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# Effects of environmental factors and soil properties on soil organic carbon stock in a natural dry tropical area of Cameroon

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**Abstract.** Researches carried out on soil organic carbon stock (SOCS) in the Sudano-Sahelian part of Cameroon are very rare. The few existing ones are mostly available in reports and concern in most cases carbon stocks in plant biomass. In order to contribute to the documentation on soils in this part of the country, the present work was designed to evaluate the SOCS in the main soil types and the influence of environmental factors and soil properties on these stocks under the natural dry tropical area of the Sudano-Sahelian zone of Cameroon. The study was undertaken in four sites, including three natural forest reserves (Laf, Zamai, Kosohon) and one national park (Mozogo), located at different latitudes. Two replications were thereafter made, thus, giving rise to three sampling points chosen per site, from 0 to 75 cm depth, for the determination of the SOCS. At each sampling point, soils were sampled using depth increments of 25 cm from the surface. The studied area is covered by Haplic Vertisols, Dystric Arenosols, Dystric Leptosols and Dystric Planosols. T-SOCS content, which refers to a depth of 75 cm, decreases with increasing latitude, with 249±26.26 $Mg.ha^{-1}$ in Vertisols at Laf forest reserve in the low latitude, 199±8.00 $Mg.ha^{-1}$ in Arenosols at Zamai forest reserve, 166±16.63 $Mg.ha^{-1}$ in Leptosols at Kosohon forest reserve and 161±8.88 $Mg.ha^{-1}$ in Planosols at Mozogo national park in the high latitude, regardless of the altitude. No significant correlation was noted between T-SOCS and the altitude. A good correlation was noted between precipitation which decreases with increasing latitude and T-SOCS, indicating the importance of climate in the distribution of T-SOCS in the study area, which directly influence the productivity of the vegetation. More than 60% of the SOCS was stored below the first 25 cm from the soil surface, a peculiarity of SOCS in the drylands. The SOCS in the Sudano-Sahelian area of Cameroon is mainly influenced by climate and vegetation.

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## 1 Introduction

Soil is the largest carbon (C) pool and its content in the first 100 cm is estimated at about 1500 Pg (1 petagram $=10^{15}$ g), which represent more C than what is currently contained in the atmosphere and vegetation combined (Boulmane et al., 2010; Lehmann and Kleber, 2015). Their organic C stocks are crucial for a wide range of ecosystem services such as climate regulation through atmospheric $CO_2$ storage (Olson et al., 2016; Greiner et al., 2017; Mayer et al. 2019). As a result of photosynthesis, vegetation fixes about 120 Pg of C per year from the atmosphere and half of it is returned to the atmosphere by plants (Bernoux and Chevalier, 2013). Part of the atmospheric C drawn by plants is stored in biomass and soil in the form of organic matter (Eswaran et al., 1993; Mayer et al. 2019). Terrestrial ecosystems thus have the potential to constitute a sink, slowing down the increase of $CO_2$ in the atmosphere (Mayer et al. 2019). The interaction between the root system and the soil profile has profound impact on soil C accumulation, where the root system can contribute to soil organic carbon (SOC) stocks (Olson and Al-Kaisi, 2015).

Soil organic matter (SOM), which contains more than 50% C weight, plays a fundamental role in the overall behaviour of soils and the ecosystems they support, especially the physical qualities of soils, the stimulation of biological activity of soil, the storage and provision of water and nutrients for plants and the regulation of pollutants (Haygarth and Ritz, 2009; Fernandez et al., 2009; Liu et al., 2012). It is generally considered as a primary indicator of soil quality for their agricultural and environmental functions (Plaza et al., 2018). Thus, loss of organic C or SOM results in the loss of soil quality and impaired associated functions including soil degradation, decline in agronomic productivity, food insecurity, malnutrition and starvation (Haygarth and Ritz, 2009; Brown and Huggins, 2012; Olson and Al-Kaisi, 2015; Olson et al., 2016; Plaza et al., 2018). Increasing the organic C or SOM, directly improves the quality of the soil, hence contributing to the resilience and sustainability of agriculture and consequently to the food security of societies as C is sequestered. But nowadays, many phenomena such as climate change, land degradation and loss of biodiversity make soils and their C storage one of the most vulnerable resources (Bardgett, 2005; Maestre et al., 2013; Soleimani et al., 2019). Currently, the storage of SOC is a topic of paramount importance in international negotiations to fight against climate change through a reduction in greenhouse gas emissions that contribute to global warming (FAO, 2017). Despite the importance of soils in C storage, decisions in the political sphere in terms of climate change mitigation have long focused on the industrial, transport and energy sectors only. The impact of forestry and agricultural activities on C sequestration has been neglected, making agriculture and SOC the poor parents of international negotiations (Bernoux and Chevalier, 2013). It was after the 2008 and 2009 food prices crises and hunger riots in Africa that international debates focused on the soil issue. SOC is now at the centre of global environmental issues, especially in the framework of the United Nations Agreements on Climate Change, Convention on Biological Diversity and the United Nations Convention on Desertification (Bernoux and Chevalier, 2013).

In dry areas, the SOC content is naturally low (less than 1% of the soil mass) whereas in the temperate zone it reaches 4 to 5% in grassland soils or under forest (Bernoux and Chevalier, 2013). This low SOC content in dry regions is due to several anthropogenic and natural factors that led to soil C losses (Lal et al., 2004; Pineiro et al., 2010; Bernoux and

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Chevalier, 2013; Wang et al., 2020). Anthropogenic factors include overgrazing and deforestation (Lal., et al., 2004; Plaza et al., 2018; Soleimani et al., 2019; Wang et al., 2020). Natural factors are poor weather conditions characterized by low rainfall and high temperatures, disappearance of vegetation, aggressive rains, erosion and high mineralization (Von Lutzow et al., 2006; Olson et al., 2016; Wang et al., 2020). Despite this low C concentration in dryland soils, it is important to study the C stock in this area for several reasons. Soils in dry areas cover about 41% of the global land area, and their C stock deserves to be studied for good environmental management and sustainability of the agroecosystem (Haygarth and Ritz, 2009; Lal et al., 2009; Hounkpatin et al., 2018). Moreover, their stock deserves to be studied because of their permanence, i.e. the duration during which C is stored in their soils in relation to wetland soils and if these soils are treated sustainably. They have the potential to sequester large amounts of C, which contributes to the adaptation and mitigation of climate change (Plaza et al., 2018). It is known that increasing C content in terrestrial ecosystems is one of the main approaches to mitigate anthropogenic production of atmospheric $CO_2$ (Jiang et al., 2019). However, SOC storage and dynamics depend on region, parent material, time, cover vegetation, topography and soil properties such as texture and the cation exchange capacity (Munoz-Rojas et al. 2012; Jiménez-González et al., 2020; Reyna-Bowen et al., 2020). Therefore, local studies are necessary to appraise the potential of soil to store carbon properly (Reyna-Bowen et al., 2020).

