Susceptibility of Coarse-textured Soils to Soil Erosion by Water in the Tropics

F.K. Salako[1]

Department of Soil Science and Agricultural Mechanisation,
University of Agriculture, Abeokuta, Nigeria

Lecture given at the
College on Soil Physics
Trieste, 3-21 March 2003

LNS0418030

[1] kolawolesalako@hotmail.com; kfsalako@yahoo.ie

Abstract

The application of soil physics for the evaluation of factors of soil erosion in the tropics received considerable attention in the last four decades. In Nigeria, physical characteristics of rainfall such as drop size and drop-size distribution, rainfall intensity at short intervals and kinetic energy of rainfall were evaluated using different methods. Thus, compound erosivity indices were evaluated which showed a similar trend in annual rainfall erosivity with annual rainfall amounts rainfall amounts. Attempts have also been made to use geostatistical tools and fractal theory to describe temporal variability in rainfall erosivity. High erosivity aggravates the vulnerability of coarse-textured soils to erosion. These soils, high in sand content were poorly aggregated and structurally weak. Thus, they were easily detached and transported by runoff. Long-term data are needed to describe factors of soil erosion in the tropics but quite often, equipment are not available or poorly maintained where available such data useful data are not collected. A greater cooperation of pure physicists, soil physicists and engineers in the developing nations is needed to improve or design equipment and methods for the characterization of factors of soil erosion in the tropics.

Soil erosion in the tropics and application of soil physics

Accelerated soil erosion is a major ecological problem in the tropics. Apart from climatic and edaphic factors which contribute to this menace, socio-economic factors bothering on land misuse are also significant contributors. As Hartemink (2002) noted, in tropical regions, important soil science themes have not changed much in the past decades, and soil science is still closely linked to agriculture and society at large. In other words, provision of food for the ever-increasing population with the highest rate occurring in the tropics, decrease in food production per capita and rapid soil degradation remain major issues for tropical soil scientists. These views were also poignantly expressed by Lal (2000) who stated that the principal issues in the tropics for the 21st century include (i) achieving food security, (ii) curtailing soil degradation and restoring degraded soils; and (iii) improving environmental quality.

The assessment of soil erosion and soil conservation needs has been a major focus in Soil Physics as a sub-discipline of Soil Science. Thus, it can be described as Applied Soil Physics. The development of Soil Physics in the tropics, has understandably been tilted toward Applied Soil Physics for reasons already discussed by Hartemink (2002), although there is a need to be involved in research which focus on non-agricultural use of land now in order to understand the contribution of the tropical environment to global climate change and environmental quality. This paper discusses the assessment of climatic and edaphic factors which influence soil erosion the tropics, with an emphasis on experiences garnered in Nigeria. The main factors discussed are rainfall and soil characteristics, with attention drawn to methodology and interpretation of results obtained.

Physical characteristics of rainfall in the tropics

Rainfall is the real agent of soil erosion by water in the tropics by virtue of its role as the source of water or the only form of precipitation contributing to the hydrologic cycle. For most of crop production in the tropical Africa, it is also the source of water, as irrigation agriculture is not well developed even in semi-arid and arid regions where such is needed.

The capacity or potential ability of the rain to cause soil erosion is called rainfall erosivity. This attribute of rainfall is linked to its physical characteristics, namely, amount, duration, drop size and drop size distribution, terminal velocity, intensity and kinetic energy.

Rainfall amount and duration

Quantity of rainfall is measured by amount in unit of depth (i.e., mm). In Nigeria, annual in the dry regions is can be less than 500 mm but in the wet coastal regions, it could be greater than 2500 mm. The amount of rainfall and how long it takes to fall influence how much of water infiltrates and runoff the soil. Problems of flooding and soil erosion are basically related to amount and duration of rainfall.

This is done with standard rain gauges which are found in many locations in the tropics. The problem, however, is that standard raingauge records are recorded on daily basis (24 h) and this obscures information on high rainfall intensity because of the long period after which measurements are taken. There are also auto-recording rain gauges (pluviographs) which record on charts which can be resolved to short-intervals of rainfall (e.g. $\leq$ 15 minutes).

Automatic weather stations exist today in which rainfall can be recorded at intervals of interest, which will be downloaded from data loggers into computers. As will be shown later, attention of meteorological stations need to be drawn to need of soil conservationists for relevant data collection for soil conservation planning or else the benefit of this advanced technique can be reduced or lost in this respect.

Rain drops

Rain falls as drops, which are assumed to have a spherical shape (Carter et al., 1974). The maximum raindrop size at which it can remain stable without breaking into pieces is 5 mm.

Measurement of raindrop size and distribution can be done with a number of methods, among which are:

- Drop-stain method in which an absorbent paper is dusted with powder which form stains as rain drops fall on them
- Flour-pellet method which basically assesses the size of drops from size of pellets (dough-balls) made after exposing baking flour to drops
- Use of pressure transducers or acoustic device or piezoelectric sensors to measure impact of raindrop
- Photographic method in which photographs of falling raindrops were used.

In Nigeria (Table 1), raindrop sizes have been measured by impact made on piezoelectric sensors (Kowal et al., 1973; Kowal and Kassam, 1976; Lal, 1998) and by the flour-pellet method (Aina, 1980; Obi and Salako, 1995; Salako et al., 1995; Akinnifesi and Salako, 1997). Kowal et al. (1973) developed a device which recorded graphically, on a time scale the amplitude of electric pulses originating from the impact of raindrops on the surface of a transducer disc, by making use of the piezo-electric effect.

The flour-pellet method (Carter et al., 1974) consists of calibrating plain baking flour with water drops of known sizes in the laboratory. Pellets formed are air-dried first and later dried further in the oven before weighing. Water drop weight: Pellet weight ratio is calculated. The average pellet weight from rain sample is obtained by exposing the calibrated flour to rainfall in a pan (31 cm diameter x 1.6 cm deep). Raindrop size is obtained by calculation:

$$d = (6W/\pi)^{1/3} \quad (1)$$

where d is raindrop diameter (mm) and W the average drop weight (mg) equal to weight ratio x average pellet weight. Apart from obtaining the range of drop sizes present in rainfall, the percentage of sample contributed by each drop size is calculated to obtain a cumulative frequency distribution. The median drop size (D_{50}) which is the size at 50% volume is calculated from the cumulative frequency distribution

Table 1. Raindrop size distribution in Nigeria

Agroecological zone	Maximum drop size (mm)	Median drop size (mm)	Source	Method
Northern Guinea Savanna	4.86	3.42	Kowal and Kassam (1976)	Direct with a piezoelectric sensor. Equipment assembled by Kowal et al. (1973)
Derived Savanna Southeastern Nigeria	3.4	1.1-2.9	Obi and Salako (1995)	Flour-pellet method
Derived Savanna Southwestern Nigeria	> 3	2.25-3	Lal (1998)	Direct with a piezoelectric sensor. DISTROMET
Humid forest Southcentral Nigeria	5.1	2.3	Salako et al. (1995)	Flour-pellet method
Humid forest Southwestern Nigeria	4.5	3.0	Aina (1980)	Flour-pellet method

The median drop sizes indicated that large drop sizes played significant role in the erosivity of rainfall in the tropics. Median drop sizes were greater than 3.5 mm at rainfall intensities above 125 mm h^{-1} in southwestern Nigeria (Aina, 1980). Lal (1998) observed that fogs and gentle rains often account for low drop sizes and energy load. Median drop size generally increases with increases in rainfall intensity up to a certain limit according to the relationship

$$D_{50} = aI^b \tag{2}$$

because the constants will vary with rain types. The relationship is also believed to be valid up to 75 mm h^{-1} rainfall intensity.

