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Analysis and synthesis of data on relationships between soil factors and soil erosion in South Africa

By

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Mini-dissertation submitted in partial fulfillment
of the requirements of the degree of
Magister Institutionis Agrariae
in Land Use Planning
In the faculty of Natural & Agricultural Science
University of Pretoria
Pretoria

December 2001

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Declaration

I declare that this mini-dissertation describes my original work, except where specific acknowledgement is made to the work of others, and has not previously in its entirety or in part been submitted for a degree to any other university

N.E Mulibana

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ACKNOWLEDGEMENTS

I would like to believe that in every single work, publication or dissertation completed by the researcher there are other people who contribute directly or indirectly, and this is not an exception. The author wishes to express his sincere gratitude towards the following people and institutions.

Prof. R.O. Barnard, my promoter, for the invaluable guidance, important remarks and encouragement during my studies. I believe that if it wasn't for him this dissertation would have been a wish.

The ARC Professional Development Programme (PDP) and ARC Institute for Soil, Climate and Water for financial assistance and for making it possible to complete this study.

Prof. M.C. Laker, for valuable guidance during the planning and initial stages of this investigation.

All the ARC-ISCW personnel: particularly Dr. D.P. Turner, Dr. C. Bühmann and Mr. G. Patterson for their time given to me in helpful discussions. Mrs. R. van Dyk for her professional assistance.

Thanyani Mulibana, your existence encourages me to work hard.

Pinki, thank you for being part of my life and I cherish all the things you have done for me.

Mr and Mrs Mulibana, thanks very much for grooming me to a person that I am today, you are the best parents I ever know in the world.

All my friends and relatives for their encouragement throughout the study.

May God bless you all.

ABSTRACT

South Africa is dominated by soils that are inherently unstable and extremely vulnerable to erosion. Soil erosion is most severe in the semi-arid regions although it is not confined solely to these areas. The Karoo, former Transkei, parts of the western Orange Free State and former Transvaal and a considerable part of Kwazulu Natal are subject to severe erosion by water.

Work reported in a number of post graduate studies on erosion in South Africa was critically evaluated and the most important results reported. Soil formation from different parent materials was investigated in a number of these studies. It was found that the stability of the soils developed on dolerite was higher than the stability of the soils on granite and sedimentary rocks. The soils on dolerite have lower stability with decreasing rainfall. The soils on granite do not show the same decrease in stability with a decrease in rainfall.

It was also found that some soils erode more easily than others under the same conditions of rainfall, vegetation and topography. This shows that the nature of the soil plays an important role in the occurrence and severity of erosion.

A distinction between hydraulic conductivity (HC) and infiltration rate (IR) was also studied. HC is usually measured under conditions where the soil surface is not disturbed. Considerable surface disturbance occurs when IR is measured, especially when precipitation is involved, leading to crust formation at the soil surface and thus to differentiated water transmission properties of the crusted layer compared with the underlying soil. The HC of the soil depends to a large extent on the exchangeable sodium percentage (ESP) of the soil and the salt concentration of the percolating solution.

Levy (1988) studied the effect of clay mineralogy and soil sodicity on the IR of soils subject to rain. The IR experiments were carried out using a laboratory rainfall simulator.

On the basis of the results presented by Levy, it is suggested that as far as clay mineralogy is concerned, the order of soil sensitivity to crust formation and its dependence on the level of soil sodicity is as follows: smectite>illitic> kaolinitic.

Threshold slope criteria were established for the dominant soils of three different pedosystems of the former Ciskei. Pilot areas were selected by means of aerial photographs and orthophoto maps.

Sumner (1957), D'Huyvetter (1985) and Bloem (1992) evaluated the effect of texture on erosion. Sumner and Bloem showed that clay content is the soil parameter with the biggest effect on surface sealing. Sumner indicated that crust formation could develop with any type of texture except sand with very low silt and clay. D'Huyvetter found that in all three main pedosystems namely, Mavuso, Keiskammahoek and Middledrift, topsoils with less than 20 per cent clay were found to be highly erodable while those which had a clay content of more than 20 per cent were found to be less erodable.

The effects of soil properties (e.g. texture, clay mineralogy, exchangeable cations, organic matter content) and experimental conditions (e.g. soil bulk density, moisture content, aging duration, prewetting rate) on rill erodibilities were investigated. Rill erosion depends on flow shear stress and stream power, the shear strength of the soil and cohesion forces between soil particles and the stream transporting capacity.

Stern (1990) studied the effects of slope and phosphogypsum (PG) application on seal formation, runoff and erosion from kaolinitic soils which vary in their response to rainfall and mechanisms which account for these variabilities were proposed. He found that PG was beneficial in increasing the IR of the soils. He also established that by diminishing chemical dispersion at the soil surface, the PG treatment slowed down seal formation and reduced removal of seal by erosion. The kaolinitic soils were divided into two categories: stable (S) and dispersive (D) soils. The presence of small amounts of swelling minerals caused dispersion, seal formation, high runoff and high soil loss on the dispersive kaolinite soils. Stable kaolinite soils did not contain swelling minerals.

Stern (1990) also studied the effects of surface application of mulch cover, PG, and polyacrylamide (PAM) on runoff during natural rainstorms. Runoff from control (bare) plots were high and during high intensities storm events most of the rain was lost through runoff. Mulch was beneficial in reducing runoff, indicating water infiltration was restricted by seal formation rather than by the hydraulic properties of the profile. PG reduced runoff to about half the amount from control plots. PAM treatment reduced runoff by three fold compared with the control treatment.

Knowledge of soil properties is essential for adequate land use planning in ensuring sustainable utilization of soil and preventing erosion and the mistakes of the past.

CONTENTS

CHAPTER 1

INTRODUCTION

1.1 THE SOIL EROSION SITUATION IN SOUTH AFRICA

South Africa is faced with problems of severe soil erosion and depletion of the soil resource base. It is estimated that in South Africa the rate of soil erosion, expressed per person of the population, is 20 times higher than the average for the world (Laker, 1990). It is of importance to note that once the topsoil is washed away it takes many years to recuperate.

Severe erosion has resulted from (a) overgrazing and (b) injudicious cultivation of land that is unstable. Soil erosion is endangering agricultural production, lowering farm incomes and escalating food prices. Overgrazing is undoubtedly the biggest cause of erosion in South Africa. It removes the dense grass cover, which is the only effective protection for vulnerable soil against erosion, because without a grass cover most of the soils tend to form a crust (D'Huyvetter, 1985). It is considered that overgrazing results from economic pressure, poor grazing management and excessive estimates of grazing capacity.

1.2 EFFECTS OF INCORRECT LAND USE PLANNING

Severe soil erosion due to incorrect land use planning in the former homelands has caused environmental and/or social disasters (Laker, 1990). In many areas the most severe soil erosion is found in cultivated plots which were identified and demarcated during "betterment" schemes and in the "rehabilitation" of the areas. Because the people involved in the planning had inadequate knowledge of how to correctly evaluate the qualities of the soil resources they used generalised norms. The result was widespread disastrous erosion despite "planning" and contouring when these areas eroded because of the inherent instability of the soils, most of which are in any case not arable because of poor cropping potential (Laker, 2000). The problem is that due to lack of basic data, a standard slope gradient criterion of 12 per cent was used to distinguish between arable and non-arable land in former homeland areas, such as the former Transkei and Ciskei in the Eastern Cape.

Hensley and Laker (1975) were the first to highlight this problem and indicated that different Eastern Cape soils vary widely with regard to their erodibility and that it is not possible to use a single slope value for all soils. Hensley and Laker (1978) also stated that "it must again be emphasized that careful attention must be given to the maximum steepness of slopes which are permitted for cultivated lands. An overall standard recommendation should not be used in all areas. The kind of soil and the climate of each area should be used to set standards for that specific area". It was not possible to define new criteria without the necessary research. This led to the research by D'Huyvetter (1985), aimed at developing appropriate slope criteria for different soil/climate combinations in the former Ciskei.

1.3 HISTORY OF SOIL EROSION RESEARCH AND SOIL CONSERVATION IN SOUTH AFRICA

Soil conservation became a big concern in South Africa just after world war II. This was because torrential rains and high winds in the early 1940's, following the very long and intense drought of the 1930's, caused extreme erosion. Soil erosion studies and soil conservation were almost exclusively in the hands of pasture scientists and engineers. They did absolutely sterling work, but inadequate knowledge of South Africa's unique's soil base hampered the development of appropriate criteria for determining the erodibilities of different soils.

South Africa's small band of soil scientists could not adequately cover all fields of soil science in the country and soil erosion was one of the fields in which they played virtually no role. A rare exception was the master's study of Sumner (1957).

From the middle 1980's to late 1990's a number of comprehensive studies on soil erosion and related topics were conducted by post-graduate students in soil science at the Universities of Fort Hare and Pretoria. These included the M.Sc. Agric. dissertations of D'Huyvetter (1985) at the University of Fort Hare and Smith (1990) at the University of Pretoria, as well as the Ph.D. theses of Levy (1988), Stern (1990) and Rapp (1998) at the University of Pretoria. Although the M.Sc. dissertation done at the University of Pretoria by Bloem (1992) dealt with overhead sprinkler irrigation, it contains much information that is relevant to a better understanding of factors determining the erodibilities of South African soils.

The dilemma of the lack of soil scientists in South Africa is well illustrated by the above list of students. Two-thirds of them were foreign students who were contracted to do the research, after which they returned to their countries. These are D'Huyvetter (Belgium) and Levy, Stern and Rapp (all Israel).

1.4 OBJECTIVES OF THE PRESENT STUDY

The above-mentioned dissertations and theses are separate loose-standing documents that are not accessible to many people. Publications from them also do not give complete, coherent pictures. The objectives of the present study were, therefore:

(a) To extract and summarize the main findings from the dissertations/theses of Sumner, D'Huyvetter, Levy, Stern, Rapp, Smith and Bloem mentioned under section 1.3. The research of Du Plessis and his co-workers from the 1980's (Agassi and Shainberg) is also reviewed.

(b) To synthesize the findings of these studies regarding the relationships between certain soil factors and the erodibilities of soils, so as to improve evaluation of the erodibilities of soils and land use planning in general.

1.5 STRUCTURE OF THE DISSERTATION

Chapter 2 gives an overview of international information on the relationships between various factors and the erodibilities of soils. One of the main objectives of this is to highlight differences between international perceptions and the realities of

South African soils as outlined in Chapters 3 and 4. Chapter 3 gives an analysis and summary of the dissertations and theses mentioned in Sections 1.3 and 1.4. Chapter 4 gives a synthesis of the results outlined in Chapter 3. Chapter 5 gives final conclusions and recommendations.

CHAPTER 2

PROCESSES AND MECHANISMS OF SOIL EROSION BY WATER

2.1 GENERAL

When rainwater reaches the soil surface it will either enter the soil or run off. Runoff occurs when the rainfall intensity exceeds the infiltration capacity of the soil. Water erosion is the result of the dispersive action of raindrops, the transporting power of water and also the vulnerability of the soil to dispersion and movement (Baver and Gardner, 1972). The process of water erosion can be separated into two components, rill and interrill erosion (Young and Onstad, 1978).

Interrill erosion (sheet erosion) is mainly caused by raindrop impact and removes soil in a thin, almost imperceptible layer (Foster, 1989). In interrill erosion the flow of water is generally unconfined, except between soil clods, and covers much of the soil surface. As the velocity of flow increases the water incises into the soil and rills form (Evans, 1980).

Rill erosion begins when the eroding capacity of the flow at some point exceeds the ability of the soil particles to resist detachment by flow (Meyer *cited by* Rapp, 1998). Soil is detached by headcut advance from knickpoints (De Ploey, 1989; Bryan, 1990), rill side sloughing and hydraulic shear stress (Foster *cited* by Rapp, 1998) as well as by slumping by undercutting of side walls and scour hole formation (Van

Liew and Saxton, 1983). These processes are usually combined into a detachment prediction equation as a function of average shear stress (Foster cited by Rapp, 1998). When rills develop in the landscape, a three to five fold increase in the soil loss commonly occurs (Moss, Green and Hutka 1982 and Meyer & Harmon 1984).

Rill erodibility depends both directly and indirectly on soil properties such as bulk density, organic carbon and clay content, clay mineralogy, cations in the exchange complex, soil pH and experimental conditions such as moisture content, aging of pre-wetted soil and quality of the eroding water (Rapp, 1998). Govers (1990) found that runoff erosion resistance of a loamy material was extremely sensitive to variation in the initial moisture content and to a somewhat lesser extent to changes in bulk density.

2.2 SPLASH EROSION

The slope gradient is an important factor governing the efficacy of splash erosion. A considerable amount of soil is eroded by the simple process of splashing. When a raindrop hits the soil surface it imparts a velocity to some of the particles, launching them into the air (Morgan, 1977). The higher the impact velocity, the greater the amount of soil splashed (Bisal, 1960).

It seems that the impact of raindrops becomes more effective when a thin film of water covers the soil surface. Maximum dispersion of soil particles occurs when the depth of water is about the same as the diameter of the raindrops (D'Huyvetter, 1985). An increase in wind speed and slope steepness also favours the process, especially on fine sandy soils, although soil particles are not moved far. Erosion and deposition of soil particles are in balance and only the surface of the soil is affected. This process is, however, still very important because it provides material which can subsequently be removed by running water (Evans, 1980).

When a raindrop hits the surface of the soil it also imparts a consolidating force, compacting the soil (Evans 1980). The consolidation force of the raindrops is best seen in the formation of a surface crust, which also results from the clogging of the pores as a result of the dispersal of fine particles from soil aggregates and their movement into the pores. This crust usually consists of a very thin (0,1 mm) non-porous layer and a zone of up to 5 mm thick of washed-in fine material (Evans, 1980).

2.2.1 Structure of the crust

Several studies have been conducted on the structure of the soil crusts resulting from rainfall. Dulay (1939) studied micrographs of crusts, obtained with an optic microscope with a magnification of X15, and found that the crust was a very thin layer, closely packed and with a higher density than the profile underneath. McIntyre (1958) found that the crust consists of two distinct parts: an upper skin seal, 0,1 mm thick, attributed to the accumulation of fine particles, and a "washed in" zone below it. The "washed in" zone was formed only in easily dispersed soil (McIntyre, 1958). Chen, Trachitzky, Broower, Morin and Banin (1980) examined scanning electron micrographs of crusts of loessal soils and also observed a thin skin seal about 0,1 mm in thickness. They did not, however, find an accumulation of fine particles in the 0,1

– 2,8 mm region as was observed by Gal, Arcan, Shainberg and Keren (1984). These researchers showed that the presence of the "washed in" zone depended on the exchangeable sodium percentage (ESP) of the soil and hence on the susceptibility of the soil to dispersion.

2.2.2 Factors affecting the formation and permeability of crusts

2.2.2.1 Effect of rain

Rain can be characterized by the following parameters: (1) rain intensity, (2) raindrop median diameter and (3) final velocity of the median drop (Levy, 1988). The relationships among these parameters were examined (Laws, 1940; Wischmeir and Smith, 1958) and it was found that drops with a large diameter reach a high final velocity and *vice versa*. Furthermore, Wischmeir and Smith (1958) observed an increase in the percentage of big drops with an increase in rain intensity. It was also noted that the volume of the median drops and their final velocity govern the rate of crust formation (Ellison, 1947). On the other hand, when the soil surface is protected by vegetation and raindrop impact thus prevented, no crust was observed at the soil surface and hardly any reduction in the permeability was noticed (Duley, 1939; Morin and Benyamini, 1977), emphasizing the vital importance of plant cover.

2.2.2.2 Effects of soil properties

2.2.2.2.1 Physical factors

Soil texture, and especially clay content, affects crusting. Betrand and Sor (1962) found that the higher the clay content the more aggregated the soil surface remained during rain, while the rate of crust formation was reduced. Conversely high sand and silt contents enhance crusting. Kemper and Noonan (1970) studied the effect of sand content and found that when sand (0,2 mm – 2 mm diameter) content was greater than 80 percent the soil maintained a high permeability. Medium textured soils (approximately 20 % clay) were found to be the most susceptible to crusting (Ben-Hur, Shainberg, Bakker and Keren, 1985).