In Cameroon, the few existing works were mainly achieved in the humid tropical part of the country (Amougou et al., 2016; Tsozué et al., 2019). Studies on the Sudano-Sahelian part of the country are scare and the few existing ones are mainly available in reports or are more concerned with C stocks in plant biomass (Ibrahim and Habib 2008; Tchobsala et al., 2014, 2016). Moreover, many studies have ignored deeper SOC in the world (Ruiz Sinoda et al, 2012; Tornquist et al., 2009). The present work aims to evaluate the SOC stock (SOCS) in the main soil types under the natural dry tropical area in the Sudano-Sahelian zone of Cameroon. More specifically, it focused on: (i) the identification of the main soil types in the natural ecosystems that constitute the study area, (ii) the assessment of the SOCS in each soil type, (iii) the influence of soil depth in the storage of SOC and (iv) the evaluation of the influence of environmental factors and soil properties on these stocks. To achieve this goal, four sites were chosen to conduct the study in the Far North region of Cameroon, namely Laf, Zamai, Kosohon, and Mozogo.

## 2 Material and methods

### 2.1 Study area

The study was conducted in the Far North region of Cameroon (10°-13°N; 13°-15°E) which covers an area of about 3424600 ha (Fig. 1). The climate is dry tropical in nature, characterized by two highly contrasted seasons with a long dry season from October to May and a short rainy season from June to September. The mean annual rainfall and temperature are 800-1000 mm and 26-28°C, respectively. The geologic substratum consists of Precambrian formations (gneiss, migmatite, micaschist, granite, syenite), volcanic formations (basalts, trachyte, rhyolite) and alluvial deposits (Ngounouno et al., 2000; Tamen et al., 2015; Gountié Dedzo et al., 2019). Two morphological units are distinguished in the study area:

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mountain (600-1000 m as an average altitude) and plain (400-600 m as an average altitude). The vegetation of the area belongs to the sudano-sahelian domain with woody arboreous or scrubby vegetation with abundant perennial grasses. The characteristic species are summarized in Table 1. However, a major part of the natural vegetation has been degraded and the soil subjected to cultivation and fallowing. Natural vegetation only exists now in dry forest reserves and national parks.

### 2.2 Experimental design and sampling

The study was undertaken in four sites, including three natural forest reserves (Laf, Zamai, Kosohon) and one national park (Mozogo), located along a latitudinal gradient and corresponding respectively to low density shrub savannah, herbaceous savannah, moderately densified shrub savannah and wooded savannah (Figs. 1 and 2). The characteristics of the studied sites are summarized in Table 1. After the description of the physical characteristics of the studied sites (relief, geologic substrate, vegetation), soil classification was done in the IUSS Working Group WRB (2015) and results are presented in Table 2. SOCS was thereafter evaluated at 75 cm depth since SOC in dryland is believed to reside in topsoil (Ciails et al., 2011; Plaza et al., 2018). Two replicates were made, giving rise to three sampling points chosen per site, separated from each other by a minimum distance of 100 m, for the determination of the SOCS. The soil samples (about 500g of disturbed samples and undisturbed clods) were taken from three soil sections of 25 cm (S1: 0-25 cm, S2: 25-50 cm and S3: 50-75 cm), corresponding to a total of 36 disturbed samples and 36 undisturbed clods, in order to establish a better comparison between the different study sites. Undisturbed clods were taken in the same vicinity as disturbed samples. They were collected for bulk density analysis. Clods were roughly first-sized in order to fit into the cell (8 x 6 x 6 cm) of the clod box.

### 2.3 Laboratory analysis

In the laboratory, after sieving the soil through 2 mm meshes, physical and chemical analyses were carried out. Bulk density was measured by paraffin-coated clods method (USDA, 2004). The particle-size distribution was analyzed by sieving and using Robinson's pipette method (USDA, 2004). Soil pH was measured with pH meter equipped with a glass electrode in 1:2.5 soil-water suspensions (Guitián and Carballas, 1976). Exchangeable cations were extracted by 1 N $NH_4OAc$ at pH 7 and their concentrations were determined by atomic absorption spectrometry Perkin Elmer for Ca and Mg, and by flame emission spectrometry for K and Na. Cation exchange capacity (CEC) was also determined using the ammonium acetate method at pH 7, by a direct continuation using a 1 N potassium chloride (KCl) saturation solution. SOC was determined by dichromate oxidation using the Walkley-Black method (Walkley and Black, 1934). SOM content was obtained using a conversion factor of 1.724 (Walkley and Black, 1934). Total Nitrogen (N) was determined with the Kjeldahl method (Bremner, 1996) and the C:N ratio calculated by dividing the SOC concentration by the N concentration. Available phosphorus was extracted from the soil using Bray-2 solution as extractant (Bray and Kurtz, 1945).

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SOC stock (SOCS), expressed for a specific depth in Mg $ha^{-1}$, was computed as the product of SOC concentration, bulk density, depth and gravels, according to IPCC (2003) as follows: SOCS = (SOC concentration x BD x d x (1 - $\delta_{2mm}$%) x 10), where SOC is the organic C content (g $kg^{-1}$), d the thickness of the control section (m), $\delta_{2mm}$ is the ratio of gravel larger than 2mm in size in the soil, and BD the soil bulk density (Mg $m^{-3}$). Total SOC stock (T-SOCS), refer to 75 cm of depth (Mg.$ha^{-1}$), was calculated according to IPCC (2003) as follows: T-SOCS = $\sum_{\text{soil section}}$ SOC $\text{Stock}_{\text{soil section}}$.

### 2.4 Statistical analysis

The normality of the distribution was studied by the Anderson-Darling normality test. Descriptive statistics (mean values, standard deviations and correlation) were used to characterize the general trends of soil properties, using Excel 2007. A Spearman rank correlation coefficient was carried out in order to assess the possible connection between the soil properties and SOC in each site. Kruskal-Wallis statistical test was used to identify the statistical significance of the differences in each soil variable, among each sampling point within the site and among the sites. Significance was considered at $p<0.05$. All analyses were performed using XSLSAT 2008.6.03 software for Excel.