Rainfall intensity

Rainfall intensity, which is referred to already in Equation 2, is calculated from rainfall amount relative to duration (amount/duration). The actual duration of rainfall within a day is not discernible from daily rainfall records of standard rain gauges but can be discerned from pluviograph or autoweather station data. Thus, short-interval intensities needed for soil conservation planning and construction of hydrologic structures can only be obtained with gauges whose records can be analyzed at less than 1 h (Figure 1).

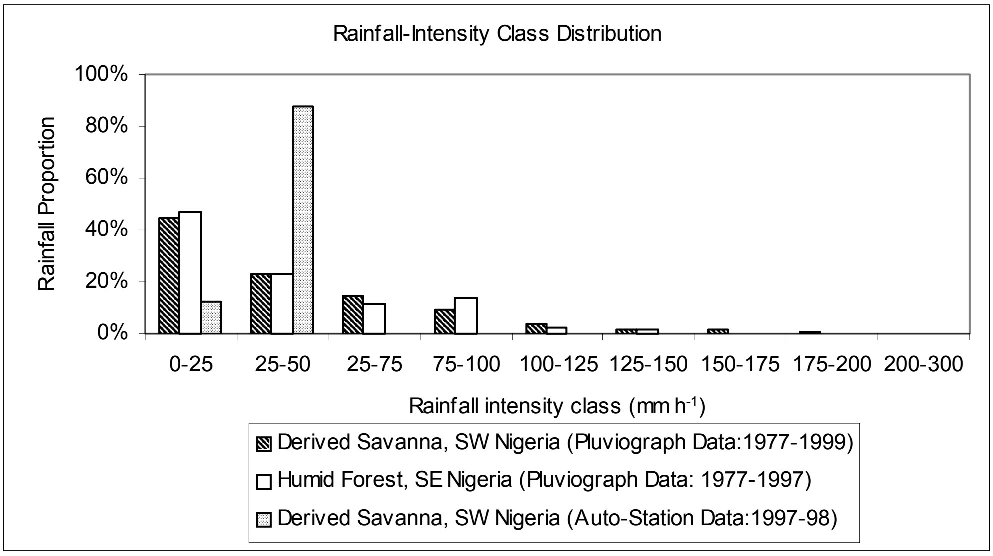

Figure 1: Rainfall-intensity class distribution and comparison of pluviograph data analyzed at 15-minute intervals with automatic-weather station data analyzed at 1-h interval. (Source: Salako, F. K. unpublished)

The pluviograph data in Figure 1 were obtained at 15-minute intervals but those of the auto-weather station which are supposed to be more accurate were obtained at 1-h interval. They suggest the following:

(i) About 50%of rainfall in the tropics will fall at intensities exceeding 25 mm h^{-1} which is considered the threshold intensity for soil erosion to occur

(ii) Intensities greater than 75 mm h-1 which would hardly occur in temperate regions occur in the tropics. According to Salako F. K. (unpublished), intensities up to 150 mm h-1 occur annually.

(iii) The data suggest that records of rainfall events made at less than 1 hour would obscure exceptional intensities exceeding 50 mm h-1. Definitely, a 24-h measurement with standard raingauge will quite often show intensities less than 25 mm h-1 which will cause poor judgment of rainfall erosivity in the region.

Terminal velocity of rainfall and kinetic energy

Terminal velocity refers to the velocity a body falling freely under the force of gravity reaches when frictional resistance of the air is equal to the gravitational force. In physics, kinetic energy and momentum is related to mass and velocity of objects. This has, therefore, generated the argument on which is appropriate for rainfall erosivity; kinetic energy or momentum. However, kinetic energy of rainfall is frequently used for evaluating rainfall erosivity.

Rainfall kinetic energy is related to rainfall drop sizes and intensities. It has, therefore, been possible to evaluate kinetic energy of rainfall directly from piezoelectric sensors which measure drop sizes as those used by Kowal et al. (1976) and Lal (1998) in Nigeria (Table 1). Lal (1998) did not however find a correlation between kinetic energy and momentum of rainfall in Ibadan, southwestern Nigeria in 1980 whereas he found a significant relationship in 1981

$$E = -1018.1 + 996.2D_{50} - 152.3D_{50}^2 \qquad r = 0.63 \qquad (3)$$

$$MV = -264.6 + 264.0D_{50} - 41.0D_{50}^2 \qquad r = 0.66 \qquad (4)$$

where E is the kinetic energy (J/m^2), D_{50} the median drop size (mm) and MV the momentum (kg·m/s).

Direct measurement of rainfall kinetic energy or momentum is rare in many locations all over the world. It is common to evaluate rainfall kinetic energy using Wischmeier and Smith (1978) empirical equation proposed for the (Revised) Universal Soil Loss Equation (RUSLE) by Renard et al. (1997):

$$e_m = 0.0119 + 0.0873 \log_{10}(i) \qquad i \leq 76\ mm{\cdot}h^{-1} \qquad (5)$$

$$e_m = 0.283 \qquad i > 76\ mm{\cdot}h^{-1} \qquad (6)$$

Another empirical equation which Brown and Foster (1987) suggested for tropical regions and adopted in RUSLE (Renard et al., 1997) is:

$$e_m = 0.29\ [1 - 0.72 \exp(-0.05i)] \qquad (7)$$

where e_m is kinetic energy of a unit rainfall ($MJ{\cdot}ha^{-1}{\cdot}mm^{-1}$ of rain) and i the corresponding rainfall intensity ($mm{\cdot}h^{-1}$). The procedures used for the evaluation of kinetic energy E from e_m have been discussed by Renard et al. (1997). Unless otherwise stated, Equations 5 and 6 are usually used for computation of E or KE.

Application of equations 5 and 6 in the tropics suffered the inherent problem in using empirical equations, which is the fact that they are best suited for similar environment to where they were developed. Unfortunately, these equations had not taken into consideration the high intensities of tropical rainfall as demonstrated in Figure 1 as it is assumed that intensities above 76 mm·h^{-1} would hardly occur. Still, they have been widely used to evaluate rainfall kinetic energy in the tropics, perhaps, because convincing alternatives have not been provided in terms of database used in generating these other empirical equations. For instance, Kowal and Kassam (1976) based on 18 rainstorms proposed the equation:

$$E = 0.414A - 1.2 \qquad r = 0.99 \tag{8}$$

where E is the kinetic energy (MJ/ha) and A the rainfall (mm). Salako et al. (1991) found that E computed with Equation 8 was more than E computed with Equation 5. As Lal (1998) reported, the relationships presented in Equations 3 and 4 were valid in 1981 but not in 1980 at the same location. Definitely, there is a need for a large database to account for temporal variability of rainfall and the large database involved in the development of RUSLE has accounted for the confidence often reposed in it for soil conservation planning.

