100)	S_2(80)	S_1(90)	S_2(75)	S_1(100)	S_1(100)
AI2P2	S_1(100)	S_1(100)	S_1(100)	S_1(100)	S_1(100)	S_I(96)	S_1(90)	S_1(90)	S_2(80)	S_1(100)	S_1(86)	S_2(70)	S_1(100)	S_1(100)
AI3P1	S_1(100)	S_1(100)	S_1(95)	N_1(37)	S_3(50)	S_I(90)	S_1(97)	S_3(40)	S_1(88)	S_2(83)	N_1(35)	S_1(89)	S_1(100)	S_1(100)
AI3P2	S_1(100)	S_1(100)	S_1(94)	N_1(37)	N_1(38)	S_I(90)	S_1(95)	S_1(87)	S_1(100)	S_2(70)	S_1(90)	N_1(30)	S_1(100)	S_1(100)
MF1P1	S_1(100)	S_1(100)	S_1(95)	S_1(100)	S_1(100)	S_I(96)	S_1(97)	S_1(89)	S_1(100)	S_1(90)	S_2(84)	S_2(75)	S_1(100)	S_1(100)
MF1P2	S_1(100)	S_1(100)	S_1(95)	S_1(100)	S_1(100)	S_I(96)	S_1(96)	S_1(100)	S_1(100)	S_2(70)	S_3(58)	S_1(90)	S_1(100)	S_1(100)
MF2P1	S_1(100)	S_1(100)	S_1(100)	S_1(100)	S_1(100)	S_I(96)	S_1(92)	S_1(100)	S_1(100)	S_1(85)	S_3(56)	S_1(100)	S_1(100)	S_1(100)
MF2P2	S_1(100)	S_1(100)	S_1(93)	S_1(100)	S_1(100)	S_I(90)	S_1(99)	S_1(100)	S_1(100)	S_3(40)	S_3(60)	S_1(93)	S_1(100)	S_1(100)
MF3P1	S_1(100)	S_1(100)	S_1(100)	N_1(36)	S_3(50)	S_I(96)	S_1(100)	N_I(30)	S_1(100)	S_1(86)	S_1(90)	S_1(85)	S_1(100)	S_1(100)
MF3P2	S_1(100)	S_1(100)	S_1(95)	N_1(36)	S_3(50)	S_I(90)	S_1(96)	N_1(31)	S_1(100)	S_1(93)	S_1(86)	S_1(100)	S_1(100)	S_1(100)

N/B: MAR= Mean annual rainfall, MAT= Mean annual temperature, CF= Coarse fragment, ACEC= Apparent CEC, SEB= Sum of exchangeable bases, OC= Organic carbon, ESP= Exchangeable sodium percent, MWD= Moderately well drained, WD= Well drained, PA= Poor aeric, PD= Poor drainable, Fo= No flooding, F3= Severe; Every year, 2-3 mths of flood.

b. Topography: The topography of the areas overlying limestone formations in Cross River State was highly suitable for oil palm cultivation with a score of S_1(100) in all the locations.

c. Wetness: Wetness land quality was represented by drainage and flooding. Soils in higher elevation ranges were well-drained and free of flooding. However, the lowest elevation ranges (IH2, AI3, MF3) were most affected by flooding and drainage. Suitability class scores for these low elevation ranges were N_1(35 to 37) for flooding characteristics and S_3(50) – N_1(38) for drainage characteristic. These characteristics cause severe limitations to oil palm cultivation in the low elevation ranges.

d. Soil physical characteristics: Soil physical quality was represented by soil texture, the coarse fraction (CF) and soil depth. Soil texture was optimum and did not constitute a major limitation in the three limestone formations for the cultivation of oil palm with a suitability class score of S_1(90 to 98). Such optimum soil textures (Tables 6 to 13) and scores (Table 39) may reduce drainage and increase nutrients and moisture retention. However, a slight limitation was observed for CF in mapping unit IH1. This further down-graded the suitability class score to S_2(76-80), resulting in slight oil palm cultivation limitations. Furthermore, other mapping units were optimum with a suitability class score of S_1(89-100) for the cultivation of oil palm and no significant limitation. Finally, soil depth was optimum for the well-drained soils in high elevations (irrespective of the limestone formations) with suitability class scores of S_1(88-100) and without a limitation. However, soils in the low elevation ranges were limited by the high water table resulting in suitability class scores of S_3(40-48) and N_1(30-31) with severe limitations.

e. Potential soil fertility: Included in the potential soil fertility were chemical properties that were not easily altered and exempt base saturation, soil pH and organic carbon, which are easily altered. The entire soils were optimum in terms of CEC with a suitability class of S_1(88-100) except AI2P2, which witnessed slight limitation and was rated S_2(80). Similarly, the Mg: K ratio and exchangeable sodium percent were optimum in the entire soils with a suitability class score of S_1(100) (Table 39).

f. Current soil fertility: Current soil fertility refers to chemical fertility when properties that are easily altered are taken into consideration alongside the requirements for potential fertility. The properties include base saturation, soil pH and organic carbon. Base saturation was optimum in the entire soils with a suitability class score of S_1(85-100) except for soils in AI2P2, AI3P1, AI3P2 and MF1P2 (S_2;70-83) and MF2P2 (S_3; 40) which had slight and severe limitations. Soil pH was optimum in IH1P2, AI1P1, AI1P2, AI2P1, AI2P2,

AI3P2, MF3P1 and MF3P2 with suitability class scores of S_1(86-93) and without obvious limitations to oil palm cultivation. Similarly, IH1P1, IH2P2 and MF1P1 were moderately suitable (S_2) and rated from 82 to 84, with slight limitation caused by soil pH, while other soils were moderately or severely limiting in pH and rated either as S_3(56-60) or N_1(35). Organic carbon was adequate for most of the soils with suitability class scores of S_1(85-100). The slight limitation was observed for IH1P1, AI1, AI2 and MF1P1, and a moderately suitable class (S_2) obtained. However, AI3P2 was not suitable (N_1; 30) in terms of organic carbon for oil palm cultivation. By the requirements of Sys *et al.* (1991b), organic carbon was slight to severely limiting for oil palm cultivation, especially in the soils formed over SL.

(ii) Major Limitations to Suitability for Oil palm

It is clear, therefore, that in the soils found over limestone formations in Cross River State, climate (temperature and rainfall), slope as well as soil characteristics such as Mg: K ratio, exchangeable sodium percent, texture, CF and CEC were optimum or nearly so for oil palm cultivation. The loss of K may cause severe loss in oil palm yield (Ogunkunle, 1993). The ratio of Mg: K in the soils was optimum, hence the moderately (S_2) to highly (S_1) suitable soils for oil palm cultivation except in the soils affected by wetness (IH2, AI3, MF3) or soil pH and base saturation. Sandy loam to sandy clay loam textures was obtained. Sandy clay loam texture has been recommended for optimum oil palm cultivation (Sys, 1985). Such textural class encourages soil water holding capacity, particularly in the upper 30 cm, where more oil palm roots are concentrated (Omoti and Ataga, 1982). Sandy and gravelly soils, when combined with poor fertility, cannot be suitable for oil palm cultivation (Ogunkunle, 1993). High rainfall (> 2000 mm/year), as well as sandy loam to loamy sand textures, may have been responsible for the sub-optimum base saturation in some of the soils.

Also, soil depth and wetness properties (drainage and flooding) were optimum in the well-drained high elevation ranges of the soils overlying the limestone formations and grossly limiting in the low elevation ranges of the study areas; hence the sub-optimal drainage and flooding condition for oil palm cultivation. With unfavorable wetness condition and shallow depth, a soil can be barely suitable for oil palm cultivation. This was the case in IH2P2, AI3 and MF3. Base saturation, organic carbon and CEC were sub-optimum for oil palm cultivation in most of the soils and optimum in a few of them. When base saturation and soil pH become marginal or sub-optimum, the soil may not be economically favourable for oil palm cultivation.

(iii) Suitability Classes

Suitability classification and scores of individual pedons are presented in Table 39. The aggregate rating was done for current and potential suitability using the parametric and non-parametric methods and summarized in Table 40.

Table 39 shows that all the pedons in the well-drained high elevation ranges (IH1, AI1, AI2, MF1, MF2) were suitable for oil palm cultivation, but to varying degrees. On the other hand, pedons in the low elevation ranges (IH2, AI3 and MF3) were either marginal or not suitable for oil palm cultivation.

a. Parametric method

With the parametric method, nine (9) pedons (IH1, AI1, AI2, MF1, MF2P1) were moderately suitable (S_2) and only one (MF2P2) was marginally suitable (S_3) for soils in the high elevation ranges, currently. Furthermore, five pedons (IH2P2, AI3 and MF3) in the poorly drained low elevation ranges were not suitable (N) and only one (IH2P1) was marginally suitable (S_3) for oil palm cultivation. Potentially, the pedons were more suitable with seven pedons (AI1P1, AI2, MF1 and MF2) being highly suitable (S1) and three pedons (IH1 and AI1P2) being moderately suitable (S_2) in the well-drained high elevation ranges. In the poorly drained soils in the low elevation ranges, most of the pedons were not suitable (N), while only IH2P1 and AI3P2 were rated as marginally suitable (S_3).

The distribution of the entire mapping units as well as their suitability classes (currently and potentially) by the parametric approach are as presented in Figs. 15 to 20. For soils overlying shale, limestone and sandstone in the Ishibori area, potential land suitability indicates that, 285.19 ha (42.0 %) met the requirements for placement into land suitability class S_2 (moderately suitable), while 375.15 ha (55.2 %) was classified as S_3 (marginally suitable) and only 18.76 ha (2.8 %) was N (not suitable) for oil palm cultivation out of a total land area of 679.1 ha.

In the soils overlying SL in the Agoi Ibami area, potential land suitability indicates that 98.92 ha (35.4 %) of the entire area was classified as highly suitable (S1) for oil palm cultivation, while 88.36 ha (31.6 %) was moderately suitable (S2) and 74.37 ha (26.6 %) was marginally suitable (S3). Furthermore, only 18.09 ha representing 6.4 % of the entire 279.74 ha was not suitable (N) for oil palm cultivation. Currently, 185.53 ha (66.3 %) was classified as moderately suitable (S2) for oil palm cultivation, while 94.25 ha (33.7 %) was not suitable for oil palm cultivation out of the total land area of 279.78 ha. This suggests that soil management procedures earlier suggested when applied, will upgrade a great proportion of

Table 40:Aggregate suitability and classification of the pedons for oil palm cultivation

	IH1P1	IH1P2	IH2P1	IH2P2	AI1P1	AI1P2	AI2P1	AI2P2	AI3P1	AI3P2	MF1P1	MF1P2	MF2P1	MF2P2	MF3P1	MF3P2
Potential																
Parametric	S_2(69.2)	S_2(73.3)	S_3(31.4)	N(22.9)	S_1(92)	S_2(72.4)	S_1(85.3)	S_1(75.9)	N(21.4)	S_3(33.5)	S_1(86.7)	S_1(93.1)	S_1(92.0)	S_1(86.8)	N(18.0)	N(18.1)
Non-parametric	S_2s	S_2s	S_3w	S_3ws	S_1	S_1	S_1	S_1	S_3ws	S_3w	S_1	S_1	S_1	S_1	S_3ws	S_3ws
Current																
Parametric	S_2(57.2)	S_2(69.1)	S_3(28.8)	N(17.5)	S_2(68.4)	S_2(72.4)	S_2(68.2)	S_2(66.4)	N(13.1)	N(16.5)	S_2(69.0)	S_2(55.4)	S_2(53.7)	S_3(36.6)	N(16.6)	N(16.8)
Non-parametric	S_2sf	S_2s	S_3wf	N_1wsf	S_2f	S_2f	S_2f	S_2f	N_1wsf	N_1wf	S_2f	S_2f	S_2f	S_3f	N_1ws	N_1ws

Definition of suitability classes: S1(100-75), S2(74-50), S3(49-25), N(24-0); Source: Ogunkunle (1993)

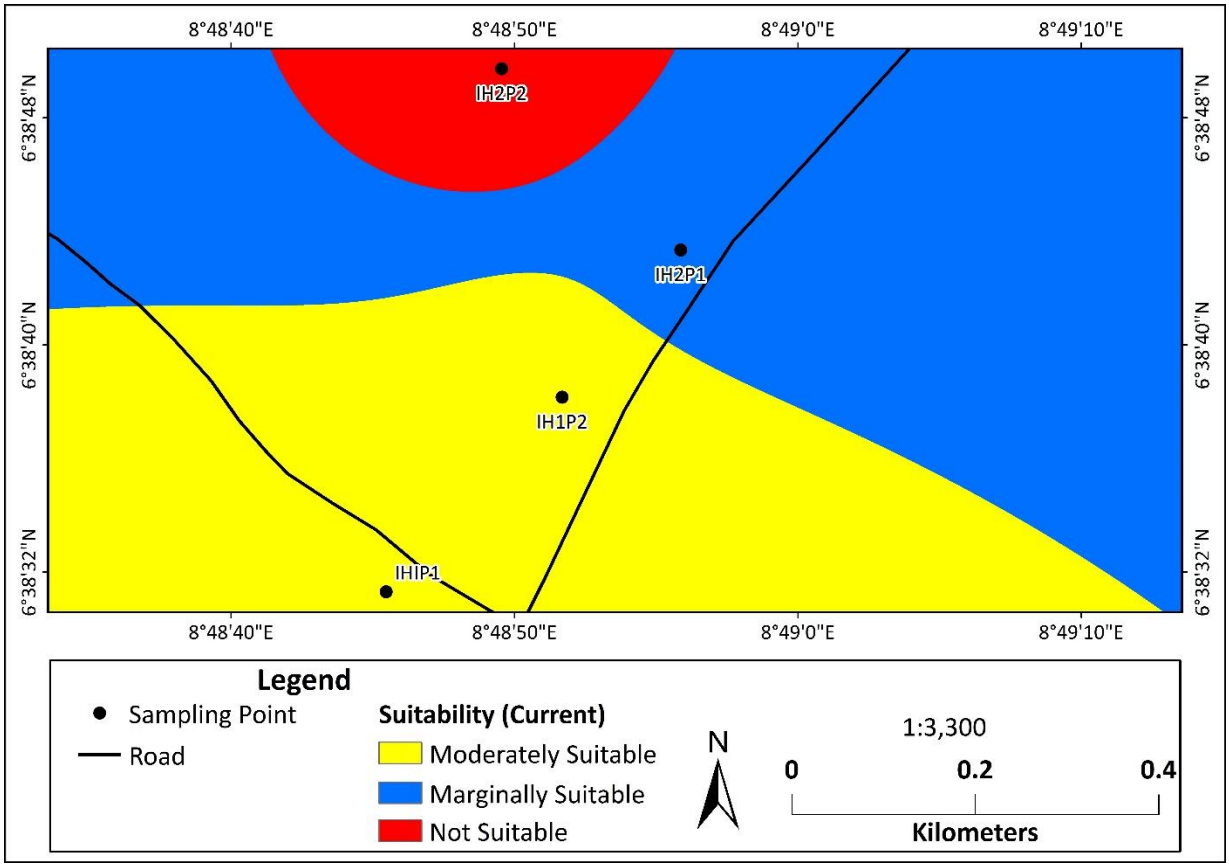

Fig. 15: Distribution of land suitability classes (current) of soils on shale, limestone and sandstone in Ishibori

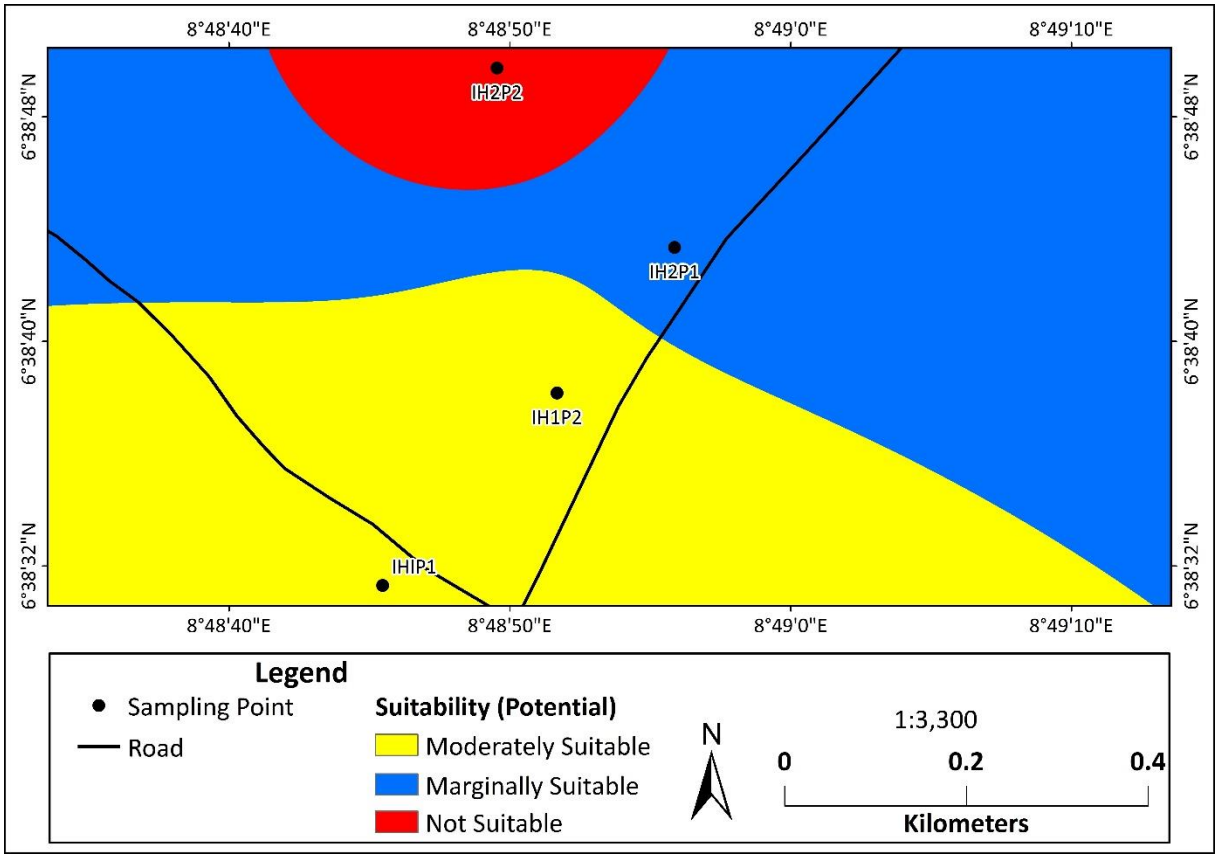

Fig. 16: Distribution of land suitability classes (potential) of soils on shale, limestone and sandstone in Ishibori