Aggregation and aggregate size distribution at the soil surface are other important factors, since crusting is related to aggregate breakdown. Well-aggregated soils must break down into fine sizes before compaction and seal formation occur (Epstein and Grant, 1973). This suggests that aggregate breakdown and seal formation would continue progressively under drop impact. Moldenhauer and Koswara (1968) stated that a rapid decrease in soil permeability is due to unstable structure. They added that this could be corrected by increasing aggregate size by using a suitable tillage practice. Moldenhauer and Kemper (1969) found that the larger the aggregates, the higher the permeability of the crust formed.

Farres (1978) suggested that initial mean aggregate size determines the thickness of the crust and that the rate of crusting increases with a decrease in mean aggregate size. The effect of water content of the soil on the rate of crusting and crust permeability has also been studied (Duley & Kelly 1941and Levy, Shainberg & Morin 1986). They found that water content and the depth of wetting front had very little effect on the permeability of the crust. It increased with a decrease in water content at the beginning of each storm, when subjected to consecutive rainstorms.

2.2.2.2.2 Chemical factors

The hydraulic conductivity (HC) of the soil depends to large extent on the exchangeable sodium percentage (ESP) of the soil and the salt concentration of the percolating solution (Quirk and Schofield, 1955). As with HC, the permeability of soil exposed to rain is affected by the exchangeable cation species (Rose, 1962) and the quality of the rain water (Oster and Schroer, 1979). Rain water, being salt-free, leaches the salts, thus decreasing the salt concentration below the flocculation value, which in turn causes clay dispersion and enhances the breakdown of aggregates at the soil surface (Levy, 1988).

The HC of a soil is correlated with soil texture, mainly clay content. The higher the clay content the lower the HC (McNeal, Layfield, Norvel and Rhoades, 1968). Clay mineralogy is also an important factor influencing the HC of the soil. Soils rich in iron and aluminium oxides maintain a high HC and prevent the combined deleterious effect of exchangeable sodium and low salt concentration in the soil solution McNeal and Coleman, 1966; McNeal *et al.*, 1968; Cass and Sumner, 1982). McNeal *et al.*, (1968) and EL- Swaify (1973) found that red soil colour together with high free iron and aluminium contents is associated with high stability and hydraulic conductivity values of soils. Further evidence for this phenomenon was given by Du Plessis and Shainberg (1985). They found that some South African red soils have very stable hydraulic conductivity properties.

Aggassi, Morin and Shainberg (1985) also studied the interaction between the physical effects of raindrops and the chemical effects of the composition and concentration of applied water. They found that in a situation where both mechanisms (i.e. rain with energy and distilled water causing chemical dispersion) were in operation, crusts with low permeability (3 mm/h) were formed, even in a soil with low ESP. When the chemical effect was diminished by using saline water, they obtained crusts with a relatively high permeability (8,7 mm/h). On the other hand, when rain with very low energy (fog-type rain) was used together with distilled water, a limited reduction in the permeability of the soil was observed. This reduction was related to the changes in HC of the soil profile in accordance with the ESP of the soil when no crust was evident at the soil surface. On the basis of these results, Agassi *et al.* (1985) concluded that, in the absence of a physical mechanism, the chemical one does not come into effect at low ESP levels. However, the chemical mechanism needs some activation energy for it to start operating at the soil surface, which in this case was provided by the impact of the raindrops.

Quirk and Schofield (1955) showed that the hydraulic conductivity of a given soil decreases with increasing exchangeable sodium percentage, provided that the electrolyte concentration is below the critical threshold value. An increase in both ESP and clay content caused a decrease in final infiltration rate. After the soils were separated as chemically dispersive or stable on the basis of ESP, clay mineralogy, organic content and calcium to magnesium ratio, the final infiltration rate of both groups correlated well with clay content.

2.2.2.2.3 Clay mineralogy

Bryan (1974) considered clay mineral type as an important factor controlling the stability of soil aggregates and hence erodibility. Soil mineralogy is often implicated in inhibition of soil water movement. In some cases, the mineralogical influence is primarily physical. In other cases, the mineralogical influence is related to the chemical properties of the minerals or to the response of particular minerals to their physical or chemical environment (McNeal and Coleman, 1966; Yaron and Thomas, 1968). The chemical dispersion of soil depends on clay mineralogy, exchangeable ion composition and the electrolyte concentration in the soil solution (Stern, Ben-Hur and Shainberg, 1991).

Smectite and illite clays are known to be more dispersive than kaolinite clays. Soils which contain pure kaolinite form stable aggregates, maintain high IR and have low erosion. Conversely, kaolinitic soils which contain small amounts of smectites are dispersive. Soils which do not contain smectite are more stable, less erodable, and less susceptible to seal formation (Stern *et al.*, 1991).

Increasing proportions of 2:1 swelling clays in the clay mineral suite of a soil reduces the stability of the soil in the presence of dispersing cations such as sodium. This is simply a function of double layer chemistry (Singer, Janitzly and Blakar, 1982). Normally soils rich in clay are regarded as being more stable than those with low clay contents.

Swelling of clay and movement and deposition of dispersed clay particles may clog the soil pores. McNeal and Coleman (1966) and Yaron and Thomas (1968) concluded that soils which are more susceptible to dispersion are those high in 2:1 layer silicates (especially montmorillonite), while those high in kaolinite and sesquioxides are less susceptible. Velasco-Molina, Swobada and Godfrey (1971) concluded that in the virtual absence of electrolyte, the order of soil dispersion at a given ESP was montmorrilonitic > kaolinitic and halloysitic >micaceous. Arora and Coleman (1979) concluded that clay mineral susceptibility to deflocculation was in the decreasing order of illite, vermiculite, smectite and kaolinite. This implies that soils with illitic clay are more dispersive than soils dominated by montmorillonitic clays, especially at relatively low sodium adsorption ratios (SAR).

2.3 SURFACE WASH

Overland flow is initiated on slopes during heavy rainstorms when the rainfall intensities exceed the local infiltration capacity (Horton overland flow) or by localised saturated conditions (saturated overland flow) (Gerrard, 1981).

Micro-topography has a big influence on the type of surface flow that occurs. Unconcentrated flow in thin sheets is only possible on fairly smooth surfaces. It is rarely in the form of a sheet of water with uniform depth, but varies greatly in the character of laminar and turbulent flow (Gerrard, 1981). Emmet (1978) noted linear concentration of flow within sheet wash.

The hydraulic character of a flow can be described by its Reynolds number (Re) which is an index for its turbulence. The greater the turbulence, the greater the erosive power generated by the flow (Morgan, 1979).

Flow velocity is an important factor in this hydraulic relationship. The velocity must attain a threshold value before it becomes erosive (D'Huyvetter, 1985). This is related to the inherent resistance of the soil. The critical velocity is dependent upon the particle size distribution. For particles smaller than 0,5 mm the critical velocity increases with grain size as shown in Figure 2.1.

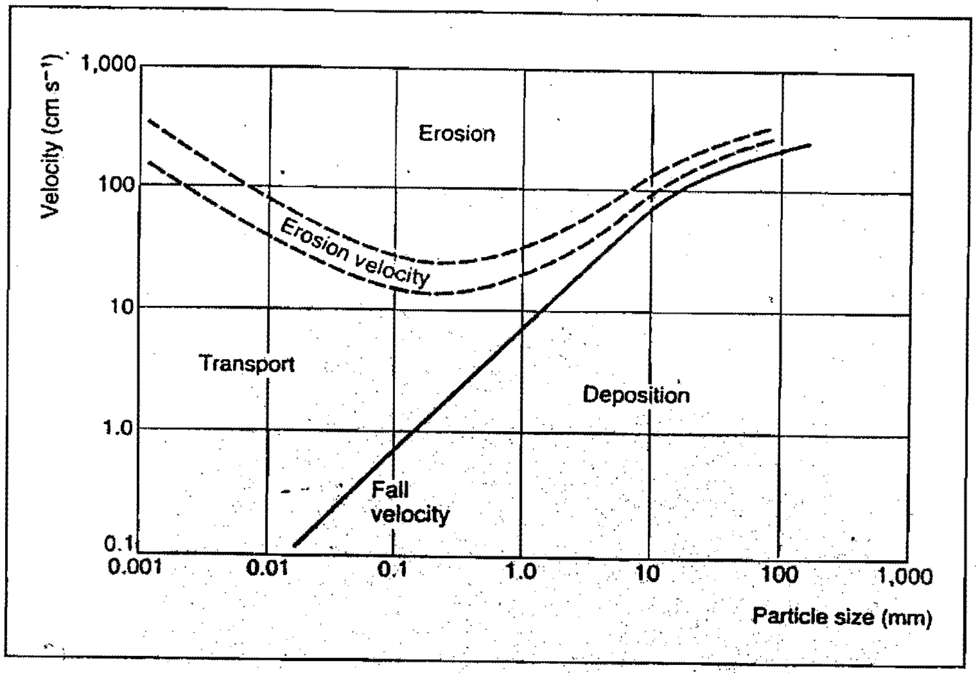

Fig 2.1 Critical water velocities for erosion, transport and deposition as a function of particle size (From: Hjulstrom, 1935).

The erosive capacity of unconcentrated flow will be slight and only very fine particles will be transported. The natural surface is usually too rough to allow substantial amounts of uniform flow. Only where local flow concentration occurs is some erosion possible (Gerrard, 1981).

Morgan (1977) also indicated that erosion does not take place uniformly across a slope. Only where water is confined between soil clods is there evidence of erosion. When rain splash and sheet erosion are acting together, both processes are much more efficient. This is because the soil particles are brought into suspension by rain splash and then transported by sheet flow.

Rills and gullies are formed when the velocity of the water increases and flow becomes more turbulent (D'Huyvetter, 1985). The increase in the hydraulic gradient can be the result of an increased slope gradient, increase in rainfall intensity or because surface storage is exceeded and incision takes place (Evans, 1980).

Gully erosion usually represents a permanent loss of soil where agricultural production proceeds without appropriate protective measures and recultivation. Gullies can be developed as enlarged rills, but their initialazisation can also be a more complex process. Gerrard (1981) noticed that rills are usually associated with silt or clay soils. The erosive character of rills and gullies is very high. They remove much larger volumes of soil per unit area than sheet wash does (D'Huyvetter, 1985).

2.4 FACTORS AFFECTING WATER EROSION

2.4.1 Rainfall factors

The interaction between raindrop size, shape, duration of a storm, and wind speed controls the erosive power of rainfall (D'Huyvetter, 1985). The erosivity of the rainfall is expressed in terms of kinetic energy and is affected by various factors.

According to Wischmeier and Smith (1965), the intensity of rainfall is closely related to the kinetic energy, according to the regression equation:

$$E = 1{,}213 + 0{,}890 \log I$$

Where:

E = the kinetic energy, ($kg.m/m^2.mm$)

I = rainfall intensity (mm/h)

Raindrop size, distribution and shape all influence the energy momentum of a rainstorm. Laws and Parson (1943) reported an increase in median drop size with increase in rain intensity. The relationship between median drop size (D_{50}) and rainfall is given by:

D_{50}: 2.23 $I^{0,182}$ (inch per hour)

Gerrard (1981), stated that the median size of raindrops increases with low and medium intensity fall, but declines slightly for high intensity rainfall.

The kinetic energy of a rainstorm is also related to the velocity of the raindrops at the time of impact with the soil (D'Huyvetter, 1985). The distance through which the raindrop must fall to attain its terminal velocity is a function of drop size. The kinetic energy of a rainstorm is related to the terminal velocity according to the equation:

$E_k = IV^2/2$

Where: E_k = Energy of the rainstorm

I = Intensity

V = Velocity of raindrop before impact

Ellison (1945) developed an equation describing the relationship between the soil detached by splashing, terminal velocity, drop diameter and rainfall intensity:

$E = KV^{4,33}\, d^{\,1.07}\, I^{0,63}$

Where: E = Relative amount of soil detached

K = Soil constant

V = Velocity of raindrops (ft/sec)

d = Diameter of raindrops (mm)

I = Rainfall intensity

Wind velocity accompanying a storm also influences its kinetic energy and hence the erosive capacity (D'Huyvetter, 1985).

2.4.2 **Soil factors**

According to Baver *et al.* (1972), the effect of soil properties on water erosion can be in two ways: Firstly, certain properties determine the rate at which rainfall enters the soil. Secondly, some properties affect the resistance of the soil against dispersion and erosion during rainfall and runoff.

An important soil property in regard to erodibility is the ***particle size distribution***. Generally it is found that erodable soils have low clay content (D'Huyvetter, 1985). Soils with more than 30 – 35% clay are often regarded as being cohesive and having stable aggregates which are resistant to dispersion by raindrops (Evans, 1980). On the other hand Evans (1980) also stated that sands and coarse loamy sands are, as a result of their high infiltration rate, not easily eroded by flowing water. In contrast soils with a high silt and/or fine sand fraction are very erodable.

The proportion of ***water-stable aggregates*** with a diameter less than 0,5 mm is a good index for erodibility. The erodibility of soil increases with the proportion of aggregates less than 0,5 mm (Bryan, 1974). Factors which contribute to aggregate stability include: organic matter content, root secretions, mucilaginous gels formed by the breakdown of organic matter, the binding of particles by sesquioxides and the presence of a high Ca concentration on the exchange sites of the of the colloids, instead of a high sodium content (D'Huyvetter, 1985).

The ***soil profile*** often determines the depth of the erosion feature (Evans, 1980). According to him soil horizons below the A horizon or plough layer are often more compact and less erodable. The texture and chemical composition of the sub-surface horizon can also have an adverse effect, however, for example:

- Soils with structured prismatic B horizons are not only poorly drained, but once they are exposed, they are very susceptible to erosion, due to dispersion resulting from the presence of a high Na concentration on the exchange sites of the clay particles (D'Huyvetter, 1985).

- Soils with a dense massive structure or well developed platy structure have impeded drainage which can cause severe erosion.

Normally deep gullies can be cut if the ***parent material*** is unconsolidated. If resistant bedrock is near the surface only rills will develop. Soil rich in surface ***stones*** are less susceptible to erosion (Lamb, 1950 and Evans, 1980). Stones protect the soil against erosion and also increase the infiltration of the flowing water into the soil.

The ***antecedent soil moisture*** and the ***surface roughness*** are both regarded by Evans (1980) as important soil factors affecting erosion. The ability of a soil to accept rainfall depends on the moisture content at the time of the rain. It will attain its final infiltration rate more quickly when the soil is already wet.

2.4.2.1 Factors affecting aggregate stability

Soil structure is determined by the shape and size distribution of aggregates. Aggregate size and strength determine the physical properties of a soil and its susceptibility to breakdown due to wind or water forces. Their stability in the field will have a decisive effect on soil physical, and thus also water conducting, properties (Lynch and Bragg, 1985). The main binding materials giving stable aggregates in the air dry state are the glueing agents in organic matter (Chaney and Swift, 1984; Tisdale and Oades, 1982) and sesquioxides (Goldberg and Glaubic, 1987).

2.4.2.1.1 Organic matter

Organic matter can bind soil particles together into stable soil aggregates. The stabilising effect of organic matter is well documented. Little detailed information is available on the organic matter content required to sufficiently strengthen aggregates with ESP values > 5-7, and containing illite or montmorrillonite, so as to prevent their dispersion in water (Smith, 1990). Van Beekom, Van den Berg, De Boer, Van der Malen, Verhoeven, Westerhof and Zuur (1953) noted that a high humus content made soil less susceptible to the unfavourable influence of sodium. Kemper and Koch (1966) also found that aggregate stability increased with an increase in the organic matter content of soil. A maximum increase of aggregate stability was found with up to 2% organic matter, after which aggregate stability increased very little with further increases in organic matter content. Unstable soil conditions are associated with decreasing organic matter contents of soils.