## 3 Results

### 3.1 Physical characteristics of the studied soils

According to the International Soil Classification System, soils of the study area show a sandy clay loamy (SCL) to sandy loamy (SL) texture. Sand is the most abundant fraction. The highest sand contents were observed in the 0-25 cm depth interval in Vertisols (71±12.12 %), Leptosols (71.33±2.88 %) and Arenosols (76.66±3.21%). It decreased from the soil surface to the base of soil in the 50-75 cm intervals (Table 3). In the Planosols however, high sand content (79±6.42 %) was observed at the middle 25-50 cm interval, in line with the classification of the studied soil. Clay content increased from the surface to the bottom of the soil, except in the Planosols where the low clay content (15±6.42%) was noted in the middle part of the soil, in the 25-50 cm interval (Table 3). Silt contents were low, ranging between 4.0±1.73 and 9.3±2.51%. Bulk density (Bd) values increased with depth except in Vertisols where similar values were observed in the two upper sections. Similarly, in Planosols, as for the sand content, the high value of bulk density (1.68±0.11 Mg $m^{-3}$) was observed in the 25-50 cm interval. With regard to soil pH values, Arenosols, Leptosols and Planosols were acidic (4.7±0.47 to 5.6±0.4), and slightly acid to neutral in Vertisols (6.8±0.59 to 7.9±0.84). However, a very acid value of 4.7±0.47 was noted in the 25-50 cm interval of Planosols (Table 3). Gravel content in the studied soil was globally low (< 3%). A value of 7% was however obtained in the Vertisols, in line with the presence of calcareous nodules in these soils. Generally, there was no significant difference between main soil physical parameters within each site and between sites.

### 3.2 Soil organic matter, nitrogen and C:N ratio

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The average SOM values varied between 21.7±6.10 and 37.9±7.10 g $kg^{-1}$ (Table 3). The average SOM values ranged between 26.50±5.84 and 37.88±7.10 $g.kg^{-1}$ in Vertisols and Arenosols and between 21.70±6.10 and 27.95±2.55 $g.kg^{-1}$ in Leptosols and Planosols. The SOM content decreased with soil depth at Zamai in Arenosols and at Kosohon in Leptosols. In Vertisols at Laf, there was a decrease of SOM from the soil surface in the 0-25 cm interval to 25-50 cm depth and an increase from this depth to the base of soils in the 50-75 cm interval, whereas in Planosols SOM exhibited an opposite trend (Table 3). Total nitrogen (TN) contents were low. TN decreased with soil depth in Vertisols from 0.39 g $kg^{-1}$ in the 0-25 cm interval to 0.21 g $kg^{-1}$ in the 50-75 cm interval at the base of the soil profile. In Arenosols, there was a decrease of TN content from the soils surface to the subsurface of the soil and an increase to the base of the soil. TN content in Leptosols and Planosols exhibited an opposite trend to TN contents in Arenosols. The C:N ratios ranged between 31±15.82 and 112±28.16 suggesting an absence of reactive and readily biodegradable SOM. Beside the high value of C:N ratio in the 25-50 cm interval of Arenosols (112±28.16), the highest ratios were observed in Vertisols while the lowest rations were noted in Planosols (Table 3). A significant negative correlation was observed between SOM and silt fraction (-0.70, $p$ <0.05) (Table 4).

### 3.3 SOC and SOCS

The highest SOC content in the study area was observed in the 0-25 cm interval in Vertisols (21.97±4.12 $g.kg^{-1}$). The lowest one was also observed in the same interval, but in Planosols (12.59±3.53 $g.kg^{-1}$) (Table 5). In fact, beside Arensols where a regular decrease of SOC with depth was observed, a zigzag evolution with depth was observed for the other three soil types, with the trend in vertisols opposing to that observed in Leptosols and Planosols (Table 5). The T-SOC was high in Vertisols and Arenosols (53.75±3.25 and 54.61±2.71 $g.kg^{-1}$ respectively) and low in Leptosols and Planosols (41.21±2.89 and 42.03±3.33 $g.kg^{-1}$ respectively). With respect to SOCS, a high content was observed in the 0-25 cm interval in Vertisols (102.04±30.59 $Mg.ha^{-1}$). This content decreased with depth reaching 70.99±21.68 $Mg.ha^{-1}$ in 25-50 cm interval and increased thereafter reaching 75.57±26.50 $Mg.ha^{-1}$ in the 50-75 cm interval (Table 5). This trend was similar in Arenosols and Leptosols, but differed from that of Planosols where SOCS content increased from the soil surface to the middle part of the soil (47.03±4.73 to 61.82±1.93 $Mg.ha^{-1}$) and decreased thereafter reaching 52.10±19.98 $Mg.ha^{-1}$ in the 50-75 cm at the base of the soil. The highest T-SOCS was obtained in Vertisols at low latitude while the lowest was obtained in Planosols at high latitude. It appears clearly that T-SOCS content decreased with increasing latitude, with 248.60±26.26 $Mg.ha^{-1}$ in Vertisols at Laf, 199.04±8.00 $Mg.ha^{-1}$ in Arenosols at Zamai, 166.52±16.63 $Mg.ha^{-1}$ in Leptosols at Kosohon and 160.95±8.88 $Mg.ha^{-1}$ in Planosols at Mozogo (Table 4 and Fig. 1), irrespective of the altitude. In fact, a very weak correlation was noted between the altitudinal gradient and T-SOCS ($R^2$=0.23), meaning that there was no influence of the altitudinal gradient on the distribution of T-SOCS in the studied area (Fig. 3). A good correlation was noted on contrary between precipitation which decreased with increasing latitude and T-SOCS (r=0.84), indicating the importance of the

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climate in the distribution of T-SOCS in the study area (Fig. 4). Globally, 58.94%, 57.87%, 64.23% and 70.77% of SOC respectively in Vertisols, Arenosols, Leptosols and Planosols was stored below the first 25 cm from the soil surface.

## 4 Discussion

### 4.1 Soil physical properties

Clay content increased from the surface to the base of the soil in Vertisols and Arenosols. This might be due to eluviation and illuviation processes as consequences of vertical movement of water through the soil profile. The clay content was almost similar in the three increments in Leptosols, in line with the weak development and differentiation of the soil. Contrastingly in Planosols, the presence of rusty and greyish spots in addition to the abrupt variation of sand and clay fractions contents had already been reported elsewhere (Driessen et al., 2001; Van Ranst et al., 2011). Bulk density is high and specially in Vertisols. Similar values were obtained by Azinwi et al. (2011). This might be due to the sudano-sahelian climate with eight to nine months of dry season which lead to soil compaction and sealing (Tsozué et al., 2014). Bulk density increases with soil depth in Vertisols and Arenosols in line with increase of clay fraction. In Planosols, the evolution of bulk density with depth follows that of sand. Bulk density is known to vary under the influence of particle size distributions in a soil layer (Carter, 1990; Tuttle et al., 1988). The increase of bulk density with soil depth regardless of the texture in Leptosols might be linked to the progressive weathering of the parent rock. High pH values are noted in Vertisols. The slightly acid to neutral nature of studied Vertisols is unfailingly attributed to the presence of free calcium from calcareous nodules or base-rich parent materials from which they were developed. Similar results were obtained by Nguetnkam (2004) in the far North region of Cameroon and Moustakas (2012) in the North Eastern of Greece. Low pH in the bleached layer of Planosols might be due to ferrolysis processes which have been suggested to explain oxides (Mn-Fe) segregations and gleying in such environments (Singh et al., 1998; Barbiero et al., 2010), textural variations (O'Geen et al., 2008; Van Ranst et al., 2011; da Silva et al., 2019). It is common for Planosols to present pH < 5.5 due to acidification and destruction of clay minerals, leaving only quartz as the major constituent (Brinkman, 1970; Spiau and Pedro, 1986; Van Breemen and Buurman, 2002).