Compound rainfall erosivity

There are three popular indices for rainfall erosivity evaluation:

(i) EI_{30} index (Wischmeier and Smith, 1978; Renard et al., 1997)
(ii) KE ≥ 25 (Hudson, 1995)
(iii) AI_m index (Lal, 1976)

These are compound erosivity indices because they incorporate more than one rainfall characteristic in their evaluation. The EI_{30} index is a product of rainfall kinetic energy (E) and maximum 30-minute intensity, I_{30}; the KE ≥ 25 is a summation of kinetic energy of rainfall exceeding 25 mm·h^{-1}, based on the premise that such rainfall events are the culprits in the soil erosion problem and the AI_m index is the product of daily rainfall amount and maximum short-term intensity (I_m). Although both the AIm and KE ≥ 25 were developed to characterize tropical rainfall, the EI_{30} index had been widely used because it is amenable to RUSLE for soil conservation planning. From published and unpublished data, the magnitude of rainfall in Nigeria follows the distribution and magnitude of annual rainfall amounts (Table 2).

In a recent study with long-term data covering 1977-1999 (Salako, F. K. unpublished; Figure 1), the annual rainfall erosivity for Ibadan with the EI_{30} was 17745 MJ·mm/ha·h and for Port-Harcourt it was 26829 MJ·mm/ha·h. The AI_m index annual value for this study was 1100 cm^2·h^{-1} for Ibadan and 1457 cm^2·h^{-1} for Port-Harcourt. The difference in these values, compared with Table 2 for same locations, was due to the use of long-term data and the inclusion of all rainfall events in the analysis. The long-term data caused increase in values because they could capture rainfall events, especially for some months with infrequent rainfall which might be omitted by short-term data. Problem of underestimation with all-inclusive long-term data is remote and it is better that planning is based on these.

Temporal variability of rainfall erosivity and its prediction from daily rainfall

It is often desirable to evaluate rainfall erosivity with long-term data (≥ 20 years) to account for temporal variability Unfortunately, such long-term pluviograph data are difficult to obtain in tropical areas. Besides, pluviographs are rare in meteorological stations of many developing nations.

Table 2. Annual rainfall erosivity in southern Nigeria

Location	Years of observation	Rainfall erosivity indices		
		EI_{30} (MJ·mm/ha·h)	AI_m ($cm^2·h^{-1}$)	KE > 25 ($MJ·ha^{-1}$)
Southeastern Nigeria (Annual rainfall between 1500 and 2900 mm)##				
Nsukka	1976-1983	12814	849	141
Enugu	1977-1980	15776	1055	235
Umudike	1974-1985	15922	1300	210
Owerri	1973-1982	18545	1300	249
Calabar	1976-1980	22168	1468	303
Ikom	1978-1980	19482	1327	297
Port-Harcourt	1973-1981	18611	1400	234
Onitsha	1977-1980	12087	1017	188
Southcentral Nigeria (Mean annual rainfall of 2000 mm) ##				
Okomu, near Benin City	1986-1990	18510	1329	216
Southwestern Nigeria (Mean annual rainfall of 1300 mm)				
Ibadan	1996-1999	12137	754	170

Mean annual rainfall from 1961 to 1990 for different locations are reported by Jagtap (1995), Armon (1984), Salako et al. (1995), and Salako, F.K., Kirchhof G., Tian, G. (Unpublished)

Spatial variability, albeit temporal variability, can be quantitatively assessed with geostatistical tools and fractal theory (Vieira et al., 1983; Webster and Oliver, 1990). This is an area which soil scientists have vigorously explored in the last decade (Burrough, 1993; Anderson et al., 1998) but less so in the tropics due to dearth of data.

Using the regionalized variable theory in geostatistics, an experimental variogram is constructed which is useful for interpolation, optimal sampling and description of temporal dependence (Figures 2 and 3). A variogram is a plot of variance (Y axis) versus lag (distance or time on X axis).

Variograms are fitted with various theoretical models which take different forms which classified as bounded variograms with finite ranges (e.g., circular and spherical models), bounded variograms with asymptotic ranges (e.g., exponential and gaussian models) and linear variogram (power and linear models) (Vieira et al., 1983; Webster and Oliver, 1990). Below are some variogram models which have been applied to long-term rainfall erosivity data in southern Nigeria (Salako, F. K., unpublished).

The Gaussian model is

$$\gamma(h) = c_o + c_1[1\text{-}\exp(-h^2/a_o^2) \tag{9}$$

where γ (h) is the semi-variance of erosivity index (units of index)^2), h the lag distance (years), c_o the nugget variance, (c_o+ c_1) the sill variance, and $a_o = (1/\sqrt{3})a$ with a being the range.

The linear model used for the HF zone was of the form

$$\gamma(h) = c_o + ch \tag{10}$$

where c is the slope. Also, the power model which fitted both the DS and HF zones was used for the calculation of fractal dimensions

$$\gamma(h) = wh^{\alpha} \tag{11}$$

or

$$\log \gamma(h) = \log w + \alpha \log h \tag{12}$$

which is the linear form of the power model where γ(h) is the semi-variance, h the lag distance in years, w the intercept (constant), α the slope = (4 - 2D) with D being the fractal dimension.

Some of these analyses can be carried out with some statistical software like GENSTAT (Lawes Agricultural Trust, 1996). Thus, the burdensomeness of a large body of data is reduced considerably.

As shown in Figures 2 and 3, the typical characteristics of variograms are the sill, the nugget and the range. The sill, which is a horizontal part of the variogram implies a lack of spatial (temporal) dependence as the variances have no defined relationship with the lag. The curve rises to the sill at a lag value called the range. The range is an aspect of the curve which shows data points that are alike or interdependent.

The nugget variance for the EI_{30} was 86 (MJ·mm/ha·h)^2, the sill variance was 217 (MJ·mm/ha·h)^2 and the range was 16.5 years. The nugget variance is an indication of existence of measurement errors or short-term variations with lags less than one year. The range in this situation indicated that within 16.5 years, there was interdependence of rainfall erosivity by which interpolation could be confidently made. These characteristics have been used to deduce that temporal variations in rainfall erosivity could be adequately accounted for at Ibadan with data spanning 17 years.

Fractal dimensions D are calculated from the sampled data using the exponential model (Burrough, 1993; Anderson et al., 1998). Given the variogram in Figure 3, the fractal dimension D using the AI_m index was 1.927. For EI_{30}, it was 1.88. Fractal dimensions close to 2 such as these indicate the importance of long-range variations (Eghball and Power, 1995; Eghball et al., 1995). Both regionalized and fractal theories have been use to deduce the processes of vegetational and climatic erosivity changes over the years at Ibadan. For about 3 decades, there has been a complex interaction of human, edaphic and climatic factors in the transformation of the forest to savanna. The variograms seemed to suggest that after about two decades of continuous change, the processes became random as a vast area of savanna had been created from the hitherto forest vegetation. Fractal dimensions were large because of the complex interactions of factors which caused these changes.