2.4.2.1.2 *Aluminium and iron oxides*

The soil used by Kemper and Koch (1966) contained relatively little free iron, although it did contribute to aggregate stability. Their data show a sharp increase of free iron from 1 to 3%. Goldberg and Glaubic (1987) concluded that Al-oxides were more effective than Fe-oxides in stabilizing soil structure against the dispersive effect of Na, probably because of the size and morphology of the oxide particles. Al-oxides have a greater proportion of sub-micrometer size particles in a sheet form as opposed to the spherical form of the Fe-particles. This implies higher surface charge densities that may bind particles together. Shainberg, Singer and Janitzky (1987) compared the effect of aluminium and iron oxides on the hydraulic conductivity of a sandy soil. They found that iron treatments were more effective than Al treatments. Effluent from HC measurements was not turbid, which indicated that the Fe and Al were both able to prevent clay dispersion.

2.4.3 Slope factors

Slope characteristics are important factors in determining the amount of runoff and erosion (D'Huyvetter, 1985). As the slope gradient increases, runoff and erosion usually also increase (Stern, 1990).

The three main components of topography which affect soil erosion processes are steepness, slope length and slope shape (D,Huyvetter, 1985). As a result of the increased downslope component of gravity, the erosion potential is greater on steep slopes and also on long slopes because of a down slope increase in surface flow (Baver *et al.*, 1972). Foster, Meyer and Onstad (1976) presented a conceptual model that showed that at lower slopes, interill transport determined erosion, while at steeper slopes, raindrop detachment determined it. The uniform or nearly flat bed characteristics of sheet-flow transport tend to be replaced by channels because of instability and turbulent flow effects (Moss, Green and Hutka, 1982). When channels are formed, rill erosion becomes the dominant mechanism in water erosion. As a result, most interill catchments slope downwards towards rill segments and supply their solids to the rill system (Stern, 1990).

There are many empirical relationships relating soil transport by surface wash to slope length and slope gradient. Zingg (1940) showed that erosion varied according to the equation:

$S = X^{1.6} \tan B^{1.4}$

Where: S = Soil transport cm/year

X = Slope length (m)

B = Slope gradient (%)

Studies conducted by Gerrard (1981), showed that plane and convex slopes did not differ significantly in the amount of soil lost by surface runoff, but concave slopes were less eroded. He found that in the upper part of a convex-concave slope the soil is usually severely eroded. In addition the lower slope is covered with slope wash material. The reverse is true if the intensity of erosion is determined on the basis of size and quantity of rills and dongas (Gerrard, 1981).

2.4.4 Vegetative factors

The effects of vegetation can be classified into three categories:

(a) The interception of raindrops by the canopy (D'Huyvetter, 1985). This has two effects: firstly, part of the intercepted water will evaporate from the leaves and stems and thus reduce runoff. Secondly, when raindrops strike the vegetation, the energy of the drops is dissipated and there is no direct impact on the soil surface. The interception percentage depends on the type of crop, the growth stage and the number of plants per unit area.

(b) A well distributed, close growing surface vegetative cover will slow down the rate at which water flows down the slope and will also reduce concentration of water (D'Huyvetter, 1985). As a result of this, it will decrease the erosive action of running water.

(c) There is also the effect of roots and biological activity on the formation of stable aggregates, which results in a stable soil structure and increased infiltration that reduces runoff and decreases erosion (D'Huyvetter, 1985). Increased permeability also reduces erosion as a result of increased water percolation due to better drainage. Stable aggregates in the topsoil also counteract crusting.

CHAPTER 3

COLLATION AND INTERPRETATION OF INFORMATION FROM THESES AND DISSERTATIONS DEALING WITH SOIL-RELATED STUDIES ON WATER EROSION IN SOUTH AFRICA

3.1 DATA EXTRACTION

Data were extracted from different theses, dissertations and other relevant material covering soil-related studies on water erosion in South Africa. The rationale was that that data from different studies on relationships between soil factors and soil erosion in South Africa are found in various theses and dissertations, but the information is scattered and uncoordinated. It was, therefore, recommended that it should be put together in one mini-dissertation so as to make it user friendly and more readily available.

The results presented by the different authors were not always in the same format and units, sometimes making comparisons difficult. As much standardization as possible has been introduced into this discussion.

The following theses and dissertations were included in the study:

Smith, HJC (1990). The crusting of red soils as affected by parent material, rainfall, cultivation and sodicity. *M.Sc.Agric., Univ. Pretoria*. (Section 3.2)

D'Huyvetter, JHH (1985). Determination of threshold slope percentage for the identification and delineation of arable land in Ciskei. *M.Sc.Agric., Univ. Fort Hare*. (Section 3.3)

Levy, GJ (1988). The effects of clay mineralogy and exchangeable cations on some of the hydraulic properties of soils. *Ph.D., Univ. Pretoria*. (Section 3.4)

Sumner, ME (1957). The physical and chemical properties of tall grass veld soils of Natal in relation to their erodibility. *M.Sc.Agric., Univ. Natal.* (Section 3.5)

Rapp, I (1998). The effects of soil properties and experimental conditions on the rill erodibilities of selected soils. *Ph.D., Univ. Pretoria.* (Section 3.6)

Stern, R (1990). Effects of soil properties and chemical ameliorants on seal formation, runoff and erosion. *Ph.D., Univ. Pretoria.* (Section 3.7)

Bloem, AA (1992). Criteria for adaptation of the design and management of overhead irrigation systems to the infiltrability of soils. *M.Sc., Univ. Pretoria.* (Section 3.8)

In this chapter the main findings from each of these studies are discussed and interpreted separately. In Chapter 4 the findings are integrated according to the main factors determining water erosion in South Africa.

3.2 SMITH, HJC (1990): THE CRUSTING OF RED SOILS AS AFFECTED BY PARENT MATERIAL, RAINFALL, CULTIVATION AND SODICITY

3.2.1 General background and objectives of the study

The study was conducted on soils from different localities in the former Transvaal. The localities of the soils sampled were Marken, Towoomba, Sabie, Krugersdorp, Irene and Amsterdam. The soils were all of the Hutton soil form (Soil Classification Working Group, 1977).

The main objectives of the study were to compare the stabilities of soils developed on basic and acidic igneous rocks respectively, as well as the effect of degree of weathering on the stability of red soils.

3.2.2 Summary of introductory remarks by Smith

Generally it has been found in South Africa that the hydraulic properties of red soils are much better than those of non-red soils, and that red soils are much more stable against erosion. The presence of high iron and aluminium oxide contents makes these soils stable (Smith, 1990). The presence of organic material in higher concentrations in soil profiles is also accepted to enhance the stability of soils. These phenomena are not shared by all red soils, however (Smith, 1990).

The importance of crust formation in South African red soils was shown by Du Plessis and Shainberg (1985). Infiltration rate values obtained from their rainfall simulation studies in crusting soils were one order of magnitude lower than the saturated hydraulic conductivity values for the same soils under the same conditions, but where no crust was formed. Du Plessis and Shainberg (1985) also found that some red soils did not have the very high stability properties usually ascribed to such soils and were in the same class as some non-red soils. Furthermore, Van der Merwe (1973) also found no decrease in hydraulic conductivity with decreasing amounts of free iron between 0,4 and 4,9 per cent in soils. Alperovitch, Shainberg, Keren, and Singer (1985) found that even soils with moderate amounts of sesquioxides were susceptible to erosion although they were red-coloured.

3.2.1.1 Soil formation on basic and acidic rocks

As a general background, Smith (1990) discussed soil formation from different parent materials.

The origin of most soils can be traced back to the transformation and changes of igneous rocks, although many soils also originate from already weathered products which are the main parent materials for the formation of sedimentary rocks, which could, with the addition of heat and/or pressure, also have been transformed when the original magma cooled (Smith, 1990). Mainly two types of magma are generally recognised and the basis of differentiation between the two is their content of silica. One is described as acidic and contains more than 60% SiO_2, has relatively high amounts of Na and K, but has low amounts of iron, magnesium and calcium (Smith, 1990). The other is basic and has less than 50% SiO_2. The basic magma is relatively rich in iron, magnesium and calcium. A further differentiation according to the silica content of the magma results in a broad classification of rocks. Rocks that contain more than 66% SiO_2 are acidic, 52-66% intermediate, 45-52% basic and less than 45% ultra-basic (Deer, Howie and Zussman, 1978). Table 3.1 shows the average compositions of abundant elements in various igneous rocks (Turekian and Wadepohl, 1961). Basic and ultra-basic rocks are recognised internationally as very important parent material from which soils of agricultural importance originate. In South Africa, dolerite is a very important example of such rock. It is widespread throughout the country in the form of dikes and sills with limited geographical extent. Dolerite is rich in bases, especially calcium and magnesium. This is because it contains 46% plagioclase and 37% augite (total: 83%) which are both rich in bases (Laker, 1990). It can thus be expected that soils developed on basic (dolerite) parent material have higher amounts of clay (higher amounts of clay forming minerals present) and also higher pH values generally (higher amounts of basic cations) as compared to soils developed on acidic parent material. This will only be true for soils formed under the same climatic conditions (Buol, Hole and McCracken, 1973).

Table 3.1 Average composition of igneous rocks in mg kg^{-1}.

Element	**Ultrabasic**	**Basic**	**High Ca Granite**	**Low Ca granite**
Si	205	230	314	347
Al	20	78	82	72
Fe	94	86	30	14
Ca	25	76	25	5
Mg	204	46	9	2
Na	4	20	28	26
K	0.04	8	25	42
Ti	0.30	14	3.4	1.2
Mn	1.6	1.5	0.54	0.4
P	0.22	1.1	0.92	0,6

3.2.2.2 Weathering in soils

Weathering can be considered as a combination of material destruction and material synthesis. Many soils have features which relate them to the parent rock from which they were formed. The resemblance is often expressed by the resistance of certain minerals (from parent rock) to weathering and which consequently still occur in the clay size fraction of soil (Smith, 1990). Weathering of rock constituents will release ions into the weathering environment to start the pedogenesis process in a soil profile. Some minerals, like quartz, muscovite, mica and some feldspars are more resistant to weathering than others. Generally these are the more acid minerals. These stable minerals will accumulate in a soil profile while the more basic minerals such as pyroxenes and amphiboles, are more easily altered by weathering and are also less commonly found in soils (Buol *et al.*, 1973). Conditions in the soil will ultimately determine the types of clay minerals formed from weathered products. When Si and basic cations such as Ca and Mg are not removed by leaching, 2:1 type clay minerals, such as smectites and micas (the latter where there is also abundant K) will dominate in the soil environment. This will occur under arid to sub-humid conditions or where soils are poorly drained. Highly weathered soils in humid tropical and subtropical areas, where intensive weathering and leaching of bases and silica has occurred, are dominated by minerals representing advanced stages of weathering, such as aluminium hydroxide (gibbsite) and mixtures of iron and manganese oxides. Kaolinite, a 1:1 type alumino-silicate clay mineral, is dominant where weathering is somewhat less intensive and silica concentrations higher than 2 mg/kg (Patterson, 1967; Curtis and Spears, 1971). The common mineralogical end points for weathering include goethite, hematite, gibbsite, 1:1 layer clay minerals and coarse-grained quartz (Jackson, Tyler, Willis, Boubeau and Pennington, 1968). According to them, this indicates relative immobilities of SiO_2, Al_2O_3 and Fe_2O_3. According to Chesworth (1973) an example of an order of mobility would be $Al^{3+} < Fe^{3+} < Si^{4+} < Fe^{2+} < K^{+} < Ca^{2+} = Mg^{2+} = Na^{+}$, although the absolute order of mobility is less important than the recognition that the relatively immobile components, expected to accumulate with increasing weathering intensities in soils, are Al_2O_3 and Fe_2O_3.

3.2.3 Research Technique

The studies were conducted on six red apedal (Hutton form) soils representing different combinations of parent material and degree of weathering (Table 3.2).

The studies made use of a laboratory scale rotating disk type rainfall simulator, similar to that used by Morin, Goldberg and Seginer (1967), to study the effect of crusting on the hydraulic properties of soils. Surface crusts are general thin (2 –3 mm) and are characterised by greater density, finer pores and lower saturated conductivity than the underlying soil. These crust properties increase erosion by reducing infiltration and thus increasing runoff. Crust formation and the resulting decrease in infiltration rates were studied and the data used for evaluating soil stability. The smaller than 4 mm size fraction of the soil samples was used and sample were packed in 300 × 500 mm boxes, 20 mm deep over a 80 mm deep layer of coarse sand. The boxes were placed in the rainfall simulator at a slope of 5% and subjected to high energy rain until a steady final infiltration rate was observed (Agassi and Du Plessis, 1984).

Adjustment of the ESP of the soils was carried out in the rainfall simulator using fog type rain. This low energy rain method was developed by Agassi *et al.*(1985). The diameter of the fog droplets is less than 0,1 mm, with maximum drop velocity of 0,1 m/s. The kinetic energy is less than 0,01 Jmm^{-1} m^{-2} with the intensity 45 mm/h. Initially 90 mm of a concentrated solution (400 $mmol/dm^{-3}$) having a given sodium adsorption ratio (SAR) was applied, followed by a 60 mm application of a more dilute solution (50 $mmol/dm^{-3}$) having the same SAR, to wash out most of the excess salts. After this pre-leaching, the soil was left overnight in the rainfall simulator and subjected to a storm with energy, using distilled water (DW), the next day.

Table 3.2 Summary of parent material, weathering intensity and localities of the soils used in the study by Smith (1990)

	Parent material					
	Basic rock			Acidic rock		
Locality	Amsterdam	Irene	Towoomba	Sabie	Krugersdorp	Marken
Degree of weathering	High	Med	Low	High	Med	Low
Annual rainfall (mm)	995	697	636	912	795	601
Humidity Max %	80	74	77	77	81	83
Min %	33	38	31	37	33	29
Average temp. °c	15	18	19	13	17	19
Height above sea level in (m)	1620	1524	1143	2118	1699	1215

3.2.3.1 Results and discussion by Smith (1990)

3.2.3.1.1 Chemical, mineralogical and physical properties of the soils

The soils used in this study were sampled in the areas shown in Table 3.2, corresponding to different weathering levels (high, medium and low), and basic or acidic parent material. Soils developed on basic parent material show a strong tendency to have higher values for CBD extractable iron and aluminium, while clay percentage is also higher than in soil from acidic parent material. These are indicated in Tables 3.3 and 3.4. This is due to the fact that the basic rocks contain larger amounts of minerals that are conducive to clay formation than the acidic rocks. The basic parent material soil group also contains higher amounts of basic cations and for this reason also has higher pH values (Table 3.3).

Smith's interpretations and discussion: Smith indicated that the basic parent material soil from the high rainfall region has the lowest CEC value of all the soils, possibly resulting from the very high concentration of pure kaolinite, while the other soils contain mixtures of different clay minerals with higher CEC values as a result. However, this conclusion of Smith is not in line with Table 3.3. In Table 3.3 the lowest value is found on acidic parent material. The mineralogical data in Table 3.5 show that the basic parent material soils consist mainly of kaolinite and quartz in the clay size fraction while the soils formed on the acidic parent material consist mainly of kaolinite, quartz and mica in the clay size fraction of the soil.

Remarks by the author (Mulibana): It is important to mention that there are some anomalous findings amongst the above. The high quartz contents in the medium weathered, and especially the highly weathered, soils are strange, since clay size quartz is very easily weathered and one of the first minerals to disappear from the clay fraction (Laker, personal communication, 2001). The quartz in the soils from dolerite is strange, because dolerite does not contain quartz and if there is any quartz in the dolerite derived soils it could have been brought there from elsewhere, for example by wind (Bühmann and Schoeman, 1995). The fairly high mica values for the soils from granite may make sense in view of the fact that many of the granites in the former Eastern Transvaal are biotite granites. One would expect the CEC per unit clay of the granite soils to be low because they have so much gibbsite (which does not have any negative charges).