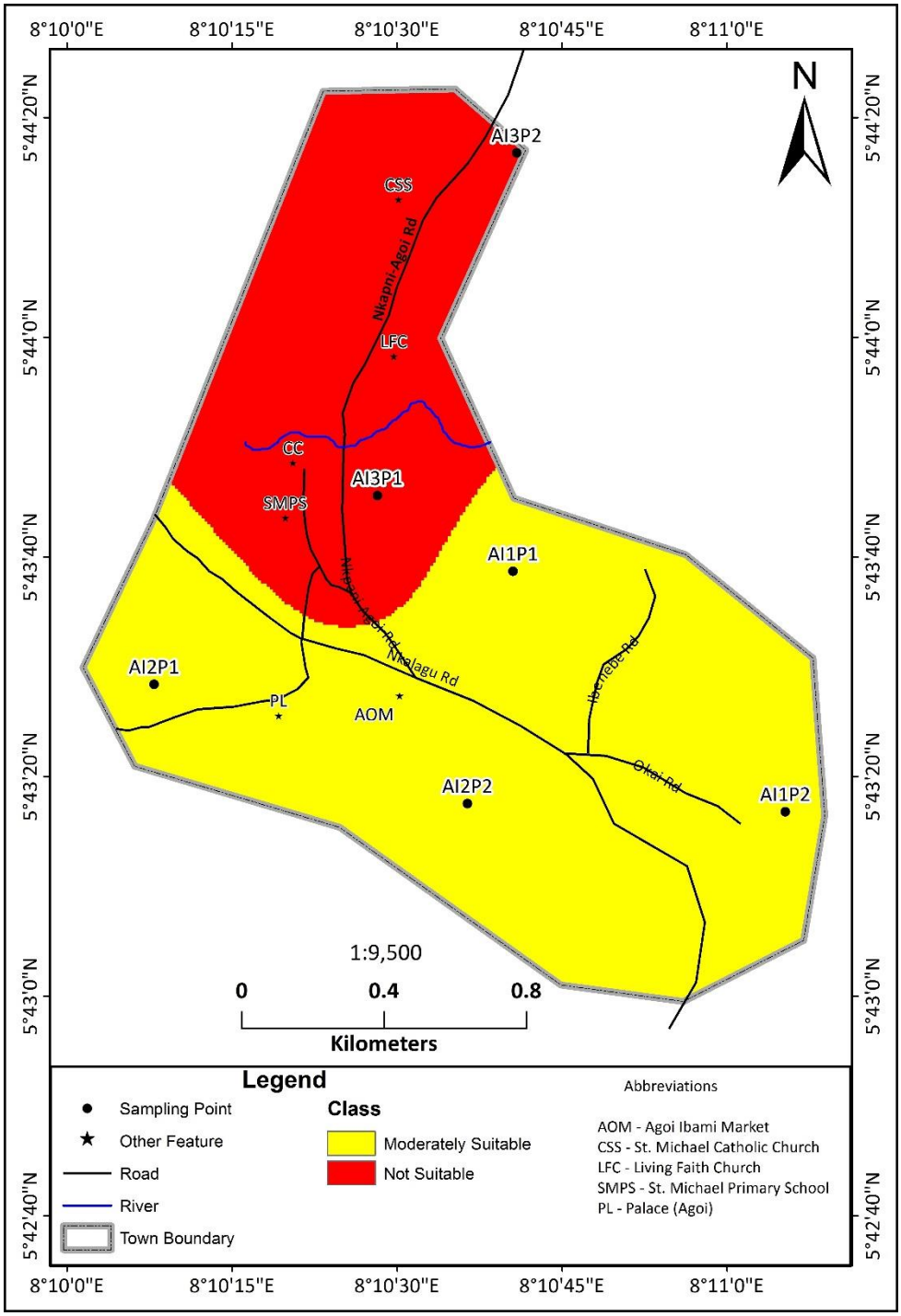

Fig. 17: Distribution of land suitability classes (current) on sandstone and limestone formation in Agoi Ibami

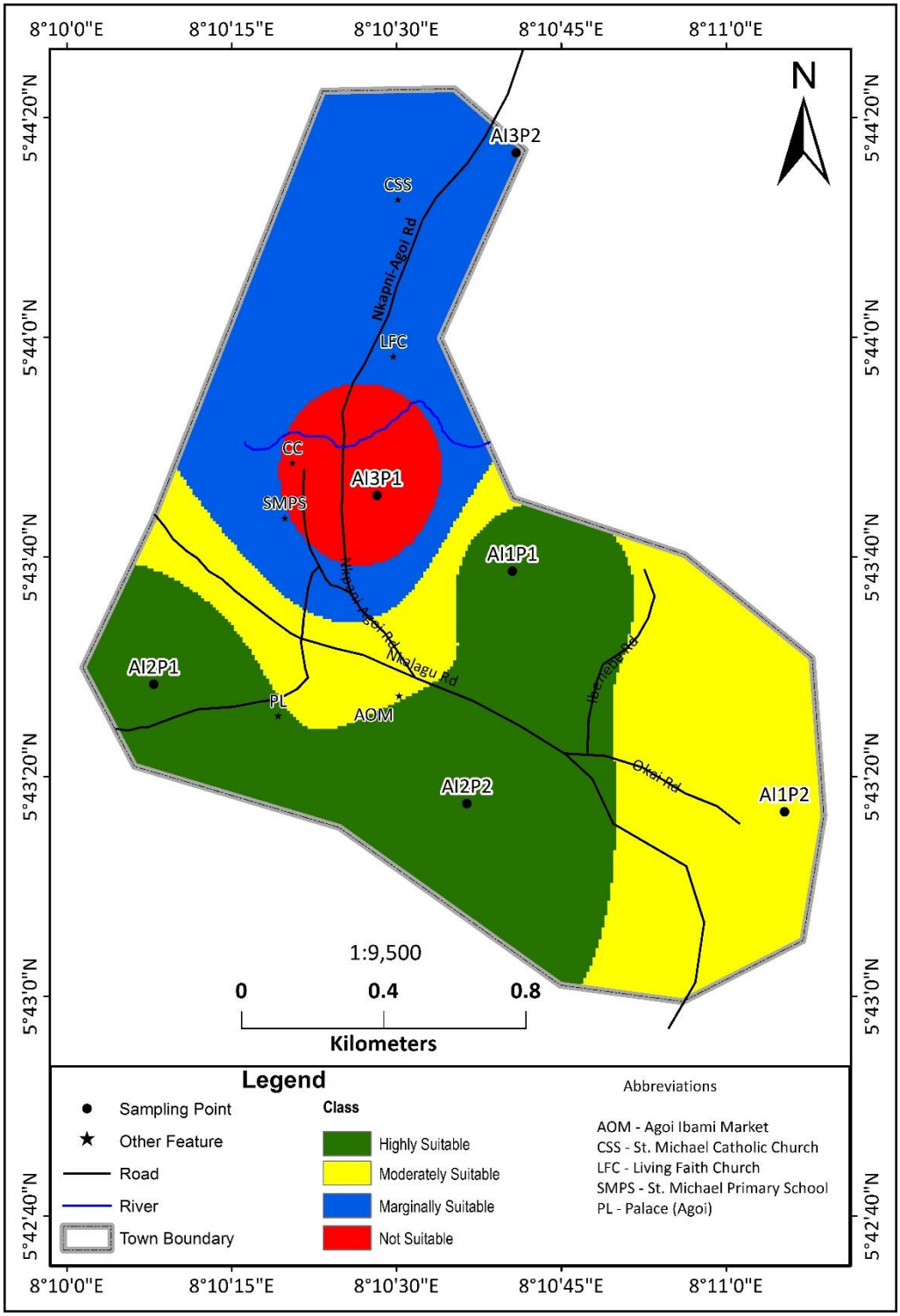

Fig. 18: Distribution of land suitability classes (potential) on sandstone and limestone formation in Agoi Ibami

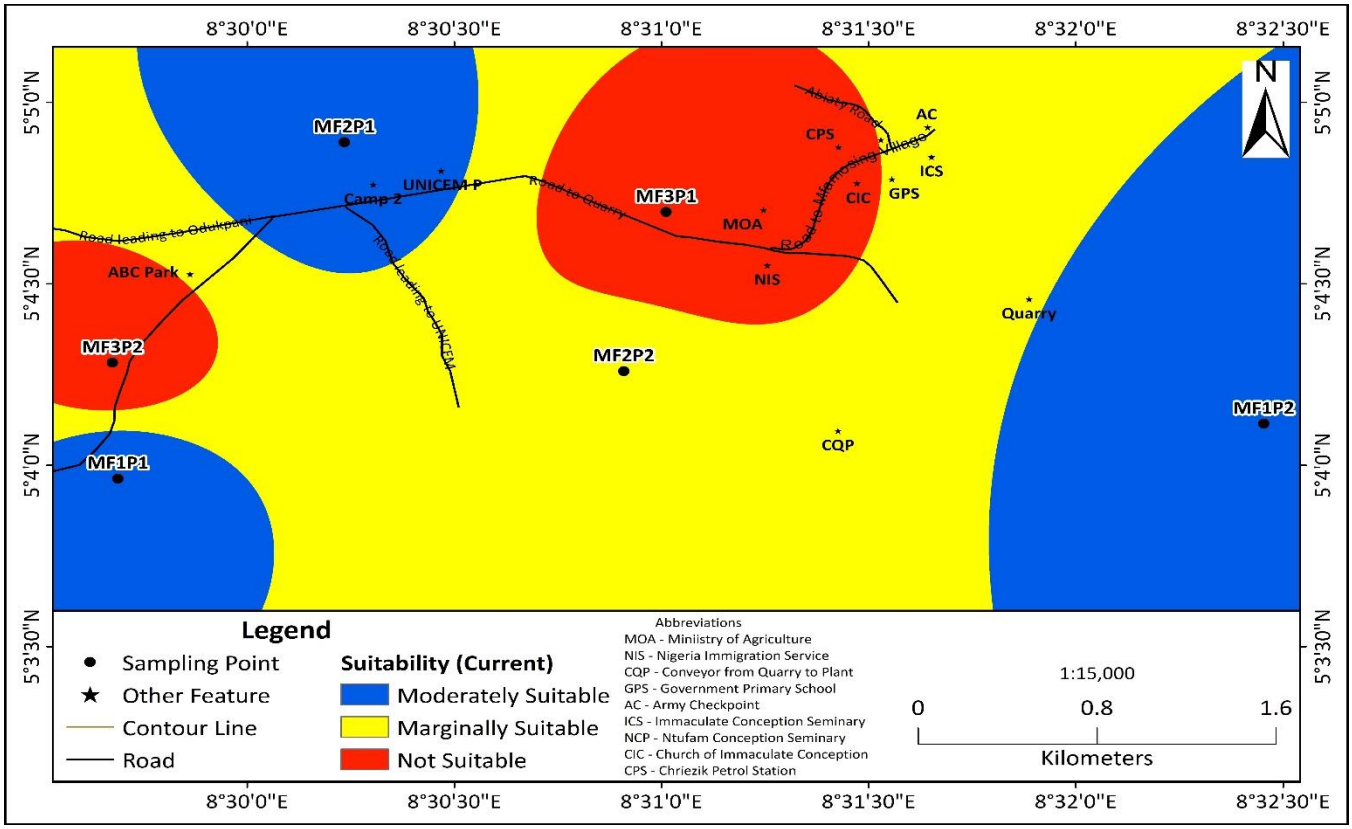

Fig. 19: Distribution of land suitability classes (current) on Shale and limestone with sandstone intercalation in Mfamosing

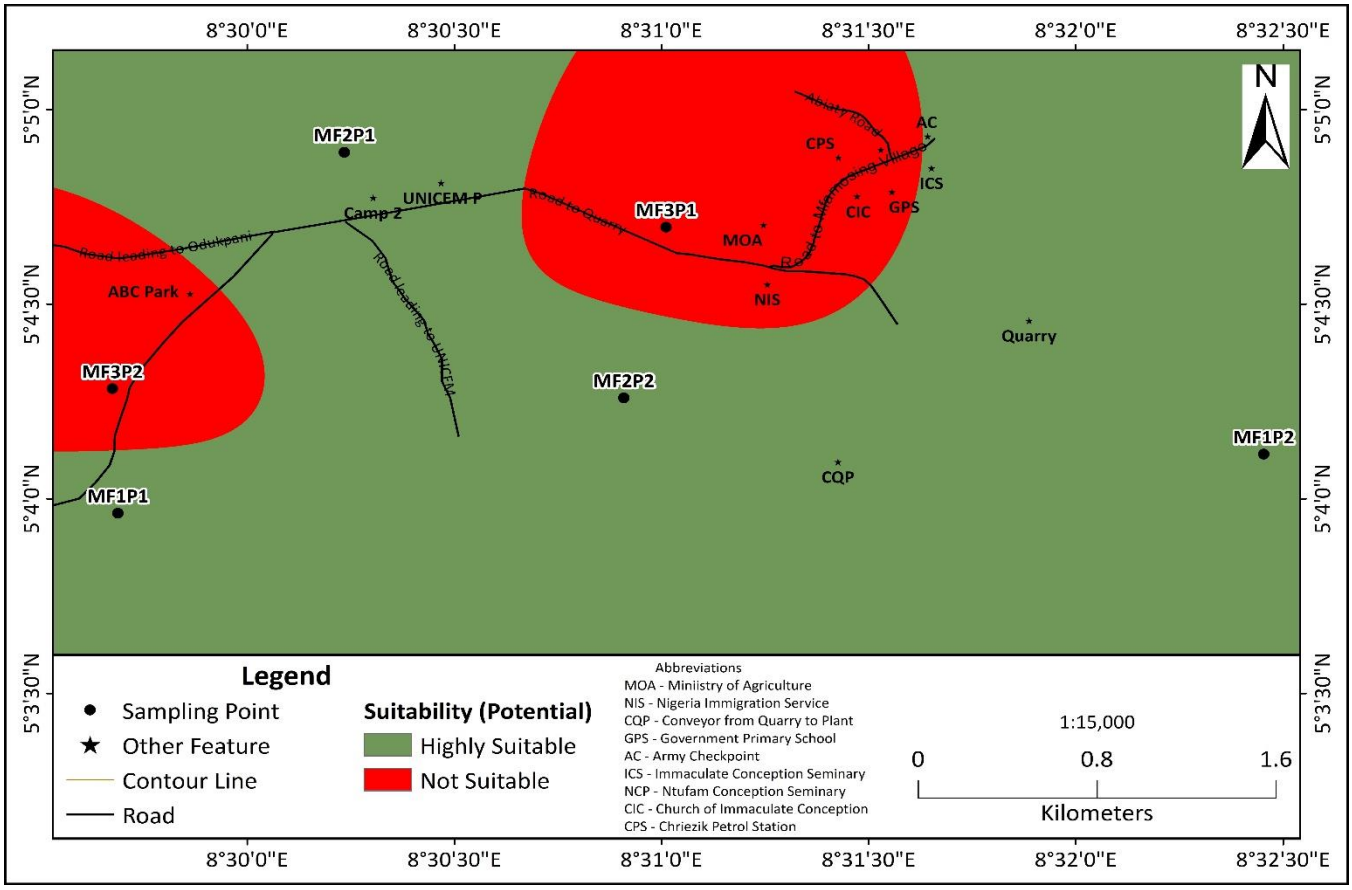

Fig. 20: Distribution of land suitability classes (potential) on Shale and limestone with sandstone intercalation in Mfamosing

the area. This is compared to the area occupied by not suitable class (6.4 %) in the potential suitability.

Soils formed over SLSi in Mfamosing occupied a total area of 2201.65 ha; 1806.68 ha (S_1; 82.1 %) of this area was highly suitable for the cultivation of oil palm, while only 394.97 ha (S_3; 17.9 %) was marginally suitable, potentially. This area of coverage was obtained after the earlier suggested land management procedures may have been applied. Currently, 826.34 ha (S2; 37.5 %) of the total area (2201.65 ha) was moderately suitable (S2), while 1129.96 ha (51.3 %) was marginally suitable (S3) and 245.38 ha (11.2 %) was not suitable for the cultivation of oil palm. Again, it will be noticed that a good percentage of soils that were currently rated marginally suitable (S3) were later upgraded, potentially to highly suitable (S1) mainly because the limiting properties were easily removed fertility limitations.

According to Ogunkunle (1993), the efficiency of each method (parametric and non-parametric) of land evaluation depends on the relevance of the most limiting characteristic to oil palm cultivation. Consequently, where the climate, soil physical condition or CEC is identified as the most limiting property, the non-parametric system may be most accurate (Ogunkunle, 1993). In this way, a limiting soil characteristic determines crop performance and reduces the effect of other soil properties. If soil depth or wetness is unsuitable as in the case of IH2, AI3 and MF3, then the land is unsuitable irrespective of the suitability status of soil chemical or other land qualities. If easily amended chemical properties are rather limiting, the parametric method may be a better approach (Ogunkunle, 1993). On this premise, the non-parametric method may be preferred in the low elevation soils of IH2, AI3 and MF3, while the parametric method is adopted for the well-drained soils in the upland of IH1, AI1, AI2, MF1 and MF2.

b. Non-parametric

By the non-parametric approach, the soils were more suitable for oil palm cultivation, potentially; more than it was obtained by the parametric method. This was made possible only after the removal of limitations except in situations where the main limitation was caused by soil physical characteristics such as CF (IH1), depth or wetness (IH2, AI3, MF3). In these situations, the cost implication of removing such limitation may be more than the risk in cultivating with such limitations. Alternatively, the land use type may be changed to a more suitable type. For instance, in IH1P1 organic carbon content was rated S_2(72) and constituted a major limitation (though slightly), while CF was rated S_2(76). IH1P1 was therefore classified as S_2sf, currently. Potentially, the organic carbon content of the soil may be improved by the use of soil organic amendments (animal droppings, plant residues,

compost etc.). However, it becomes difficult to remove limitations caused by CF; hence, its rating as S_2s, potentially. Similarly, the slight limitation caused by soil pH (S_2; 84) in IH2P1 can be removed by liming, while poor wetness condition may be removed by the installation of appropriate drainage facilities. This however, is a difficult task and may not be economically feasible, especially for tree crops. The soil qualified as S_3wf (limitation due to 'f' can be removed), currently and as S_3w, potentially. The suitability class (S_3) did not change as the main limitation (wetness) was not removed in the long run. In forest and fallow plots in southeast Nigeria, Asadu *et al*. (2020) rated soil properties except cation exchange capacity as highly suitable for oil palm cultivation (currently).

For each of IH2P2 (N_1wsf), AI3P1 (N_1wsf), AI3P2 (N_1wf), MF3P1 (N_1ws) and MF3P2 (N_1ws), the soils were either limited by wetness (w), depth (s), soil pH, base saturation or organic carbon (f); currently. Potentially, all the soils were upgraded to S_3ws as fertility limitations due to pH were removed by liming, suitable mineral fertilizer (base saturation) and organic soil amendments like compost and animal droppings (organic carbon). On the other hand, AI1P1, AI1P2, AI2P1, AI2P2, MF1P1, MF1P2 and MF2P1 were classified as S_2f in the land suitability subclass, while MF2P2 was classified as S_3f, currently. Interestingly, all the soils were limited by one kind of fertility characteristic or the other. It was therefore easier to remove limitations due to organic carbon by the use of poultry droppings, compost or by reducing bush burning and spreading palm fronds between the rows of oil palm trees. Limitations due to soil pH and base saturation were removed by the careful use of liming materials and suitable mineral fertilizers. The soils were therefore upgraded to highly suitable class (S_1).