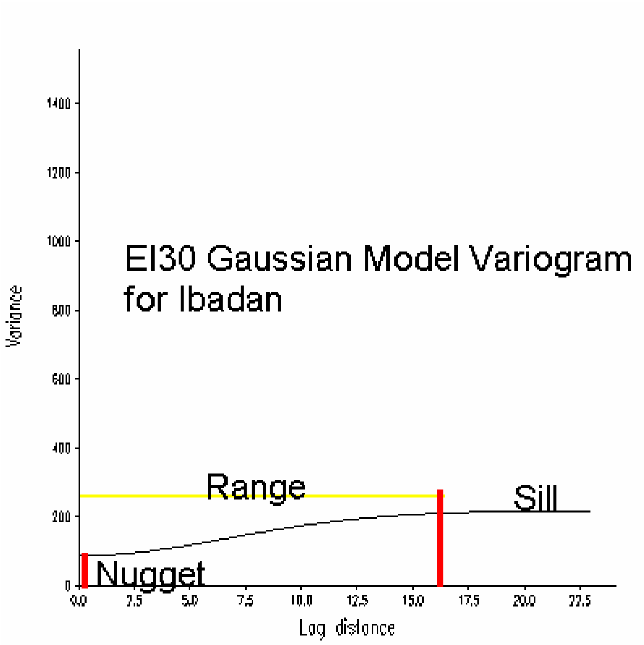

Figure 2: Experimental variogram describing temporal variability of EI_{30} (MJ·mm/ha·h) index.

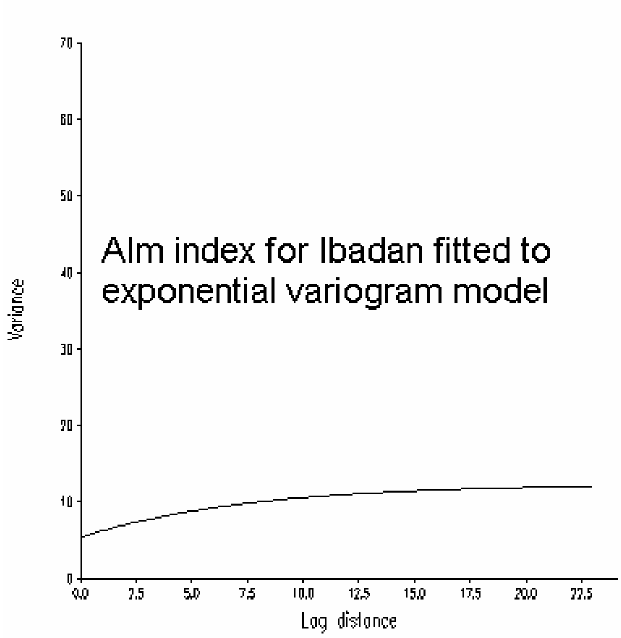

Figure 3: Experimental variogram describing temporal variability of AI_m ($cm^2 \cdot h^{-1}$) index.

Long-term data of daily rainfall amount are easier to come by than pluviograph or auto-weather station data in the tropics. This has encouraged development of relationships between rainfall erosivity, especially EI30 and daily rainfall amount. The relationship of the form below has been found widely applicable (Elsenbeer, 1993; Yu and Rosewell, 1996; Salako, F. K. unpublished) for tropical rainfall:

$$R = aA^b \tag{13}$$

where R is the rainfall erosivity, A the daily rainfall amount, and a and b are constants. Values of a and b calculated for a region in southern Nigeria are given in Table 3.

Table 3. Regression parameters of the model, $R = aA^b$, for the prediction of rainfall erosivity (dependent variable) from daily rainfall amount (mm) (independent variable) in southern Nigeria

Rainfall erosivity index R	Intercept a	Slope b	Coefficient of determination r^2
AI_m ($cm^2 \cdot h^{-1}$)	0.03	1.76	0.91
EI_{30}-WS (MJ·mm/ha·h)	0.41	1.85	0.91

Soil characteristics

Susceptibility of soil to erosion is termed soil erodibility. It is strictly a soil attribute while erosivity is a rainfall attribute. Soil susceptibility to erosion is influenced by properties such as texture, aggregate stability, water transmission characteristics, and organic matter content. The process of soil erosion involves detachment, transportation and deposition. Thus, soils which are susceptible to detachment and transportation are can be easily eroded. This explains the reason why characteristics measured in the laboratory or in the field to evaluate soil erodibility focus on splash erosion and transportation of soil materials. Soils, which are weakly aggregated, can be easily detached or dispersed by water while fine materials such as clay are easily transported once detached.

Particle size distribution

Soil texture refers basically to the distribution of sand, silt and clay. These are components of materials less than 2 mm size, called the fine earth fraction. Materials greater than 2 mm in size are called gravel, cobbles and stones. By the United States Department of Agriculture classification, the size ranges are 0.05-2 mm for sand, 0.002-0.05 mm for silt and < 0.002 mm for clay. In southwestern Nigeria, a vast area of the Alfisol formed on basement complex contains particles > 2 mm (i.e., gravel and stones) (Table 4).

Table 4. Particle size distribution, organic carbon, calcium and bulk density values for a forest profile in Ibadan, southwestern Nigeria (Salako F. K., unpublished)

Depth (cm)	Total sand	VCS (1-2 mm)	CS (0.5-1mm)	MS (0.25-0.5mm)	FS (0.1-0.25mm)	VFS (0.05-0.10mm)	Clay	Silt	Gravel content of field soil (g/kg)	Org. C (g/kg)	Ca (cmol/kg)	Bulk density (g/cm^3)
0-7	710	90	130	130	280	70	90	210	40	18.6	6.73	1.46
7-15	770	140	220	180	190	40	110	120	120	9.5	2.46	1.54
15-36	420	120	130	80	70	20	220	360	310	6.0	2.79	1.53
36-84	430	130	130	80	60	30	400	160	430	4.2	2.73	1.41
84-132	400	120	120	70	60	30	480	120	150	1.2	3.15	1.36
132-190	460	110	130	80	80	40	410	140	110	2.1	3.09	1.39

Particle size distribution ($g\ kg^{-1}$) = 0.1%

Particle size distribution of soil samples ≤ 2 mm is determined by pipet or hydrometer method (Gee and Bauder, 1986). Pre-treatment of samples ensures that samples are properly dispersed before determination. Variations exist from laboratory to laboratory on the pre-treatments. For instance, the removal of organic matter is skipped when it is known that this is very low in some soils while removal of iron oxides is usually considered optional.

The particle size distribution in Table 4 was determined by pipet method although it is often determined by hydrometer method in most laboratories. It is often not a common practice to separate sand in many laboratories but this is highly desirable for the evaluation of soil erodibility (Renard et al., 1997) and the application of fractal theory (Tyler and Wheatcraft, 1992; Anderson et al., 1998; Salako et al., 1999).