The soils studied were collected from areas which receive seasonal summer rainfall. Mean summer temperatures are usually high. It is thus expected that areas with higher rainfall will also be the areas where the older, well weathered soils occur (Jenny, 1941; Loughnan, 1969). The mineralogical data (Table 3.5) show that the soils from high rainfall areas contain mainly some of the end-products of weathering and resistant clay-size minerals i.e., kaolinite, vermiculite, mica and gibbsite. The soils from medium and low rainfall areas contain mainly kaolinite but also smaller amounts of 2:1 clays and mica as rainfall increases. The amounts of 2:1 minerals decreases with increasing rainfall on both the acidic and basic parent material soil samples. The acidic parent material soil group is also the only soil group that contains interstratified clay minerals in the clay fraction of the low rainfall area. This is also in accordance with the fact that fine crystalline structured rocks (basic rocks) are much more resistant to weathering than coarse crystalline rocks (acidic rocks) (Schroeder, 1984).

SiO_2: (Al_2O_3 + FeO_3) ratios for both the soil groups are given in Table 3.8. Smith found that high rainfall on both parent material groups resulted in a decrease of the ratio. The ratios for the acidic parent material soil group are higher as a result of the high amounts of quartz in the acidic parent material. Jenny (1941) found the same trends for these ratios and is of the opinion that soils developed in low rainfall areas have undergone less intensive weathering than the soils that developed in high rainfall areas on the same type of parent material. It was found that clay percentages increase as rainfall increases, which is due to the higher weathering rates in the wetter climates (Jenny, 1941). Smith indicated that the amounts of free iron and aluminium also increase as weathering increases, with the basic parent material soil group having higher values for both (higher concentration in the primary minerals of the basic rocks).

Table 3.3 Some chemical properties of the uncultivated soils studied (Smith, 1990)

Parent material	**Basic igneous rock**			**Acidic igneous rock**		
Degree of Weathering	High	Medium	Low	High	Medium	Low
Fe%	16,3	7,3	5,9	3,3	3,2	1,65
Al%	1,5	0,15	0,09	0,20	0,09	1,04
C%	3,3	3,2	2,0	3,7	1,8	1,04
pH (H_2O)	5,4	7,0	6,0	4,7	4,8	5,4
K cmol/kg	2,8	5,7	10,1	2,0	2,3	7,0
Na "	0,8	1,5	1,9	<1	<1	<1
Ca "	11,2	60,4	35,1	1,6	15,2	18,1
Mg "	12,2	27,3	11,3	2,0	6,9	13,1
CEC (soil) "	4,7	10,3	6,35	9,0	2,83	3,56
(clay) "	9,59	46,8	28,86	32,14	14,15	39,5

Table 3.4 Particle size analysis of the soils studied (Smith, 1990).

Particle size classes						
		Coarse sand	Medium sand	Fine sand	Silt	Clay
Parent mateial	Weathering Status	2 – 0,5 mm	0,5 – 0,2 mm	0.2 – 0,02 mm	0,02 - 0002 mm	<0,002 mm
(Per cent)						
Uncultivated						
Basic	High	9	5	14	17	49
	Medium	8	14	30	21	22
	Low	5	21	42	8	22
Acidic	High	16	18	24	11	28
	Medium	16	20	28	13	20
	Low	19	17	39	13	9
Cultivated						
Basic	High	16	7	13	25	45
	Medium	10	11	27	21	26
	Low	5	19	50	6	18
Acidic	High	15	18	27	11	27
	Medium	12	20	43	12	12
	Low	25	14	29	12	18

Table 3.5 X-ray diffraction data of the clay fractions of the soils studied (Smith, 1990)

Parent material	**Basic**			**Acidic**		
Degree of weathering	High	Medium	Low	High	Med	Low
Uncultivated soils						
Smectite	-	1	1	-	1	1
Kaolinite	5	5	5	5	5	5
Vermiculite	1	-	-	-	-	-
Quartz	2	3	3	1	4	4
Mica	2	-	-	1	3	-
Hornblende	-	-	-	3	-	-
Gibbsite	-	-	-	3	-	-
Cultivated soils						
Smectite	-	-	-	-	1	2
Kaolinite	5	5	5	5	5	5
Vermiculite	1	-	-	-	-	-
Quartz	3	3	3	3	4	4
Mica	2	-	-	1	3	3
Hornblende	-	-	-	-	-	-
Gibbsite	-	-	-	5	-	-
Interstratified minerals	-	-	-	-	-	-

1 = 0-15%   3 = 25-50%   5 = 75-100%

2 = 15-25%   4 = 50-75%

Values 1 – 5 indicate peak intensities only and not percentages of minerals present.

3.2.3.1.2 Infiltration data

Smith (1990) found significant differences between the effects of uncultivated vs cultivated soils developed on basic vs acidic parent material and ESP changes on these soils. He found that the basic parent material soil group is thus much less susceptible to crust formation or other factors reducing infiltrability despite a general finer texture.

A comparison of the infiltration curves of the basic parent material group of soils against acidic parent material soils (Figures 3.1a and 3.1b) shows that differences occur both in the final infiltration rate (FIR) as well as in the rate of crust formation. The rates of infiltration decreased more rapidly in the acidic parent material group than in the basic parent material soil group.

Smith found that the basic parent material soils were more stable than those from acidic parent material. This may be explained by the higher free iron, aluminium and carbon contents of the basic parent material soils as compared with the acidic parent material group (Table 3.3). The basic parent material soils also contain higher amounts of clay which will increase aggregate stabilities due to cementing action by the clay (Ben Hur *et al.*, 1985). Table 3.4 contains the particle size analysis data. Smith found that the very stable highly weathered soil in the basic soil group is the only soil that contains a low amount of fine sand relative to the other soils and high amounts of iron, aluminium, carbon and clay. A high proportion of stabilising factors occur in this soil as compared to all the other soils, where much higher percentages of fine sand, which is considered to be an indicator as to a soil's resistance to erosion (Evans, 1980), are found. The basic soils as a group have the advantage of higher amounts of free iron, aluminium, carbon and clay, than the acidic soil group. These properties of the basic soil group cause the basic soils to be more resistant to crust formation and thus higher final infiltration rate (FIR) and cumulative infiltration (CI) values will result in simulated rainfall experiments.

It is clear from Figure 3.1a that there is a very sharp reduction in the FIR for the basic parent material soil group as the weathering degree (and rainfall) decreases from high to medium. From medium to a low degree of weathering (and rainfall) there is only a minor further decrease. The same cannot be said about the FIR values for the soils developed on acidic parent material (Figure 3.1b). On the other hand, CI values or the area under curves indicate the same trend for the soils that developed on the acidic parent material (Table 3.9). Smith found that CI values decrease from 67 mm for the high rainfall area to 43, 45 and 34,6 mm for the medium and low rainfall areas on basic parent material soils, respectively. The corresponding values for the soils on acidic parent material soils in the same order were 43; 32 and 24 mm respectively. The decrease in CI in the case of the basic parent material soil group from high rainfall to medium rainfall area was higher than that of the corresponding acidic parent material soils. The decrease from the medium to low rainfall areas in both soil groups, was in the same order, 11mm as opposed to 8 mm.

It is clear from Table 3.3 that decreases occur in both the organic matter and free iron and aluminium contents, from the high to low weathering degree soils in both the acidic and the basic parent material soil groups. Smith found that the decreases are much bigger in the basic parent material soil group, where the biggest reduction in the FIR values occurred. The acidic parent material soil group, on the other hand, shows much smaller differences in these values and also no big differences in the FIR's. The clay contents of the basic parent material soil group also decreases markedly from the high rainfall to the medium rainfall sites (Table 3.4). This correlates with the decrease of the FIR values of the soils.

Table 3.6 and 3.7 show that both the FIR and CI values differ significantly for the three weathering degrees. An increase in rainfall (weathering) also increases FIR values for all soil, from 7.6 to 9.1 and 14.8 mmh^{-1} from the low to the high rainfall (weathered) areas respectively. The CI values are 29.6, 37.6 and 54.9 mm respectively for the same sequence of areas, indicating that soil genesis in high rainfall areas on any type of parent material will create soils that are more stable against crust formation than soils formed under medium or low rainfall.

Figure 3.2 a and Figure 3.2 b contain the FIR and CI results as functions of the SiO_2: $(Al_2O_3 + Fe_2O_3)$ ratios. It is clear that FIR and CI decreased as the SiO_2: $(Al_2O_3 +$

Fe_2O_3) increased under less intensive weathering conditions. Smaller ratios are thus associated with higher stability against crust formation on the soil surface.

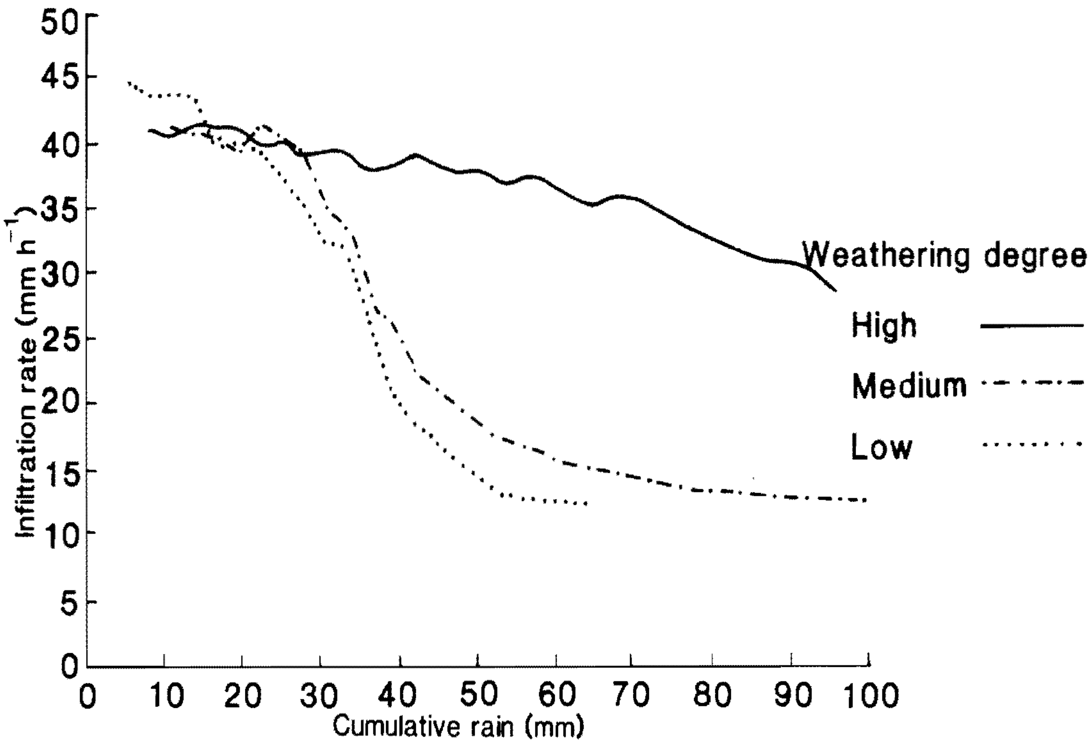

FIG 3.1 (a) Infiltration curves for the basic parent material soils.

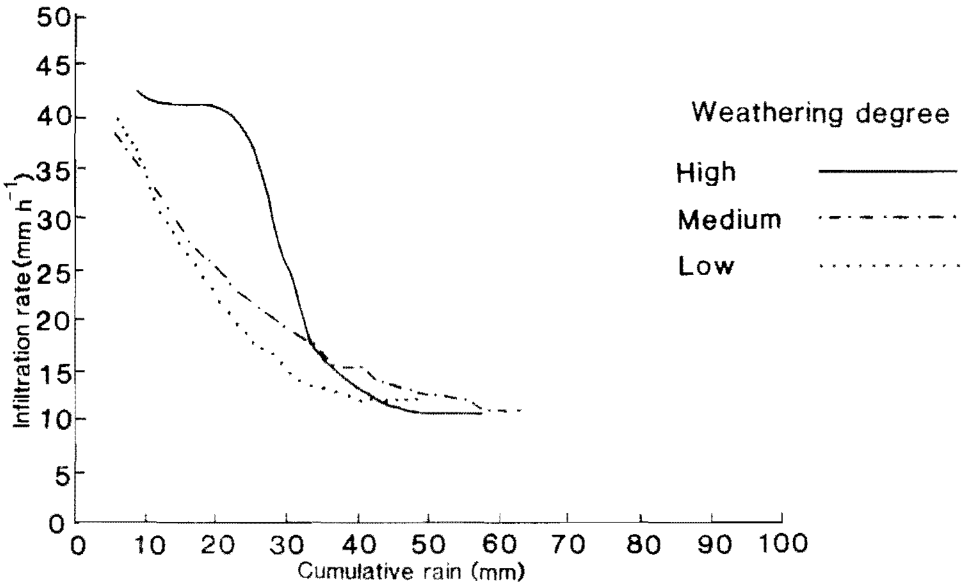

FIG 3.1(b) Infiltration curves for the acidic parent material soils.

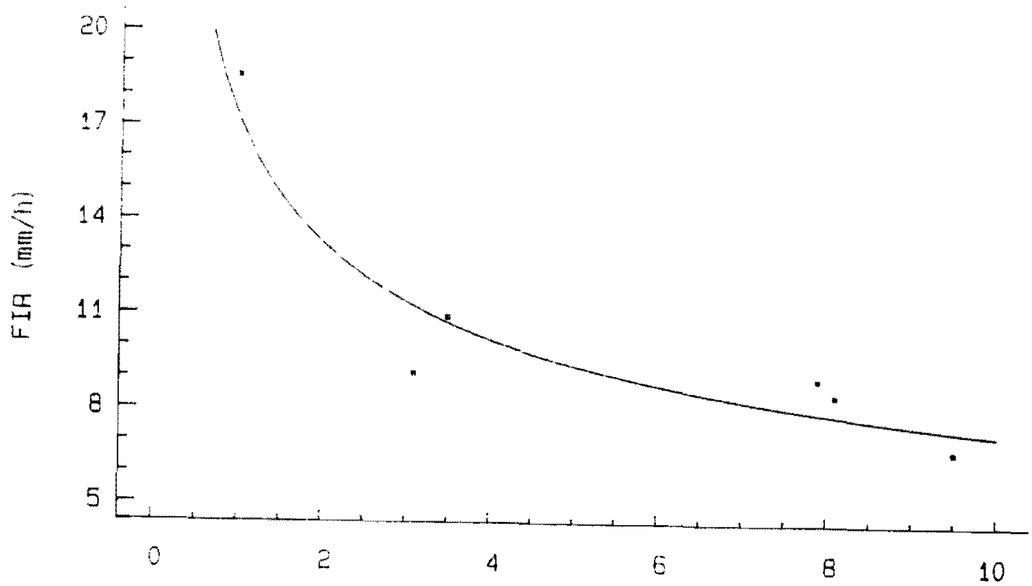

FIG. 3.2 (a) FIR as a function of SiO_2: $Al_2O_3 + Fe_2O_3$

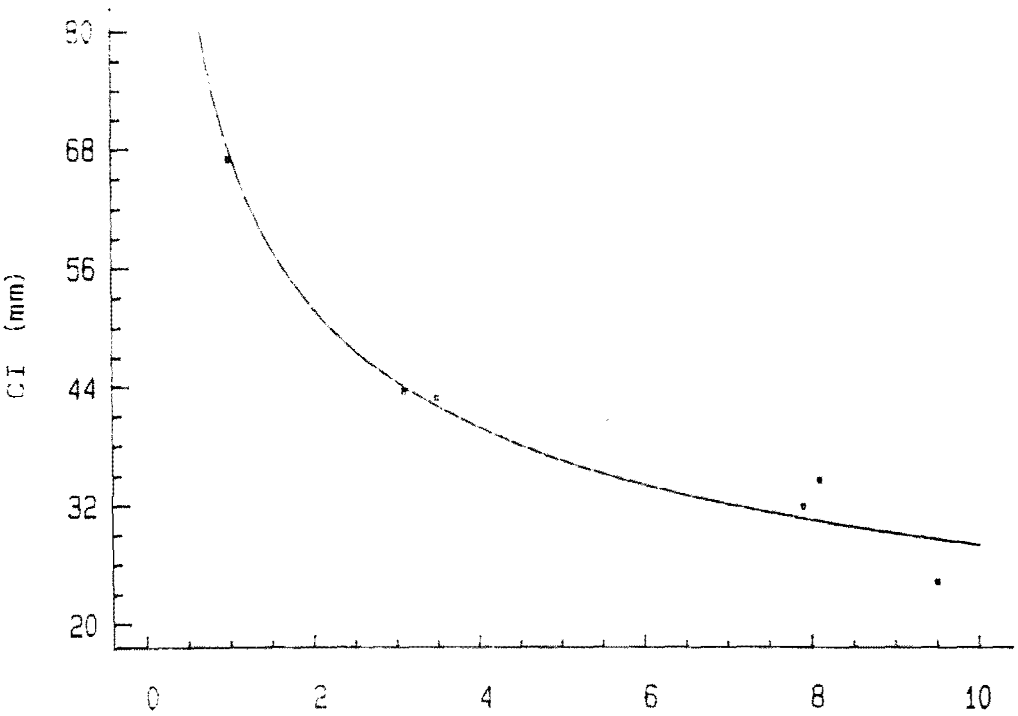

FIG. 3.2 (b) CI as a function of SiO_2: $Al_2O_3 + Fe_2O_3$

Table 3.6 Multiple range analysis for FIR values.