Ranking the pedons for oil palm cultivation by their scores using the parametric approach (Table 39) indicated that the best ten (10) pedons were well-drained and found in higher elevation ranges, while the worst six (6) of them were poorly drained and influenced by regular flood, and all located in the lowest elevation range. This suggested that, wetness (drainage and flooding) quality was a major constraint to oil palm cultivation in the soils overlying limestone formations in Cross River State. Though the tracts of land had not degraded to permanently not suitable subclass, it is recommended that the land use type of the soils in the lowest elevation range (IH2, AI3, MF3) be changed and used for more water tolerant crops. This is so because of the expenses that will be incurred during the installation of drainage facilities, if it must be used for the cultivation of oil palm. The economic implication of the installation may be more than the benefits that will accrue from the installation process.

CHAPTER FIVE
SUMMARY AND CONCLUSIONS

5.1 Summary

Three areas underlain by various formations of limestone, ranging from SLS in the Ishibori area, SL in Agoi Ibami and SLSi in the Mfamosing area, were identified. A detailed survey of soils overlying these limestone formations was carried out using a digital elevation model (DEM), and maps were produced at scales ranging from 1:4,000 in Ishibori, 1:9,500 in Agoi Ibami to 1:15,000 in Mfamosing. Two elevation ranges were identified in soils developed from shale-limestone-sandstone, hence two mapping units (IH1 and IH2). Three elevation ranges were obtained, each in soils developed from sandstone-limestone formation (AI1, AI2 and AI3) and shale and limestone with sandstone intercalation (MF1, MF2 and MF3). Two mapping units in Ishibori against three in other locations was due to the narrow gap in elevation range. There was only slight difference in relief between the two mapping units. For instance, elevation ranges of 51-64 m, 54-106 m and 18-70 m were obtained for Ishibori, Agoi Ibami and Mfamosing areas, respectively.

Dark reddish brown (5YR 3/3) and black (5YR 2.5/1) topsoil colours and very dark grayish brown (10YR 3/2) to gray (7.5YR 5/1) colours were obtained in the subsurface soils overlying SLS in the high elevation ranges. However, the low elevation soils (IH2) were mainly dark brown (7.5YR 3/2) and gray (2.5Y (5/1-6/1)) with mottles of various colours recorded in the endopedons. Soils overlying SL were dark brown (7.5YR 3/2) to very dark gray (10YR 3/1) and brown (7.5YR4/2) in the epipedons. Furthermore, the surface soils overlying SLSi were very dark gray (7.5YR 3/1) to dark gray (5YR 4/1) and black (5YR 2.5/1). Of no less importance were the presence of Fe and Mn concretions and nodules in high to moderate elevation ranges, while weathered secondary carbonates were observed in IH1, AI2P2 and MF1P1 with clay and Fe cutans obtained on ped faces. Cracks of variable widths were observed in AI1, AI2 and MF1 mapping units.

Sandy clay loam and sandy loam textures dominated soils overlying SLS, while loamy sand, sandy loam and sandy clay loam textures dominated soils over SL and SLSi. Irrespective of elevation ranges, bulk density values of 0.71-1.66, 0.98-1.66 and 0.87-1.62 g/cm^3 were obtained in soils over SLS, SL and SLSi, respectively.

The high % contribution of $\theta_{30}=\theta_{90}>\theta_{60}>\theta_{v30}$>SMC>Capillary porosity>BD>Total porosity>θ_{v90} to principal component 1 (PC1) explains the relative importance of the properties to soil variation between the lithologies.

Soil pH (H_2O) values varied from 5.2 to 7.3, 5.2 to 7.7 and 5.1 to 6.7 for soils over SLS, SL and SLSi, respectively. Organic carbon values were comparatively higher in the low elevation areas. Consequently, ranges of 1.37-86.64, 1.03-22.16 and 0.69-46.34 g/kg were obtained in the soils overlying SLS, SL and SLSi, respectively. Exchangeable Ca was generally rated low to medium depending on elevation range and limestone formations. On the other hand, exchangeable Mg was rated medium in the soils over SLS and low to medium in other limestone formations. However, exchangeable K was generally low in the entire soil. Cation exchange capacity (NH_4OAc) was low to moderate in the soils overlying SL and high in soils of other limestone formations with correlation values that indicate that organic matter may be responsible for its trend. The values of calcium carbonate equivalent ranged from 17.7 to 38.8, 15.8 to 30.0, and 6.1 to 45.7 g/kg in the soils developed from SLS, SL and SLSi, respectively.

Principal component 1 (PC1) contributed 25.7 % of the 55.6 % contributed by PCs 1-3 to soil variation between lithologies. PC1 was dominated by loadings from organic carbon (93 %), exchangeable Na (91 %) and K (89 %), with exchangeable Ca (65 %) and calcium carbonate equivalence (6 %) contributing very low to PC1. This is perhaps due to the similarity in lithological make-up.

High values of immobile oxides such as TiO_2, MnO_2, Fe_2O_3, Al_2O_3 and ZrO_2 obtained by XRF in the soils indicate that the soils are highly weathered. However, lower values of mobile K_2O in the same soils is linked to active leaching process. The dominance of SiO_2 in the soils fortells the preponderance of kaolinite and quartz in the mineralogy of the soils.
Principal component 1 contributed 31.42 % to the overall variation between lithologies. Factor loadings to PC1 were influenced by K_2O, TiO_2, MnO_2 and Fe_2O_3 with loadings of 0.878, -0.790, -0.619 and -0.836, respectively.

Ruxton index, weathering index and molar ratio identified soils developed from SLSi in southern agricultural zone as the most weathered. Values of MR less than 3.0 indicates tha the soils contain kaolinite and gibbsite. Molar ratio was strongly positively correlated with R (r= 0.85) and WI (r= 0.477), and indicated good level of compatibility. However, R and WI were not correlated significantly. Dominant soil forming processes in the areas irrespective of limestone formations include: lessivation or argilluviation, calcification-decalcification, gleization, lateritization and latosolization, faunal pedoturbation, mineralization and immobilization. The mineralogy of clay fraction indicated the dominance of kaolinite, quartz, montmorillonite, vermiculite and palygorskite.

Soil classification by the USDA System of Soil Taxonomy was correlated with WRB for Soil Resources System and indicated that Loamy Skeletal, Mixed, Superactive, Isohyperthermic, Typic Rhodustults was correlated with Ferric Rhodic Acrisol (Ochric), while Loamy Skeletal, Mixed, Superactive, Ishyperthermic, Typic Paleustult matched Ferric Acrisol (Cutanic) in IH1. In IH2, Fluvaquentic Humaquept and Humic Endoaquept were correlated with Gleyic Cambisol (Humic) and Stagnic Gleyic Cambisol (Humic), respectively. Sandy, Mixed, Superactive, Isohyperthermic, Rhodic Paleudults was correlated with Ferric Acrisols (Ochric), while Fine loamy, Mixed, Superactive, Isohyperthermic, Plinthic Paleudults correlated with Plinthic Acrisols (Differentic, Hyperdystric, Ochric) for AI1P1 and AI1P2, respectively. Loamy, Mixed, Superactive, Isohyperthermic, Vertic Paleudult (AI2P1) and Coarse-loamy, Smectitic, Semiactive, Isohyperthermic, Vertic Paleudalf (AI2P2) correlated with Haplic Acrisols (Differentic, Hyperdystric, Ochric, Profondic) and Vertic Ferric Luvisols (Differentic, Loamic, Ochric, Profondic), respectively. Poorly drained and low elevation soils of AI3 were classified as Humic and Typic Endoaquepts which correlated with Dystric Gleyic Cambisols (Humic, Loamic) and Dystric Cambisols (Arenic, Ochric).

In the soils overlying SLSi, MF1P1 met requirements for Coarse loamy, Siliceous, Superactive, Isohyperthermic, Typic Paleuhumult (Haplic Acrisol; Humic, Hyperdystric, Loamic, Profondic), while MF2P2 qualified as Coarse Loamy, Kaolinitic, Superactive, Isohyperthermic, Typic Plinthohumult (Haplic Acrisol; Differentic, Humic, Hyperdystric). MF2P1, MF2P2, MF3P1 and MF3P2 were classified as Coarse-loamy, Siliceous, Superactive, Isohyperthermic, Typic Dystrudept, Sandy, Mixed, Superactive, Isohyperthermic, Typic Palehumult, Aeric Fluvaquent and Fluvaquentic Endoaquept in the USDA System which correlated with Dystric Cambisol (Humic, Loamic), Chromic Acrisol (Humic, Hyperdystric), Dystric Regosol (Loamic, Ochric) and Dystric Fluvic Cambisol (Humic), respectively.

Land capability classification indicates that the uplands of all the limestone formations except AI1P1 were class II soils, and limited by gravels or soil texture (s) and/or base saturation; in combination or in combination or alone. On the other hand, soils in the poorly drained IH2, AI3 and MF3 were classified as class III or IV and variously limited by wetness, soil physical properties and fertility characteristics. In the land suitability system, soils in mapping units IH1, AI1, AI2, MF1 and MF2P1 qualified as moderately suitable (S_2) for the cultivation of oil palm (currently), while AI1P1, AI2, MF1 and MF2 were highly suitable (S_1); potentially by the parametric system. On the other hand, IH2, AI3 and MF3

were either not suitable (N) or marginally suitable (S_3) by the parametric system. By the non parametric approach, most of the soils were better rated, potentially and currently, compared to the parametric approach. IH2P1 was marginally suitable (S_3), currently and potentially, and limited by wetness and fertility properties.

5.2 Conclusion

Soil survey maps were produced at scales of 1:4,000 in Ishibori, 1:9,500 in Agoi Ibami and 1:15,000 in Mfamosing resulting in eight soil mapping units identified *via* digital elevation model.

Concretions and nodules of Fe and Mn were more in the endopedons of soils developed on SLS and SL than it was obtained in the epipedons. Cracks were prominent in the endopedons on high and medium elevation ranges of the soils overlying SL and few in SLSi. Finer soil textures were obtained in the soils overlying SLS than it was recorded in soils over limestone formations in other areas. The epipedons of the well-drained soils over SLS were thick compared to the thin epipedons in soils over SL and SLSi.

Soil particle size distribution was dominated by sand in the entire soils with comparatively higher clay in soils over SLS. Bulk density was less than 1.8 g/cm^3 in the entire soils, while saturated hydraulic conductivity (Ksat.) had wide variation of 0.49-256.54 cm/h with higher values occurring in the epipedons. Ksat. correlated strongly but negatively with soil bulk density. Soil water content at the various tensions, capillary porosity and bulk density were among the most influential physical variables that contributed to the loading of PC1 and soil variation between the lithologies.

Low elevations had higher organic carbon than soils in the uplands which gave low to moderate values. Low to moderate exchangeable Ca and Mg, and low exchangeable K and Na in the entire soils suggest the occurrence of active leaching process. Calcium carbonate equivalent had a direct relationship with exchangeable Ca and clay. The first PC was dominated by contributions from soil organic carbon, exchangeable Na and K, with exchangeable Ca (65 %) and calcium carbonate equivalence (6 %) contributing very low to PC1. This is due to the similarity between the various limestone formations.

The ratio Fe_o/Fe_d in the solum is low and indicates low degree of activation with free iron oxides at an advanced stage of crystallinity or aging. However, co-migration of clay indicates relatively active pedogenic processes in the soils developed from SL and SLSi. About 48 % of variation between lithologies was contributed by PC1. PC1 was influenced by Fed, Ald, Alo, crystalline Fe and Al as well as co-migration of Fe and Al, with over 80 % loading of each property to variation between the soils.

Higher Al and Fe, and lower Si oxides in the epipedons of soils over SLS, than those over SL and SLSi suggest active leaching of SiO_2 and abundance of Fe_2O_3 and Al_2O_3 in the soils over SLS, relative to soils over SL and SLSi. Only 31.4 % of the total variation was accounted for by PC1, with 87.8, 79.0, 61.9 and 83.6 loadings from K_2O, TiO_2, MnO_2 and Fe_2O_3, respectively. Ruxton index, molar ratio and weathering index identified soils developed from SL and SLSi as the most developed. Quartz, kaolinite and montmorillonite as well as vermiculites and palygorskite were distributed in substantial amounts across the studied soils.

Lessivation, calcification-decalcification, gleization, lateritization, faunal pedoturbation, as well as mineralization and immobilization were among the dominant soil forming processes in the study locations.

Taxonomically, the soils were classified as Ultisols and correlated with Acrisols in the WRB for soil resources system. However, poorly drained IH2, AI3, MF2P1 and MF3P2 qualified as Inceptisols and correlated with Cambisols. AI2P2 was classified as Alfisols (Luvisols), while MF3P1 was qualified as Entisols (Regosols).

The soils variously met the criteria for land capability classes II, III and V. In the soils overlying SLS, 37.2 % were class II soils, while 61.7 % were class III and only 1.1 % qualified as class IV. For soils overlying SL, 49 % were class II soils, 35.3 % class III and 15.7 % were classified as class IV soils. In the soils overlying SLSi, 71.8 % of the soils qualified as class II, while 28.2 % qualified as class III soils.

Land suitability evaluation for oil palm cultivation indicates suitability classes ranging from S_1 to S_2, S_3 and N by the parametric and non parametric approaches. In the soils overlying SLS, the parametric approach indicated that 42 % of the soils were S_2 and 55.2 % were S_3, while only 2.8 % were not suitable for oil palm cultivation; potentially. For soils developed from SL, 35.4 % of the soils were S_1, while 31.6 % were S_2 and 26.6 % were S_3. However, only 6.4 % of the soils were N for oil palm cultivation, potentially. Currently, 66.3 % of the soils were S_2, while 33.7 % were N for the land use. In the soils overlying SLSi, the parametric approach indicated that 82.1 % of the soils were S_1, while only 17.9 % were S_3, potentially. The potential suitability classes witnessed significant improvement from the current suitability which indicated 37.5 %, 51.3 % and 11.2 % for S_2, S_3 and N, respectively.

5.3 Recommendations

1. An intensive fertility evaluation study for the limestone soils in the state is recommended for a site specific nutrient recommendation for commonly grown crops.

2. Blanket application of fertilizers should as much as possible be avoided as the soils over limestone formation in the humid tropics do not match the characteristics of similar soils in moisture stressed environments.
3. The poorly drained soils in the low elevations are not economical for cultivation with oil palm. Soils with comparative advantages for oil palm cultivation should be properly managed for best results.

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APPENDICES

Appendix 1

Long-hand profile description of the soils

General site information for IH1P1 (Pedon I)

Location: Icaptang - Ishibori oil palm plantation – Ogoja (N06º38.53′, E008º48.75 ′, 64 m).

Soil parent material: Shale, limestone and sandstone formation

Geology: Sedimentary Basin

Geomorphology: Nearly level, 2-4 %

Topography: Upper slope

Vegetation/land use: Southern guinea savannah cultivated with oil palm

Soil erosion hazard: Slight sheet erosion

Drainage: Moderately well drained

Depth to water table: None encountered at 195 cm

Depth to impenetrable layer: None encountered at 195 cm

Date sampled: December 11, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 37	Dark reddish brown (5YR 3/3) clay loam; moderate medium sub-angular blocky structure; slightly sticky (wet) friable (moist); many medium pores; common medium and coarse roots; common medium Ants and Worms; many Fe and Mn concretions; clear smooth boundary.
Bt	37 – 64	Dark red (10R 3/6) clay loam with common fine to medium distinct very pale brown (10YR 7/3) mottles; moderate medium sub-angular blocky structure; sticky (wet) firm (moist); few moderate clay and Fe cutans on ped faces; common medium pores; common medium roots; few Ants; few Fe and Mn concretions; clear smooth boundary.
Btg	64 – 130	Brown (7.5YR 5/3) clay with many coarse distinct gray (5Y 6/1) mottles; moderate medium angular blocky structure; sticky (wet) firm (moist); very few thin clay cutans on ped faces; few very fine pores; few to common fine roots; few Mn nodules; clear wavy boundary.
BC	130 – 195	Very dark grayish brown (10YR 3/2) sandy clay loam with few fine to medium faint gray (5Y 6/1) mottles; moderate medium granular structure; slightly sticky (wet) friable (moist); few and common fine pores; evidences of Fe^{3+}.

Plate 1: Soil profile IH1P1

General site information for IH1P2 (Pedon 2)

Location:	Urban ward 1 - Ishibori Teak Plantation – Ogoja (N06^{o}38.643′, E008^{o}48.866 ′, 65 m).
Soil parent material:	Shale, limestone and sandstone formation
Geology:	Sedimentary Basin
Geomorphology:	Nearly level, 2-4 %
Topography:	Crest
Vegetation/land use:	Southern guinea savannah cultivated with Teak
Soil erosion hazard:	None
Drainage:	Moderately well drained
Depth to water table:	None encountered at 150 cm
Depth to impenetrable layer:	150 cm
Date sampled:	December 11, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 42	Black (5YR 2.5/1) sandy loam; weak medium granular structure; non sticky (wet) loose (moist); many medium pores; many coarse roots; many Ants; few Fe nodules; clear wavy boundary.
BC	42 – 70	Light brown (7.5YR 6/4) sandy clay; moderate medium sub-angular blocky structure; slightly sticky (wet) friable (moist); few thin clay and Fe cutans on ped faces; common fine pores; very fine and medium roots; very few Ants; many Fe nodules; gradual wavy boundary.
Crkt	70 – 150	Gray (7.5YR 5/1) clay with common medium distinct reddish yellow (7.5YR 6/8) mottles; moderate coarse angular blocky structure; sticky (wet) firm (moist); very few thin clay cutans on ped faces; few very fine pores; thin layer of weathered $CaCO_3$ crossing the layer at mid-way

Plate 2: Soil profile IH1P2

General site information for IH2P1 (PEDON 1)

Location:	Nwar Rice field – Ishibori, Ogoja (N06^{o}38.711′, E008^{o}48.930 ′, 54 m).
Soil parent material:	Alluvium influenced by limestone
Geology:	Sedimentary Basin
Geomorphology:	Nearly level, 0-2 %
Topography:	Valley floor
Vegetation/land use:	Southern Guinea savannah cultivated with Rice
Soil erosion hazard:	None
Drainage:	Poorly drained
Depth to water table:	145 cm
Depth to impenetrable layer:	None encountered at 145 cm
Date sampled:	December 11, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 26	Dark brown (7.5YR 3/2) clay loam; moderate medium crumb structure; slightly sticky (wet) friable (moist); common fine and medium pores; many fine roots; many Worms and Ants; clear smooth boundary.
Cg1	26 – 84	Gray (2.5Y 6/1) clay with many medium and coarse prominent brown (7.5YR 4/4) mottles; moderate medium sub-angular blocky structure; sticky (wet) firm (moist); few fine pores; few very fine roots; few Ants; few fine charcoal; gradual irregular boundary.
Cg2	84 – 145	Gray (2.5YR 6/1) clay with common medium distinct brown (7.5YR 4/4) mottles; massive structure; very sticky (wet) very firm (moist); few Fe nodules; common fine charcoal.