### 4.2 Soil organic matter, nitrogen and C:N ratio

The studied soils are the most common under dryland ecosystems (FAO, 2004). Their average SOM values are low, ranging between 21.70±6.10 and 37.88±7.10 $g.kg^{-1}$, compared to those obtained in soils of humid-region soils by Tsozué et al. (2019), which range between 15.10 and 180.50 $g.kg^{-1}$. The low SOM concentrations in the study area might be due to the semi-arid climate and the sandy textures (Gallardo et al., 2000; Gonzalez and Candas, 2004; Parras-Alcántara et al., 2015). This is in agreement with observations already made in dryland soils (Lal, 2004; Plaza et al., 2018). SOM influences almost all physical, chemical, and biological properties and processes in such ecosystems (Lal, 2004, 2009; Adoum et al., 2017). It

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promotes soil aggregation, which improves soil structure, porosity, and moisture-holding capacity, thus reducing the severity of water scarcity and protecting soils from erosion and compaction (Ruiz Sinoga et al., 2012; Yüksek, 2012). Nitrogen (N) is an essential nutrient used in relatively large amounts by all living organisms. Generally, the concentration of N is high in areas where SOM and SOC are high (Sakin, 2012), indicating that N nutrition of plants greatly depends on the maintenance of SOM and SOC level. In the study area, no significant correlation was noted between N and SOM. Except in Planosols where the evolution of SOC with soil depth is modelled on that of N, the absence of correlation between N and SOM implies that in the studied area, N and SOC act independently. The C:N ratios are high, ranging between 31.33±15.82 and 112±28.16. The wide variation of C:N ratios across the studied area might be due to the tropical dry climate, natural vegetation, intrinsic soil properties and soil drainage (Yoh, 2001; Cools et al., 2013; Zinn et al., 2018). Since C:N ratio can reflect the degree of decomposition of SOM, this might indicate that the C fractions in the studied soils are not homogeneous materials, but that they are still composed of various C components with different properties (Chen et al., 2020).

**4.3 SOC and SOCS**

The highest SOC content is observed in the 0-25 cm interval in Vertisols, Arenosols and Leptosols. This is in line with Batjes (1996) who reported that the amount of SOC in the topsoil was higher than in subsoil. SOM could be concentrated and the mechanisms of C mineralization and immobilization are more active in the soil's top 30 cm (Hiederer, 2009). Similar results were also obtained by Reyna-Bowen et al. (2020) in Hinojosa del Duque in Spain where they noted that trees' influence on SOC is strongly related to their roots in the A horizon. In Planosols, there was an increase in SOC content in the bleached layer. This is in agreement with observations made by da Silva et al. (2019) in three Brazilian Planosols. The low SOC content in the 0-25 cm interval of Planosols might be due to the fact that these soils are periodically flooded. In fact, in periodically flooded soils, aerobic periods promote higher microbial activity and more rapid decomposition of SOM, which is also poorly stabilized by sorption in the sandy surface horizons (Vepraskas, 2001; da Siva et al., 2019). In the study area, a high T-SOC value is observed in Vertisols and Arenosols (53.75±3.25 and 54.61±2.71 $g.kg^{-1}$ respectively) and a low one is observed in Leptosols and Planosols (41.21±2.89 and 42.03±3.33 $g.kg^{-1}$ respectively). The T-SOC content in the studied Vertisols (53.75 $g.kg^{-1}$) is 50% of the T-SOC content obtained in the semi-evergreen medium forest Vertisols (95.20 $g.kg^{-1}$), but 200% of the quantity obtained in the flooded low rainforest zone Vertisols (28.60 $g.kg^{-1}$), all in the warm subhumid equatorial climate of Yucatan Peninsula in Mexico characterized by dry winters (Tsao, 2017). It is however, lower than that obtained in Spain by Muñoz-Rojas *et al*. (2015). This difference can be attributed to their richness in clay and the climatic conditions. In fact, climatic conditions in the study area are close to those of Yucatan Peninsula in Mexico (tropical climate), but in Spain, low temperature would slowdown the microbiological activities leading to an increase in SOC contents. Also, wetting and drying cycle plays a crucial role in aggregation and C stabilization in Vertisols (Rahman et al., 2018). On the other hand, the T-SOC content in Arenosols is 54.61±2.71 $g.kg^{-1}$. This SOC content is higher than that found by Batjes (2008) on sandy marine sediments on freely drained Arenosols in the lower Congo. This difference

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can be attributed not only to parental materials but also to drainage conditions (freely drained sediments) that favour the departure of SOM and SOC (Batjes, 2008; Torn et al., 2009). The T-SOC value of 41.21±2.89 $g.kg^{-1}$ in Leptosols here is lower than that obtained in Umbric Leptosols of the temperate zone (52.40 ± 7.05 $g.kg^{-1}$) but higher than the value obtained in Mollic Leptosols under similar climatic conditions in Southern Spain (38.95 ± 6.41 $g.kg^{-1}$) studied by Parras-Alcantara et al. (2015), meaning that climate is not responsible for the variation of SOC contents in Spanish Leptosols.

