Physical properties of tropical sandy soils: A large range of behaviours

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Key words: sandy soil, bulk density, structure, porosity, particle size distribution, clay, hydraulic properties, compaction

Abstract

Sandy soils are often considered as soils with physical properties that are easily defined, however they are far from being simple. This is particularly the case for sandy soils in the tropics where they are subjected to a cycle of wetting and drying associated with seasonality. In this respect small changes in composition lead to significant differences of physical properties. One of the major soil characteristics to be taken into account is the size distribution of the sand grains. If fine sand induces greater porosity, water retention and resistance to penetration than coarse sand, they exhibit lower permeability. Porosity decreases when the heterogeneity of the sand grain distribution increases leading to an increase in resistance to penetration and decreases in permeability. The presence of silt particles leads to similar consequences. Thus, silty sands are more compact than sandy soils, most silt particles occupying the voids between sand grains thereby reducing porosity and consequently permeability. Size distribution and mineralogy of silt and clay sized particles that are associated with sand grains are also responsible for variations in physical properties of tropical sandy soils. Under tropical environments, sandy soils undergo significant weathering to depth thus resulting in a mineralogy where quartz the dominant mineral in the sand and silt fraction and forms a significant proportion of the clay sized fraction. On the other hand, sandy soils can be present in the lower part of the landscape where clays or salts form during the dry season. As a consequence, sandy soils with similar particle size distribution but due to differences in mineralogy of the clay sized fraction that represents not more than a few percent of the soil mass, show very different physical properties. Finally, in sandy soils unlike other soils, the elementary fabric is easily affected by tillage practices. If greater porosity can be produced through tillage operations, the stability of these systems is very weak and compaction by wheels or other actions can in return produce a dense structure. Thus, compaction results from a variation of the structure at all scales, i.e. from the macroscopic to microscopic scales.

Introduction

Sandy soils are characterized by less than 18 % clay and more than 68 % sand in the first 100 cm of the solum. In the World Reference Base (WRB) soil classification system (ISSS Working Group R.B. 1998), sandy soils may occur in the following Reference Soil Groups: Arenosols, Regosols, Leptosols and Fluvisols. These soils have developed in recently deposited sand materials such as alluvium or dunes. They are weakly developed and show poor horizonation. Soils characterized by a high proportion of sand in the first 100 cm can also correspond to the upper part of highly developed soils formed in weathered quartz-rich material or rock, as evidence by the development of a highly depleted horizon. In the following discussion consideration will be given to a range of soils including sandy soils of the WRB and those with sandy horizons in the upper 100 cm of the profile.

Sandy soils are often considered as soils with physical properties easy to define: weak structure or no structure, poor water retention properties, high permeability, highly sensitivity to compaction with many adverse consequences. However, analysis of the literature shows that their physical properties are far from simple. This is particularly true in the tropics where sandy soils are subjected to a cycle of wettting and drying, with a wet seasons that greatly affects the soil and small differences in composition leads to significant differences of physical properties.

Structure, porosity and bulk density

Sandy soils are characterized by a lack of structure or it is weakly development. Coquet (1995) measured the shrinkage properties of two soils in Senegal with different texture. On the sandy soil, results obtained in the field and in the laboratory (on cores originating from the same horizons), showed very small shrinkage: bulk volume variation was only 0.05 %. When they dried, the sandy soils develop only very few thin cracks organised in a loose network. The poor shrinkage properties of these soils are related to the low clay content and the high proportion of low activity clays of many tropical sandy soils.

A large range of porosity

Sandy soils in the tropics show a large range of porosities and consequently bulk density (D_b). Porosity ranges from 33 % (D_b = 1.78g cm^{-3}) to 47 % (D_b = 1.40 g cm^{-3}) are commonly recorded (Figure 1). The porosity in sandy soils is usually smaller than in clayey and silty soils.

Very small porosities are generally observed in sandy soils of the tropics. Lamotte *et al.* (1997a and b) observed a porosity of 28 % (D_b = 1.91g cm^{-3}) between 35 and 45 cm depth in the Northern Cameroon's in ancient cultivated soils. Lesturgez (2005) measured a porosity of 28 % between 20 and 40 cm depth in cutivated soils belonging to the Warin and Satuk series in Northeast Thailand. Deeper in the soil, similar small porosities were recorded by Burt *et al.* (2001) in sandy soils developed in a saprolite derived from granitic rocks in Zimbabwe. These small porosities were recorded in sandy soils with no gravel or stones thus indicating a close packing of elementary soil particles in soils that have been subjected to continuous cultivation.