Main effect	**Average FIR mm h^{-1}**
Uncultivated	12,52 a
Cultivated	8,40 b
Basic soil group	12,02 a
Acidic soil group	8,90 b
Untreated	12,89 a
ESP 5	9,23 b
ESP 10	9,38 b
High degree of weathering	14,77 a
Medium degree of weathering	9,06 b
Low degree of weathering	7,65 c

[a]Figures followed by the same letter within a column for each treatment, do not differ significantly at 95 % confidence level.

Table 3.7 Multiple range analysis for CI values after 100 mm of rain.

Main effect	Average CI mm
Uncultivated soil group	47,12 a
Cultivated	34,05 b
Basic soil group	48,13 a
Acidic soil group	33,03 b
Untreated	48,81 a
ESP 5	35,96 b
ESP 10	37,38 b
High degree of weathering	54,90 a
Medium degree of weathering	37,58 b
Low degree of weathering	29,56 c

Table 3.8 Weathering degree as indicated by $SiO_2:Al_2O_3 + Fe_2O_3$ ratios

Parent material	Rainfall	Locality	Weathering degree	$\frac{SiO_2}{Al_2O_3 + Fe_2O_3}$
Basic	995 697 636	Amsterdam Irene Tawoomba	High Medium Low	1.0 3.1 8.1
Acidic	912 795 601	Sabie Krugersdorp Marken	High Medium Low	3.5 7.9 9.5

Table 3.9 Calculated cumulative infiltration (CI mm) after 100 mm of rain and Measured final infiltration rates (FIR mmh^{-1})

Parent material	Degree of weathering	CI (mm)	FIR ($mm\ h^{-1}$)
Basic	Low	34.59	8.49
	Medium	43.45	9.11
	High	66.95	18.60
Acidic	Low	24.32	6.76
	Medium	31.95	9.00
	High	42.85	10.94

3.3 D'HUYVETTER, JHH (1985): DETERMINATION OF THRESHOLD SLOPE PERCENTAGES FOR THE IDENTIFICATION AND DELINEATION OF ARABLE LAND IN CISKEI

3.3.1 General background and objectives of the study

The erosion extent in Ciskei had been increasing at an alarming rate, especially in cultivated maize fields of so-called "betterment schemes" or "rehabilated areas". Hensley and Laker (1978) identified the main reason as being the use of a standard slope criterion of **12%** for arability by "planners", without taking the inherent differences between the erodibilities of different soils into account. It was therefore crucial to develop criteria and models by means of which arable land could be identified and delineated. D'Huyvetter focussed on establishing threshold slope criteria for the dominant soils of three different pedosystems of the former Ciskei. Attempts were made to derive predictive which could be used to predict threshold slopes from soil-slope-erosion data.

3.3.2 Research procedures

3.2.2.1 Areas studied

Three pilot areas were selected in three main pedosystems of the area formerly known as Ciskei, namely Mavuso, Keiskammahoek and Middledrift. The pedosystems were delineated and described by Hensley and Laker (1978). Pilot areas were selected by means of aerial photographs and orthophoto maps. The areas studied in each pedosystem ranged between approximately 2000 and 2400 ha in extent. Only "planned" areas on which small-scale farmers grew (or had grown) maize were included in the study.

The climate was outlined by Marais (1978). The climate of all three areas is described as temperate warm, with rainfall peaks in spring/early summer and late summer/autumn and a pronounced mid-summer drought. Rain is mainly in the form of intense thunder storms.

3.3.2.2 Determination of slope parameters

The topographical factors which received attention were: slope gradient, slope form (convex, concave or plane), and slope length above the point of study. These have all been identified as important parameters in regard to erodibility.

With the help of high quality large-scale (1:5 000) orthophoto maps accurate measurements of the important slope factors length and gradient were obtained. It also provided an excellent method for the determination of slope form and the transition between different types of slopes.

3.3.2.3 Determination of extent and degree of erosion

In the study the emphasis was on gully erosion, because its extent and degree is relatively easy to determine. Furthermore, it was the dominant type of erosion in the area. It was determined by means of stereoscopic analysis of contact aerial photographs, which proved to be a very efficient method. Some field verification was also done.

Four erosion classes were defined. A small number of classes was used, so as to have more observations per class, thus facilitating statistical analysis of the data.

3.3.2.4 Soil parameters

Detailed soil surveys were conducted, using 1:5 000 orthophoto maps as base maps. Soils were classified according to the South African binomial soil classification system (Macvicar *et al.*, 1977). Soil data were collected by means of field sampling and laboratory analysis of both topsoils and subsoils. The soil parameters that were selected were those that have been identified by other researchers as being most important in regard to erodibility and were feasible to determine within a short time space. These included:

(1) Morphological features: diagnostic horizon, depth to limiting layer, total depth, colour, etc.
(2) Parent material.
(3) Particle size distribution.
(4) Chemical properties: CEC, exchangeable bases, organic carbon.

3.3.3 Results

3.3.3.1 Mavuso Pedosystem

3.3.3.1.1 Field observation and soil classification

Soils were divided into three groups. The first group included the very stable red soils of the Shortlands and Hutton forms, derived from dolerite. The second group of soils included soils of the Glenrosa, Oakleaf, Westleigh and Swartland forms, of which the Glenrosa form is the most common soil in the Mavuso pedosystem. The third group included soils of the Sterkspruit, Valsrivier, Vilafontes and Estcourt forms. These highly erodable soils occur on footslopes and valley bottoms. Two areas were chosen. The one at Mabandla's location was characterized by stable soils and virtually no visible erosion, having predominantly dolerite as parent material. The adjacent area was very unstable and so badly eroded that most of the cultivated fields had been abandoned and hardly any form of further cultivation was possible. The soils of this area formed on grey to blue mudstone of the lower Beaufort group. Average annual rainfall ranges between 500 mm and 600 mm. The most common soil forms and series, classified according to the South African binomial system (Macvicar *et al.*, 1977), are given in Table 3.10.

Table 3.10 Classification of soils in the pilot areas of the Mavuso pedosystem (D'Huyvetter, 1985)

Soil form	**Soil series**
Bonheim	Bonheim
Glenrosa	Williamson
Hutton	Shorrocks
Mayo	Mayo
Mispah	Mispah
Oakleaf	Jozini
Shortlands	Kinross Glendale Shortlands
Swartland	Hogsback
Valsrivier	Arniston Herschel
Vilafontes	Blythdale

3.3.3.1.2 Type(s) of erosion that occurred

Topsoil removal by sheet erosion occurred to some extent in both Shortlands and Hutton form soils, but did not result in a significant deterioration of the fields. Damage from sheet flow was extensive and linear erosion features such as rills and "dongas" were observed on Oakleaf, Westleigh and Swartland form soils, but this happened on steeper slope segments. Severe erosion occurred on Sterkspruit, Valsrivier and Escourt form soils.

3.3.3.1.3 Calculation of threshold slope percentages

To determine whether significant correlations exist between the degree of erosion observed in the field and slope gradient for the different soil forms, linear, exponential, geometric and n^{th} order regressions were carried out. In cases where simple linear and geometric regressions were used, best fitting curves were produced. Since the maximum possible erosion intensity has a value of four, each curve should level off at that value. D'Huyvetter did not include all values of slope percentage and degree of erosion, e.g. only one value for Arcadia soils was recorded in his dissertation.

The regression equations for the relationships between slope gradient and degree of erosion for the most important soils of the Mavuso pedosystem are listed in Table 3.11. The relationships are illustrated in Figures 3.3 to 3.6.

Table 3.11 Relationship between the degree of erosion and slope percentage for different soil of the Mavuso pedosystem (D'Huyvetter, 1985).

Soil form	Regression	R^2	sig.	Number of observations
Hutton	Y = 0,44 + 0,148x	0,62	0,01	13
Shortlands	Y = 0,634 + 0,105x	0,64	0,01	37
Estcourt Sterkspruit Valsrivier Vilafontes	Y = 0,689 + 0,289x	0,40	0,05	12
Swartland	Y = 0,564 + 0,26x	0,79	0,01	7

Where Y = Degree of erosion

X = Slope percentage

New threshold percentages for the different soils were established (Table 3.12) using the equations in Table 3.11. A value of 2 for the degree of erosion (Y) is regarded as being critical by the researchers,and it was therefore used as the maximum allowable erosion value.

Table 3.12– New maximum threshold slope percentages for arability for different major soils of the Mavuso pedosystem (D'Huyvetter, 1985)

Soil form	**Threshold slope %**
Glenrosa	6,0 %
Hutton	10,6 %
Oakleaf	6,1 %
Shortlands	13,0 %
Swartland	5,5 %
Estcourt Sterkspruit Valsrivier Vilafontes	4,5 %

From Table 3.12 it is clear that cultivation, with proper contouring, up to a slope of 12% would be safe only on the stable Shortlands soils. On Hutton soils few areas would have erosion problems. For the unstable Swartland, Estcourt, Sterkspruit, Valsrivier and Vilafontes soil erosion starts at much lower slope gradients. This explains the severe erosion found in so many of the areas included for cultivation by the "planners", using a blanket criterion of recommending soils on slopes up to 12% gradient for cultivation.

3.3.3.1.4 Correlations between soil factors and erosion

A highly significant correlation was found between the ESP of the soil and degree of erosion. The correlation was found to be stronger for A horizons than for B horizons. The highest ESP values were found in soils of Valsrivier, Vilafontes and Westleigh forms, all of which have high erodibilities. Soils with lower ESP values, such as Hutton, Shortlands and Oakleaf are less erodable. A relatively high ESP in the A horizon induces colloidal dispersion, resulting in the formation of a dense surface crust. This crust reduces infiltration of water into the soil.

Soils of the Bonheim and Swartland forms and some Glenrosa form soils were exceptions to the general rule. It is important to note that although low ESP values were recorded for both A and B horizon of these soils, they were strongly eroded. This clearly indicates that in some cases ESP does not play a dominant role in determining erosion. Erosion in other cases may be caused by chemical and physical factors and/or topographical factors.

D'Huyvetter also indicated the correlation between the S-value, i.e. (sum of exchangeable bases, (Ca^{++} + Mg^{++} + K^{+} + Na^{+}). He found a strong correlation between S-value per 100g clay and degree of erosion. These correlations were valid for both A and B horizons. The S-value/100g clay is mainly a reflection of the type of clay mineral in the soil- a high value being an indication of 2:1 swelling type (smectite) clays. Degree of leaching of bases affects it to a lesser extent.

Increased S-value/100g clay was strongly correlated with increased ESP. A combination of relatively high S-value/100g clay (i.e. a high proportion of 2:1 swelling clay) and relatively high ESP values will indicate an unstable soil.

Texture- normally, soils rich in clay are regarded as being more stable than those with a low clay content. This trend was found in the Mavuso pedosystem. D'Huyvetter (1985) found that at a clay content of less than 20% for the A horizon there was a very sharp increase in the erodibility of soil. At a clay content higher than 20% the soils were found to be quite stable and differences in clay had practically no effect on erodibility in this range.

3.3.3.2 Keiskammahoek Pedosystem

3.3.3.2.1 Field observations and soil classification

The Amatola basin was selected as pilot area for this pedosystem and all the accessible cultivated plots in the basin were studied. This area was chosen because a detailed soil survey was available for it and, therefore, no further time-consuming soil survey was required. The parent materials for the majority of soils in the basin consist of grey and/or green mudstone (rich in silt size particles) of the lower Beaufort group, ferricretes (also giving rise to soils rich in silt), dolerite and some sandstone. The climate of the area is relatively humid compared with the largest part of the former Ciskei. The mean annual precipitation is about 650 mm and higher. The dominant soil forms and soil series in the area are listed in Table 3.13.

Table 3.13 Classification of soils of the Keiskammahoek pedosystem (Hensley and Laker, 1978)

Soil form	Soil series
Arcadia	Arcadia
Glenrosa	Saintfaiths Williamson
Hutton	Makatini Shorrocks
Mispah	Mispah
Oakleaf	Jozini (Modal phase) Jozini (Mudstone phase) Zozini (Plinthite phase)
Shortlands	Glendale Shortlands
Swartland	Swartland
Valsrivier	Waterval Sunnyside Herschel Valsrivier

3.3.3.2.2 Type(s) of erosion that occurred

The stability of soils in Amatola Basin showed wide differences between the soil forms.

The ***Valsrivier*** soil form was observed as the least stable form in the basin. It occurs as red or non-red alluvial soil on the lowest terraces of the Amatola river, or as a red colluvial soil on mid- and footslope positions. Erosion varies from sheet erosion to the presence of well-developed rill and /or gully systems.

Oakleaf soils in the Amatola basin are members of the Jozini series. A subdivision was made between a Jozini (modal phase), Jozini (plinthite phase) and Jozini (mudstone phase) by Du Preez and Botha (1980). The Jozini (modal phase) is formed from colluvial material from weathered mudstone and dolerite. The mudstone phase has formed from colluvial material from mudstone. The plinthite phase has formed from weathered ferricrete and mudstone and contains non-diagnostic Fe and Mn concretions. Only the modal phase is relatively stable.

Soils of the ***Shortlands and Hutton*** forms are the most stable soils of the Amatola basin. They are all formed on mixed colluvial material from dolerite and mudstone.
The doleritic parent materials impart stability to these soils. Sheet erosion is present, but it does not cause irreparable damage to cultivated fields. Fields steeper than 18% are all strongly subjected to sheet erosion (D'Huyvetter, 1985)

Clayey ***Arcadia*** soils have formed on colluvial material from weathered dolerite in lower topographic positions. Compared to other soils in the Amatola basin, the Arcadia soils appear to be fairly stable. Sheet and rill erosion was not observed, and well developed dongas are scarce in the Arcadia form soils.

3.3.3.2.3 Calculation of threshold slope percentage for the Amatola Basin (Keiskammahoek pedosystem)

Significant correlations were found between slope gradient (%) and degree of erosion for the different soil forms of the Amatola basin (Table 3.14)

Table 3.14 Relationships between the degree of erosion and the slope percentage for the different soils of the Amatola basin.