The trend of SOCS contents with depth in the studied area follows that of SOC, in line with observations of Batjes (1996), Hiederer (2009) and da Silva et al. (2019). SOCS in the upper 30 cm in the studied Vertisols is 96.87 Mg $ha^{-1}$. This value is higher than that obtained in the upper 30 cm of Vertisols (56.4 Mg $ha^{-1}$) by Tornquist et al. (2009) in Brazil. It is also relatively higher than those obtained by Tsao (2017) in different land covers in Yucatan Peninsula in Mexico, with 70.70 Mg $ha^{-1}$ in flooded low rainforest zone Vertisols, 80.03 Mg $ha^{-1}$ in semi-deciduous low rainforest, 88.43 Mg $ha^{-1}$ in Semi-evergreen low rainforest and 154.16 Mg $ha^{-1}$ in semi-evergreen medium forest Vertisols. High T-SOCS in the studied Vertisols might be due to the fact that Vertisols contain smectitic clay minerals, with a high cation exchange capacity dominated by $Ca^{2+}$ ions, which favours SOC stabilization and accumulation (Muneer and Oades, 1989; Tornquist et al., 2009). This is in line with high correlation between CEC and SOC commonly demonstrated worldwide. T-SOCS values are higher in Arenosols and Leptosols than those obtained in Mexico by Tsao (2017), irrespective of altitude. It might also be related to the nature of clay mineral as montmorillonite in Vertisols. They are also higher than the values documented in Europe by De Vos et al. (2015), with 102 Mg $ha^{-1}$ in Arenosols and 134 Mg $ha^{-1}$ in Leptosols. Another comparison could be made with Andosols from the eastern part of the African continent (Jones et al., 2013) and Andosols from the humid mountainous zone of south Cameroon (Tsozué et al., 2019), where locally they present elevated stocks reaching respectively 150 Mg $ha^{-1}$ and 302 Mg $ha^{-1}$. Soluble substances and labile compounds of litter are rapidly degraded in the early stages of decomposition by fast-growing microorganisms that might lead to the loss of SOC. Also, cellulose and lignin, the most abundant components of litter in Europe for example, are decomposed slowly (Fioretto et al., 2005) and might negatively impact the SOC content. In the studied Planosols on the contrary, SOCS was high in the 0-25 cm interval and the T-SOCS value was low in the study area (160.95±8.88 Mg $ha^{-1}$). This T-SOCS value, although low in the study area, was higher than that obtained by Batjes (1996) in the world (77 Mg $ha^{-1}$ with coefficient of variation of 0.56), De Vos et al. (2015) in Europe (67 Mg $ha^{-1}$ with coefficient of variation of 0.35) and Batjes (2002) in Central and Eastern Europe (108 Mg $ha^{-1}$ with coefficient of variation of 0.18). The high T-SOCS value in the studied Planosols might be attributed to the very low slope gradient (2%) and the possible presence of biocrusts which take C and N from the atmosphere and might protect soil from wind and water erosion in the studied area (Plaza et al., 2018). Globally, the obtained T-SOCS values in the studied area are close to those obtained by Adoum et al. (2017) (173.9 to 241.0 Mg $ha^{-1}$) in polder soils under semiarid climate in Lake Chad.

**4.4 Effect of environmental factors on SOCS**

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There was a decrease of SOCS with latitude in the studied area. This agrees with Jenny's (1930) findings, which revealed that the major reservoirs of soil C change with latitude. At low latitudes, very little fraction of the T-SOC is stored in surface detritus and most of the C is in the mineral soil. At high latitudes, slow litter decay leads to large accumulations of detrital organic material, and relatively little part of the organic C is in the mineral soil. This is in agreement with a high T-SOCS content obtained in the Laf Vertisols (248.60±26.26 Mg $ha^{-1}$) at low latitude, and a low T-SOCS obtained in the Mozogo Planosols (160.95±8.88Mg $ha^{-1}$) at high latitude in the studied area. In fact, there is a decrease in total precipitations and an increase in the mean air temperature as one moves away from the equator, and thus with increasing latitude in the tropical zone. This implies that T-SOCS storage responds negatively to increasing temperature and decreasing precipitations along with the latitude gradient. Similar results were obtained by Tang et al. (2020) in China. Climate appears to be the principal factor explaining the low SOC contents in the studied soils (Plaza et al., 2018), with precipitations explaining 71.04% of the variability of T-SOCS ($R^2$= 0.7104). Similar observations were noted by Lozano-García et al. (2017) in the semiarid Mediterranean part of South Spain and Jiménez-González et al. (2020) in the humid subtropical, Mediterranean and oceanic temperate part also of Spain. Moreover, 58.94%, 57.87%, 64.23% and 70.77% of SOC respectively in Vertisols, Arenosols, Leptosols and Planosols are stored below the first 25cm from the soil surface. This is a perculiarity of SOCS under dryland ecosystems. It is essentially climate dependent (Gray et al., 2016, Plaza et al., 2018). Globally, temperature and precipitation negatively and positively (respectively) affected SOCS because they affect the balance between C inputs from plant residues and C outputs caused by microbial decomposition of SOM (Post et al., 1982; Wang et al., 2020).

Spearman correlation between different soil parameters of the study area show that there was no correlation between the different soil parameters and SOM apart from the negative relationship with silt contents. The lack of correlation between SOM and particle size fractions in different soils can be explained by the fact that the alteration in dry intertropical zone is essentially geochemical. The influence of the parent rock in the accumulation of SOM in the study area therefore seems negligible, in line with Plaza et al. (2018) observations. However, the availability of polyvalent cations such as $Ca^{2+}$ from the weathering of a parent rocks in the soil, might be an important factor in the chemical protection of SOC (Briedis et al., 2012). In fact, $Ca^{2+}$ is frequent in the soil solution and precipitates in the form of calcareous nodules at the base of soils in the study area.

The vegetation cover and/or the type of vegetation influences the amount of SOC stored in soils (Torn et al. 2009; Reyna-Bowen et al., 2020). The site of Laf located at low latitude had the highest T-SOCS while the site of Mozogo located at high latitude had the lowest T-SOCS. The difference in T-SOCS contents might be due to the type of vegetation in the various sites. The Laf studied site would therefore have more productive vegetation due to high precipitations and thus provide more SOC than the other sites. This is also valid for the site of Kosohon which, although located at higher altitude (865 m a.s.l.) and therefore likely to have a high T-SOCS, shows a lower T-SOCS than that obtained in Vertisols at Laf. The influence of the vegetation might occur through deep root distribution and the chemical nature of litter (Bird and Torn 2006; Zhang et al., 2008; Torn et al., 2009; Reyna-Bowen et al., 2020). Generally, decomposition of SOM involves multiple microbial processes catalyzed by various enzymes (Zhang et al., 2020). Changes in microbial community and enzyme

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activity are therefore expected to influence the decomposition rate (Chenu et al., 2019; Wang et al., 2020; Zhang et al., 2020). It is demonstrated that soil acidification depresses the decomposition of SOM, both by decreasing microbial activity and by increasing protection of SOC by mineral phases (Zhang et al., 2020). Soil pH decreases with latitude and SOCS, reinforcing thus the fact that the productivity of the vegetation due to precipitations is a main factor controlling the distribution of SOC in the study area.

With respect to topography, T-SOCS contents decrease with increasing latitude, regardless of the altitude. Moreover, no significant correlation was noted between T-SOCS and the altitude. This implies that topography has no impact on the T-SOCS in the studied sudano-sahelian zone of Cameroon.

The role of topography and parent rock in the storage of SOC seems to be negligible in the studied area. The availability of polyvalent cations such as $Ca^{2+}$ is linked to geochemical processes induced by the climate. The distinguishing factors affecting the SOCS are thus the climate and the vegetation. According to Plaza et al. (2018), these two factors are the main soil-forming factors driving the SOM storage and consequently the SOCS in such dryland ecosystems.