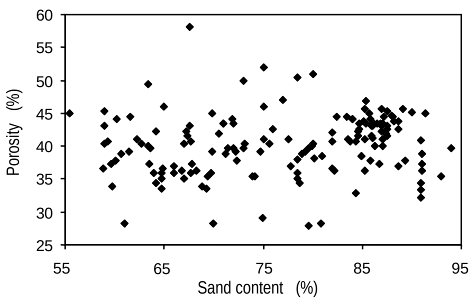

Figure 1. Variation of porosity according to the sand content in tropical sandy soils (after Nicou, 1974 and 1976; Chauvel, 1977; Coquet, 1995; Lamotte *et al.*, 1997a; Burt *et al.*, 2001; Nyamangara *et al.*, 2001; Feng *et al.*, 2002; Bortoluzzi, 2003; Bruand *et al.*, 2004; Lesturgez, 2005; Osunbitam *et al.*, 2005)

However, under native vegetation with intense biological activity or after recent tillage operation (wheel tracks excluded), greater porosity close to 60 % (D_b = 1.10g cm^{-3}) have also been recorded (Bortoluzzi, 2003; Lesturgez, 2005). Such a large porosity is related to the presence of numerous macropores that results from both faunal activity and root development. Bruand *et al*. (2004) have also observed great porosity in the subsoil of intensively cultivated soils, this greater porosity was related to a loose assemblage of elementary soil particles unaffected by farming practices.

Significance of sand and silt grain size distribution

Porosity varies with time after tillage operations thus making it difficult to attribute to soil composition alone. Osunbitan *et al*. (2005) showed a continuous decrease in the porosity of the 0–5 cm layer of a Nigerian loamy sand soil. Porosity ranged from 47.7% (D_b = 1.30 g cm^{-3}) to 60.4 % (D_b = 1.05 g cm^{-3}) according to the tillage system and time after tillage. If we exclude the topsoil horizons from the set used in Figure 1, Figure 2 indicates that the fine sand:coarse sand ratio ranges from 0.5 to 6.1 for the data collected in the literature and the porosity tends to decrease when that ratio increases (R^2 = 0.40, n = 55). The fine sand particles occupying the voids resulting from the packing of the coarse particles, would result in the porosity decreasing when the proportion of fine sand particles increases up to a value that would correspond to the total infilling of that void. Then, for greater proportion of fine sand, the porosity would start again to increase. An increase in the silt–sand ratio would also result in a decrease in the porosity as discussed by Agrawal (1991) for Indian loamy sand and sandy loam soils.

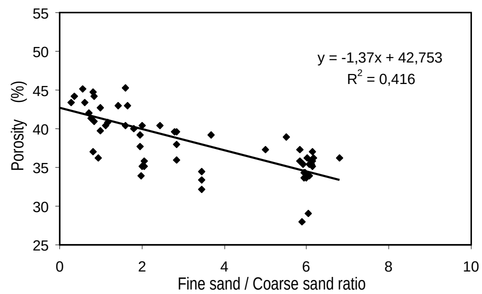

Figure 2. Relationship between the fine sand and coarse sand ratio (after Nicou, 1974; Chauvel, 1977; Coquet, 1995; Lamotte *et al*., 1997a; Nyamangara *et al*., 2001; Feng *et al*., 2002; Bortoluzzi, 2003; Bruand *et al*., 2004; Lesturgez, 2005)

These data recorded with soils samples are consistent with those obtained earlier with models and artificial mixtures in the laboratory (Fiès, 1971; Fiès *et al*., 1972; Panayiotopoulos and Mullins, 1985). Fiès *et al*. (1972) studied the porosity of granular binary mixture and modelled porosity according to the proportion of coarse and fine fractions. They showed that for a mixture of 20–50 µm material with grains 2mm in diameter, the porosity is minimum ($P \approx 0.20$) for 20–50 µm material content close to 25 % (Figure 3). That proportion of fine material is consistent with the theory developed by Westman and Hugill (1930). They calculated an optimum ratio of 3.46 parts by mass of coarse sand to one part of very fine sand (i.e. 22.4 % on mass basis of very fine sand) was required to obtain a mixture with the lowest porosity. Fiès and Stengel (1981) showed good concordance between porosity measured on small aggregates resulting from soil fragmentation and theoretical porosity computed with a model of binary mixtures. In particular, they showed that the porosity was at a minimum for a mixture of 2–20 µm and 200–2000 µm when 2–20 µm content was close to 20 %