Soil form	Regressions	R^2	significance	Number of observations
Arcadia	$Y = 0,201 + 0,148x$	0,81	0,01	21
Glenrosa	$Y = 0,520 + 0,13x$	0,46	0,01	38
Hutton	$Y = 0,252x^{0,789}$	0,85	0,01	15
Shortlands	$Y = 0,169 + 0,136x$	0,77	0,01	49
Swartland	$Y = 0,559 + 0,132x$	0,66	0,01	13
Valsrivier	$Y = 0,664 + 0,166x$	0,59	0,01	39

Where: Y = Degree of erosion
X = Slope %

By means of these regression equations D'Huyvetter (1985) calculated new threshold slope percentages for the different soil forms of the Amatola basin (Table 3.15), considering Y = 2,0 as the maximum permissible degree of erosion. Best fitting curves were constructed by means of these regression equations (Figure 3.7 – 3.11).

3.3.3.2.4 Correlation between soil factors and erosion

D'Huyvetter did not elaborate much on the correlation between soil factors and erosion in the Amatola basin. It is imperative therefore to mention that he found that most of the soils in the Amatola basin were rich in silt size particles. However, it is fair to indicate that in Arcadia and Valsrivier soil forms clay content was dominating. In general most of the soils in this basin were fairly stable against erosion.

Table 3.15 New threshold slope percentage for the Amatola basin (Keiskammahoek pedosystem).

Soil form	**Threshold slope %**
Arcadia	12,2%
Glenrosa	12,0%
Hutton	13,9%
Shortlands	13,5%
Swartland	10,9%
Valsrivier	8,0%

For the Shortlands soil form a significant correlation was also found between the degree of erosion and a combination of slope gradient and the slope length above a certain point of observation. This multiple regression can be written as:

$$Y = 0{,}111 + 0{,}131x_1 + 2{,}016\ 10^{-4}\ x_2$$
$$r^2 = 0{,}72 \text{ (Sign. at 0,01 level)}$$

Where: Y = Degree of erosion
X1 = Slope gradient (%)
X2 = Slope length above point of observation (m)

By using this equation new threshold slope percentages could be calculated for Shortlands soils as a function of slope length above the point of observation (Table 3.16).

Table 3.16 Variation in threshold slope percentage with distance from the top of a hill (slope) for Shortlands soils in the Amatola Basin.

Distance from top	Threshold slope %
0 m	14,4%
100m	14,2%
250m	14,0%
500m	13,6%
1000m	12,8%

The stronger impact of surface runoff in lower lying areas is reflected in the above table. It indicates that somewhat steeper threshold slope percentages are permissable in topslope positions compared to mid-slope and bottom-slope positions, but that even large great slope lengths have unexpectedly small effects on erosion of these very stable soils. It is very important to note that if a threshold slope percentage of 12% was used as upper limit for arability, significant areas of good arable land at steeper slopes on these stable soils would be excluded from cultivation. This can actually be seen in parts of the former Ciskei. Meanwhile highly erodable soils from less stable forms were included for cultivation just because they are on slopes of less than 12%, leading to the severe erosion seen in many parts.

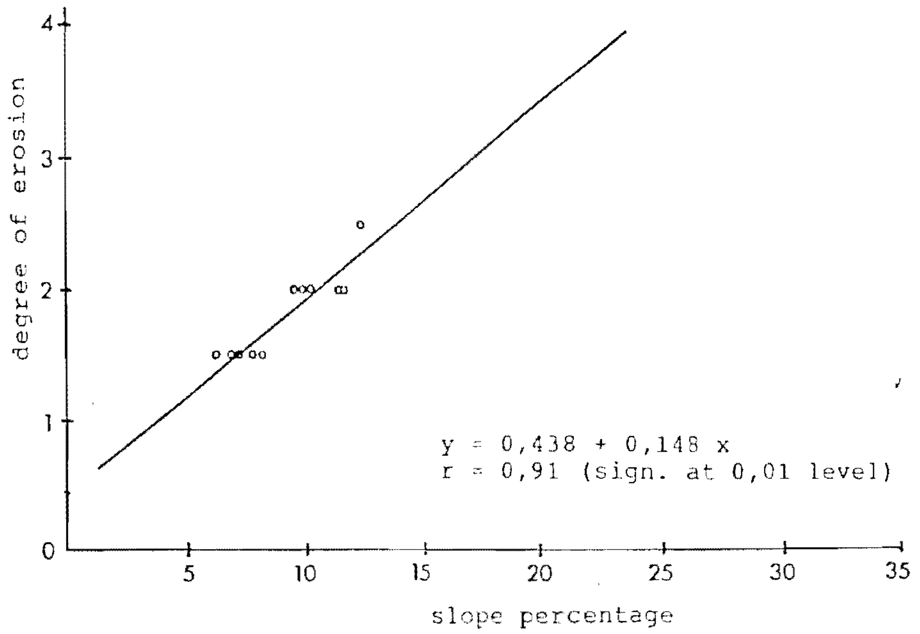

FIG 3.3 Relationship between slope percentage and degree of erosion Hutton soils of the Mavuso pedosystem (D'Huyvetter, 1985)

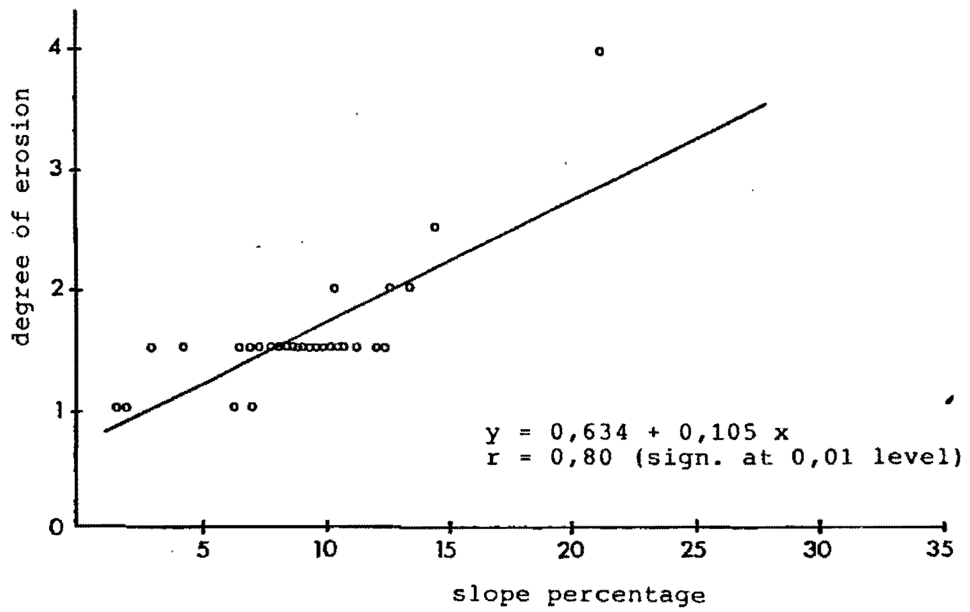

FIG. 3.4 Relationship between slope percentage and degree of erosion for Shortlands soils of the Mavuso pedosystem (D'Huyvetter, 1985)

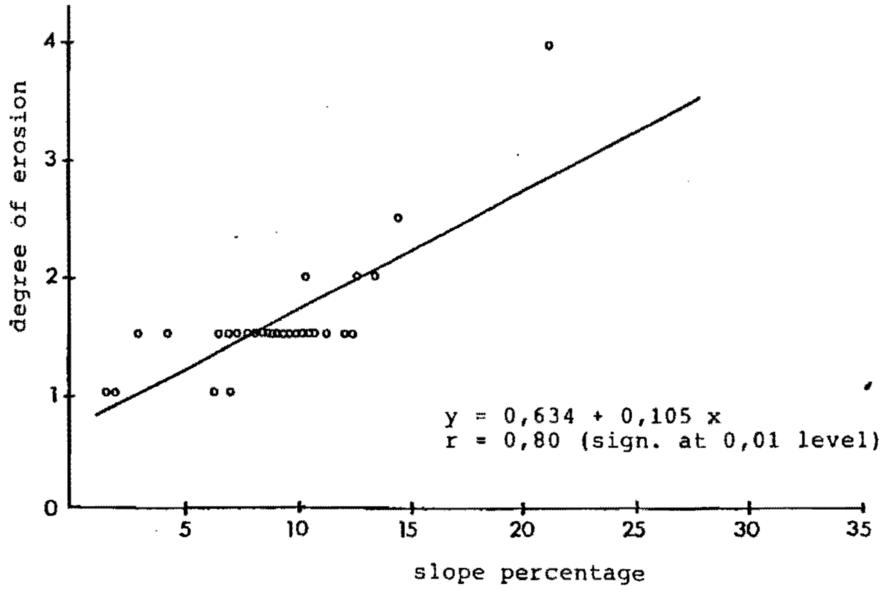

FIG. 3.5 Relation between slope between slope percentage and degree of erosion for Swartland soils of the Mavuso pedosystem (D'Huyvetter,1985).

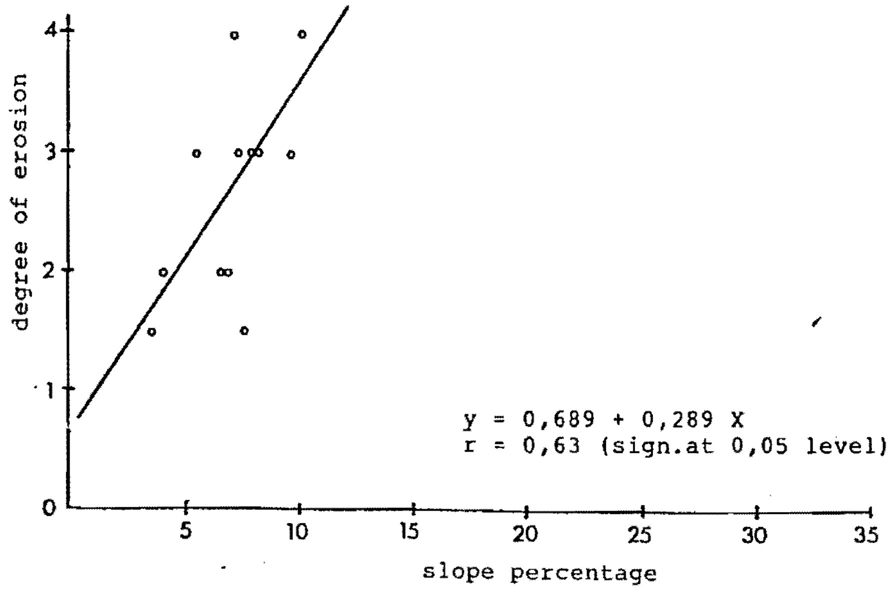

FIG. 3.6 Relation between slope percentage and degree of erosion for Valsrivier soils of the Mavuso pedosystem (D'Huyvetter, 1985)

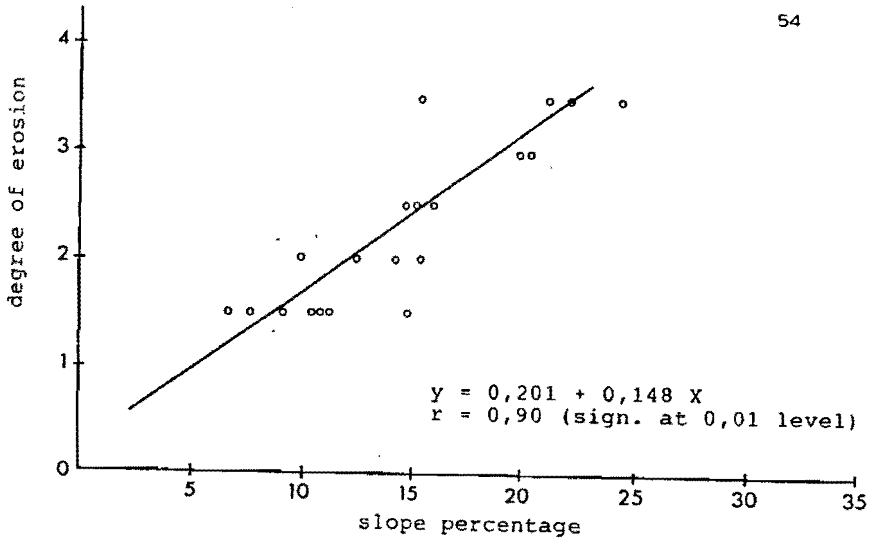

FIG 3.7 Relationship between slope percentage and degree of erosion for Arcadia soils in the Amatola Basin (D'Huyvetter,1985).

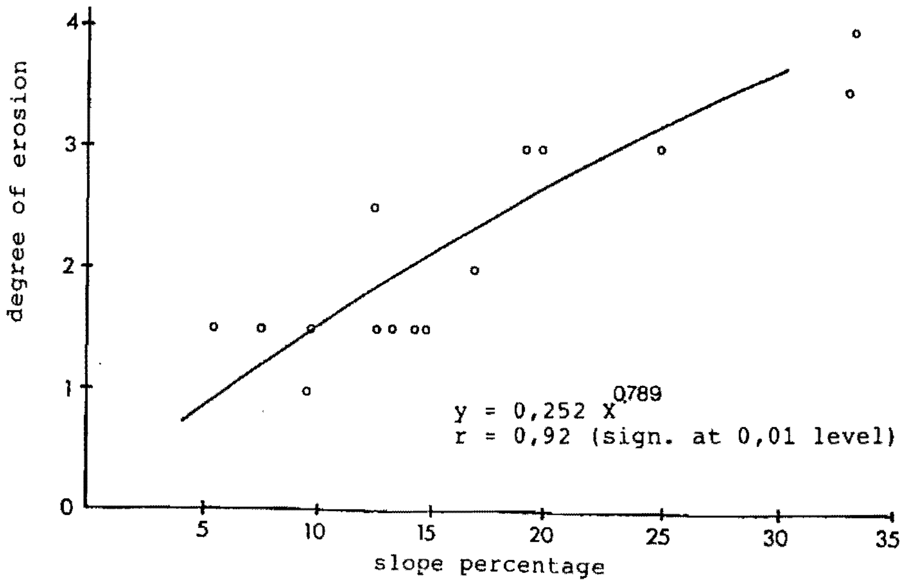

FIG. 3.8 Relationship between slope percentage and degree of erosion for Hutton soils in the Amatola basin (D'HuyVetter, 1985)

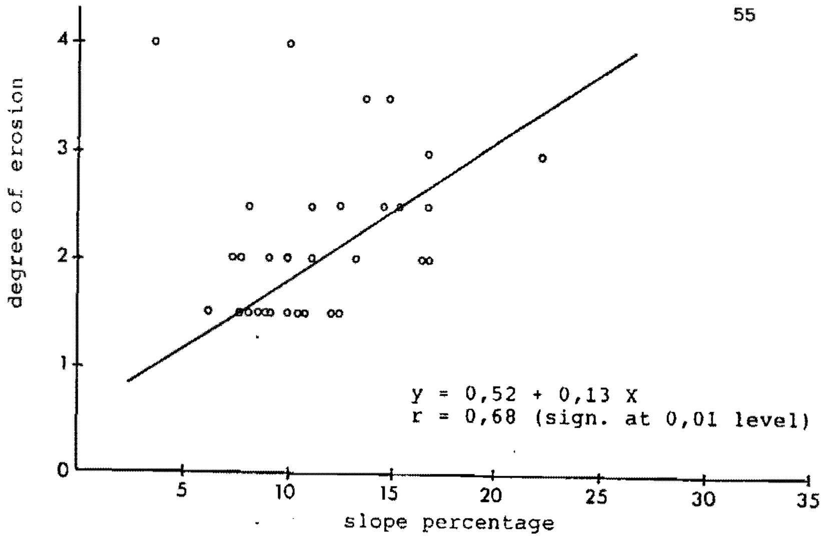

FIG. 3.9 Relationship between slope percentage and degree of erosion for Glenrosa soils in the Amatola basin (D'Huyvetter, 1985).