Particle size distribution is basic to the behavior of soils in terms of their chemical and physical properties. In spite of the higher clay content of the subsoil in Table 4, gravel concentration is higher. Thus when the topsoil is eroded, it is not just the chemical fertility that is reduced, there are more grave consequences with stones exposed which would make cultivation either with hand hoe or machinery difficult. Besides, the subsoils often exhibit more hardsetting properties (Mullins et al., 1990).

Aggregate stability and soil structure

In soils, the particles are aggregated and arranged in various forms. Soil aggregation is as a result of flocculation of clay and cementation of the particles through some chemical, biological and physical mechanisms. The arrangement of these particles into peds or aggregates is referred to as soil structure. Kay (1990) stated that soil structural form is the heterogeneous arrangement of solid and void space that exist at a given time while the stability is the ability to retain its arrangement of solid and void spaces when exposed to different stresses.

Aggregate stability is a measure of the resistance of the aggregates so formed to disruptive forces such as impact of rainfall. Aggregate stability with regard to water erosion can be measured by measuring resistance of aggregates wet-sieving (Yoder, 1936; Angers and Mehuys, 1993). In this method, aggregates placed on sieves are oscillated in a cylinder of water to disrupt the bonds cementing the soil particles. This is to simulate aggregates vulnerability to breakdown under the influence of runoff. Also, resistance ofaggregates to water drops are measured by bombarding the aggregates with water drops from a drop former (Bruce-Okine and Lal, 1975), which could be as simple as a buret.

Aggregates occur in soils in hierarchical size orders (Dexter, 1988). Soil aggregates can be broadly classified as macroaggregates (> 250 μm) and microaggregates (< 250 μm). Wet-sieving and water drop impacts are carried out to test macroaggregate (> 250 μm). Microaggregates (< 250 μm) are tested for stability usually by dispersion techniques, which in the simple form involves comparison of water-dispersible clay (and silt) with clay (and silt) dispersed in a dispersing agent like calgon. There are variations from laboratory to laboratory on pre-treatment of aggregates, period of we-sieving, nest of sieves selected, method of dispersion, etc.

Wet-sieving can be done by a single sieve or multiple sieves. When multiple sieves are used, the data are amenable to calculation of aggregate stability indices such as the mean-weight diameter MWD and geometric diameter GMD. For instance in some studies carried out in southwestern Nigeria (Salako et al., 1999; Salako and Hauser, 2001), wet sieving of 4-10 mm aggregate samples was carried out in a nest of sieves with 4, 2, 1, 0.25, 0.125 and < 0.125 mm openings. To calculate MWD and GMD,

$$MWD = \sum x_i WSA \tag{14}$$

where WSA is the proportion of water stable aggregates retained on sieves, and x_i the mean aggregate size obtained from mean sieve size (e.g., the mean sieve or aggregate size for aggregates retained on the 2 mm sieve is (4+2)/2) which is 3).

The geometric mean diameter GMD is another index based on the observation that distribution of aggregates after wet-sieving follows a log-normal distribution (Gardner, 1956).

$$GMD = \exp[\sum(w_1-w_2) \log (x_i/W) \tag{15}$$

where w_2 is the water stable aggregates after wet-sieving, w_1 the coarse materials in aggregates and W the initial sample mass

The time period soil aggregates are subjected to sieving vary as many other pre-treatments from laboratory to laboratory. In order to improve efficiency in laboratories, it is necessary to understand the adequacy of period used for sieving. Salako and Hauser (2001) found that there was no significant relationship in aggregate stability of coarse-textured Alfisols and Ultisols of southern Nigeria when sieved between 5 and 35 minutes (Table 5) because of the inherently week aggregation due to the dominance of sand. This was consistent with the position of Raine and So (1997) who showed that unstable soils might not be sensitive to increases in the amount of energy applied to break down their aggregates

Table 5. Variations of soil aggregate stability with we-sieving period for Alfisols and Ultisols in southern Nigeria*

	Alfisol (0-30 cm depth)			Ultisol (0-15 cm depth)		
Wet-sieving period (min)	MWD (mm)	GMD (mm)	WSA > 0.25 mm	MWD (mm)	GMD (mm)	WSA > 0.25 mm
5	3.55	1.20	57	3.55	1.09	57
10	5.09	1.21	76	3.80	1.11	63
15	3.05	1.12	51	4.11	1.23	64
20	5.10	1.20	76	3.81	1.11	61
25	2.98	1.10	46	3.08	1.05	55
30	3.59	1.19	58	2.34	1.08	43
35	3.82	1.06	61	3.11	1.04	51

* Source: Salako and Hauser (2001).

The study by Salako and Hauser (2001) showed that prolonged wet-sieving analysis (> 10 minutes) was not necessary for coarse-textured soils. Therefore, it was recommended that wet-sieving analyses for such soils be carried out within 10 minutes (5-minute wetting by immersion followed by 5-minute wet sieving). The study further showed that soil aggregate stability differed between soil types, land uses and soil management and its characterization should therefore be site specific.

Soil aggregate stability varied with season for the coarse-textured soils of southwestern Nigeria (Figure 4; Salako et al., 1999). The MWD was higher in the dry season (January) than the wet season (July and September) due to closer association of the soil particles in the dry season and their dispersion in the wet season. The intermediate period (March) between the dry and wet seasons had similar MWD to both seasons.

Dispersion ratio, which is the ratio of (silt + clay) in water to (silt + clay) in a dispersant such as calgon is an index of structural stability inasmuch as its magnitude is related to the stability the soil microaggregates. If the ratio is higher than 50% and could be up to 87% for the coarse textured Alfisol of southwestern Nigeria (Salako, 2001), the soils are structurally weak.

Describing particle size distribution and aggregate stability with fractal theory

There is a review of the application of fractal theory to soil studies by Anderson et al. (1998). Fractal theory has been applied to a wide range of topics in natural and physical sciences but appear to be used more by soil physicists among soil scientists, especially with regards to particle soil distribution and soil structural characterization.

Salako et al. (1999) and Salako (2001) used the number-size distributions of soil aggregates as described by a power law relationship to obtain fractal dimensions:

$$N_{>x} = k\, X^{-D}$$

where X is the mean sieve size obtained by finding the average value of successive sieve sizes, $N_{>x}$ the cumulative number of aggregates and D the fractal dimension. A log-log regression analysis between N and X provides values of k and D. Salako et al. (1999) reported that fractal dimensions D for 4-10 mm ranged from 2.29-2.72 with low D values associated with more stable aggregates under fallow and high D values associated with unstable aggregates under cultivation. Furthermore, the coarseness of soil influenced the D values with higher D values being observed for fine-textured soils. The increase in D after the removal of coarse fragments indicates the higher irregularity of aggregate shapes. Such separation of coarse and fine materials in soils occurs with soil erosion, and the implications can be described using the fractal theory.

Water transmission characteristics

Water entry into the soil by infiltration is a major determinant of how much rain will runoff and cause soil erosion. Coarse-textured soils have high water infiltration rates and hydraulic conductivity. For soils in southwestern Nigeria, saturated hydraulic conductivity could be as high as 525 $cm \cdot h^{-1}$, and after 2 hours, water infiltration rate may still exceed 72 $cm \cdot h^{-1}$ (Salako, 2002). Water infiltration rate is usually measured with a double-ring infiltrometer having an outer ring of 60 cm diameter and an inner ring of 30 cm diameter, both having a depth of 25 cm. The water head maintained in such a method is usually between 5 and 12 cm.