(Figure 4). These data indicate that the loose relationship shown in Figure 2 between the porosity and the fine sand and coarse sand ratio would be valid for a limited range of fine sand and coarse sand ratio.

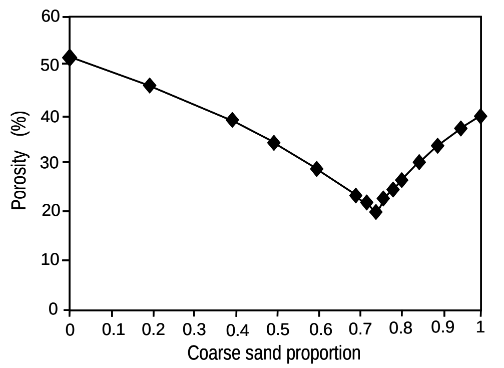

Figure 3. Porosity recorded for a mixture of silt (20–50 µm) and coarse sand (2 mm) particles according to the coarse sand proportion (modified after Fiès *et al*., 1972).

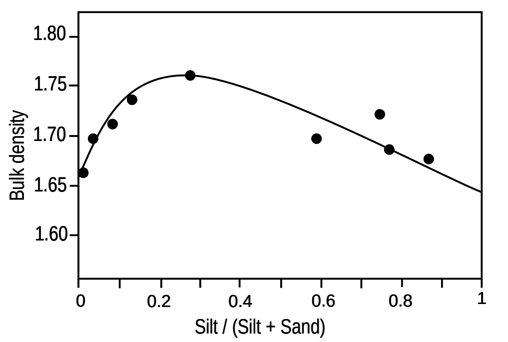

Figure 4. Bulk density recorded by Fiès and Stengel (1981) according to the proportion of silt (20–50 µm) relative to the silt and sand particles (200–2000 µm) in soils with clay content < 20 % and no macroporosity.

Role of the clay fraction characteristics

The large range of porosity (Figure 1) is related to the small cohesion forces between elementary particles thus enabling the formation of a large range of assemblage from very loose to very compact. This is specific to sandy soils because of the small amount of clay that can act as inter-grain cement. In the tropics, clay is often low activity clay (mainly kaolinite) and for similar clay content, tropical sandy soils show usually much smaller inter-grain cohesion than sandy soils in temperate and Mediterranean regions (van Wambeke, 1992). In these deeply weathered soils, the clay-sized fraction may in part consist of quartz as recorded by Hardy (1993) in soils that developed in sandy colluvial deposits in Northern Vietnam. Hardy (1993) showed that 10 to 40 % of the <2 µm fraction was quartz in the soils studied. Bruand *et al.* (2004) studied sandy soils belonging to the Nam Phong series in Northeast Thailand and found that 25 to 35 % of the <2 µm fraction was quartz. The presence of quartz in the <2 µm fraction contributes to its low activity. In some sandy soils, the presence of smectitic clays can lead to very different physical soil properties. In semi-arid tropics, Lamotte *et al*. (1997) studied soil hardening in sandy soil with contrasting loose topsoil and underlying hard horizons. The horizon had similar particle size distributions and the hardness was closely related to a fabric with clay coatings on the sand grains and clay wall-shaped bridges linking the latter. This induced a strong continuity of the solid phase with only a minimum clay content of 6 %.