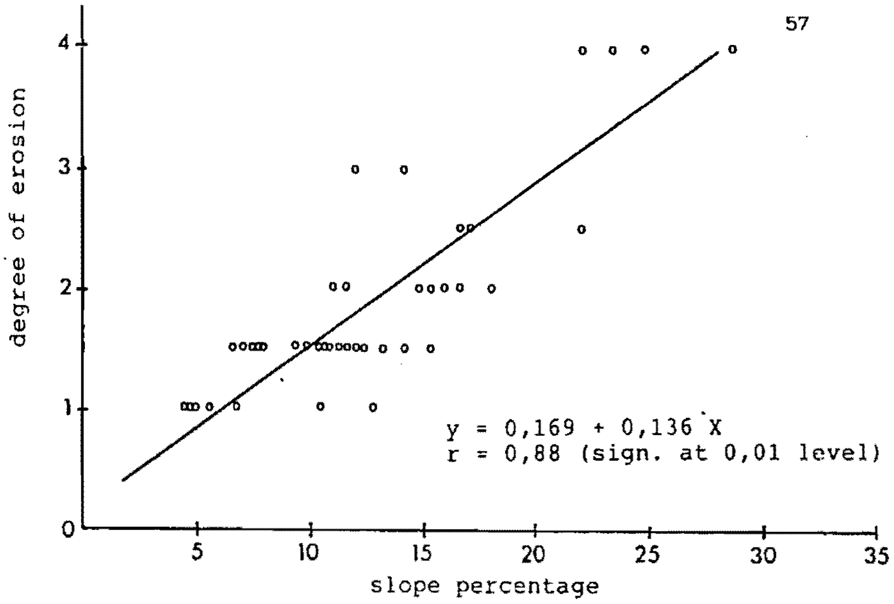

FIG. 3.10 Relationship between slope percentage and degree of erosion for Shortlands soils in the Amatola Basin (D'Huyvetter, 1985).

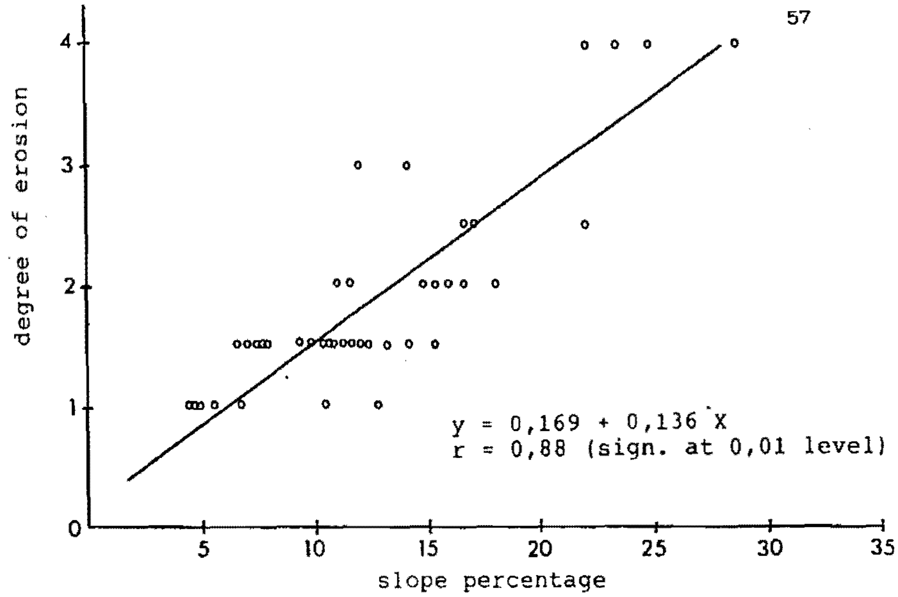

FIG. 3.11 Relationships between Slope percentage and degree of Erosion for Swartland soils in the Amatola Basin (D'Huyvetter, 1985)

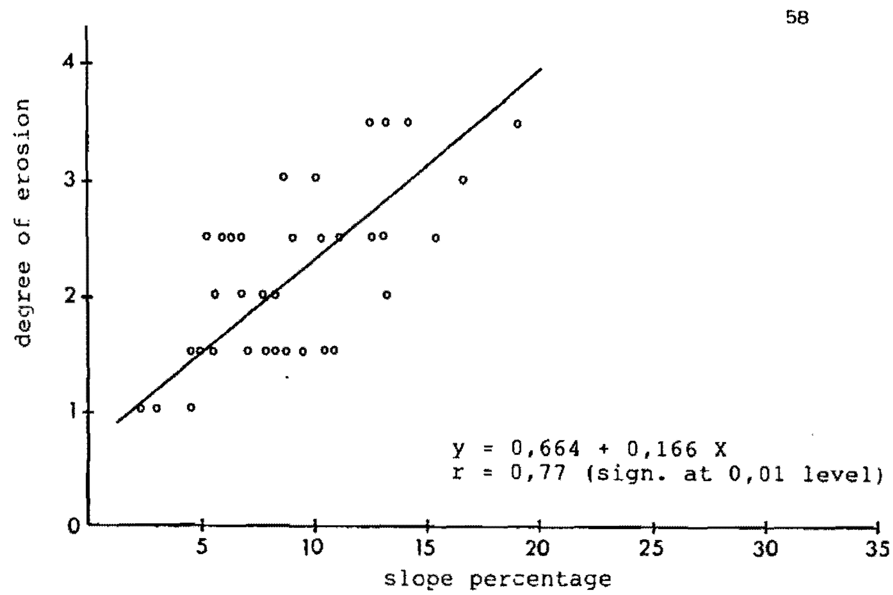

FIG. 3.12 Relationships between slope percentage and degree of erosion for Valsrivier soils in the Amatola Basin (D'Huyvetter,1985).

3.3.3.3 Middledrift pedosystem

3.3.3.3.1 Field observation and soil classification

Cultivated soils in the Middledrift pedosystem belonged to five series of four soils forms according to the South African Binomial System (MacVicar *et al.*, 1977) (Table 3.17).

Table 3.17 Taxonomic classification of the dominant soils in the Middledrift Pedosystem (Hensley and Laker, 1978)

Binomial Classification[1]		Approximate Soil Taxonomy Equivalent[2]
Form	Series	
Glenrosa	Williamson	Lithic Ustochrepts
Oakleaf	Jozini Limpopo	Typic Ustochrepts Typic Eutrochrepts
Shortlands	Shortlands	Typic Rhodustalfs
Valsrivier	Lindley	Typic Haplustalfs

[1]MacVicar *et al.*, (1977)
[2]U.S.D.A. (1975)

The main features of these soils are:

Williamson series: A shallow (usually 200 to 250 mm thick) grey apedal Ap horizon over a layer dominated by partly weathered mudstone or shale with tongues of soil into it. The clay content of the A-horizon is about 20%.

Jozini and Limpopo series: Deep (more than 2 metres) dark grey to dark brown soils with a small gradual increase in clay content with depth. These are soils developed in alluvial or colluvial drift. Clay content of the B-horizon is about 30 per cent.

Shortlands series: A structured clayey soil with a red structured B diagnostic horizon, i.e. with a homogeneous red colour (MacVicar *et al.*, 1977). The clay content increases gradually from about 45 per cent in the topsoil to about 60 per cent in the subsoil.

Lindley series: A pseudo-duplex soil consisting of a thin (about 150 mm) grey apedal Ap horizon with 15% to 20% clay over a dark grey clayey subsoil (45% to 50% clay). The topsoil of all these soils, except the Shortlands series, has very high (very fine sand + silt) content.

3.3.3.3.2 Type(s) of erosion that occurred

The Valsrivier, Vilafontes, Kroonstad and shallow Glenrosa soils are characterised by high to extremely high erodibilities, often resulting in a dense network of deep rills and dongas. Except for the shallow Glenrosa soils, which occur on convex topslope positions, the rest are mainly found in mid- or bottomslope positions. All four soils are formed in colluvial mudstone material or by *in situ* weathering of mudstone.

The Oakleaf soil form was subdivided into two, a stable dark brown Jozini and strongly eroded dark grey Jozini. Sheet and rill erosion was observed in both Jozini soil phases. Well developed gully erosion is mainly restricted to the dark grey phase. Shallow dongas were observed in the Shortlands soil form that was found on slope segments steeper than 13%.

3.3.3.3.3 Calculation of threshold slope percentages of Middledrift pedosystem

Significant correlations were found between slope gradient (%) and degree of erosion for different soil forms of the Middledrift pedosystem (Table 3.18).

Table 3.18 Relationship between the degree of erosion and the slope percentage for the different soils of the Middledrift pedosystem

Soil form	Regression	R^2	sig.	Number of observations
Glenrosa (sh.) Glenrosa (deep) Kroonstad	$Y = 0,912\ X^{0,631}$ $Y = 0,823 + 0,418x$ -	0,62 0,36 -	0,01 0,05 -	24 11 -
Oakleaf (d.br.) Oakleaf (grey)	$Y = 0,441\ x^{0,74}$ -	0,61 -	0,01 -	17 -
Shortlands	$Y = 0,099X^{1,24}$	0,69	0,01	14
Valsrivier	$Y = 0,99\ X^{0,642}$	0,59	0,01	34

Where: Y = Degree of erosion

X = Slope %

By means of these regression equations D'Huyvetter (1985) calculated new threshold slope percentages for the different soil forms of the Middledrift (Table 3.19). Best fitting curves were constructed by means of these regression equation (Figure 3.13).

Table 3.19 New threshold slope percentages for the Middledrift pedosystem

Soil form	Threshold slope
Glenrosa (shallow) Glenrosa (deep) Kroonstad	3,5% 6,8% -
Oakleaf (dark brown) Oakleaf (dark grey)	7,8% -
Shortlands	11,3%
Valsrivier	3,2%

It is important to note that for the Valsrivier soil form, a significant correlation was found between the degree of erosion and a combination of the slope gradient and slope length above the point of observation.

$$Y = 1{,}365 + 1{,}88x_1 + 9{,}58\ 10^{-4}\ x_2$$

$$r^2 = 0{,}67 \text{ (sign. At 0,01)}$$

where:

Y = Degree of erosion

X1 = Slope gradient %

X2 = Slope length above the point of observation

By using this equation new threshold slope percentages could be calculated for the Valsrivier soil form as a function of the slope length above the point of observation (Table 3.20).

Table 3.20 Variation in threshold slope percentage with distance from the top of a hill (slope) for Valsrivier soils in the Middledrift pedosystem.

Distance from top	Threshold slope %
0 m	3,38%
100 m	2,87%
200 m	2,36%
300 m	1,85%
400 m	1,34%
500 m	0,83%
750 m	0,00%
1000 m	0,00%

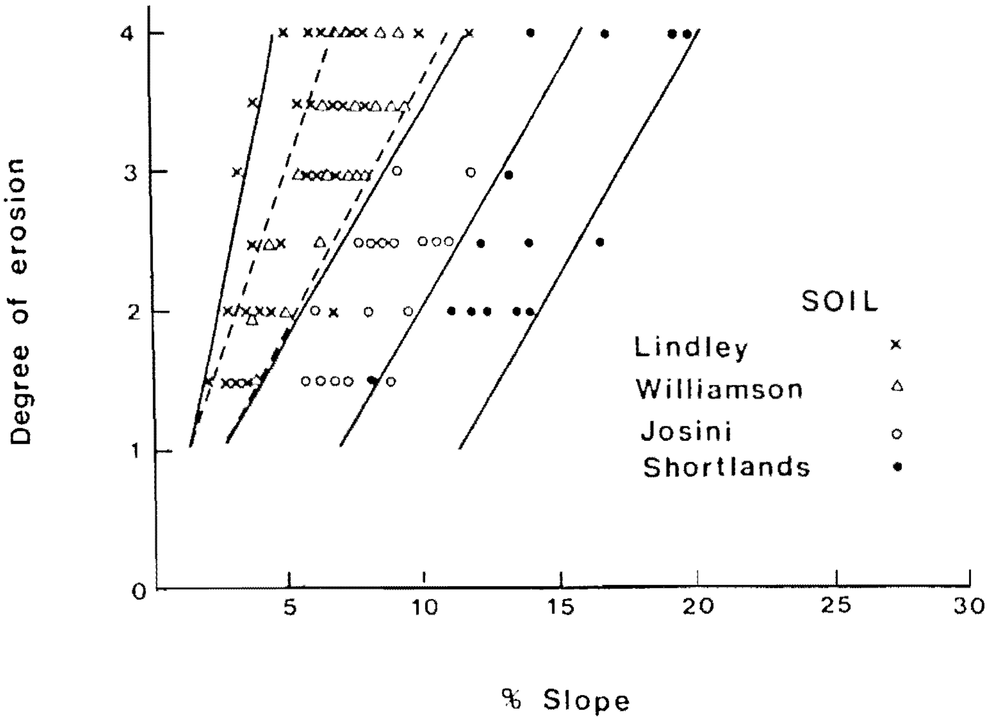

Figure 3.13 Relationship between slope percentage and degree of erosion for different soils of the Middledrift pedosystem. (Laker & D'Huy-vetter, 1988)

3.3.3.4 Correlations between soil factors and erosion

A significant correlation between ***ESP*** of the A horizon and degree of erosion was found. D'Huyvetter found that the ESP of the B horizon also gave a significant, but much poorer, correlation. Relatively (but not absolutely) high A horizon ESP values and high B horizon ESP values in the Valsrivier and Kroonstad soils correspond to severe erosion. Low A and B horizon ESP values, such as found in the Oakleaf and Shortlands soils, were associated with weak to moderate degrees of erosion. The Glenrosa soil, having low ESP values in both A and B horizon while showing severe erosion, was an exception.

D'Huyvetter found that ***S-value/100 g clay*** of the A horizon was significantly, but not well, correlated with degree of erosion. S-value/100 g of the B horizon was very well correlated with degree of erosion. Mg^{++}/Ca^{++} and $(Mg^{++} + Na^{+}) / Ca^{++}$ ratios did not give statistically significantly correlations with the degree of erosion.

***Texture*:** The fine sand content of the A horizon was found to be significantly correlated with the degree of erosion. D'Huyvetter indicated that high fine sand levels partly explain the high degree of erodibility of the Kroonstad and Glenrosa soils.

3.3.4 Conclusions

The role of parent material in the erodibility of soils was clearly indicated. Soils that are derived from dolerite showed higher stability against erosion compared with the majority of other soils, mainly those from Beaufort mudstone.

A significant correlation was found between the ESP of the soil and degree of erosion. One should note that the general rule is that soil with lower ESP values would be less erodable and those with high ESP would be more erodable. However in some cases it was found that in some soil forms even soils with low ESP, were strongly erodable. This implies that other soil chemical and physical factors and/or topographical factors are responsible for their susceptibility to erosion.

It was also found that the erodibility of a soil from corresponding soil forms decreases with increasing rainfall between the different pedosystems. With increasing rainfall in the order Middledrift < Mavuso < Keiskammahoek, critical threshold slope percentages increased in the same order.

For most of the soil forms, clay content plays a significant role in their stability against erosion. Soils which were dominated by fine sand were highly susceptible to erosion.

3.4 LEVY, GJ (1988): THE EFFECTS OF CLAY MINERALOGY AND EXCHANGEABLE CATIONS ON SOME OF THE HYDRAULIC PROPERTIES OF SOILS

3.4.1 General background and objective

The study was undertaken to determine:

1. The effect of clay mineralogy, specifically kaolinite and illite, on crust formation and infiltration rate (IR), as opposed to hydraulic conductivity (HC), and interaction thereof with soil sodicity.
2. The homogeneity in structure and permeability of crusts.
3. The effects of exchangeable Mg and K on IR as opposed to their effects on HC.

The present study focusses mainly on the findings of Levy in regard to (1) and (3).

In his literature review, Levy indicates that the ***structure of the crust*** differs from that of the bulk of the soil and hence affects soil properties such as water penetration (McIntyre, 1958). Obviously if water does not penetrate into the soil, runoff takes place (Brester and Kemper, 1970), with the increased runoff leading to accelerated erosion.