## 5 Conclusion

The present work was designed to evaluate the SOCS in the main soil types under the natural dry tropical area in the Sudano-Sahelian zone of Cameroon. The main soil types are Haplic Vertisols, Dystric Arenosols, Dystric Leptosols and Dystric Planosols. The C:N ratios are high, suggesting the absence of reactive and readily biodegradable SOM. Vertisols recorded the highest T-SOCS, followed by Arenosols, Leptosols and Planosols. More than 60% of SOCS are stored below the first 25cm from the soil surface. The T-SOCS is controlled by climate and vegetation, with precipitations explaining 71.04% of the variability of T-SOCS. The influence of the climate is manifested by the decrease in the SOCS as the latitude increases, in line with the increase of the temperature and the decrease of the rainfall as one move away from the equator. The climate affects the primary productivity of the vegetation, which influences the quality and quantity of inputs to the SOM, as attested by the good correlation between precipitation and T-SOCS. Furthermore, there is no influence of altitude gradation in the repartition of T-SOCS in the studied area.

## 6 Author contribution

**Désiré Tsozué:** Conceptualization; Funding acquisition; Investigation; Methodology; Supervision; Validation; Writing-original draft; Writing - review & editing. **Nérine Mabelle Moudjie Noubissie:** Conceptualization; Funding acquisition; Investigation; Methodology; Writing-original draft. **Estelle Lionelle Tamto Mamdem:** Conceptualization; Investigation; Methodology; Writing-original draft. **Simon Djakba Basga:** Conceptualization; Investigation; Methodology; Writing-original draft. **Dieudonné Lucien Bitom Oyono:** Conceptualization; Investigation; Methodology; Visualization; Validation.

## Competing interests

The authors declare that they have no conflict of interest.

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630 **Table 1.** Physical characteristics of the study sites.

| Site | Altitude (m a.s.l.[a]) | Climate | | Slope gradient (%) | Vegetation[b] | | | | | n[c] |
|---|---|---|---|---|---|---|---|---|---|---|
| | | Precipitation (mm) | Temperature (°C) | | Species | stems/ha | Basal area ($m^2$/ha) | Below ground biomass (Kg MS/ha) | Root density/root depth | |
| Laf | 459 | 855 | 27.4 | 2 | *Acacia hockii, Albizia chevaleri, Balanites aegyptiaca, Bauhinia rufescens, Combretum aculeatum, Dichrostachys cinera, Ziziphus mauritiana, Piliostigma reticulatum, Strychnos spinosa, Ximenia americana, Acacia gerrardii, Anogeissus leiocarpus, Acacia senegal, Terminalia avicennoides* | 312 | 152.52 | 1674.54 | Presence of millimeter roots and rootlets exclusively in the 0-45cm interval | 9 |
| Zamai | 608 | 800 | 28.0 | 2 | *Sclerocaria birrea, Anogeisus leiocarpus, Adansonia digitata, Acacia albida, Parkia biglobosa, Balanites eagyptiaca, Ziziphus, Celtis integrifolia, Cassia siamea, Daniella olivera, Khaya senegalensis, Tamarindus indica, Terminalia brownii, Acacia nilotica, Acacia seyal, Acacia atxacanta* | 275 | 126.87 | 2316.76 | Presence of millimeter roots and rootlets in the 0-33cm interval and very rare below | 9 |
| Kosohon | 865 | 802 | 27.5 | 5 | *Isoberlinia doka, Erythrina sigmoidea, Alchornea cordifolia, Trichodesma africanum, Ficus abutilifolia, Faidherbia albida, Ziziphus mauritiana, Ficus dicranostyla, Vitex doniana, Garcinia afzelii, Haematostaphis barteri, Antidesma venosum, Boscia salicifolia, Celtis integrifolia, Ficus polita, Ficus platyphylla, Ficus gnaphalocarpa, Ficus cordata, Aspidotis schimperi. Piliostigma thonningii, Ziziphus mauritiana, Anogeissus leiocarpus, Boswellia dalzielii, Lannea fruticosa* | 1575 | 86.63 | 928.82 | Presence of millimeter roots and rootlets in the 0-17cm interval and very rare below | 9 |
| Mozogo | 475 | 726.2 | 28.7 | 2 | *Acacia ataxacantha, Anogeissus leiocarpus, Balanites aegyptiaca, Boscia senegalensis, Capparis fascicularis, Celtis integrifolia, Grewia bicolor, Stereospermum kunthianum, Opilia celtidifolia, Hexalobus monopetalus, , Grewia barteri, Dalbergia melanoxylon, Tamarindus indica, Cissus quadrangularis, Piliogstima reticulatum, Diospyros mespiliformis, Combretum fragrans* | 2694 | 120.78 | 2528.12 | Presence of millimeter roots and rootlets in the 0-38cm interval and very rare below | 9 |

[a] Meters above sea level.

[b] Letouzey (1985).

[c] number of sample.

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**Table 2.** Main soil characteristics and soil classification of the study area.

| Sites | Soil characteristics | Soil classification |
|---|---|---|
| Laf | In dry season, many large cracks are observed in the soil surface, drawing a set of polygonal figures. Clay contents are more than 30%. Wedge-shaped soil aggregates and slickensides are observed from 9 cm of the soil surface to the base of the soil profile. Calcareous nodules (7%) are present at the base of the soil profile. The CECclay ranges between 39 to 70 cmol(+)/kg. The sum of exchangeable bases is low and base saturation varies between 10.5 and 15.2 %. | Haplic Vertisols |
| Zamai | The soils are dominated by sand fraction, with 75% in the upper 25 cm of the soil profile. The CEC is low, ranging between 7.2 and 11.6 cmol(+)/kg in the upper 100 cm of the soil. Base saturation values range between 22.0 and 31.4%. | Dystric Arenosols |
| Kosohon | The surface humiferous horizon is 17cm thick. It is followed beneath by an underlying weathering continuous rock, characterized by a well preserved bedrock structure. The fine earth (silt+clay) content is about 32%. Base saturation is low, ranging between 8.7 and 10.3%. | Dystric Leptosols |
| Mozogo | The soils are characterized by the presence of rusty and greyish spots in the yellowish red surface horizon, an indurated subsurface horizon between 20 and 38 cm and abrupt textural difference noted in sand contents. There are whitish calcareous nodules (2%) at the base of the soil profile. The CEC varies between 42 and 45 cmol(+)/kg and base saturation is < 5% in all the horizons. | Dystric Planosols. |

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**Table 3.** Main soil physical and chemical properties (Mean ± SD*) in the study area.