Hydraulic properties

Water retention properties

Sandy soils retain little water at high water potentials and water content decreases rapidly with the water potential. Panayiotopoulos and Mullins (1985) studied the water retention properties of pure sand materials varying in size (Figure 5). They showed that most water was released between –0.1 and –1 kPa for a coarse sand (2000–710 µm) and between –15 and –30 kPa for a very fine sand (125–45 µm). The small water release recorded for the very fine sand between saturation and –0.5 kPa was not discussed by Panayiotopoulos and Mullins (1985). Mullins and Panayiotopoulos (1984) showed that the water retention curve was only very slightly affected by the clay content for a clay content <20 %. The clay used was a kaolinite. With sandy soils, two thirds of the water present at saturation is usually released at –30 kPa as recorded by Obi and Ebo (1995) in a sandy soil in Southern Nigeria. Water contents ranging from 0.20 to 0.30 $cm^3\,cm^{-3}$ and from 0.04 to 0.12 $cm^3\,cm^{-3}$ are often recorded at –33 and –1500 kPa, respectively in tropical soils belonging to the sand, loamy sand and sandy loam textural class (Hodnett and Tomasella, 2002). In sandy soils, there is very little water available at matric potential <–100kPa. Kukal and Aggarwal (2004) measured a water content of 0.16 and 0.10 $cm^3\,cm^{-3}$ at –33 and –1500 kPa, respectively in a sandy loam topsoil (%clay = 10 %) in India. The water content significantly increased with a slight increase in the clay content and was 0.22 and 0.13 $cm^3\,cm^{-3}$ –33 and –1500 kPa, respectively when the clay content was 14 %. Osunbitam *et al*. (2005) showed in Nigeria an averaged water loss in sandy soils of 0.006 $m\,m^{-3}$ between –100 and –150 kPa, the water content at – 150 kPa being 0.017 $m\,m^{-3}$. Several studies have shown that the available water increases with the silt content (Kapilevich *et al*., 1987; Agrawal, 1991).

Tomasella and Hodnett (1998) compared the measured volumetric water content at different matric potentials and those estimated with the pedotransfer functions (PTFs) developed by Rawls *et al*. (1992) from the USDA soil data base. They showed that these PTFs greatly overestimate the volumetric water content when applied to sandy soils of Brazilian Amazonia (Figure 6). The available water capacity measured between –5 and –1500 kPa by Nyamangara *et al.* (2001) in Zimbabwe for topsoils with a sand content close to 90% ranged from 0.159 to 0.174 $m^3\ m^{-3}$ according to cattle manure management options.

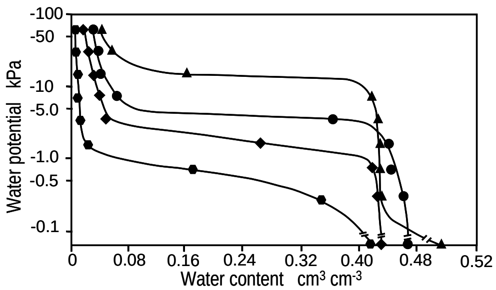

Figure 5. Water retention curves recorded for a coarse sand (hexagon, 2000–710 µm), medium sand (diamond, 500–180 µm), fine sand (circle, 220–105 µm) and very fine sand (triangle, 125–45 µm) (modified after Panayiopoulos and Mullins, 1985).

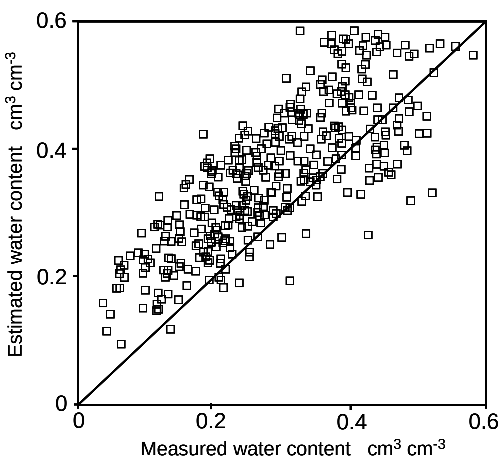

Figure 6. Comparison between measured values of volumetric water content at –33 kPa water potential and those estimated by the pedotransfert functions of Rawls *et al*. (1982) (modified after Tomasella and Hodnett, 1998).

Hydraulic conductivity

The saturated hydraulic conductivity (K_s) of sandy soils in the tropics varies within a range of values covering several orders of magnitude ($10^{-7}<K_s<10^{-3}$ m s^{-1}). In a Brazilian sandy soil with very low clay contents (average content in the whole profile of 0.25 %), Prevedello *et al.* (1995) measured $1.1\times10^{-6}<K_s<7.5\times10^{-5}$ m s^{-1}. Contrasting this in another Brazilian sandy soil with only a slight change in clay content (average content in the whole profile of 6%), Faria and Caramoni (1986) measured $1.5\times10^{-5}<K_s<2.8\times10^{-4}$ m s^{-1}.