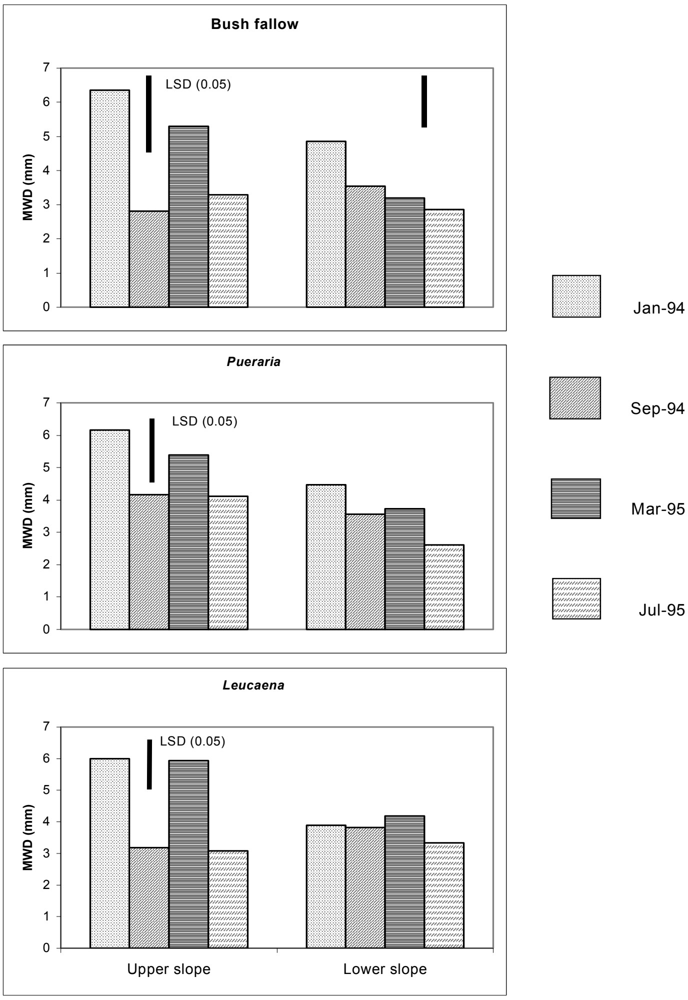

Figure 4. Temporal and spatial variations of soil aggregates in southwestern Nigeria (Source: Salako et al., 1999)

There have been recent advances in which both saturated and unsaturated flow characteristics can be measured with less strenuous equipment such as the disc permeameter. Preliminary studies using the disc permeameter in southwestern Nigeria showed that the sorptivity values may not actually be different from those obtained by double-ring nfiltrometer (Table 6) but the equations for calculating parameters such as the hydraulic conductivity would need to be revised for coarse-textured and heterogeneous tropical soils because unreasonable negative values were obtained for the hydraulic conductivity (Salako F. K. and Kirchhof, G., unpublished).

Table 6. A preliminary study comparing of sorptivity ($cm \cdot h^{-0.5}$) measured with double-ring and disc-permeameter on a coarse surface soil at Ibadan, southwestern Nigeria (Salako F. K. and Kirchhof G., unpublished)

Method	Sorptivity
Double-ring	135
Disc permeameter	104
LSD	NS

Evaluating soil erodibility and soil loss

Soil erodibility is actually defined as the susceptibility of soils to erosion. Particle size distribution, soil dispersion and aggregate stability have been used for many years as indices of soil erodibility (Bryan, 1968). Therefore, the foregoing discussion on soil characteristics are relevant to this sub-topic, which however, will be discussed in the context of the (Revised) Universal Soil Loss Equation , RUSLE (Renard et al., 1997).

Average annual soil erosion expected from field slopes is computed from the USELE or RUSLE as follows:

$$A = RKLSCP \tag{16}$$

where A is the computed spatial average soil loss and temporal average soil loss per unit of area ($t \cdot ha^{-1} \cdot yr^{-1}$), R the rainfall-runoff erosivity factor, K the soil erodibility factor which is the soil loss rate per erosion index unit for a specified soil as measured on a standard plot, which is defined as a 22.1 m length of uniform 9% slope in continuous clean fallow, L the slope length factor - the ratio of soil loss from the field slope length to soil loss from a 22.1 m length under identical condition, S the slope steepness factor - the ratio of soil loss from the field slope gradient to soil loss from a 9% slope under otherwise identical conditions, C the cover-management factor - the ratio of soil loss from an area with specified cover and management to soil loss from an identical area in tilled continuous fallow, and P the support practice factor - the ratio of soil loss with a support practice like contouring, strip cropping or terracing to soil loss with straight-row farming up and down the slope.

The USLE or RUSLE has been widely criticized (e.g., Hudson, 1993) for the fact than an empirical equation cannot be universally applicable. The review of USLE leading to RUSLE is an obvious attestation to this fact. Nonetheless, the RUSLE has proven to be a useful tool in many developing nations where there have been major constraints in establishing long-term experiments.

Soil erodibility can be evaluated using (i) runoff plots, (ii) rainfall simulators in laboratory or field plots and/or (iii) nomograph (Hudson, 1993; Lal, 1994). Runoff plots are very expensive to establish and maintain, and may not yield meaningful results for watershed management as the plots are often unrepresentative of watersheds (Hudson, 1993). Rainfall simulators vary in sophistication, and this would determine the cost involved in their use. The nomograph requires data on (i) percent silt + very fine sand, (ii) percent sand (0.10-2 mm), (iii) percent organic matter, (iv) soil structural class and (v) permeability. These properties have earlier been discussed. Comparison of soil erodibility values from runoff plots in southeastern Nigeria with the nomograph for its evaluation in the USLE showed that significant differences existed between the two methods (Vaneslande et al., 1984; Obi et al., 1989), suggesting some inadequacies in the nomograph. Such inadequacies could be due to the dominance of medium to coarse sand in the tropical soils whereas the nomograph was developed with soils having substantial fine particles. Obi et al. (1989) found that actual measurement of soil erodibility was 0.007 t·h/MJ·mm compared with 0.012 t·h/MJ·mm using the nomograph and 0.03 t·h/MJ·mm using rainfall simulator for a sandy loam soil in southeastern Nigeria. Table 7 gives a comparison of the nomograph with rainfall simulator estimates for different parent materials.