| Soil Types | Depth (cm) | Gravel (%) | Sand (%) | Silt (%) | Clay (%) | Texture | OC ($g.kg^{-1}$) | OM ($g.kg^{-1}$) | N ($g.kg^{-1}$) | C :N | pH | BD ($Mg\ m^{-3}$) |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Vertisols (Laf) | 0-25 | 1±0.00a | 71±12.12a | 3.66±1.25a | 25.66±10.78a | S-C-L | 21.97±4.12a | 37.88±7.10a | 0.39±0.12a | 58±14.73a | 7.21±0.89a | 2.18±0.11a |
| | 25-50 | 1±0.00a | 59.33±12.34a | 8±1.0a | 32.66±11.32a | S-L | 15.36±3.39a | 26.50±5.84a | 0.25±0.04a | 61±15.52a | 6.81±0.59a | 2.17±0.20a |
| | 50-75 | 7±0.00a | 58.33±11.84a | 7±1.73a | 34.66±10.11a | S-C-L | 16.42±2.25a | 28.31±3.88a | 0.21±0.06a | 81.33±26.76a | 7.92±0.84a | 2.55±0.18a |
| Arenosols (Zamai) | 0-25 | 1±0.00a | 76.66±3.21a | 4±1.73a | 19±3.0a | S-L | 21.33±2.59a | 36.77±4.46a | 0.38±0.04a | 55.33±15.82a | 5.63±0.40a | 1.55±0.05a |
| | 25-50 | 1±0.00a | 70±6.24a | 4±3.60a | 26±7.21a | S-C-L | 16.85±2.25a | 29.05±3.88a | 0.15±0.01a | 112±28.16a | 5.42±0.28a | 1.56±0.18a |
| | 50-75 | 1±0.00a | 67.66±0.57a | 4.33±1.52a | 28±1.73a | S-C-L | 16.43±3.28a | 28.32±5.65a | 0.31±0.13a | 56.33±15.82a | 5.27±0.06a | 1.89±0.30a |
| Leptosols (Kosohon) | 0-25 | 1±0.00a | 71.33±2.88a | 6.33±3.78a | 22±6.08a | S-C-L | 14.29±3.53a | 24.63±6.12a | 0.30±0.15a | 60.66±42.09a | 5.3±0.56a | 1.67±0.20a |
| | 25-50 | 1±0.00a | 69.33±6.11a | 9.33±2.51a | 21.66±4.04a | S-C-L | 13.44±2.22a | 23.20±3.83a | 0.52±0.29a | 31.33±15.82a | 5.36±0.39a | 1.73±0.10a |
| | 50-75 | 1±0.00a | 67.66±65a | 9±2.0a | 23.33±5.85a | S-C-L | 14.08±2.93a | 24.30±5.05a | 0.24±0.02a | 57.33±7.02a | 5.59±0.29a | 1.98±0.29a |
| Planosols (Mozogo) | 0-25 | 1±0.00a | 64.33±4.04a | 5.33±1.15a | 30±5.29a | S-C-L | 12.59±3.53a | 21.70±6.10a | 0.28±0.04a | 46.33±19.85a | 5.74±0.88a | 1.54±0.12a |
| | 25-50 | 1±0.00a | 78.66±6.42a | 6±0.00a | 15.33±6.42a | S-L | 16.21±1.48a | 27.95±2.55a | 0.49±0.31a | 56±54.58a | 4.71±0.47a | 1.68±0.11a |
| | 50-75 | 2±0.00a | 48.33±31.56a | 7.33±2.30a | 43.66±32.33a | S-C-L | 13.23±4.97a | 22.81±8.60a | 0.44±0.35a | 40.66±28.98a | 5.22±0.67a | 1.58±0.70a |

*: Standard deviation; BD: Bulk density; S-C-L: Sandy clay loam; S-L: Sandy loam. Numbers followed by different lower-case letters within the same column have significant differences ($p < 0.05$) at different depths, considering the same topographic position.

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**Table 4.** Spearman correlation matrix for relationships between selected soil parameters in the study area.

| Variables | Gravel | Sand | Silt | Clay | OM | N | C :N | pH | Bd |
|---|---|---|---|---|---|---|---|---|---|
| Gravel | 1.00 | | | | | | | | |
| Sand | -0.64* | 1.00 | | | | | | | |
| Silt | 0.25 | -0.47 | 1.00 | | | | | | |
| Clay | 0.64* | -0.91* | 0.14 | 1.00 | | | | | |
| OM | -0.13 | 0.47 | -0.71* | -0.23 | 1.00 | | | | |
| N | -0.11 | 0.33 | 0.06 | -0.43 | -0.14 | 1.00 | | | |
| C :N | 0.05 | 0.00 | -0.22 | 0.22 | 0.53 | -0.77* | 1.00 | | |
| pH | 0.12 | -0.27 | -0.13 | 0.31 | 0.30 | -0.50 | 0.43 | 1.00 | |
| BD | 0.24 | -0.25 | 0.28 | 0.20 | 0.28 | -0.12 | 0.43 | 0.40 | 1.00 |

* Significant at $p<0.05$

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**Table 5.** Soil organic carbon (SOC) content and soil organic carbon stock (SOCS) (Mean ± SD*) in the study area.

| Soil types | Sections | Depth cm | SOC g $kg^{-1}$ | T-SOC g $kg^{-1}$ | SOCS Mg $ha^{-1}$ | T-SOCS Mg $ha^{-1}$ |
|---|---|---|---|---|---|---|
| Vertisols (Laf) | S1 | 0-25 | 21.97±4.12a | 53.75±3.25 | 102.04±30.59a | 248.60±26.26 |
| | S2 | 25-50 | 15.36±3.39a | | 70.99±21.68a | |
| | S3 | 50-75 | 16.42±2.25a | | 75.57±26.50a | |
| Arenosols (Zamai) | S1 | 0-25 | 21.33±2.59a | 54.61±2.71 | 83.82±3.45a | 199.04±8.00 |
| | S2 | 25-50 | 16.85±2.25a | | 55.57±10.84a | |
| | S3 | 50-75 | 16.43±3.28a | | 59.65±9.71a | |
| Leptosols (Kosohon) | S1 | 0-25 | 14.29±3.53a | 41.21±2.89 | 59.54±15.99a | 166.52±16.63 |
| | S2 | 25-50 | 13.44±2.22a | | 51.30±13.60a | |
| | S3 | 50-75 | 14.08±2.93a | | 55.68±20.29a | |
| Planosols (Mozogo) | S1 | 0-25 | 12.59±3.53a | 42.03±3.33 | 47.03±4.73a | 160.95±8.88 |
| | S2 | 25-50 | 16.21±1.48a | | 61.82±1.93a | |
| | S3 | 50-75 | 13.23±4.97a | | 52.10±19.98a | |

*: Standard deviation. Numbers followed by different lower-case letters within the same column have significant differences ($p < 0.05$) at different depths, considering the same topographic position.