In soils, K_s varies according to the development of the macroporosity. As a consequence, K_s variation is more closely related to the macroporosity development rather than to soil texture. Thus, most studies try to relate K_s to part of the macroporosity that is called effective porosity (Φ_e) and defined as total porosity, Φ minus the water content at –33kPa (Ahuja *et al*., 1984). K_s and Φ_e are related as following:

$$K_s = B(\Phi_e)^n$$

with *B* and *n*, are two parameters varying with the soil characteristics. These parameters were obtained by Tomasella and Hodnett (1997) for Brazilian tropical soils. They found log B = 4.752 and n = 4.536 for the soils studied by Predevello *et al.* (1995) and log B = 4.758 and n = 4.532 for another soil studied by Faria and Caramoni (1986). These soils were also used by Tomasella and Hodnett (1997) to derive parameters of the Brook-Corey/Mualem model for unsaturated hydraulic conductivity.

However when cultivated, the macroporosity of sandy soils is very unstable and collapses rapidly in the presence of water. Thus, the measurement of K_s becomes difficult to perform without modifying the macroporosity that have a tremendous effect on K_s. This probably explains why many studies do not report a large range of K_s variation between field experimental treatments and with time as expected. Thus Osunbitan *et al*. (2005) recorded $5.5<K_s<7.5\times10^{-5}$ m s^{-1} in a topsoil under different tillage treatments. The K_s difference recorded by these authors between the different treatments (1×10^{-5} m s^{-1}) was small and similar to the differences recorded under the same tillage treatment over a period of 8 weeks.

Unsaturated hydraulic conductivity (K_θ) of Brazilian sandy soils was also discussed by Tomasella and Hodnett (1997) (Figure 7).

Surface crusting and water infiltration

Because of the very small inter-particle cohesion that results in a very small aggregate stability, sandy soils are highly sensitive to surface crusting, thus explaining the large number of papers in this area (e.g. Chartres, 1992; Casenave and Valentin, 1992; Isbell, 1995; Bielders and Baveye, 1995a; Valentin and Bresson, 1998; Malan Issa *et al*., 1999; Duan *et al.*, 2003; Eldridge and Leys, 2003; Janneau *et al*., 2003; Goossens, 2004;). Crusts protect the soil surface from wind and interrill erosion but they also favour runoff and consequently rill and gully erosion (Valentin and Bresson, 1998). Two main types of structural crusts were recognised in sandy soils depending on the dominant forming process (Casenave and Valentin, 1992; Valentin and Bresson, 1992; Janeau *et al*., 2003): (i) sieving crusts made of well sorted micro-layers with average infiltrability of approximately 30 mm h^{-1}, (ii) and packing crusts made of sand grains closely packed of with average infiltrability of 10mm h^{-1}. Bielders and Baveye (1995b) studied in the laboratory the processes of structural crust formation on coarse textured soils. They proposed that the formation of clay-band in sieving structural crusts would be initiated by the displacement of micro-aggregates or other small particles from the above washed-out layer, followed by their accumulation due to mechanical straining. Erosion crusts that result from smoothening and erosion of structural crusts and depositional crusts that result from sedimentation were also described (Valentin and Bresson, 1998). They exhibit more restrictive infiltrability (2–5 mm h^{-1}) than structural crusts.

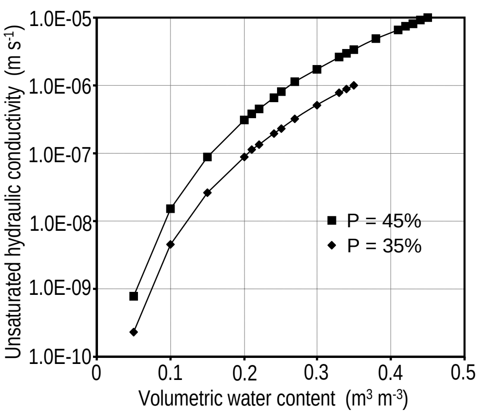

Figure 7. Unsaturated hydraulic conductivity for a sandy soil with a porosity of 45 and 35% (computed after the data published by Tomasella and Hodnett, 1995).