Plate 3: Soil profile IH2P1

General site information for IH2P2 (PEDON 2)

Location:	Nwar Rice field 2 – Ishibori, Ogoja (N06^{o}38.836′, E008^{o}48.824 ′, 51 m).
Soil parent material:	Alluvium influenced by limestone
Geology:	Sedimentary Basin
Geomorphology:	Level, 0-2 %
Topography:	Valley floor
Vegetation/land use:	Southern guinea savannah cultivated with Rice
Soil erosion hazard:	None
Drainage:	Very poorly drained
Depth to water table:	87 cm
Depth to impenetrable layer:	None encountered at 87 cm
Date sampled:	December 12, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 13	Dark brown (7.5YR 3/2) clay loam; weak medium granular structure; slightly sticky (wet) very friable (moist); many medium pores; many medium and coarse roots; many Ants and few earth Worms; clear smooth boundary.
AC	13 – 46	Gray (2.5Y 5/1) clay with common fine to medium distinct red (2.5YR 4/8) mottles; moderate medium sub-angular blocky structure; sticky (wet) firm (moist); very few thin clay cutans in pores; few fine to medium pores; common fine and medium roots; few Earth worms; clear wavy boundary.
Cg	46 – 87	Very dark greenish gray (Gley1 3/1) clay with few fine to medium distinct red (2.5YR 4/8) mottles; massive structure; very sticky (wet) very firm (moist).

Plate 4: Soil profile IH2P2

Plate 5: Gleization process evident in IH2P2

General site information for AI1P1 (PEDON I)

Location: Adagha – Agoi Ibami – Yakurr (N05^{o}43.435′, E008^{o}09.029 ′, 93 m).
Soil parent material: Sandstone and limestone formation
Geology: Sedimentary Basin
Geomorphology: Gently sloping, 2-4 %
Topography: Mid slope
Vegetation/land use: Tropical rainforest cultivated with oil palm and Cassava
Soil erosion hazard: None
Drainage: Well drained
Depth to water table: None encountered at 200 cm
Depth to impenetrable layer: None encountered at 200 cm
Date sampled: December 19, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 9	Dark brown (7.5YR 3/2) loamy sand; weak fine granular structure; non sticky (wet) loose (moist); many medium and coarse pores; many fine and medium roots; few fine and medium Ants; clear smooth boundary.
BA	9 – 54	Dark yellowish brown (10YR 4/6) sandy loam; weak medium sub-angular blocky structure; slightly sticky (wet) friable (moist); common fine and medium pores; many coarse roots; few Ants; cracks extending to the Bt horizon; gradual smooth boundary.
Bt	54 – 122	Yellowish red (5YR 4/6) sandy clay loam with few medium faint light yellowish brown (2.5Y 6/4) mottles; moderate medium sub angular blocky structure; slightly sticky (wet) friable (moist); few thin clay cutans on ped faces; few common pores; few very fine roots; many medium Worm casts; clear wavy boundary.
Cr	122 – 200	Yellowish red (5YR 4/6) clay loam with many medium and coarse prominent light yellowish brown (2.5Y 6/4) mottles; moderate medium sub angular structure; sticky (wet) friable (moist); few thin clay cutans on ped faces; few very fine pores; few fine roots; many Mn-Fe concretions with quartz.

Plate 6: Soil profile AI1P1

General site information for AI1P2 (PEDON 2)

Location: Rukri-Kokem Road, Agoi Ibami – Yakurr (N05^{o}44.134′, E008^{o}10.029 ′, 106 m).
Soil parent material: Sandstone and limestone formation
Geology: Sedimentary Basin
Geomorphology: Gently sloping, 2-4 %
Topography: Mid slope
Vegetation/land use: Tropical rainforest cultivated with oil palm and Cassava
Soil erosion hazard: Slight sheet erosion
Drainage: Moderately well drained
Depth to water table: None encountered at 210 cm after augering
Depth to impenetrable layer: None encountered at 210 cm after augering
Date sampled: December 19, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 15	Very dark gray (10YR 3/1) sand; weak fine granular structure; non sticky (wet) loose (moist); common to many medium and coarse pores; many fine to coarse roots; many Ants; clear smooth boundary.
AB	15 – 67	Brown (7.5YR 5/4) loamy sand; weak medium granular and sub-angular blocky structures; non sticky (wet) very friable (moist); many fine pores; many medium and coarse roots; few Ants and Worm casts; clear smooth boundary.
Btv	67 – 146	Very pale brown (10YR 7/3) clay loam with many medium and coarse distinct dark red (7.5R 3/8) mottles; moderate medium sub angular blocky structure; sticky (wet) firm (moist); few thin clay cutans on ped faces; few very fine pores; few fine roots; common quartz; few Ants; common cracks; clear wavy boundary.
Crt	146 – 180	Light brown (7.5YR 6/4) clay with common fine faint gray (2.5Y 6/1) mottles; moderate medium sub angular structure; very sticky (wet) firm (moist); few thin clay cutans on ped faces; few very fine pores; many Fe nodules and quartz; common to many charcoal.

Plate 7: Soil profile AI1P2

GENERAL SITE INFORMATION FOR AI2P1 (PEDON 1)

Location: Avassi, Agoi Ibami – Yakurr (N05º43.496′, E008º09.399 ′, 70 m).
Soil parent material: Sandstone and limestone formation
Geology: Sedimentary Basin
Geomorphology: Nearly level, 0-2 %
Topography: Crest
Vegetation/Land Use: Tropical rainforest under fallow with Bamboo and Independent plant and few old Cassava stands
Soil Erosion Hazard: None
Drainage: Moderately well drained
Depth to water table: None
Depth to impenetrable layer: 120 cm
Date sampled: December 19, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 12	Brown (7.5YR 4/2) sandy loam; moderate medium granular structure; non sticky (wet) loose (moist); many fine and medium pores; many fine and medium roots; clear smooth boundary.
BA	12 – 59	Reddish brown (5YR 4/4) sandy clay loam; weak medium sub-angular blocky structures; slightly sticky (wet) very friable (moist); common fine and coarse pores; few fine and coarse roots; few Ants and Worm casts; common cracks in the horizon; gradual wavy boundary.
Bt	59 – 120	Yellowish brown (10YR 5/4) clay with few fine distinct red (10R 4/8) mottles; moderate medium sub angular blocky structure; sticky (wet) firm (moist); common moderate clay cutans on ped faces; few fine to medium pores; few very fine roots; common Fe-Mn concretions; few charcoal and cracks.

Plate 8: Student delineating soil profile AI2P1

General site information for AI2P2 (PEDON 2)

Location: Community Secondary School, Agoi Ibami – Yakurr (N05^{o}44.386′, E008^{o}10.660 ′, 80 m).
Soil parent material: Sandstone and limestone formation
Geology: Sedimentary Basin
Geomorphology: Nearly level, 0-2 %
Topography: Crest
Vegetation/land use: Tropical rainforest with oil palm and gmelina
Soil erosion hazard: Slight sheet erosion
Drainage: Well drained
Depth to water table: None encountered at 132 cm
Depth to impenetrable layer: 132 cm
Date sampled: December 20, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 13	Brown (7.5YR 4/2) sandy loam; moderate medium granular structure; non sticky (wet) loose to very friable (moist); many medium and coarse pores; many medium and coarse roots; many Ants; clear smooth boundary.
Btv	13 – 60	Yellowish red (5YR 4/6) sandy clay loam with common medium distinct red (2.5YR 4/6) mottles; moderate and strong medium sub-angular blocky structures; slightly sticky (wet) friable (moist); common moderate clay and Fe cutans on ped faces; common fine to medium pores; common fine roots; few Ants; few quartz and Fe nodules; clear smooth boundary.
Bt	60 – 82	Brown (7.5YR 5/4) sandy clay loam with few fine faint brownish yellow (10YR 6/6) mottles; moderate medium sub angular blocky structure; slightly sticky (wet) friable (moist); few thin clay cutans on ped faces; common fine pores; few fine roots; few Fe nodules; clear wavy boundary.
Crk	82 - 132	Grayish brown (2.5Y 5/2) clay with few fine and medium distinct yellowish red (5YR 5/6) mottles; strong coarse sub angular and angular blocky structure; very sticky (wet) firm (moist); few medium pores; few Fe-Mn concretions; many cracks between 1.0 and 1.7 cm wide.

Plate 9: Soil profile AI2P2

Plate 10: Soil profile AI2P2 showing cracked surfaces

General site information for AI3P1 (PEDON 1)

Location: Arekpen stream, Agoi Ibami – Yakurr (N05^{o}43.725′, E008^{o}10.214′, 54 m).
Soil parent material: Alluvium over limestone
Geology: Sedimentary Basin
Geomorphology: Nearly level, 0-2 %
Topography: Valley floor
Vegetation/land use: Tropical rainforest with Bamboo and Banana
Soil erosion hazard: None
Drainage: Poorly drained
Depth to water table: None at 80 cm
Depth to impenetrable layer: 80 cm
Date sampled: December 20, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 21	Dark reddish brown (5YR 3/2) loamy sand; weak fine granular structure; non sticky (wet) loose (moist); many medium and coarse pores; many medium and coarse roots; many Ants and few beetles; gradual smooth boundary.
AB	21 – 80	Dark reddish gray (5YR 4/2) loamy sand with few fine to medium distinct red (10R 4/8) mottles; weak medium sub-angular blocky structures; slightly sticky (wet) very friable (moist); few fine pores; common fine and medium roots; few Ants and Worm casts; few Fe nodules.

Plate 11: Soil profile AI3P1

General site information for AI3P2 (PEDON 2)

Location: CSS Downstream, Agoi Ibami – Yakurr (N05^{o}44.316′, E008^{o}10.707 ′, 62 m).
Soil parent material: Alluvium over limestone
Geology: Sedimentary Basin
Geomorphology: Nearly level, 0-2 %
Topography: Valley floor
Vegetation/land use: Tropical rainforest with oil palm, teak and gmelina
Soil erosion hazard: None
Drainage: Poorly drained
Depth to water table: 124 cm
Depth to impenetrable layer: None at 124 cm
Date sampled: December 20, 2018.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 50	Dark gray (5YR 4/1) sand; weak fine granular structure; non sticky (wet) loose (moist); many medium and coarse pores; many fine to medium roots; few Ants and Termites; many charcoal; clear smooth boundary.
CB	50 – 94	Gray (5YR 5/1) sand; weak medium granular structure; non sticky (wet) loose (moist); common medium pores; common medium roots; few Fe nodules; clear wavy boundary.
Cg	94 – 124	Gray (7.5YR 5/1) sandy clay loam with many medium distinct strong brown (7.5YR 5/8) mottles; weak medium sub angular blocky structure; slightly sticky (wet) very friable (moist); few fine pores; few very fine and medium roots; few Fe nodules.

Plate 12: Soil profile AI3P2

General site information for MF1P1 (PEDON 1)

Location: Oil palm and plantain plantation opposite Community Park, Mfamosing – Akamkpa (N05^{o}04.714′, E008^{o}30.541 ′, 58 m).
Soil parent material: Shale and limestone with sandstone intercalation
Geology: Sedimentary Basin
Geomorphology: Gently sloping, 2-4 %
Topography: Crest
Vegetation/land use: Tropical rainforest with oil palm and plantain
Soil erosion hazard: Slight sheet erosion
Drainage: Well drained
Depth to water table: None encountered at 121 cm
Depth to impenetrable layer: 121 cm
Date sampled: February 8, 2019.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 17	Very dark gray (7.5YR 3/1) silt loam; moderate medium sub angular blocky structure; slightly sticky (wet) firm (moist); few to common medium and coarse pores; few fine roots; Termite cast and few Ants; clear smooth boundary.
Bt1	17 – 56	Light yellowish brown (10YR 6/4) clay; strong medium and coarse angular blocky structures; very sticky (wet) very firm (moist); very few thin clay cutans on ped faces; many coarse pores; few fine roots; wide cracks of about 3-7 cm; clear wavy boundary.
Bt2	56 – 121	Pale brown (10YR 6/3) clay with common coarse distinct gray (2.5Y 6/1) mottles; strong medium and coarse angular blocky structure; very sticky (wet) very firm (moist); common thin clay cutans on ped faces; common to many medium and coarse pores; few Ants; many coarse boulders of limestone covering about 75 % of pit.

Plate 13: Soil profile MF1P1

General site information for MF1P2 (PEDON 2)

Location: Wilmer oil palm Estate, Mfamosing – Akamkpa (N05^{o}04.120′, E008^{o}32.440 ′, 70 m).
Soil parent material: Shale and limestone with sandstone intercalation
Geology: Sedimentary Basin
Geomorphology: Gently sloping, 2-4 %
Topography: Crest
Vegetation/land use: Tropical rainforest; fallow
Soil erosion hazard: Slight sheet erosion
Drainage: Well drained
Depth to water table: None encountered at 153 cm
Depth to impenetrable layer: 153 cm
Date sampled: February 11, 2019.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 14	Black (5YR 2.5/1) sand; weak fine granular structure; non sticky (wet) loose (moist); many medium and coarse pores; many fine and medium roots; few charcoal; thick and closely interwoven fine to medium roots; clear smooth boundary.
BC	14 – 43	Reddish brown (5YR 4/3) sandy clay; weak fine to medium sub-angular blocky structures; slightly sticky (wet) very friable (moist); many medium pores; many medium and coarse roots; few Earthworms and Termites; gradual smooth boundary.
Bt	43 – 81	Reddish brown (5YR 4/4) sandy clay; weak to moderate fine and medium sub angular blocky structure; slightly sticky (wet) very friable (moist); few thin clay cutans on ped faces; few medium pores; few fine and coarse roots; few charcoal; many boulders; few plinthites; clear wavy boundary.
Crv	81 - 153	Red (10R 4/6) sand clay with few fine and medium distinct brownish yellow (10YR 6/6) mottles; moderate fine and medium sub angular blocky structure; sticky (wet) firm (moist); very few thin clay cutans on ped faces; few fine pores; rock covering over 35 % of the surface with plinthites and hard pan at this depth.

Plate 14: Soil profile MF1P2

General site information for MF2P1 (PEDON 1)

Location: UNICEM Park Canteen, Mfamosing – Akamkpa (N05^{o}04.828′, E008^{o}30.389 ′, 44 m).
Soil parent material: Shale and limestone with sandstone intercalation
Geology: Sedimentary Basin
Geomorphology: Gently sloping, 2-4 %
Topography: Mid slope
Vegetation/land use: Tropical rainforest; oil palm
Soil erosion hazard: Slight sheet erosion
Drainage: Well drained
Depth to water table: None encountered at 161 cm
Depth to impenetrable layer: 161 cm
Date sampled: February 8, 2019.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 16	Very dark gray (7.5YR 3/1) clay; weak fine sub angular blocky structure; very sticky (wet) very firm (moist); many medium and coarse pores; many fine and medium roots; many Ants; few Termites; few charcoal; clear smooth boundary.
Bt	16 – 44	Brown (7.5YR 5/3) clay; weak medium sub-angular blocky structures; very sticky (wet) firm (moist); very few thin clay cutans on ped faces; common fine and medium pores; common medium roots; few Fe nodules; common Termite cast and few Ants; few charcoal; clear wavy boundary.
Cr1	44 – 97	Gray (7.5YR 5/1) clay with common medium distinct red (2.5YR 5/8) mottles; weak and moderate medium platy structure; sticky (wet) firm (moist); few fine and medium pores; few very fine roots; gradual smooth boundary.
Cr2	97 – 161	Gray (7.5YR 5/1) clay with common medium distinct red (2.5YR 5/8) and reddish black (2.5YR 2.5/1) mottles; moderate medium platy structure; sticky (wet) firm (moist); common fine and medium pores; Duripan

Plate 15: Soil profile MF2P1

General site information for MF2P2 (PEDON 2)

Location: Gmelina plantation before UNICEM, Mfamosing – Akamkpa (N05^{o}04.009′, E008^{o}29.634 ′, 43 m).
Soil parent material: Shale and limestone with sandstone intercalation
Geology: Sedimentary Basin
Geomorphology: Gently sloping, 2-4 %
Topography: Upper slope
Vegetation/land use: Tropical rainforest; Gmelina with Cassava
Soil erosion hazard: Slight sheet erosion
Drainage: Well drained
Depth to water table: None encountered at 200 cm
Depth to impenetrable layer: None encountered at 200 cm
Date sampled: February 9, 2019.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 16	Very dark gray (7.5YR 3/1) loamy sand; weak fine granular structure; non sticky (wet) loose (moist); common fine and medium pores; many fine, medium and coarse roots; many large Ants; few fine quartz grains; clear smooth boundary.
BA	16 – 45	Brown (7.5YR 4/3) sandy clay loam; weak fine and medium sub-angular blocky structures; slightly sticky (wet) very friable (moist); common fine pores; common coarse roots; few large Ants; few charcoal; clear wavy boundary.
Bt1	45 – 107	Brown (7.5YR 4/3) sandy clay loam; moderate medium sub angular blocky structure; slightly sticky (wet) friable (moist); few thin clay cutans on ped faces; few very fine pores; few very fine roots; many fine to medium quartz grains; many medium charcoal with boulders of sandstone outcrops; gradual wavy boundary.
Bt2	107 - 200	Strong brown (7.5YR 5/8) sandy clay; moderate medium sub angular blocky structure; slightly sticky (wet) friable (moist); few thin clay cutans on ped faces; few very fine pores; few fine and coarse roots; very few charcoal with cobbles of sandstone.

Plate 16: Soil profile MF2P2

General site information for MF3P1 (PEDON 1)

Location: Swamp before NIS Check point, Mfamosing – Akamkpa (N05^{o}04.664′, E008^{o}30.973 ′, 18 m).
Soil parent material: Alluvium
Geology: Sedimentary Basin
Geomorphology: Level, 0-2 %
Topography: Valley floor
Vegetation/land use: Tropical rainforest; Maize
Soil erosion hazard: None
Drainage: Very poorly drained
Depth to water table: 48 cm
Depth to impenetrable layer: None
Date sampled: February 9, 2019.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 15	Dark gray (5YR 4/1) clay with common medium distinct red (2.5YR 4/8) mottles; moderate medium sub angular blocky structure; sticky (wet) firm (moist); common fine and medium pores; common fine roots; few charcoal; clear smooth boundary.
Cg	15 – 48	Yellowish brown (10YR 5/4) clay with common medium distinct strong brown (7.5YR 5/8) mottles; massive structure; sticky (wet) firm (moist); few very fine pores; few very fine roots; few charcoal.

Plate 17: Soil profile MF3P1

General site information for MF3P2 (PEDON 2)

Location: Before UNICEM, Mfamosing – Akamkpa (N05^{o}04.274′, E008^{o}29.695 ′, 23 m).
Soil parent material: Alluvium
Geology: Sedimentary Basin
Geomorphology: Level, 0-2 %
Topography: Valley floor
Vegetation/land Use: Tropical rainforest with Maize
Soil erosion hazard: None
Drainage: Very poorly drained
Depth to water table: 49 cm
Depth to impenetrable layer: None
Date sampled: February 11, 2019.