Table 7. Soil erodibility measured with erodibility nomograph and under rainfall simulator for soils from different parent materials in southeastern Nigeria

	Erodibility (t·h/MJ·mm)	
Parent material	Nomograph	Simulated rainfall
Sandstone	0.008	0.09
	0.002	0.02
	0.008	0.006
	0.03	0.08
	0.012	0.03
	0.058	0.05
Shale	0.013	0.006
Shale/Sandstone	0.042	0.03

Source: Obi et al. (1989)

Soil loss tolerance T is the maximum rate of annual soil loss that will permit plant productivity to be maintained economically and indefinitely. In the USA, upper limit T is 11.2 t·ha^{-1}·y^{-1} (Hudson, 1995; Renard et al., 1997) but for major soils in the humid and subhumid tropics it is less than 2 t·ha^{-1}·y^{-1} (Lal, 1985; Igwe, 1999). Either by estimation using USLE or by field plot measurements, soil loss or soil erosion from the coarse textured soils exceed a tolerance limit of about 2 t·ha^{-1}·y^{-1} for tropical soils. Obi et al. (1989) reported that soil loss from bare soils with very poor aggregation and tilled up and down the slope could be up to 59 t·ha^{-1}·y^{-1}. Lal (1997) reported annual soil losses in relation to slope length under conventional tillage in Ibadan southwestern Nigeria as 9.59 t/ha for 60 m long slope, 9.88 t/ha for 50 m, 6.84 t/ha for 40 m, 5.69 t/ha for 30 m, 1.27 t/ha for 20 m and 2.19 t/ha for 10m slopes. The slope length L and erosion Y relationship fitted a polynomial function:

$$Y = c + aL + bL^2 \tag{17}$$

where a, b and c depend on soil, slope, rainfall regime, and management practices.

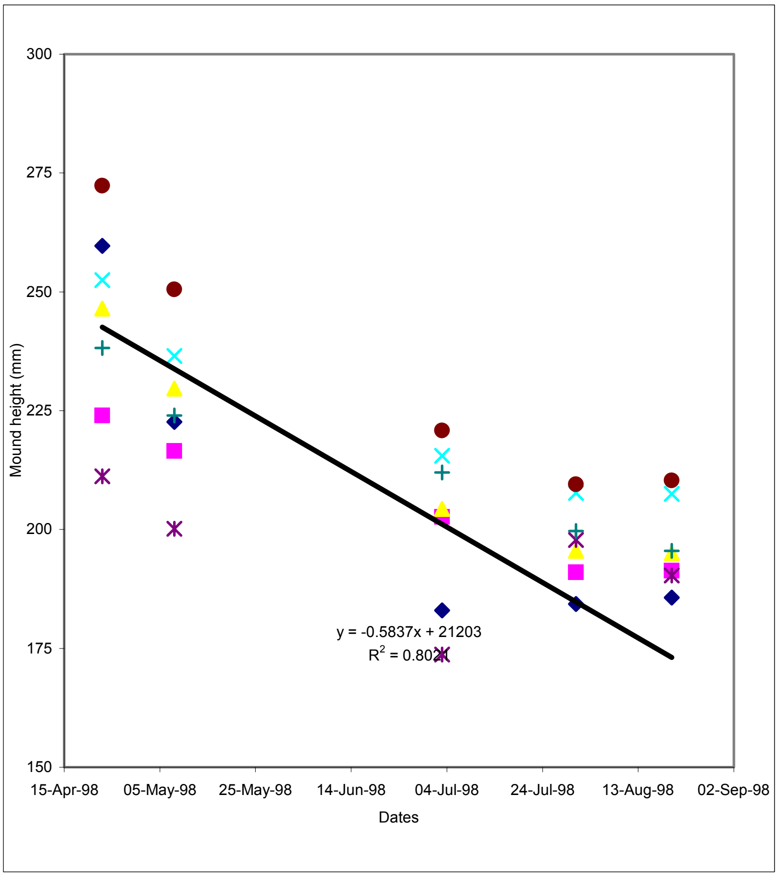

Figure 5. Changes in mound height as a measure of soil erosion, monitored during the rainy season with a profile meter in Ibadan, southwestern Nigeria in 1998 (Salako, F. K. and Kirchhof, G. unpublished data)

Profilemeters measure small changes in surface level along a cross section (Hudson, 1995) and can be found useful in detecting problems of sheet erosion, which can go unnoticed by ordinary visual observation. Soil erosion from mounds has been observed for the coarse-textured Alfisols in southwestern Nigeria (Figure 5, Salako, F. K. and Kirchhof G., unpublished). Mound tillage is a common practice for cultivation of tuber and root crops in the tropics and soil erosion studies have not focused adequately on this particular tillage practice. When depth of an area eroded is known with a profile meter, the volume of soil eroded can be calculated for a given area, and the mass of soil eroded obtained from the bulk density of the soil.

Other models and their application in the tropics

There are some other empirical models like the Soil Loss Estimation of Southern Africa (SLEMSA) (Elwell, 1977):

$$A = KCX \tag{18}$$

where A is the predicted mean annual soil loss, K the mean annual soil loss from a standard field plot on a 4.5% slope, C the crop management factor and X the slope length factor.

Process-based models now exist, requiring mathematical analyses to explain the physical processes involved in soil erosion. They are usually run with computing aids. The application of such models is rare in the tropics because of expertise and availability of data.

General discussion and recommendations

The various studies reported in this paper indicated a few areas where physics can be applied in Soil Science. Basic soil physics needs to grow along with applied soil physics but emphasis is currently on the latter because soil science is still strongly linked with food production and inter or multidisciplinary approach to research is not encouraged in many national institutions. Multi-disciplinary researches involving physicists, soil physicists and engineers like those of Kowal et al. (1973), Kowal and Kassam (1976), Lal (1976) and Lal (1992) have always led to simultaneous fabrication or design of equipment and use of the equipment for data collection. Very high rainfall intensities falling on coarse-textured soils in the tropics still cause considerable soil erosion in spite of high water infiltration rates. Often, this is due to weak soil structure. The processes involved are not yet adequately explained.

References

Aina, P. O. 1980. Drop characteristics and erosivity of rainfall in southwestern Nigeria. *Ife Journal of Agriculture* 2: 35-43.

Akinnifesi, F. K. and Salako, F. K. 1997. Growth of *Gmelina arborea* and *Senna siamea* and their effects on microclimate in southern Nigeria. *Tropical Ecology* 38: 55-63.

Anderson, A. N., McBratney, A. B. and Crawford, J. W. 1998. Applications of fractals to soil studies. *Advances In Agronomy* 63: 1-76.

Angers, D. A. and Mehuys, G. R. 1993. Aggregate stability to water. In: M. R. Carter (ed.) Soil Sampling and methods of analysis. Canadian Society of Soil Science/Lewis Publishers, Boca Raton, Florida, USA. pp. 651-657.

Armon, M.N. 1984. Soil erosion and degradation in southeastern Nigeria in relation to biophysical and socio-economic factors. Ph.D. Thesis, Department of Agronomy, University of Ibadan, Ibadan, Nigeria

Brown L.C. and Foster G.R., 1987. Storm erosivity using idealized intensity distributions. *Trans. ASAE* 30: 379-386.

Bruce-Okine, E. and Lal, R. 1975. Soil erodibility as determined by raindrop technique. *Soil Science* 119: 149-157.

Bryan, R. B. 1968. The development, use and efficiency of indices of soil erodibility. *Geoderma* 2: 5-26.

Burrough, P.A., 1993. Soil variability: a late 20th century view. *Soils and Fertilizer* 56: 529-562.

Carter C. E., Greer, J. D., Braud, H. J. and Floyd, J. M. 1974. Raindrop characteristics in south central United States. *Transactions of the ASAE* 6: 1033-1037.