The development of crusts leads to runoff that can be quite significant. Sombatpanit *et al*. (1995) measured between 300 and 400 mm of runoff that corresponded to about 35 % of the rainfall on bare sandy soils in Thailand with 25 to 70 t ha^{-1} of soil loss. Runoff was still between 10 and 20 % of the rainfall under different agricultural treatments.

Surface infiltrability can also be reduced in sandy soils by repellency. Indeed, sandy soils are particularly susceptible to water repellency and susceptibility increases with the duration of the dry season. Repellency is responsible for vertical fingered flow in sandy soils because of the presence of repellent soil volumes with hydrophobic organic matter (Roberts and Crabon,1972; Dekker and Ritsema, 1994; Ritsema and Dekker, 1994). Study of repellent soils in arid and humid climates showed that repellency would be much more related to the type of organic matter than to the duration of the dry period (Jaramillo *et al*., 2000).

Compaction

Sensitivity to soil compaction

Unlike other soils, the structure of sandy soils can be easily affected by mechanical compaction over a large range of scales. Usually mechanical compaction preferentially affects large pores (i.e. macropores that result from tillage and biological activity) but in sandy soils its affects these large pores down to the small pores that result from the arrangement of the skeleton particles (sand and silt) within the clay fraction. That re-arrangement when submitted to mechanical compaction is possible because of the small cohesion between the skeleton particles. For narrowly graded pure sand materials, Panayiotopoulos and Mullins (1985) showed that these air-dry and nearly saturated sands were always found to pack more closely under a given load than the same sand at any water content.

Very small porosity can be recorded under wheel tracks and just underneath the tilled layer. Thus, Bennie and Botha (1986) recorded $1.7<D_b<1.8$ g cm^{-3} in the 0-20 cm layer under wheel tracks and in the 20-40 cm layer. Because of this small inter-particle cohesion, high bulk density is also recorded when sandy soils are puddled in rice-wheat cropping systems. Thus, Aggarwal *et al*. (1995) recorded $1.75<D_b<1.82$ g cm^{-3} in the 15-20 cm layer after several years of high puddling in a sandy loam soil.

Smith *et al*. (1997a) studied the effect of soil compaction on a large range of South African forestry soils. They showed on soil cores in the laboratory that the porosity after compaction of sandy loam and loamy sand soils was related to the size distribution of the sand fraction and tended to decrease with the increase in the clay and silt content. Smith *et al*. (1997b) showed for loamy sand that increases in compaction were almost independent of the water content and then almost entirely due to increasing applied pressure alone. Smith *et al*. (1997c) also recorded a high compactibility for sandy soils derived from sandstone, granite and aeolian sands. The maximum bulk density was related to the loss in mass after ignition at 450°C.

Penetration resistance

Bulk density increase results in an increase in the penetration resistance with significant consequences for root development although there no clear relationship with the penetration resistance (Mullins *et al*., 1987; Bengough and Mullins, 1991). Critical values that severely restrict root growth have been estimated to vary from <1 to >4 MPa depending on the soil, water content and crop type (Greacen *et al*., 1968). Indeed, the penetration resistance varies within a large range of values according to the soil water content without any variation of the other soil characteristics (e.g. particle size distribution, mineralogy, porosity, assemblage of the elementary particles). It is significantly inversely related to water content. Many penetration resistances published in the literature for sandy soils range between 0.1 and 0.8 MPa but the water content at which they were determined often remains unclear (Osunbitan *et al*., 2005). Bruand ***et al.*** (2005) recorded a penetration resistance ranging from 0.35 to 0.55 MPa in the subsoil of a sandy soil in Thailand when the water content ranged from 0.03 to 0.09 kg kg^{-1} (Figure 8). In South African sandy soils that developed in aeolian sand, Du Preez *et al*. (1981) measured penetration resistance >1.5 MPa, at field water capacity, in a ploughed layer under wheel tracks. In these soils Bennie and Botha (1986) confirmed the presence of such values of resistance to penetration in the subsoil and showed that they result from compaction because of traffic that leads to an increase in the penetration resistance that restricts root development for wheat and maize. Kukal and Aggarwal (2003) measured much greater penetration resistance in a sandy loam soil after puddling because of subsurface compaction. Indeed, these authors recorded at field capacity a penetration resistance ranging from 3.0 to 4.5 MPa in the compacted layer.