Horizon descriptions

Horizon	Depth (cm)	Description
Ap	0 – 16	Black (5YR 2.5/1) clay loam; weak medium granular structure; slightly sticky (wet) very friable (moist); common medium pores; many fine and medium roots; many charcoal; highly ramified by fine and medium roots; clear smooth boundary.
Cg1	16 – 28	Reddish yellow (7.5YR 6/6) clay with few fine faint red (2.5YR 5/8) mottles; moderate medium subangular blocky structure; sticky (wet) firm (moist); common medium pores; common fine roots; few charcoal and many crabs; clear wavy boundary.
Cg2	28 – 49	Dark gray (5YR 4/1) sand; single grained structure; non sticky (wet) loose (moist); few very fine pores; few fine and medium roots; few charcoal and many crabs.

Plate 18: Soil profile MF3P2

Appendix 2: Correlation matrix of physical and chemical properties

					Porosity				Soil pH					Exchangeable Bases						Exch. Acidity			
	Silt	Sand	BD	K_{sat}	Total	Air P	Capil	M.con	H_2O	$CaCl_2$	OrgC	TN	C/N	Na^+	K^+	Mg^{2+}	Ca^{2+}	CEC	BS	Al^{3+}	H^+	Bray 1	CCE
Clay	0.547*	-0.887*	0.328*	-0.138	0.121	0.21	-0.121	-0.1	0.281	-0.079	-0.099	-0.033	-0.227	-0.126	-0.173	0.4*	0.544*	0.274	0.231	-0.108	-0.175	-0.099	0.387*
Silt		-0.872*	0.053	-0.065	0.122	-0.029	0.071	0.048	0.389*	0.089	0.184	0.185	0.064	0.126	0.094	0.449*	0.633*	0.261	0.415*	-0.116	-0.21	0.213	0.189
Sand			-0.221	0.117	-0.138	-0.106	0.032	0.032	-0.379*	-0.003	-0.044	-0.083	0.097	0.004	0.049	-0.482*	-0.668*	-0.304	-0.364*	0.127	0.218	-0.06	-0.331
BD				-0.712*	-0.637*	-0.01	-0.68*	-0.916*	0.254	-0.048	-0.789*	-0.635*	-0.618*	-0.76*	-0.759*	-0.204	-0.245	-0.304	0.113	-0.015	-0.035	-0.577*	0.064
Ksat					0.57*	0.268	0.433*	0.757*	-0.103	0.104	0.639*	0.326	0.58*	0.684*	0.693*	0.027	0.197	0.138	-0.207	-0.211	-0.213	0.482*	0.169
Total						0.502*	0.676*	0.833*	0.001	0.025	0.585*	0.227	0.462*	0.55*	0.518*	0.315	0.376*	0.121	-0.01	-0.305	-0.353	0.448*	0.102
Air sp.							-0.198	0.238	-0.167	-0.25	0.023	0.029	-0.066	0.079	0.052	-0.004	-0.072	-0.03	-0.169	-0.145	-0.173	-0.07	0.009
Capill								0.701*	0.071	0.169	0.59*	0.182	0.553*	0.525*	0.534*	0.255	0.386*	0.14	0.016	-0.239	-0.271	0.552*	-0.015
M.con									-0.11	0.055	0.851*	0.509*	0.66*	0.845*	0.818*	0.302	0.406*	0.318	0.071	-0.15	-0.166	0.571*	0.05
H_2O										0.752*	0.021	-0.14	0.022	0.055	0.035	0.27	0.534*	-0.019	0.29	-0.376*	-0.394*	0.159*	0.349
$CaCl_2$											0.16*	-0.023	0.222	0.174	0.178	0.007	0.295	-0.21	0.276	-0.356	-0.309	0.377*	0.171
Org.C												0.591*	0.892*	0.959*	0.962*	0.444*	0.499*	0.387*	0.041	-0.165	-0.191	0.626*	0.129
TN													0.353	0.552*	0.572*	0.363*	0.3	0.596*	-0.117	0.324	0.297	0.341	-0.006
C/N														0.84*	0.88*	0.294	0.285	0.169	-0.015	-0.243	-0.289	0.596*	0.107
Na^+															0.987*	0.401*	0.463*	0.413*	-0.047	-0.139	-0.165	0.578*	0.119
K^+																0.366*	0.413*	0.391*	-0.094	-0.136	-0.165	0.58*	0.076
Mg^{2+}																	0.686*	0.445*	0.396*	-0.009	0.066	0.125	0.23
Ca^{2+}																		0.429*	0.436*	-0.301	-0.301	0.369*	0.413*
CEC																			-0.351	0.612*	0.455*	0.081	0.154
BS																				-0.432*	-0.253	0.096	0.122
Al^{3+}																					0.91*	-0.33	-0.166
H^+																						-0.352	-0.101
Bray 1																							0.192

*significant at 5 % level of significance; BD=Bulk density; Ksat=Saturated hydraulic conductivity; Air P= Air filled porosity; Capil= Capillary porosity; M.con= moisture content; TN= Total nitrogen; CCE= Calcium carbonate equivalent; Bray 1= Available P by Bray 1 method

Appendix 3: Correlation matrix of forms of Fe and Al with elemental oxides and weathering indices

	Al_d	Fe_o	Al_o	Fe_d-Fe_o	Al_d-Al_o	Fe_o/Fe_d	Al_o/Al_d	Al_2O_3	SiO_2	P_2O_5	K_2O	CaO	TiO_2	MnO_2	Fe_2O_3	ZrO_2	R	WI	MR
Fe_d	0.712	0.223	0.096	0.987	0.728	-0.097	-0.354	-0.193	-0.334	0.222	-0.685	0.336	0.865	0.496	0.816	0.255	-0.042	-0.7	-0.472
Al_d		0.228	0.336	0.691	0.979	-0.068	-0.522	-0.049	-0.546	0.038	-0.755	0.472	0.587	0.342	0.729	-0.124	-0.289	-0.729	-0.635
Fe_o			0.197	0.062	0.198	-0.041	-0.06	-0.206	-0.361	0.716	-0.445	0.044	0.23	0.624	0.442	-0.096	-0.057	-0.459	-0.269
Al_o				0.065	0.139	-0.006	0.328	-0.101	-0.225	0.224	-0.428	-0.17	0.155	0.199	0.3	-0.2	-0.063	-0.456	-0.227
Fe_d-Fe_o					0.712	-0.093	-0.352	-0.163	-0.282	0.109	-0.627	0.337	0.848	0.404	0.762	0.277	-0.034	-0.64	-0.439
Al_d-Al_o						-0.07	-0.619	-0.03	-0.526	-0.008	-0.702	0.533	0.584	0.317	0.703	-0.088	-0.29	-0.669	-0.619
Fe_o/Fe_d							-0.009	0.314	-0.071	-0.141	0.131	-0.103	-0.114	0.175	-0.031	-0.111	-0.193	0.11	-0.092
Al_o/Al_d								-0.332	0.529	0.172	0.212	-0.373	-0.251	-0.083	-0.39	0.141	0.571	0.182	0.661
Al_2O_3									-0.384	-0.475	0.222	-0.299	-0.139	-0.063	-0.064	-0.448	-0.845	0.169	-0.532
SiO_2										-0.111	0.446	-0.034	-0.152	-0.341	-0.696	0.56	0.785	0.445	0.884
P_2O_5											-0.375	0.098	0.262	0.506	0.307	0.285	0.256	-0.366	0.031
K_2O												-0.312	-0.624	-0.475	-0.718	0.11	0.09	0.987	0.445
CaO													0.254	0.128	0.102	0.216	0.14	-0.181	-0.031
TiO_2														0.47	0.648	0.465	0.024	-0.654	-0.387
MnO_2															0.544	0.035	-0.096	-0.492	-0.369
Fe_2O_3																-0.11	-0.335	-0.757	-0.736
ZrO_2																	0.595	0.114	0.404
R																		0.12	0.85
WI																			0.477

*significant at 5 % level of significance; ‘d’= dithionite, o= oxalate, R= Ruxton index; WI= Weathering index; MR= Molar ratio

Appendix 4: Calculations involving index of productivity

IP = A √B/100 * C/100 * D/100* E/100* F/100

IH1P1

IPc = 72 √94/100 * 91/100 *97/100*76/100* 100/100 = 57.2

IPp = 76 √94/100 * 91/100 *97/100*100/100* 100/100 = 69.2

IH1P2

IPc = 80 √94/100 * 92/100 *97/100*89/100* 100/100 = 69.1

IPp = 80 √94/100 * 92/100 *97/100*100/100* 100/100 = 73.3

IH2P1

IPc = 35 √94/100 * 95/100 *90/100*84/100* 100/100 = 28.8

IPp = 35 √94/100 * 95/100 *90/100*100/100* 100/100 = 31.4

IH2P2

IPc = 35 √94/100 * 95/100 *48/100*58/100* 100/100 = 17.5

IPp = 35 √94/100 * 95/100 *48/100*100/100* 100/100 = 22.9

AI1P1

IPc = 72 √100/100 * 94/100 *100/100*96/100* 100/100 = 68.4

IPp = 94 √100/100 * 100/100 *96/100*100/100* 100/100 = 92

AI1P2

IPc = 75 √100/100 * 97/100 *100/100*96/100* 100/100 = 72.4

IPp = 75 √100/100 * 97/100 *100/100*96/100* 100/100 = 72.4

AI2P1

IPc = 75 √100/100 * 94/100 *100/100*88/100* 100/100 = 68.2

IPp = 88 √100/100 * 94/100 *100/100*100/100* 100/100 = 85.3

AI2P2

IPc = 70 √90/100 * 100/100 *100/100*100/100* 100/100 = 66.4

IPp = 80 √90/100 * 100/100 *100/100*100/100* 100/100 = 75.9

AI3P1

IPc = 35 √100/100 * 95/100 *37/100*40/100* 100/100 = 13.1

IPp = 37 √100/100 * 95/100 *40/100*88/100* 100/100 = 21.4

AI3P2

IPc = 30 √100/100 * 94/100 *37/100*87/100* 100/100 = 16.5

IPp = 37 √100/100 * 94/100 *87/100*100/100* 100/100 = 33.5

MF1P1

IPc = 75 √100/100 * 95/100 *100/100*89/100* 100/100 = 69.0

IPp = 89 √100/100 * 95/100 *100/100*100/100* 100/100 = 86.7

MF1P2

IPc = 58 √100/100 * 95/100 *100/100*96/100* 100/100 = 55.4

IPp = 95 √100/100 * 100/100 *96/100*100/100* 100/100 = 93.1

MF2P1

IPc = 56 √100/100 * 100/100 *100/100*92/100* 100/100 = 53.7

IPp = 92 √100/100 * 100/100 *100/100*100/100* 100/100 = 92.0

MF2P2

IPc = 40 √100/100 * 93/100 *100/100*90/100* 100/100 = 36.6

IPp = 90 √100/100 * 93/100 *100/100*100/100* 100/100 = 86.8

MF3P1

IPc = 30 √100/100 * 100/100 *36/100*85/100* 100/100 = 16.6

IPp = 30 √100/100 * 100/100 *36/100*100/100* 100/100 = 18.0

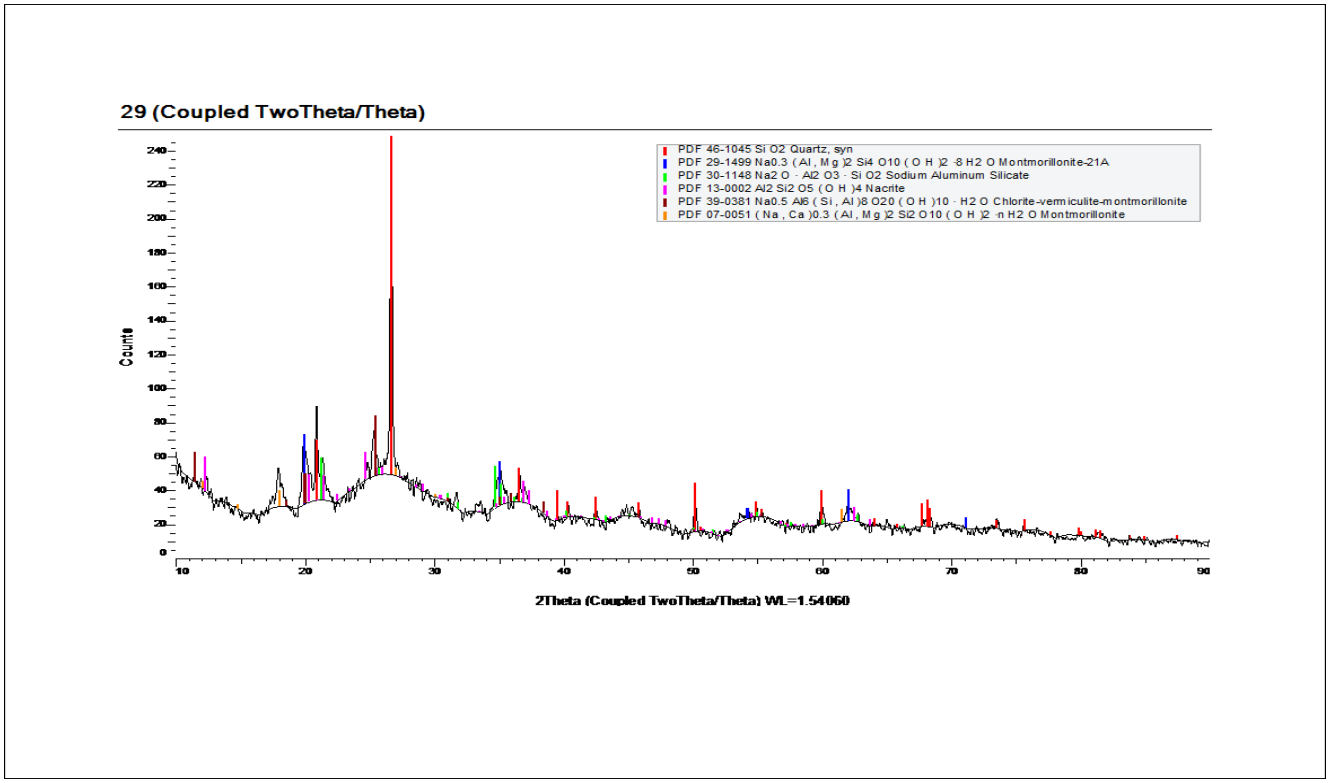
MF3P2

IPc = 31 √100/100 * 95/100 *36/100*86/100* 100/100 = 16.8

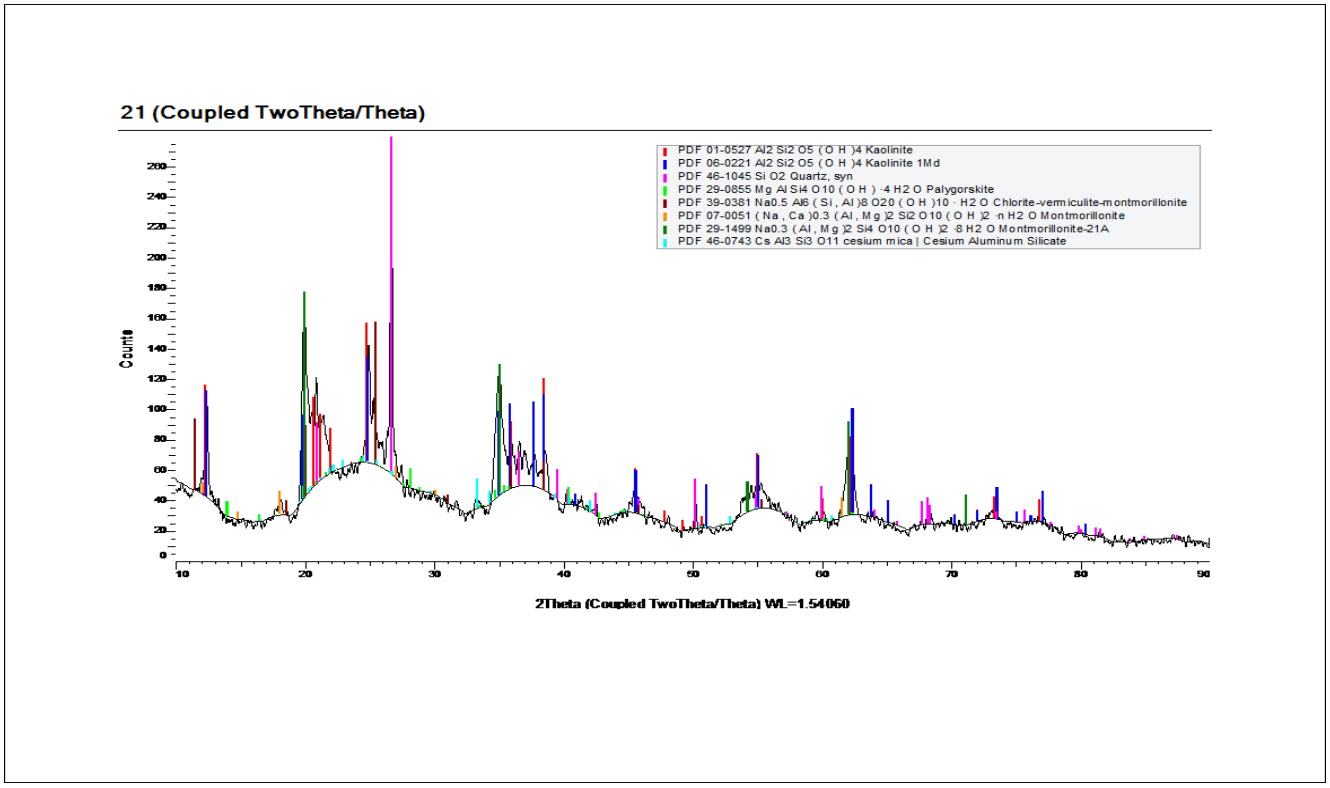
IPp = 31 √100/100 * 95/100 *36/100*100/100* 100/100 = 18.1

Where; IPp = Potential index of productivity. Base saturation, pH (H_2O) and organic carbon are not part of the “f” group since they are easily altered. IPc = Current or actual index of productivity. In this case, BS, pH (H_2O) and organic carbon are part of the “f” group.