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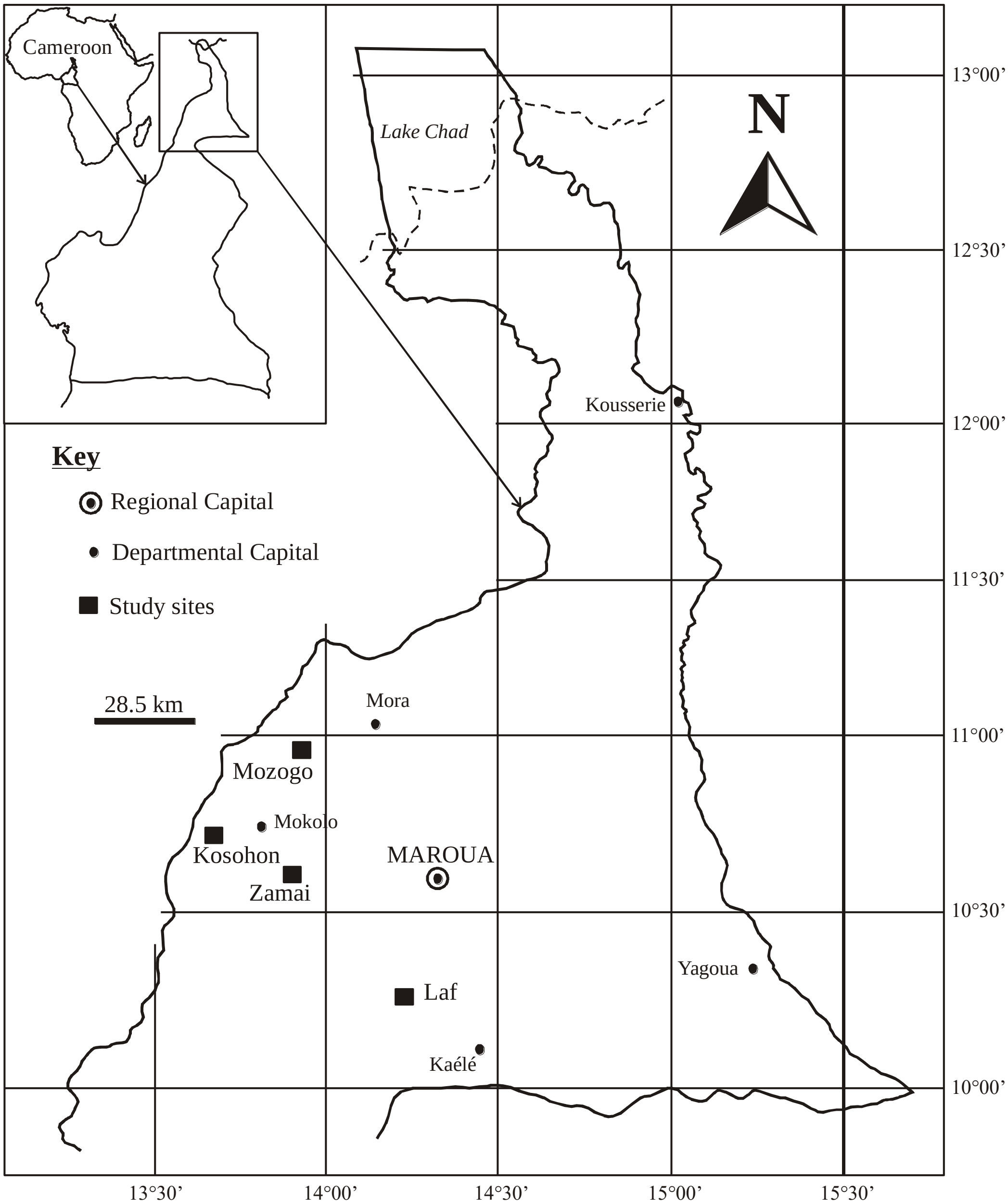


Fig. 1. Location of the study sites.

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Fig. 2. From left to right clockwise: Low density shrub savannah (Laf), herbaceous savannah (Zamai), moderately densified shrub savannah (Kosohon) and wooded savannah (Mozogo).

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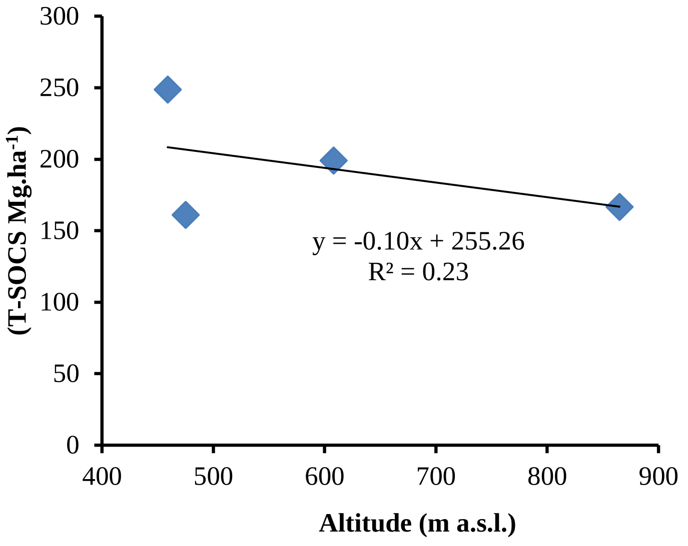


705 Fig. 3. Plots of the total soil organic carbon stock (T-SOCS) versus altitudinal gradient in the study area.

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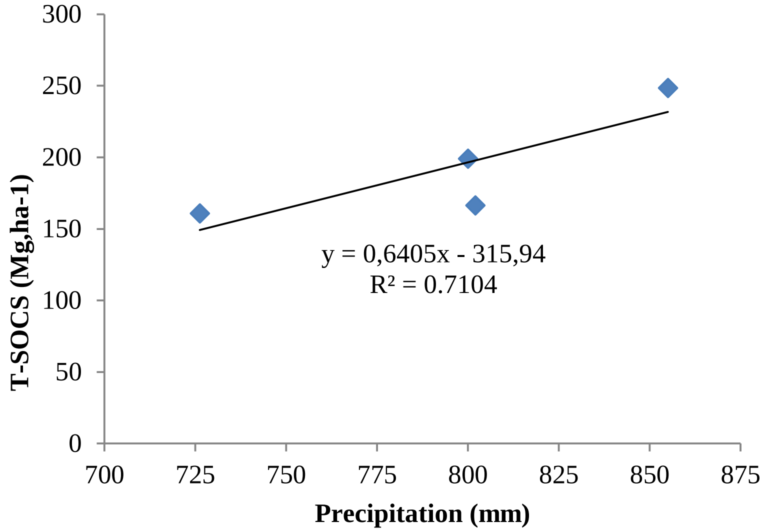


Fig. 4. Plots of the total soil organic carbon stock (T-SOCS) versus precipitations in the study area.