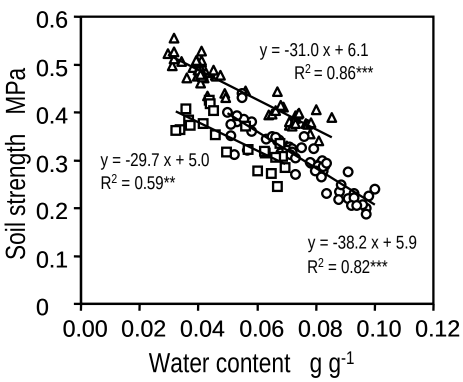

Figure 8. Resistance to penetration expressed as unconfined strength with respect to water content in the Ap (square), E (circle) and Bt (triangle) horizons. (modified after Bruand *et al.*, 2005).

In their study, Smith *et al*. (1997a) showed on compressed soil cores only small differences in strength development across a wide range of water content for a loamy sand soil. This would be related to the contribution of frictional rather than cohesion forces to resistance to penetration (Smith *et al*., 1997a). On the other hand, results also showed somewhat different behaviour for a sandy loam soil, resistance to penetration increasing from 1 to 5 MPa over a range of water content of only 4 % by mass. This large range of resistance to penetration would result from the contribution of cohesion forces that are partly related to the water content. Thus, a decrease in water content would increase frictional and cohesion forces from field capacity to intermediate water content, smaller water contents increasing the frictional forces alone, the cohesion forces disappearing, thus explaining the results recorded in Nigeria by Ley *et al*. (1995) on a large range of soils including sandy soils.

Effect of combined deep tillage and controlled traffic on penetration resistance and its consequences for root growth has been studied in several countries. In their one-year study, Bennie and Botha (1986) showed that deep ripping and controlled traffic led to a significant increase in rooting depth, rooting density in the subsoil, water use efficiency and a yield increases of 30 % for maize and 19 % for wheat. Increased yields were recorded in many earlier studies after deep tillage (e.g. Reicosky *et al*., 1976; Bennie *et al*., 1985) but the duration of the positive effects of deep tillage is still under discussion. Slotting is also an efficient technique to loosen the subsoil and promote rooting (Jayawarrdane *et al*., 1995). Hartmann *et al*. (1999) showed that rooting depth and yield of various crops were increase increased for two successive years after slotting in a sandy soils of Northeast Thailand. Lesturgez (2005) recorded a significant increase in root density in the slot that enables a better subsoil prospecting. Lesturgez *et al*. (2004) also investigated the potential use of forage legume *Stylosanthes hamata* (stylo) to ameliorate the structure of compact layers in sandy soils of Northeast Thailand. They showed that after 24 months of continuous stylo, roots were able to penetrate the compact subsoil, resulting in an improvement of its macroporosity. They also showed that a subsequent maize crop developed a deep and extensive root system using the macropores.

Hardsetting in sandy soils

Many tropical sandy soils are potentially hardsetting soils, i.e. they can become compact, hard with apparently apedal condition prevailing on drying (Northcote, 1979). In these soils, a significant increase in soil strength is recorded over very narrow water content changes within the plant available range of soil water potential (-10^2 to -10^3 kPa) with resulting adverse effects on root growth and crop production (Mullins *et al*., 1990). Thus, Chan (1995) measured strength characteristics in sandy loam hardsetting soil in the semi-arid region of Australia and showed that in the cultivated soil strength increased from 0.02 to 0.09 MPa with water content decreasing from 0.11 to 0.04 kg kg^{-1} when there was no strength variation under permanent pasture. McKyes *et al*. (1994) studied the cohesion and friction in two sandy-loam hardsetting soils from Zimbabwe. They showed that the cohesion changed from nearly zero at saturation to well over 0.1 MPa in the field dry state. Young and Mullins (1991) suggested that the amount of <60 µm particles rather than solely the <2 µm is important in causing the development of hardsetting properties.