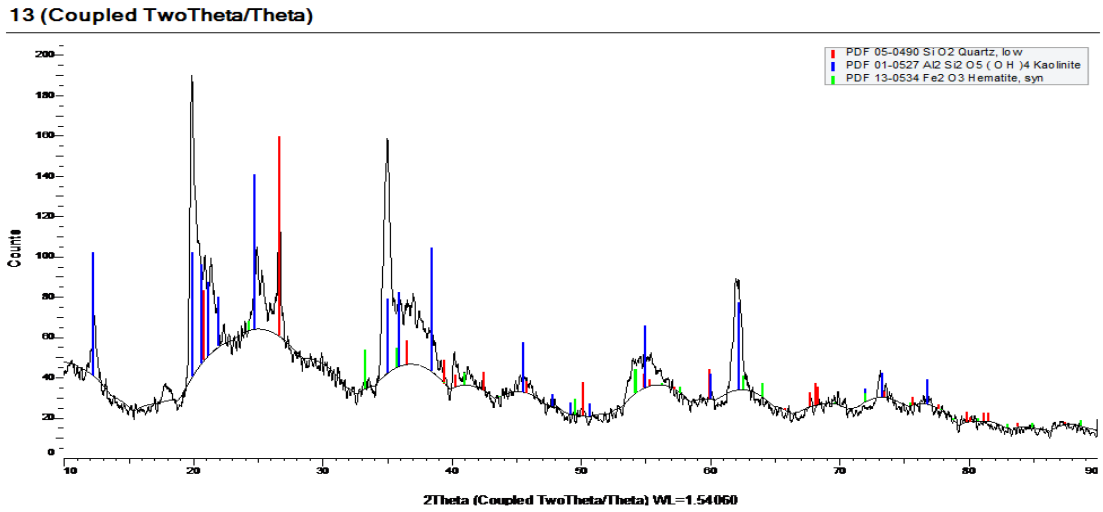
Appendix 5: X-ray diffractograms of soils formed from limestone formations
AI1P1: 0-9 cm