Dexter, A. R. 1988. Advances in characterization of soil structure. *Soil & Tillage Research* 11: 199-238.

Eghball, B. and Power, J.F., 1995. Fractal description of temporal yield variability of 10 crops in the United States. *Agronomy Journal* 87: 152-156.

Eghball, B., Binford, G.D., Power, J.F., Baltensperger, D.D. and Anderson, F.N., 1995. Maize temporal yield variability under long-term and fertilizer application: Fractal analysis. *Soil Science Society American Journal* 59: 1360-1364.

Elseenbeer, H., Cassel, D.K. and Tinner, W., 1993. A daily rainfall erosivity model for western Amazonia. *Journal of Soil & Water Conservation* 48: 439-444.

Elwell, H. A. 1977. Soil loss estimation for southern Africa. Research Bulletin 22. Department of Conservation and Extension: Harare, Zimbabwe.

Gardner, W. R. 1956. Representation of soil aggregate-size distribution by a logarithmic-normal distribution . *Soil Science Society of America Proceedings* 20: 151-153.

Gee, G. W. and Bauder, J. W. 1986. Particle-size analysis. In: Klute A. (ed.) Methods of soil analysis. Part 1. Physical and mineralogical methods. 2nd edition ASA-SSSA, Madison, WI, USA. pp. 383-411.

Hartemink, A. E. 2002. Soil science in tropical and temperate regions-Some differences and similarities. *Advances in Agronomy* 77: 269-292.

Hudson, N, W. 1993. Field measurement of soil erosion and runoff. *FAO Soils Bulletin* 68:139 pp.
Hudson, N., 1995. Soil conservation. 3rd edition. Iowa State University Press, Ames, 391 pp.
Igwe, C.A. 1999. Land use and soil conservation strategies for potentially highly erodible soils of central-eastern Nigeria. *Land Degradation and Development* 10: 425-434.
Jagtap, S.S., 1995. Changes in annual, seasonal and monthly rainfall in Nigeria during 1961-1990 and consequences to agriculture. Discovery and Innovation 7: 337-348.
Kay, B. D. 1990. Rates of change of soil structure under different cropping systems. *Advances in Soil Science* 12; 1-52.
Kowal, J. M., Kijewski, W. and Kassam, A. H. 1973. A simple device for analyzing the energy load and intensity of rainstorms. *Agricultural Meteorology* 12: 271-280.
Kowal, J. M. and Kassam, A. H. 1976. Energy load and instantaneous intensity of rainstorms at Samaru, Northern Nigeria. *Tropical Agriculture* 53:185-197.
Lal, R. 1976. Soil erosion problems on Alfisols in western Nigeria and their control. Monograph No. 1. IITA, Ibadan.
Lal, R. 1985. A soil suitability guide for different tillage systems in the tropics. *Soil & Tillage Research* 5: 179-196.
Lal, R., 1992. Tropical agricultural hydrology and sustainability of agricultural systems. The West Bank project (1978-1988). Ohio State University/International Institute of Tropical Agriculture, 303 pp.
Lal, R. (editor) 1994. Soil erosion research methods. 2nd edition. 339 pp.
Lal, R. 1997. Soil degradative effects of slope length and tillage methods on Alfisols in western Nigeria. I. Runoff, erosion and crop response. *Land Degradation & Development* 8: 201-219.
Lal, R. 1998.Drop size distribution and energy load of rain storms at Ibadan, western Nigeria. *Soil & Tillage Research* 48: 103-114.
Lal, R. 2000. Physical management of soils of the tropics: Priorities for the 21st Century. *Soil Science* 165: 191-207.
Lawes Agricultural Trust, 1996. GENSTAT 5 Release 3.2, Rothamsted, UK
Mullins, C. E., MacLeod, D. A., Northcote, K. H., Tisdall, J. M. and Young, I. M. 1990. Hardsetting soils: Behavior, occurrence and management. *Advances in Soil Science* 11: 37-108.
Obi, M. E., Salako, F. K. and Lal, R. 1989. Relative susceptibility of some southeastern Nigeria soils to erosion. *CATENA* 16: 215-225.
Raine, S. R. and So, H. B. 1997. An investigation of the relationships between dispersion, power and mechanical energy using the end-over-end shaking and ultrasonic methods of aggregate stability assessment. *Australian Journal of Soil Research 35*: 41-53.
Renard, K.G., Foster, G.R., Weesies, G.A., McCool, D.K., and Yoder, D.C., 1997. Predicting soil erosion by water: A guide to conservation planning with the

revised universal soil loss equation (RUSLE). USDA, Agriculture Handbook No. 703, Washington DC, 404 pp.

Salako, F. K. 2001. Structural stability of an Alfisol under various fallow management practices in southwestern Nigeria. *Land Degradation & Development* 12: 319-328.

Salako, F. K. 2002. Ponded water flow in an Alfisol under different intensities and fallow systems in southwestern Nigeria. *Asset Series A* 2:156

Salako, F. K., Obi, M. E. and Lal, R. 1991. Comparative assessment of several rainfall erosivity indices in southern Nigeria. *Soil Technology* 4: 93-97.

Salako, F. K., Ghuman, B. S. and Lal, R. 1995. Rainfall erosivity in south-central Nigeria. *Soil Technology* 7: 279-290

Salako, F. K., Babalola, O., Hauser, S. and Kang, B. T. 1999. Soil macroaggregate stability under different fallow management systems and cropping intensities in southwestern Nigeria. *Geoderma* 91: 103-123.

Salako, F. K. and Hauser, 2001. Influence of different fallow management systems on stability of soil aggregates in southern Nigeria. *Communications in Soil Science and Plant Analysis* 32: 1483-1498

Tyler, S. W. and Wheatcraft, S. W. 1992. Fractal scaling of soil particle-size distributions: Analysis and limitations. *Soil Science Society of America Journal 56*: 362-369.

Vanelslande, A., Rousseau P., Lal, R., Gabriels, D. and Ghuman, B. S. 1984. Testing the applicability of a soil erodibility nomograph: In: Challenges in African Hydrology and Water Resources. Proceedings of Harare Symposium. July 1984. International Association of Hydrological Sciences Publication 144: 463-473.

Vieira, S. R., Hartfield, J. L., Nielsen, D. R. and Biggar, J. W., 1983. Geostatistical theory and application to variability of some agronomical properties. *Hilgardia* 51: 1-51.

Webster, R. and Oliver, M. A. 1990. Statistical methods in soil and land resource survey. Oxford University Press, New York, USA. 316 pp.

Wischmeier, W. H. and Smith, D.D., 1978. Predicting rainfall erosion losses: a guide to conservation planning. USDA Handbook No. 537. Washington DC, 58 pp.

Yoder, R. E. 1936. A direct method of aggregate analysis of soils and a study of the physical nature of erosion losses. *Journal of American Society of Agronomy* 28: 337-351.

Yu, B. and Rosewell, C.J., 1996. An assessment of a daily rainfall erosivity model for New South Wales. *Australian Journal of Soil Research* 34: 139-152.