High soil strength in sandy soils can be also partly related to the development of silica precipitation ranging from globules and silica flower to surface of contact and quartz neogenesis (Lesturgez, 2005). These precipitations would not be responsible for a complete cementation of the

Commentaire [GL1] : Jayawardane (1995) est un papier general sur le slotting (pas specific à la Thaïlande ni au sols sableux). Le papier de Christian est par contre une application aux sols sableux du NE Thailandais.

sand and silt grains but would lead to an increase in particle contacts and frictions, thus explaining the increase in soil strength recorded.

Controlled compaction

Compaction in sandy soils was also discussed as a possible water and nutrient management to improve water retention properties and reduce nutrient leaching in Indian sandy soils (Agrawal and Kunar, 1976; Agrawal *et al*., 1987; Agrawal, 1991; Arora *et al*., 2005). Indeed, according to these authors, compaction that reduces the volume and continuity of large pores, would increase water retention and reduce water infiltration and saturated hydraulic conductivity in highly permeable deep sandy soils. Compaction would save irrigation water by 15–36 % and increase productivity by 30–50 %.

Puddling is also used to reduce high percolation losses of irrigation water and nutrient leaching when cultivated for rice production (Aggarwal *et al*., 1995; Arora *et al*., 1995). Sharma and Bhagat (1993) showed that puddling was effective in reducing percolation losses when sand was less than 70%, and finer fractions were dominated by clay (13–20%). Puddling has been reported to decrease saturated hydraulic conductivity of puddle layer (0–10 cm) of sandy loam soil from 1.8 10^{-7} m s^{-1} in unpuddled to 4.2 10^{-8} m s^{-1} with medium puddling and 2.5 10^{-8} m s^{-1} with high puddling (Kukal and Aggarwal, 2002). Kukal and Aggarwal (2003b) showed in a sandy loam soil that puddling reduced percolation losses by 14–16% with the increase in puddling intensity from medium to high, whereas the amount of irrigation water required decreased by 15–25%. Similar results were recorded by Kukal and Sidhu (2004) in another sandy loam soil in India.

However, puddling in rice–wheat cropping systems leads to some adverse effects for the following wheat crop that requires to be managed with appropriate tillage techniques (Aggarwal *et al*., 1995; Kukal and Aggarwal, 2003a; Arora *et al*., 2005). Then a yield decline of wheat is often recorded because of subsurface compaction ($1.70 < D_b < 1.75$ g cm^{-3}) at 14–20 cm depth under normal puddling at normal depth. Kukal and Aggarwal (2003a) showed in India for a sandy loam soil that puddling at shallow depth (5–6 cm) led to the development of a compact layer at 10–12 cm depth that was loosened ($1.50 < D_b < 1.55$ g cm^{-3}) during normal cultivation for wheat seedbed preparation.

Changes in soil physical attributes

Increases in the total porosity of soils was observed with the application of co-composted materials at rates > 60 t ha^{-1}. Soil receiving non-composted waste bentonite did not exhibit an increase in porosity. The soil moisture content at permanent wilting point (θpwp) increased slightly with an increase in the rate of application of co-composted bentonite wastes. This is attributed in part to an increase in the specific surface area and CEC of the soil. This relationship is illustrates in Figure 1.

Conclusion

In the tropics, physical attributes of sandy soils are particularly sensitive to both the sand and silt size distribution and mineralogy of the clay fraction. Because of the presence of low activity clay in most sandy soils, the assemblage of elementary skeleton particles is highly unstable resulting in a high instability with respect to structure from the microscopic to macroscopic scale. When for a variety reasons, the assemblage collapses, the resulting porosity and penetration resistance would be all the greater as the skeleton particle distribution is heterometric.

In contrast, in sandy soils unlike other soils, the elementary fabric can be easily loosened by tillage practices. Thus greater porosity can be produced easily by tillage but its stability is very weak and compaction by wheels or other actions can in return produce easily a dense structure and adverse physical properties. This leads to a decrease in the water retention properties and hydraulic conductivity, to an increase in the resistance to penetration and sensitivity to surface crusting.

More generally, tropical sandy soils, more than other soils, require careful management in an environmentally friendly manner. Indeed, even if most physical degradation processes are more easily reversible in tropical sandy soils than in other soils, the physical fertility of these soils is weak. These soils require very little tillage operation in the wrong way to produce significant adverse consequences for plant development and consequently for crop yield and environment.

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