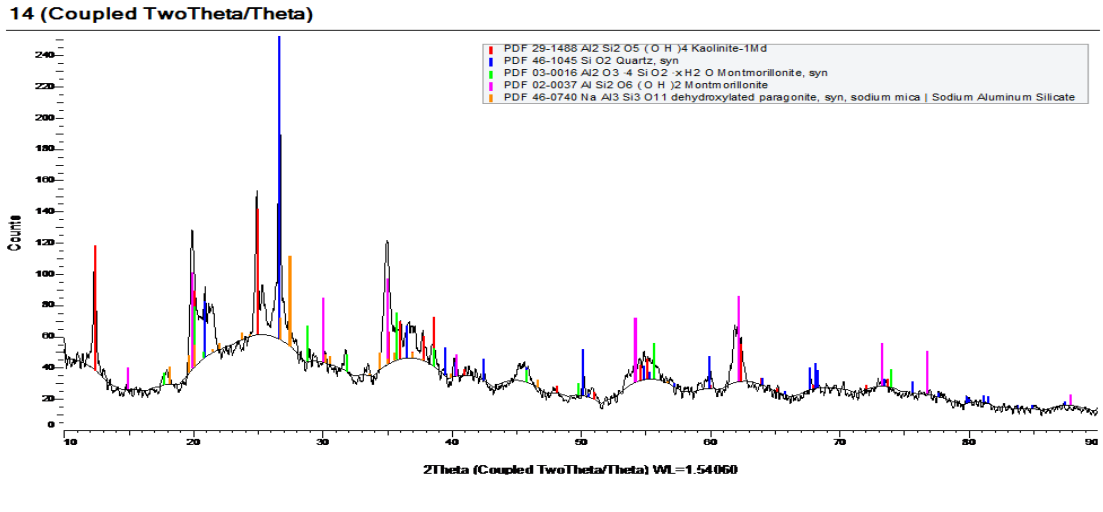
AI1P1: 9-54 cm



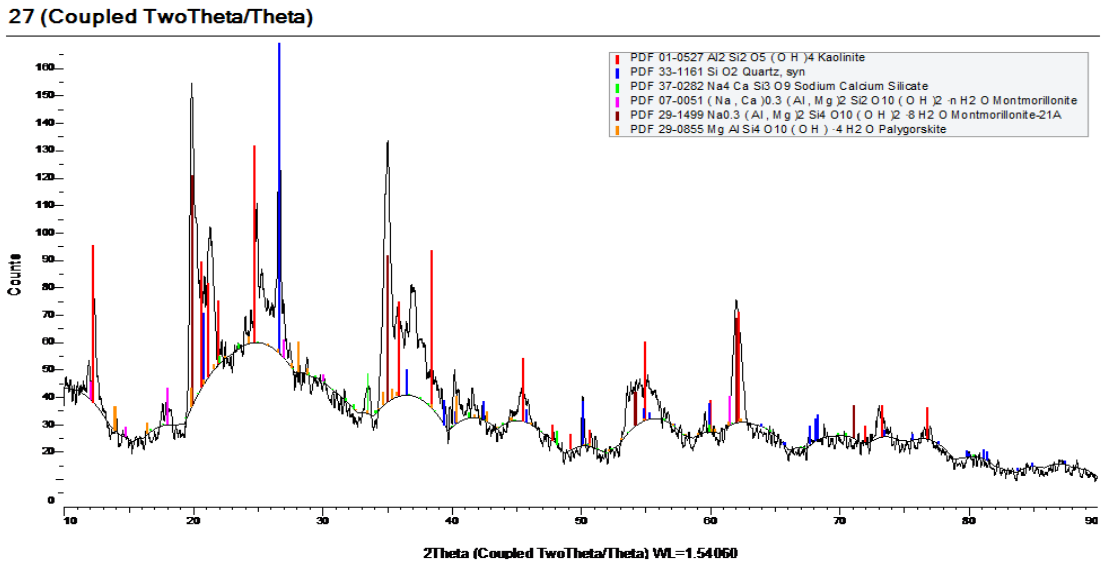
AI1P1: 122-200 cm



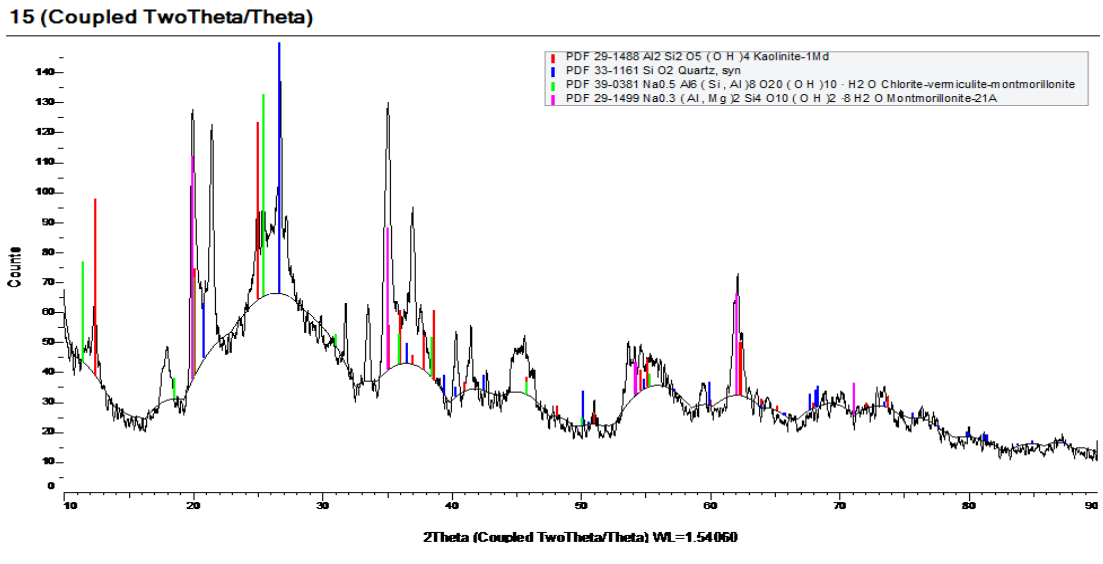
AI1P2: 0-15 cm



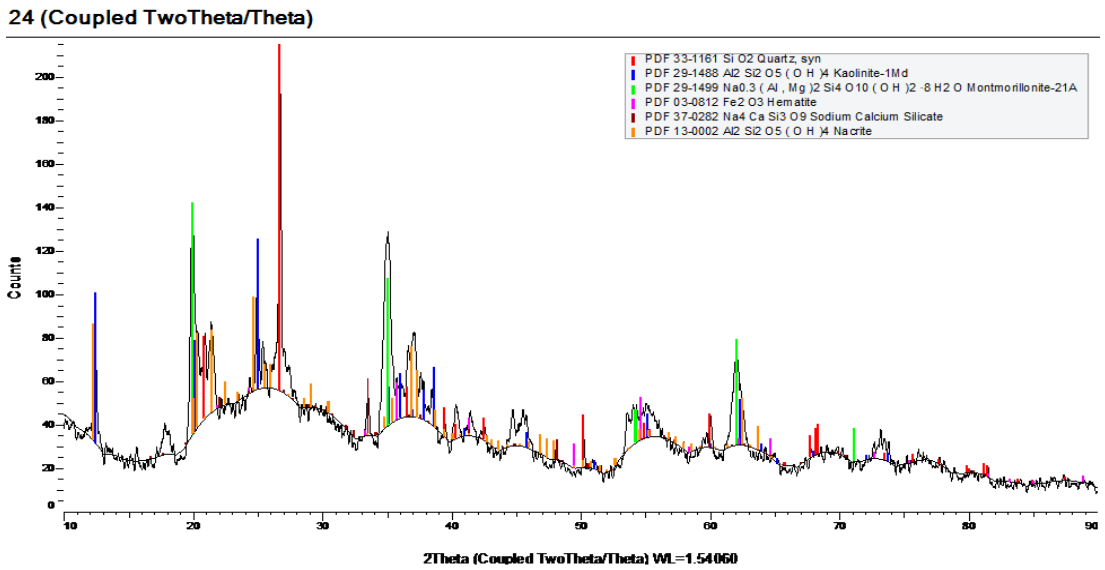
AI1P2: 67-146 cm



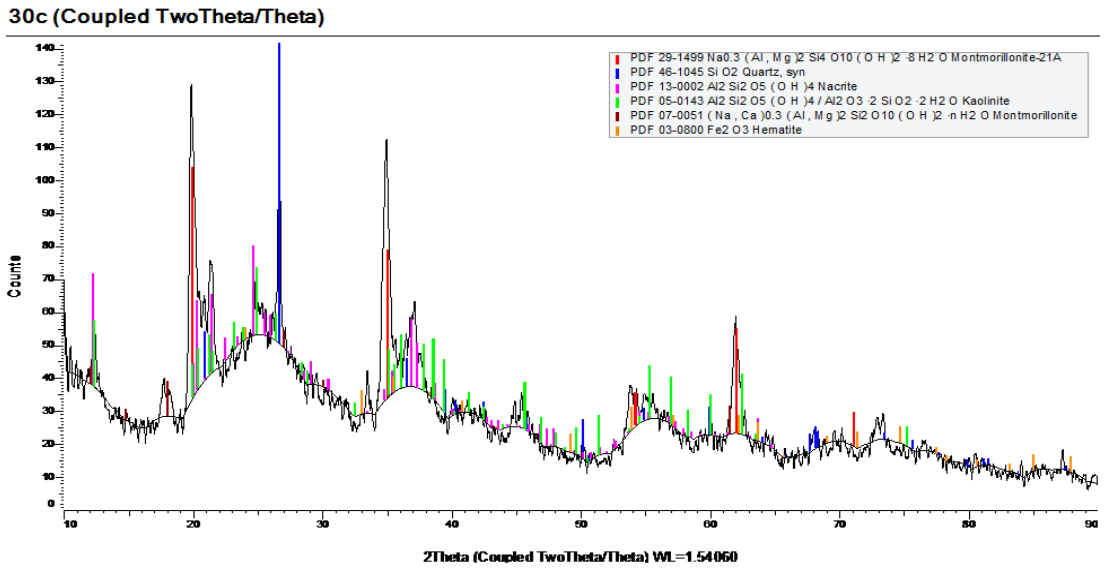
AI1P2: 146-180 cm



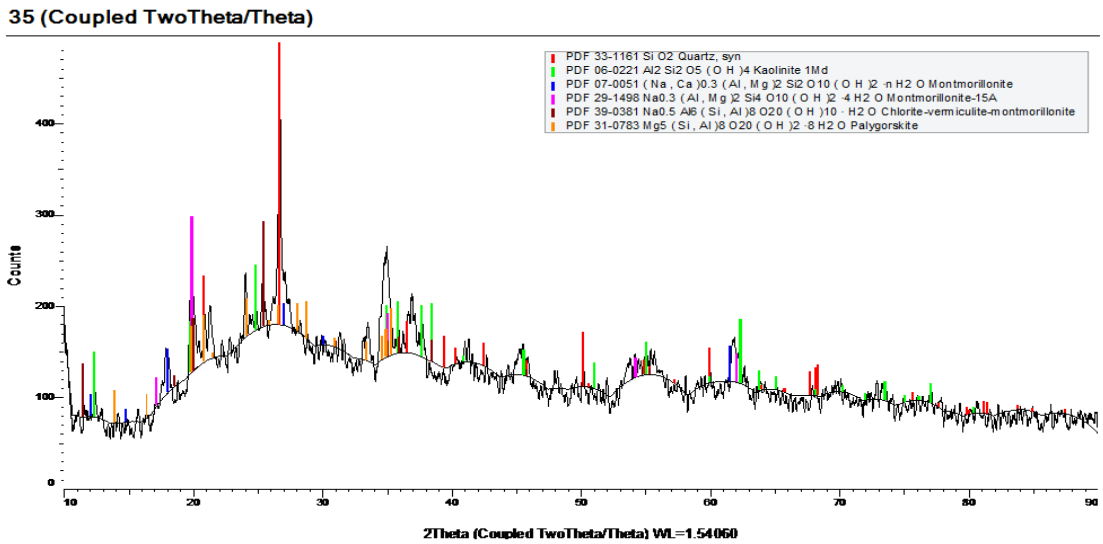
AI2P1: 0-12 cm



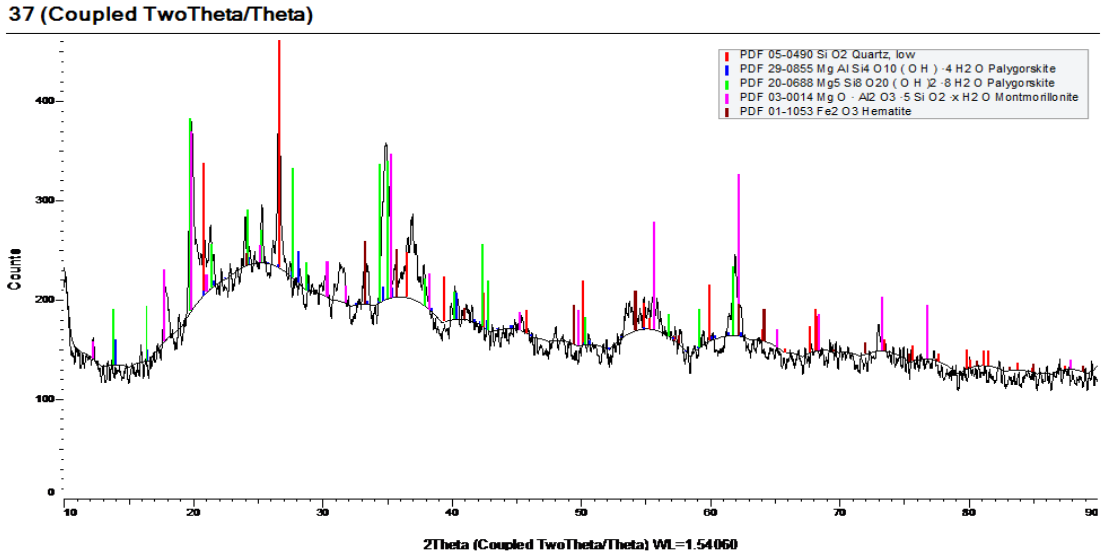
AI2P1: 59-120+ cm



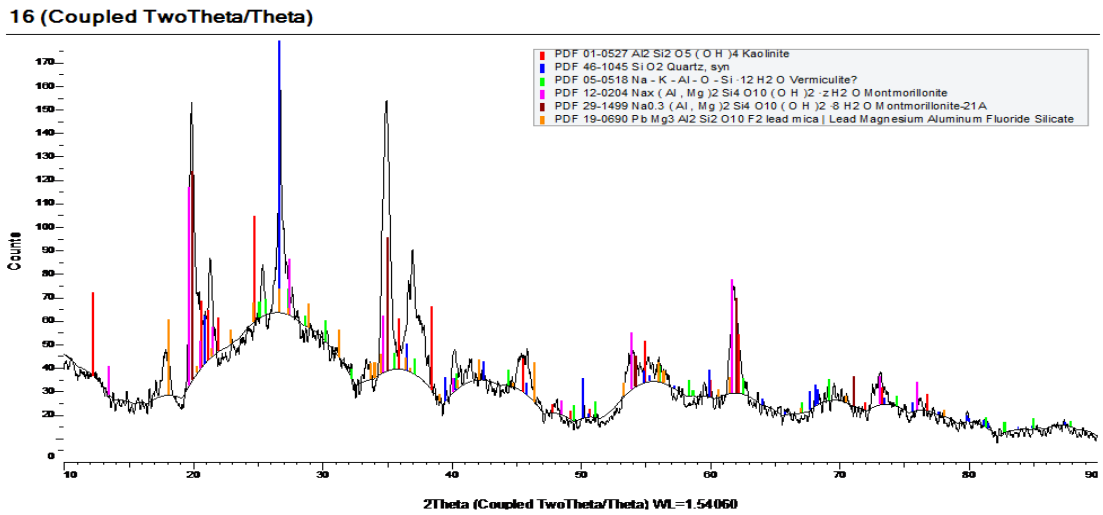
AI2P2: 0-13 cm



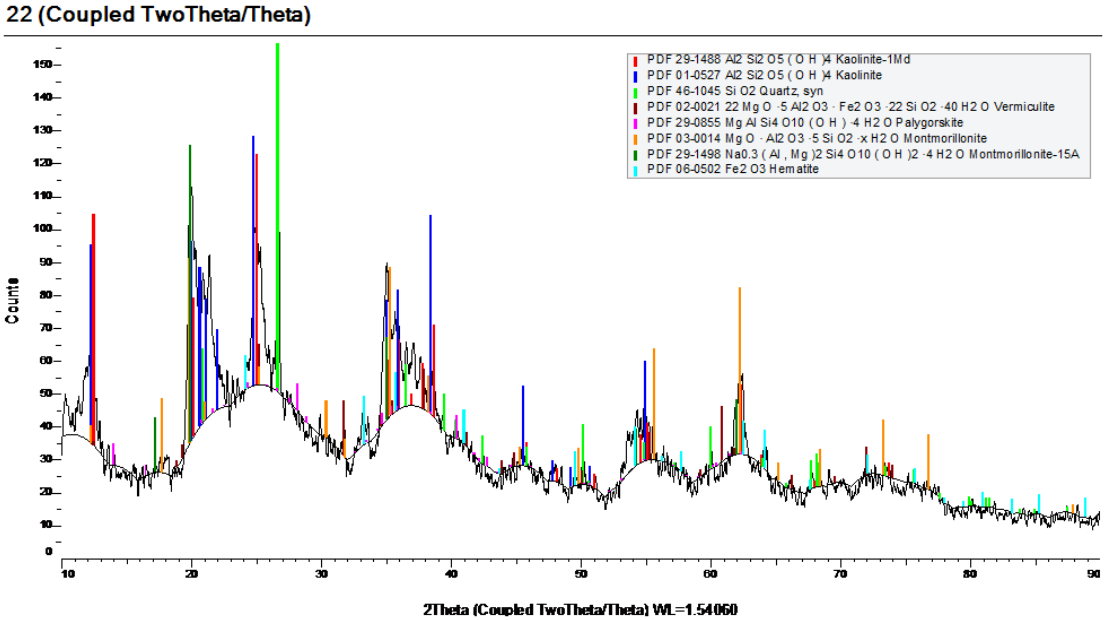
AI2P2: 60-82 cm



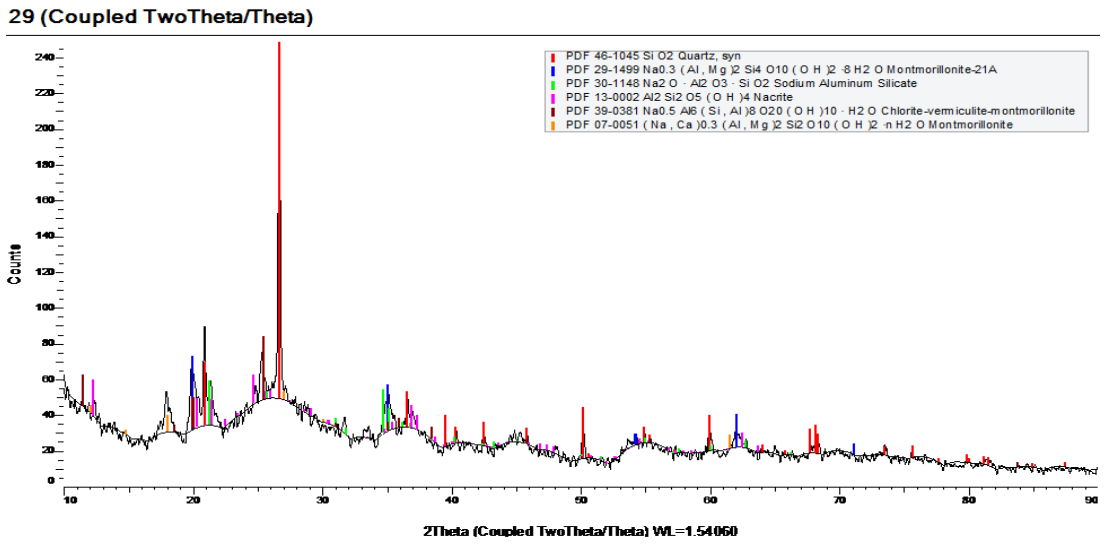
AI2P2: 82-132 cm



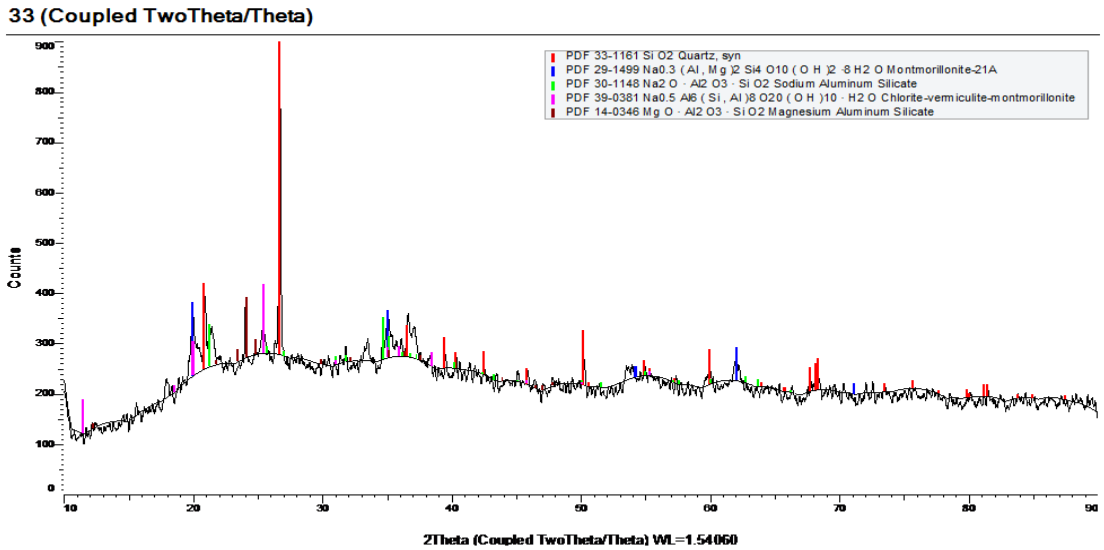
IH1P1: 0-37 cm



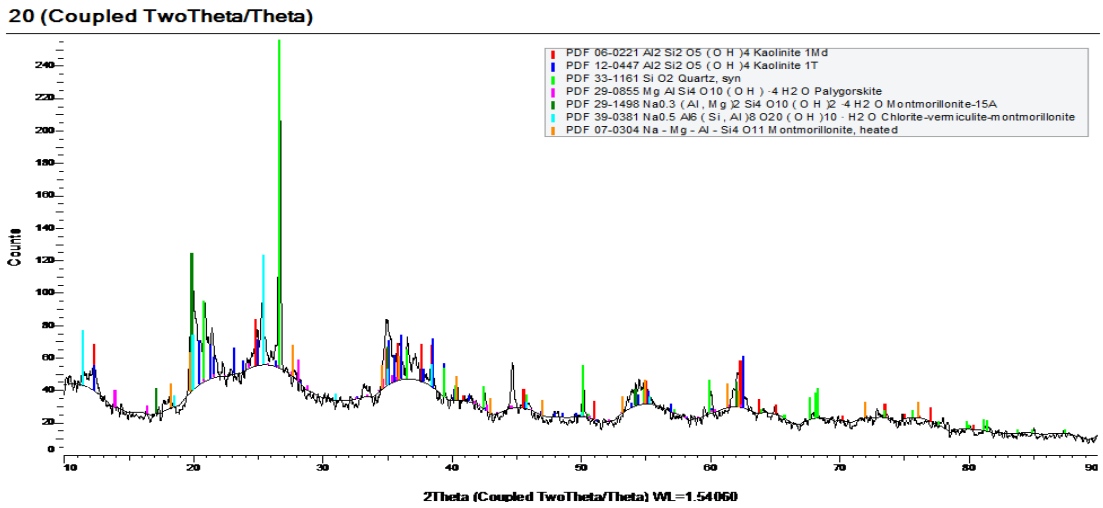
IH1P1: 64-130 cm



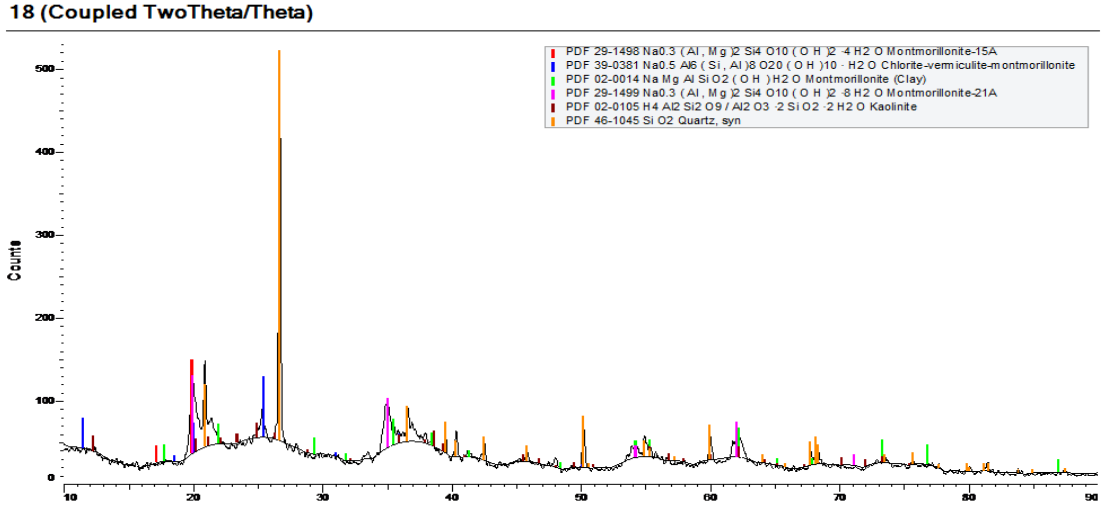
IH1P1: 130-195 cm



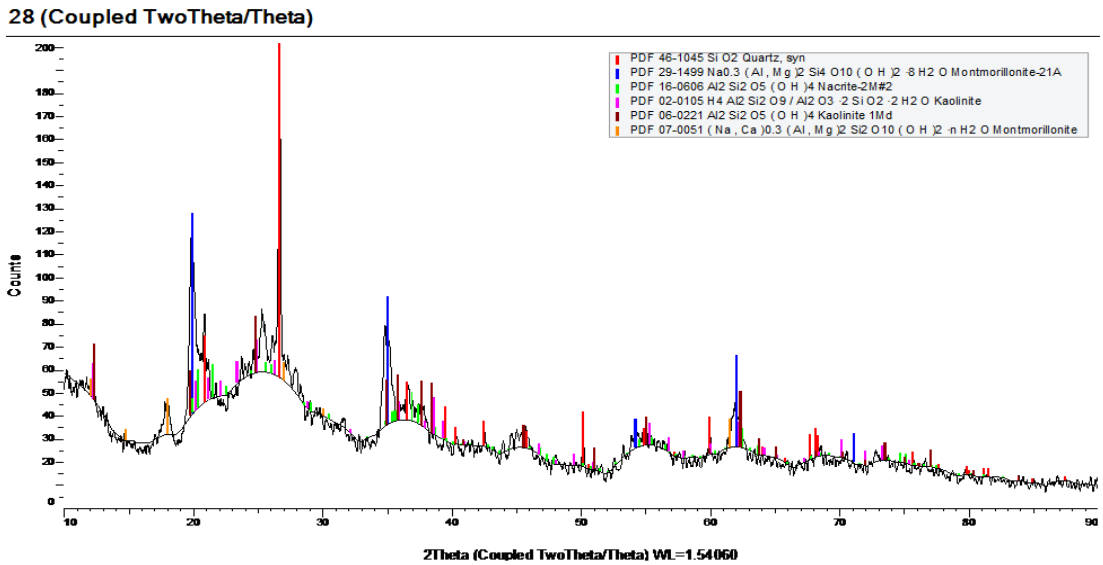
IH1P2: 0-42 cm



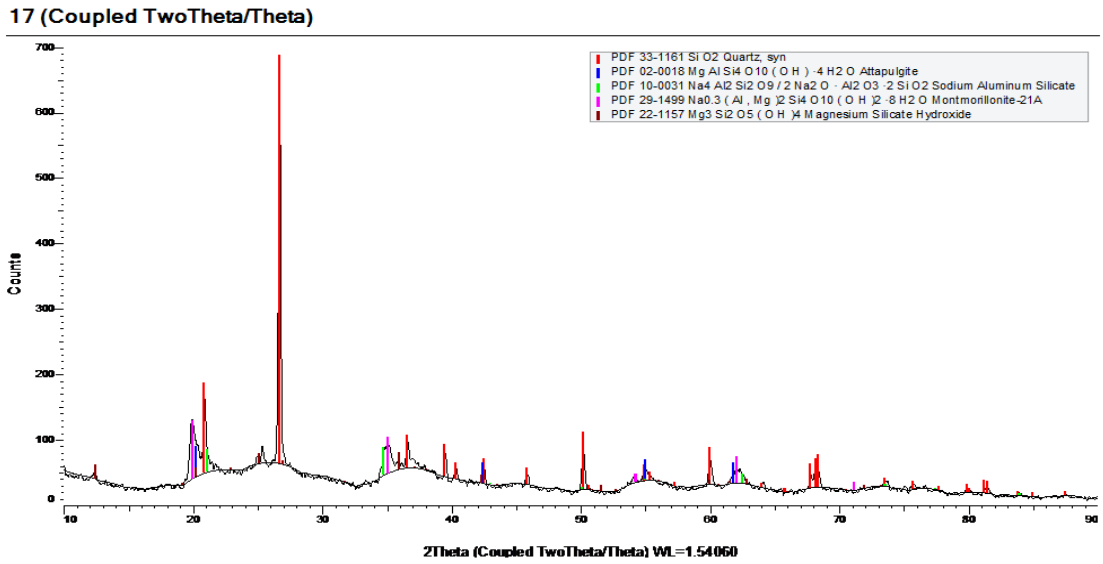
IH1P2: 42-70 cm



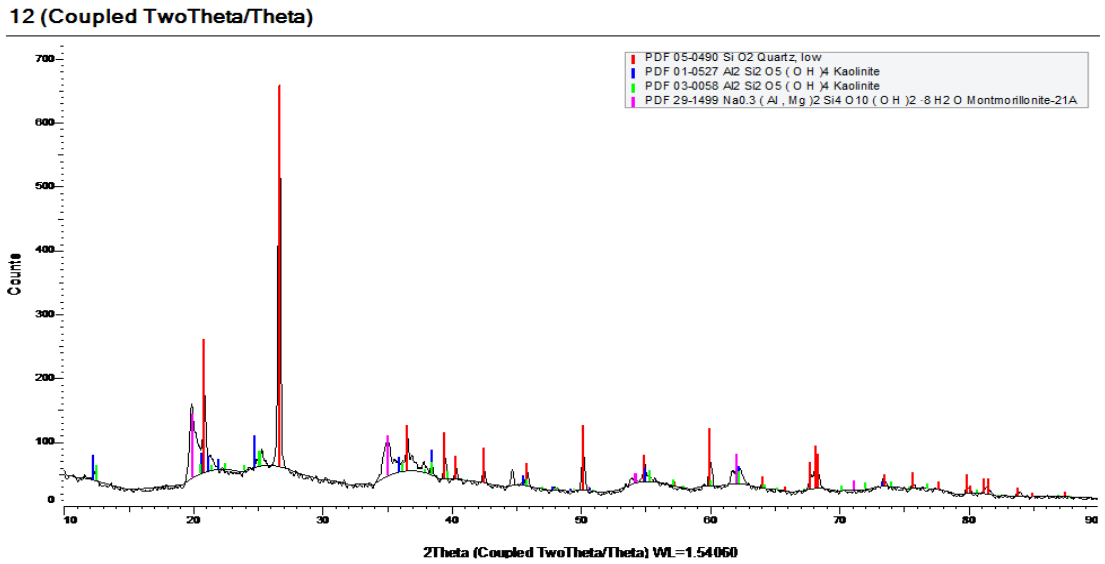
IH1P2: 70-150 cm



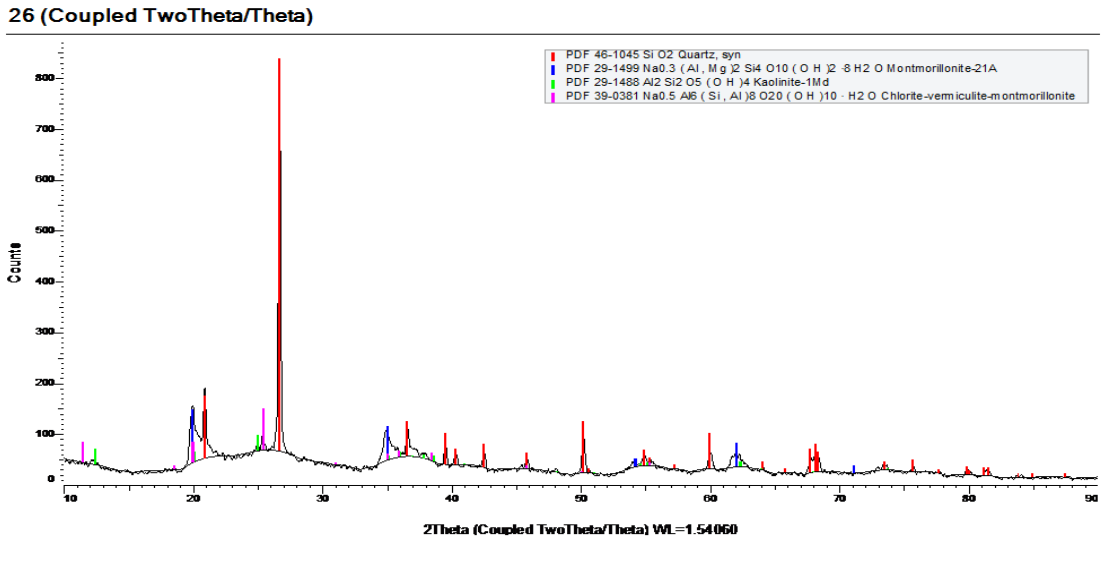
MF1P1: 0-17 cm



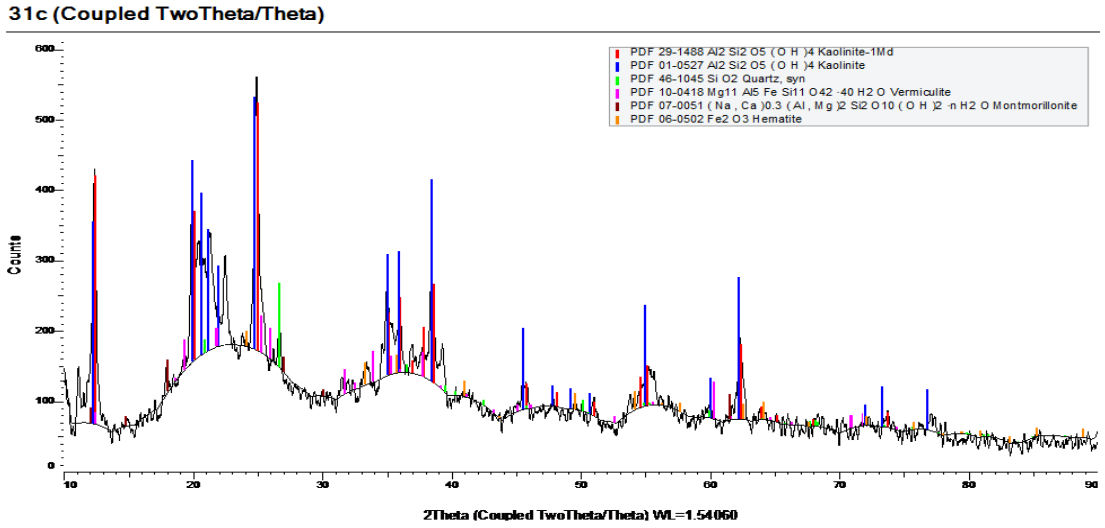
MF1P1: 17-56 cm



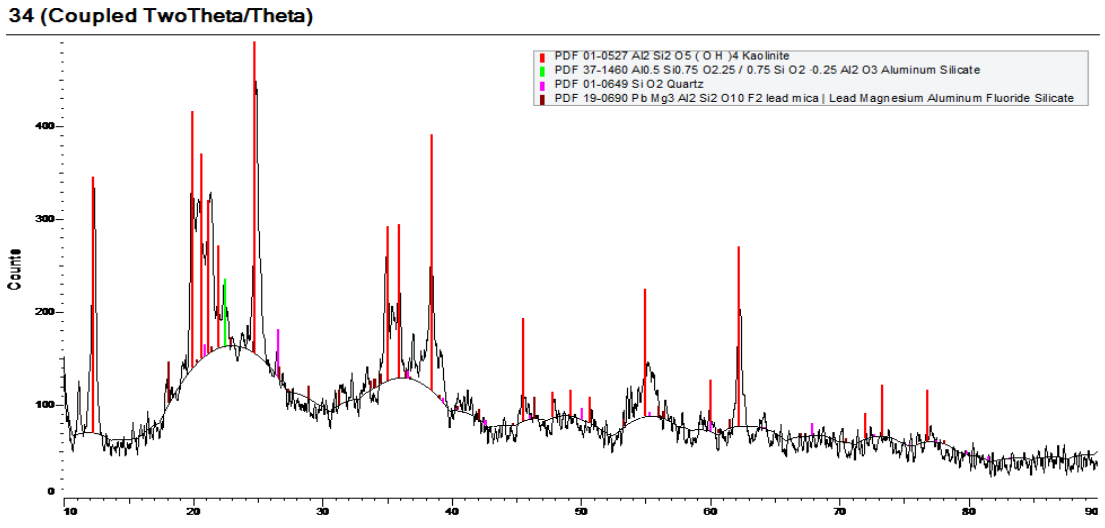
MF1P1: 56-121 cm



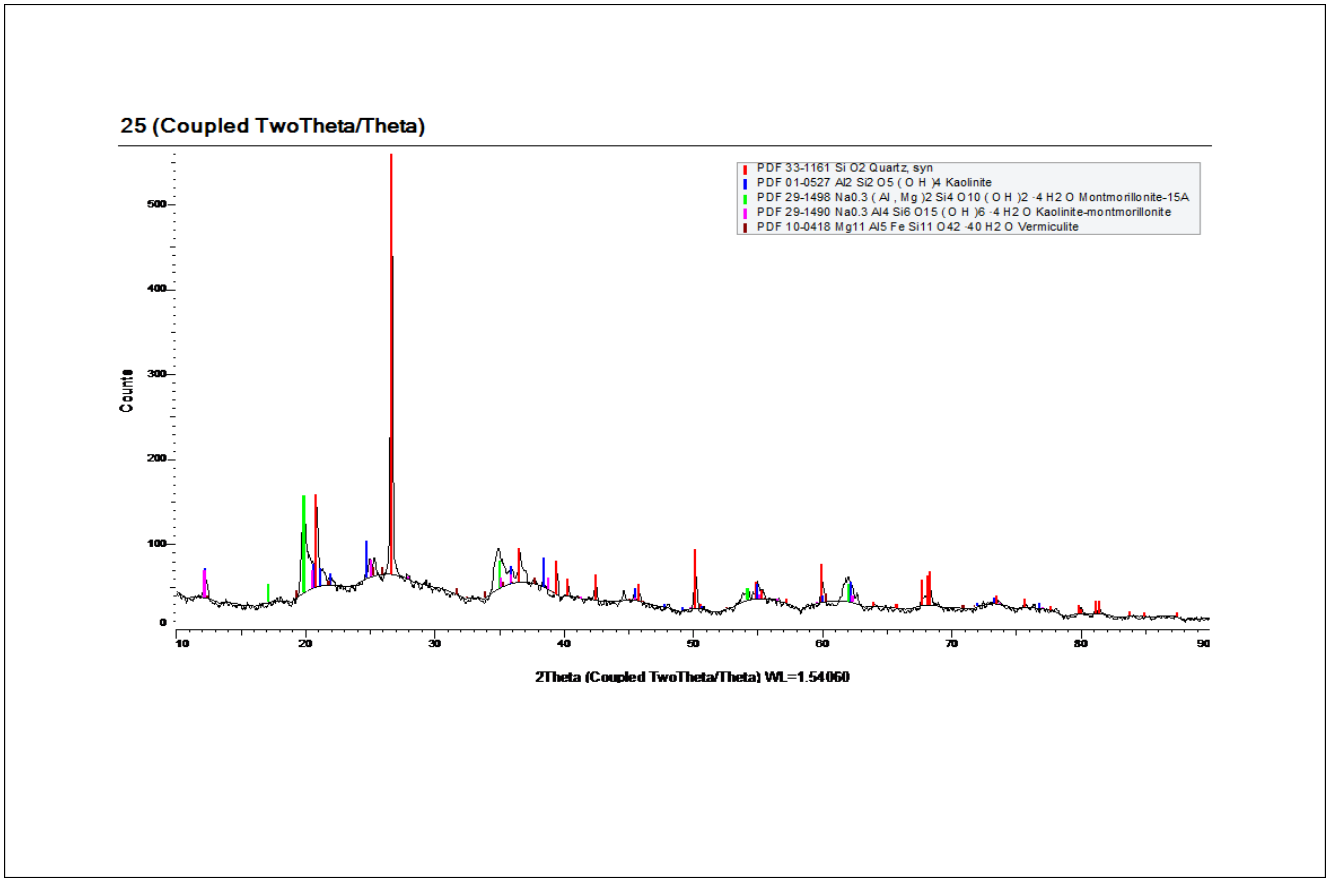
MF1P2: 0-14 cm



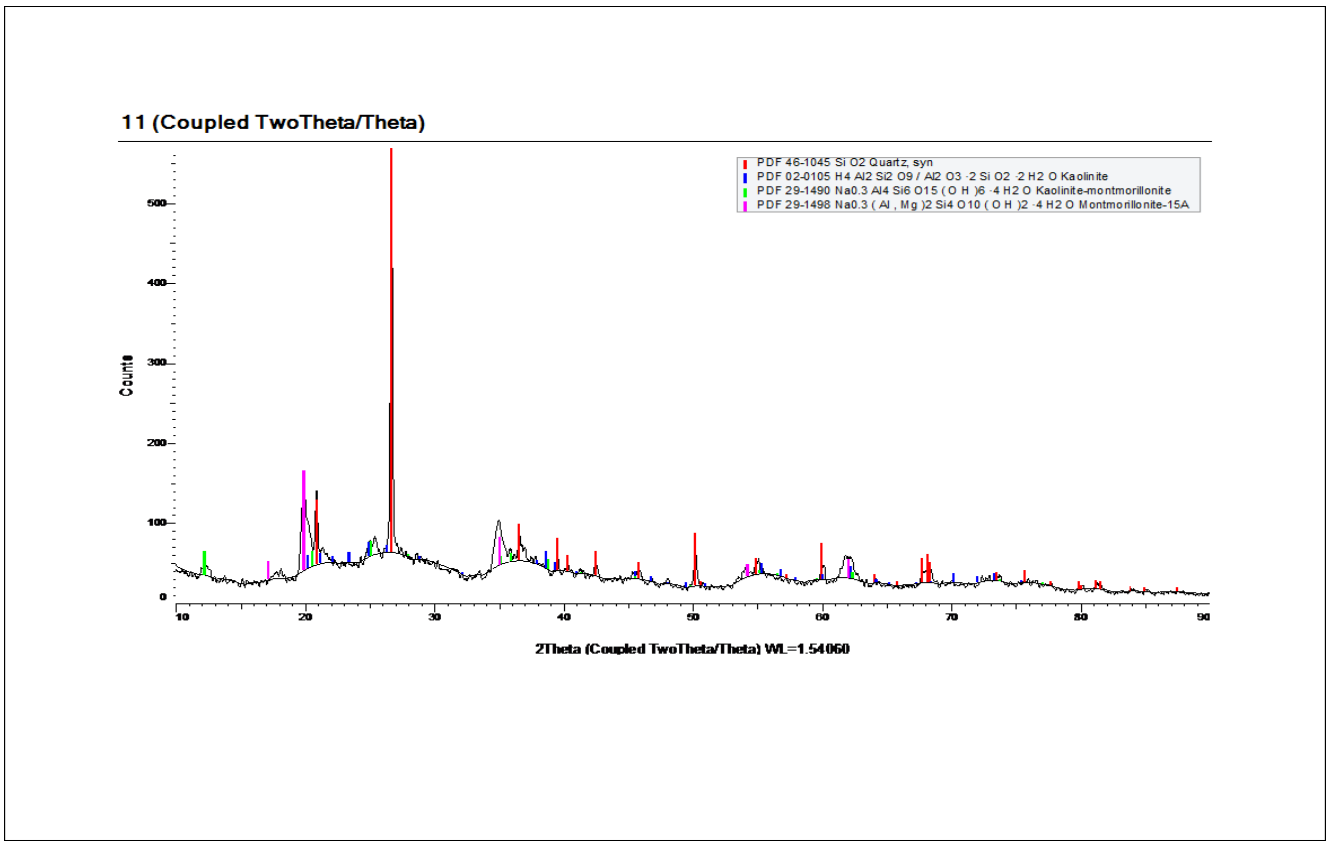
MF1P2: 81-153 cm



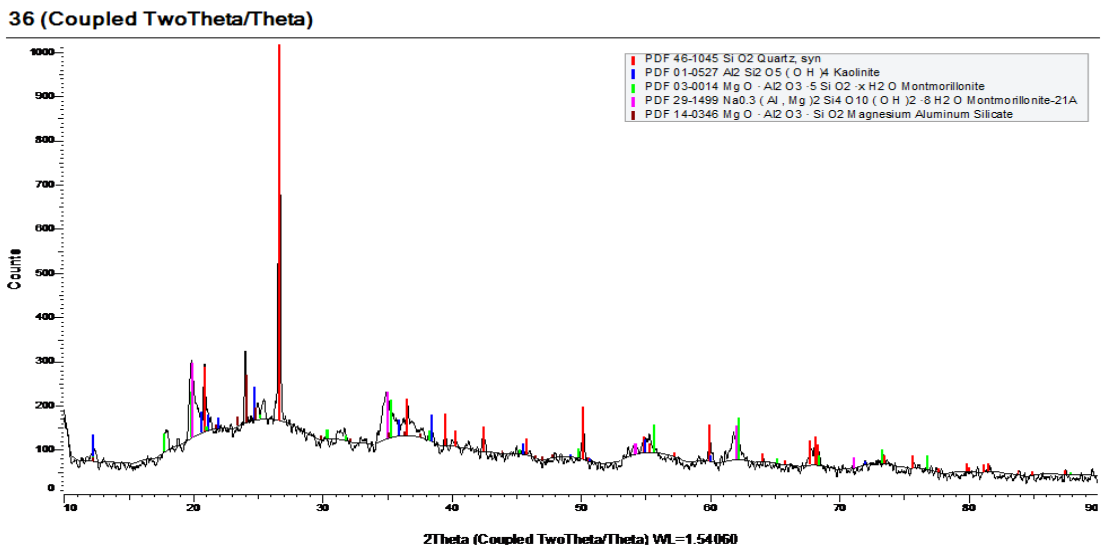
MF2P1: 0-16 cm



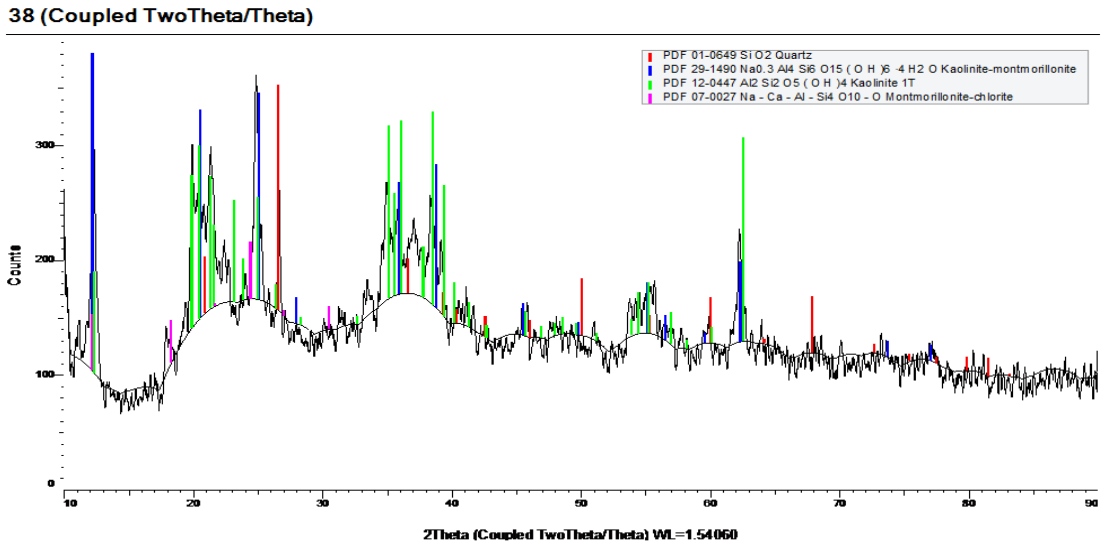
MF2P1: 44-97 cm



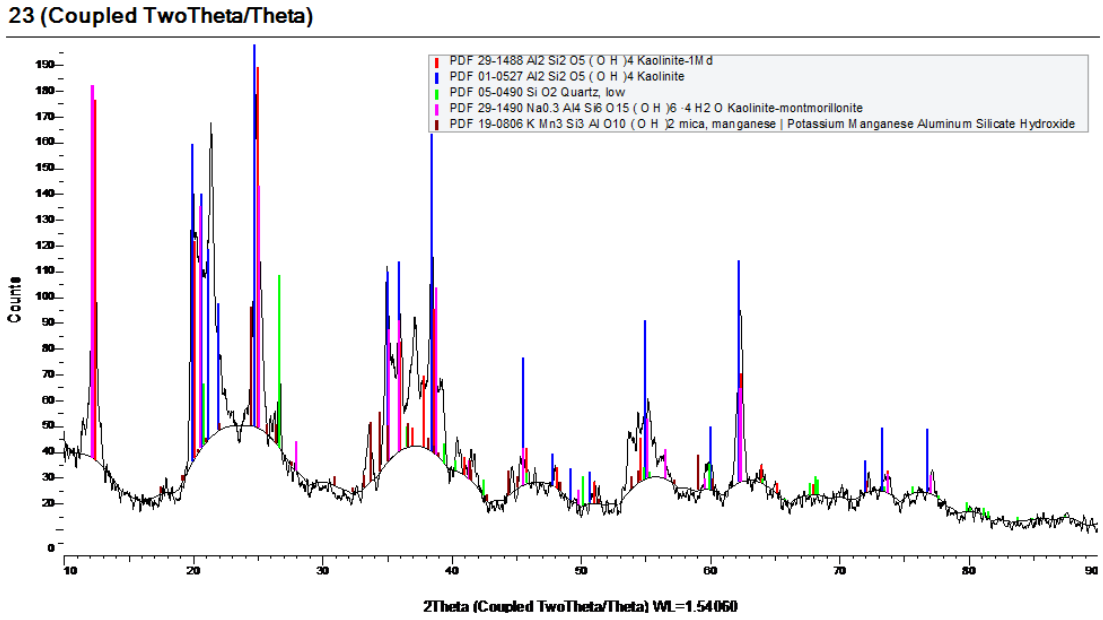
MF2P1: 97-161



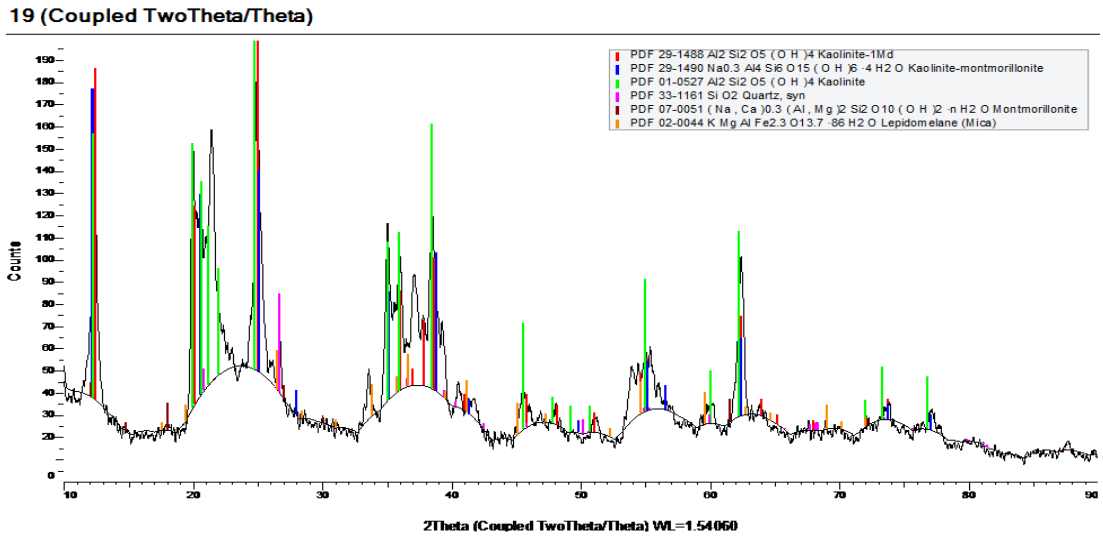
MF2P2: 0-16 cm



MF2P2: 45-107 cm



MF2P2: 107-182 cm



Appendix 6

Rating of Soil Properties

i. Coefficient of variation

Low	0-15 %
Moderate	16-35 %
High	> 36 %

Source: Wilding, 1985

ii. Soil pH

Extremely acid	< 4.5
Very strongly acid	4.5-5.0
Strongly acid	5.1-5.5
Moderately acid	5.6-6.0
Slightly acid	6.1-6.5
Neutral	6.6-7.3
Slightly alkaline	7.4-7.8
Moderately alkaline	7.9-8.4
Strongly alkaline	8.5-9.0
Very strongly alkaline	> 9.0

iii. soil organic carbon

Very low	< 4 g/kg
Low	4-10 g/kg
Medium	10-15 g/kg
High	15-20 g/kg
Very high	> 20

iv. Total nitrogen

Very low	< 1 g/kg
Low	1-2 g/kg
Medium	2-5 g/kg
High	5-10 g/kg
Very high	> 10 g/kg

v. Available phosphorus

Low	< 8 mg/kg
Medium	8-20 mg/kg
High	> 20 mg/kg

vi. Exchangeable cations (cmol/kg)

	Ca	Mg	K	Na
Very low	< 2	< 0.3	< 0.2	< 0.1
Low	2-5	0.3-1.0	0.2-0.3	0.1-0.3
Moderate	4-10	1-3	0.3-0.6	0.3-0.7
High	10-20	3-8	0.6-1.2	0.7-2.0
Very high	> 20	> 8.0	1.2-2.0	> 2.0

vii. Cation exchange capacity (cmol/kg)

Very low	< 6.0
Low	6.0-12
Moderate	12-25
High	25-40
Very high	> 40

viii. Percentage base saturation (%)

Very low	< 20
Low	20-40
Moderate	40-60
High	60-80
Very high	> 80

Source: Holland *et al*., 1989