The hydraulic conductivity (HC) and the IR of the soil are both known to depend, among others, on the ***clay mineralogy*** of the soil and ***exchangeable cation composition***. HC studies showed that unstable soils were those high in 2:1 layer silicates, whereas the more stable ones were those high in kaolinite (McNeal and Coleman, 1966; Yaron and Thomas, 1968). IR and crusting were mostly studied on montmorillonitic soils since they are dominant in arid and semi-arid areas where crusting is a severe problem. It is well known that calcium has favourable effects on the physical properties of soil whereas sodium has a deleterious effect. Many South African soils contain kaolinite or illite as the dominant clay mineral as compared to the Israeli soils. It is suggested that in comparison to smectitic soils, kaolinitic and illitic soil have a greater resistance to the impact of raindrops and are probably less susceptible to chemical dispersion of the clay, which supplements the physical mechanism in crust formation.

Du Plessis and Shainberg (1985) studied the effect of soil sodicity and phosphogypsum application on the permeability of some South African soils with different clay mineralogies. Some of the soils were stable when subjected to rain but others were susceptible to crusting, which became more pronounced as the ESP increased. This shows the differences that exist in the behaviour of the soils.

The hydraulic conductivity (HC) of kaolinitic and illitic soils has been studied extensively, yet opinions differ regarding the effect of these clays on HC. McNeal and Coleman (1966) found that under low salt and high sodium conditions soils high in kaolinite appeared to be stable, whereas smectitic soils were unstable. Soils with clays containing predominantely 2:1 layer-silicates but having only moderate amounts of smectite and high amounts of illite were intermediate in their behaviour. Similar results with regard to the stabilizing effect of kaolinite on the HC were found by El-Swaify and Swindale (1969).

Two main mechanisms have been proposed to explain a reduction in HC during water flow through soil. Firstly, Quirk and Schofield (1955) suggested that under a given condition, *the swelling of clay particles* could result in a blocking of the conducting pores. McNeal, Norvell and Coleman (1966) found a linear relationship between HC reduction and macroscopic swelling of soil clays. *Dispersion* was proposed as the second mechanism (Quirk & Schofield, 1955). Dispersion occurs when the forces of attraction between particles are no longer strong enough to oppose repulsive forces, and thus the plates can be moved by an external force. Hence the clogging of soil pores by dispersed clay particles reduces the HC. This mechanism was recognised by Rhoades and Ingvalson (1969), Van der Merwe and Burger (1969), Frenkel, Goertzen and Rhoades (1978) and Pupisky and Shainberg (1979).

3.4.2 Experimental procedures

Four soil samples, varying in their clay content and clay mineralogy, were collected from different sites in South Africa. Two samples were from the then Transvaal

(Potchefstroom, and Halfway House), one from the then Natal (Hluhluwe) and one from the then Cape Province (Somerset East). The soils were classified according to the South African binomial soil classification system (MacVicar *et al.*, 1977) at form level (Table 3.21)

Table 3.21 Classification of soils used in the study (Levy, 1988)

Site	**Soil form**
Potchestroom	Avalon
Halfway House	Hutton
Hluhluwe	Hutton
Somerset East	Swartland

Profile descriptions and exact site locations of the soils are given by Levy (1988).

Infiltration and crust formation were studied using the same laboratory scale rainfall simulator that was later used by Smith (1990) for the research discussed in Section 3.2 hereof.

The rainfall simulator consists of two parts;viz (1) the rain application system and (2) the soil box carousel.

Soil samples of < 4 mm size fraction were packed in the boxes in a layer 20 mm thick over an 80 mm layer of coarse sand. It was found that it is unnecessary to use soil layers thicker than 25,0 mm since infiltration rates are unaffected even if thicker layers are used (Morin *et al.*, 1967). A piece of cloth was placed between the coarse sand and the soil to ensure continuous flow of water from the soil to the sand. The soil and the coarse sand were saturated from below using tap water (EC = 1 mS/m), thus the soil was saturated when rain application commenced (Levy, 1988).

The ESP of the soil was adjusted to four different levels giving ESP values of about 1%, 2,5%, 5% and 10% respectively, in the rainfall simulator using fog-type rain. The soil was subjected to 90 mm of a 400 $me.dm^{-3}$ $NaCl$-$CaCl_2$ solution of an appropriate sodium adsorption ratio (SAR), followed by 60 mm of a more dilute solution (50 me dm^{-3}) having the same SAR. Thereafter it was exposed to a storm with energy, using distilled water (to simulate rain water). During each storm the volume of water percolating through the soil was recorded. Four replicates were used in each treatment. Following the storm, samples were taken and the ESP was then determined (Levy, 1988).

Levy (1988) compared the clay mineralogy and final infiltration rate (FIR) values from his study with the clay mineralogy and FIR values obtained by Ben-Hur *et al.*, (1985) for four soils with a similar range of clay contents from the coastal plans of Israel. Soils with similar ESP values were selected for the comparison. The same type of rainfall simulator was used in both studies, but the rain intensity differed, viz. 45 mm/h and 31,6 mm/h in the SA and Israeli studies respectively. It was found that within this range, the rain intensity has no significant effect on the FIR (Morin and Benyamini, 1977).

3.4.3 Results

3.4.3.1 Relationship between clay mineralogy and final infiltration rate (FIR)

Low FIR values were obtained for all the soils after subjecting them to rain with distilled water, with the exception of the Zwartfontein series (the Hutton (Z) soil) (Table 3.22). Levy (1988) indicated that Hutton (H), Avalon and Swartland showed clear examples of crust formation at the surface. The results indicated that for any clay content the FIR values for the SA soils were always higher than FIR values for the Israel soils. The kaolinitic soils from SA maintained FIR values in the range of 5 – 23 mm/h, which were in agreement with the results presented by Miller (1987), who studied infiltration and soil loss on three kaolinitic soils. The FIR values of the SA soils were at least twice as high as the FIR of the corresponding smectitic Israeli soils. The dominant clay minerals in the SA soils were either kaolinite or illite with smectite and interstratified material as the secondary minerals while smectite was the dominant clay mineral with kaolinite the secondary mineral in the region of Israel where the Israeli samples were taken (Levy *et al.*, 1986).

Table 3.22 Clay mineralogy, exchangeable sodium percentage (ESP) and final infiltration rate (FIR) for the four South African soils studied and four Israeli soils

Soil	Country[a]	Dominant clay[b] minerals	ESP	C.E.C[c] $cmol_{(+)}$/ kg clay	Clay content %	FIR mm/h
Hutton (Z)	SA	K(4), Isd(2), St(1)	2,4	72,8	7,0	22,8
Typic Rhodoxeralf	IL	St,K	2,2	65,4	7,8	4,4
Swartland	SA	I(4), Is[e](3)	2,1	43,6	14,9	4,2
Typic Rhodoxeralf	IL	St,K	2,5	81,7	15,9	2,6
Avalon	SA	K(4),I(2),St(2)	2,7	40,5	20,0	5,5
Typic Rhodoxeralf	IL	St,K	2,2	59,3	19,2	2,0
Hutton (H)	SA	K(5), I(1)	2,9	9,5	29,6	6,5
Typic Rhodoxeralf	IL	St,K	2,5	76,0	32,0	3,2

[a]Country of origin of the soil: SA = South Africa, IL = Israel.

[b]I = Illite, is = interstratified material, K = Kaolinite, St = Smectite,

(1) = very weak, (5) = very strong.

[c]C.E.C = Cation exchange capacity of the clay fraction of the soils.

[d]IS = material containing 1:1 clay components.

[e]IS = material containing swelling components.

The Israeli soils were richer in iron than the four South African soils included in the study, containing approximately 3% free iron (Frenkel, 1970), compared with <1,1% in the SA soils. This factor would have been expected to increase the relative stability of the Israeli soils. The organic carbon content of all the soils studied, was very low (<1%), which could be a major contributing factor to their instability.

According to Levy (1988) it seems reasonable to assume that the differences in the FIR values between the two groups of soils, with the possible exception of the Hutton (Z) soil, could be ascribed to the differences in their clay mineralogy. The presence of large proportion of medium sand (30,6 %) in the Hutton (Z) soil, combined with its low clay and silt contents, probably limited crusting in this soil and contributed to its high FIR.

3.4.3.2 Effect of ESP on crusting and FIR

Graphs of infiltration rate (IR, mmh^{-1}) vs cumulative rain (mm) derived from subjecting the soil to high energy rain and as a function of three different levels of ESP (approx. 1,5 ; 5 and 10) for four SA soils are given in Figures 3.14-3.16. Soils were classified according to the response to ESP. They were classified into the following three groups, viz.

Group 1: Hardly affected by ESP.

Group 2: Affected at high ESP levels only.

Group 3: Affected at all ESP levels

According to Levy (1988) the Hutton (H) soil belongs to Group 1, i.e. it was hardly affected by sodicity up to an ESP level of 9 (Figure 3.14). A crust did form at the soil surface as can be seen from the decrease in IR with increase in cumulative rain (Figure 3.14). Levy found that the FIR decreased only slightly from 6,7 mm h^{-1} to 5,4 mm h^{-1} with an increase in ESP from 1,2 to 8,9, but no significant differences could be detected in the cumulative infiltration (after 100 of rain as computed from equations) which maintained relatively high values (approxi. 39 mm). Since cumulative infiltration is an integrated measure of the IR curve, Levy suggested that small differences in FIR might not affect it, especially in stable soils where a fair amount of rain is needed in order to reach the FIR. Schofield and Samson (1954), indicated that kaolinite does not disperse readily under sodic conditions due to attraction of the positive charges on the edge faces for the negative charges in planar surfaces of the crystals. Hence the physical impact of the raindrops was the dominant mechanism in crust formation and the reduction in the permeability of the Hutton (H) soil, while the chemical dispersion of clay due to sodicity, which is an important factor in determining infiltration into smectitic soils, apparently had only a small effect on this kaolinitic soil. Thus, increasing sodicity did not have a pronounced effect on the characteristics of the crust formed in this group of soils.

Levy (1988) indicated that the Hutton (Z) soil belongs to Group 2, i.e.. The deleterious effect of exchangeable Na came into effect only at high ESP levels (Figure 3.14). The Hutton (Z) soil was found to be very stable in the lower ESP range. There was no measurable drop in the IR for the first 50 mm of the rain storm and therefore the rain intensity (45 mm h^{-1}) determined the IR. However, at a later stage IR decreased moderately until a final, yet high, IR of 21,3 mm h^{-1} was obtained. The effect of ESP on this soil was noticeable only at the relatively high ESP level of 13,0 where the decrease in IR started sooner (after 30 mm of rain) and dropped to a FIR of 11,0 mm h^{-1}, which is still relatively high (Figure 3.14). Due to these high FIR values, which indicate that the crust formed was quite permeable, the cumulative infiltration values were extremely high: 85,2 mm for the soils with low ESP and 63,4 mm for the soil with ESP of 13,0 (Table 3.23). Levy (1988) indicated that the stability of this soil when exposed to rain, could be ascribed partly to kaolinite being the dominant clay mineral as well as to the large fraction of medium size sand. Yet, the adverse effect of exchangeable Na on clay dispersion and crusting, enhanced probably by the presence of significant amounts of smectites and interstratified clay minerals in this soil, could be detected only at an ESP level of 13,0.

According to Levy (1988) Group 3 consisted of the Swartland and Avalon soils, which were the most susceptible to increasing soil sodicity. In both soils the IR dropped sharply from the beginning of the rainstorm even at a moderate ESP level (4 to 5). Thereafter each increase in ESP caused a further decrease in FIR. Comparing the Avalon and Swartland, Levy indicated that the Avalon soil was the one that was most affected by ESP. This was unexpected because Swartland soils are usually amongst the most unstable soils in South Africa and Avalon soils amongst the most stable. The Avalon soil contained predominantly kaolinite, but also a significant amount of smectite in its clay fraction (Table 3.22). The findings in respect of the Avalon could therefore be the result of an interaction between the opposing effects of kaolinite and smectite on the stability of the aggregates at the soil surface. Du Plessis and Shainberg (1985) found that IR was similarly affected by ESP in some other SA soils with similar clay mineralogy. If one compares Figure 3.15 and Figure 3.16, these conclusions appear incorrect. The Swartland soil was, in fact, affected more than the Avalon soil, in the sense that the infiltration rate dropped faster for the Swartland than for the Avalon soil.

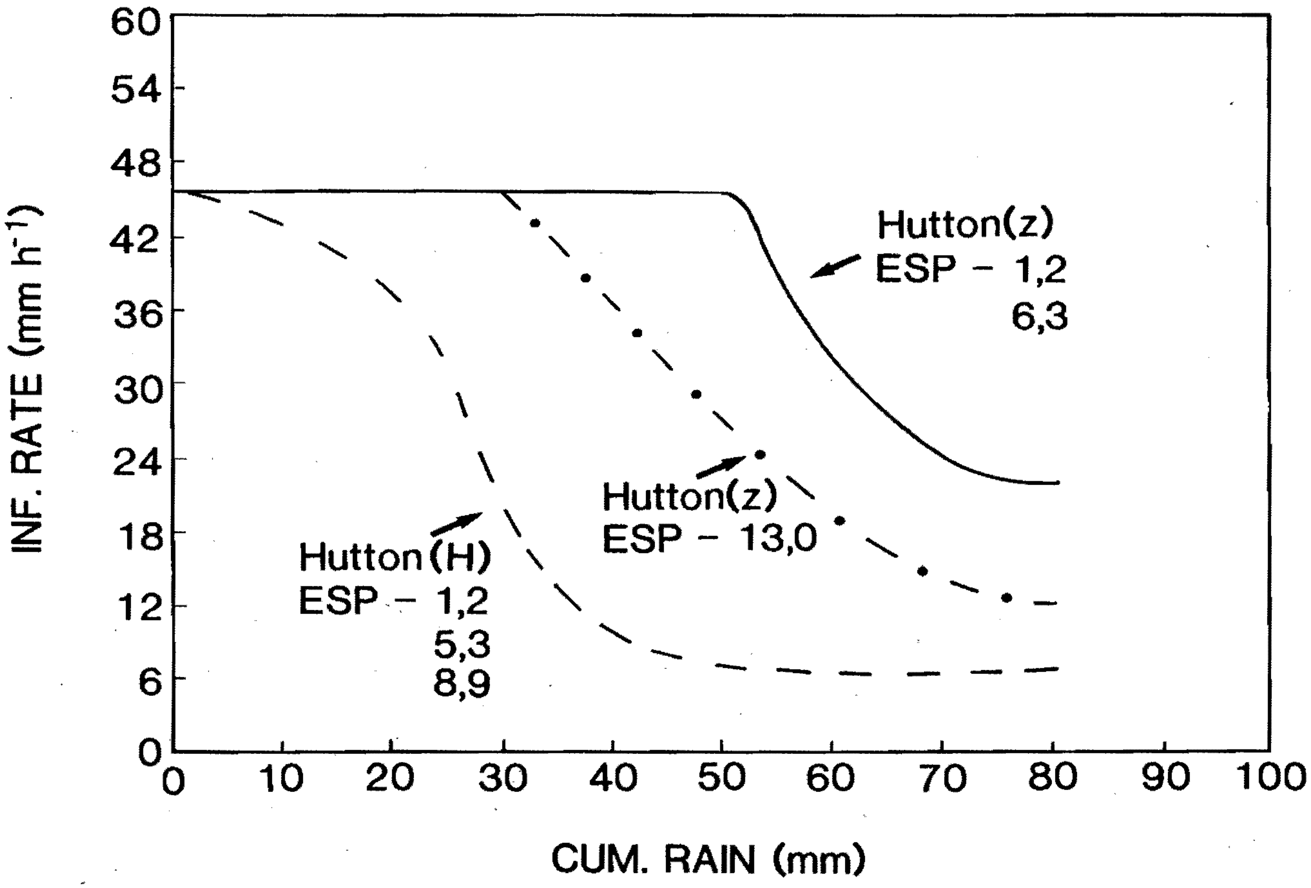

FIG. 3.14 The effect of soil ESP on infiltration rates of the two relatively stable kaolinitic soils.

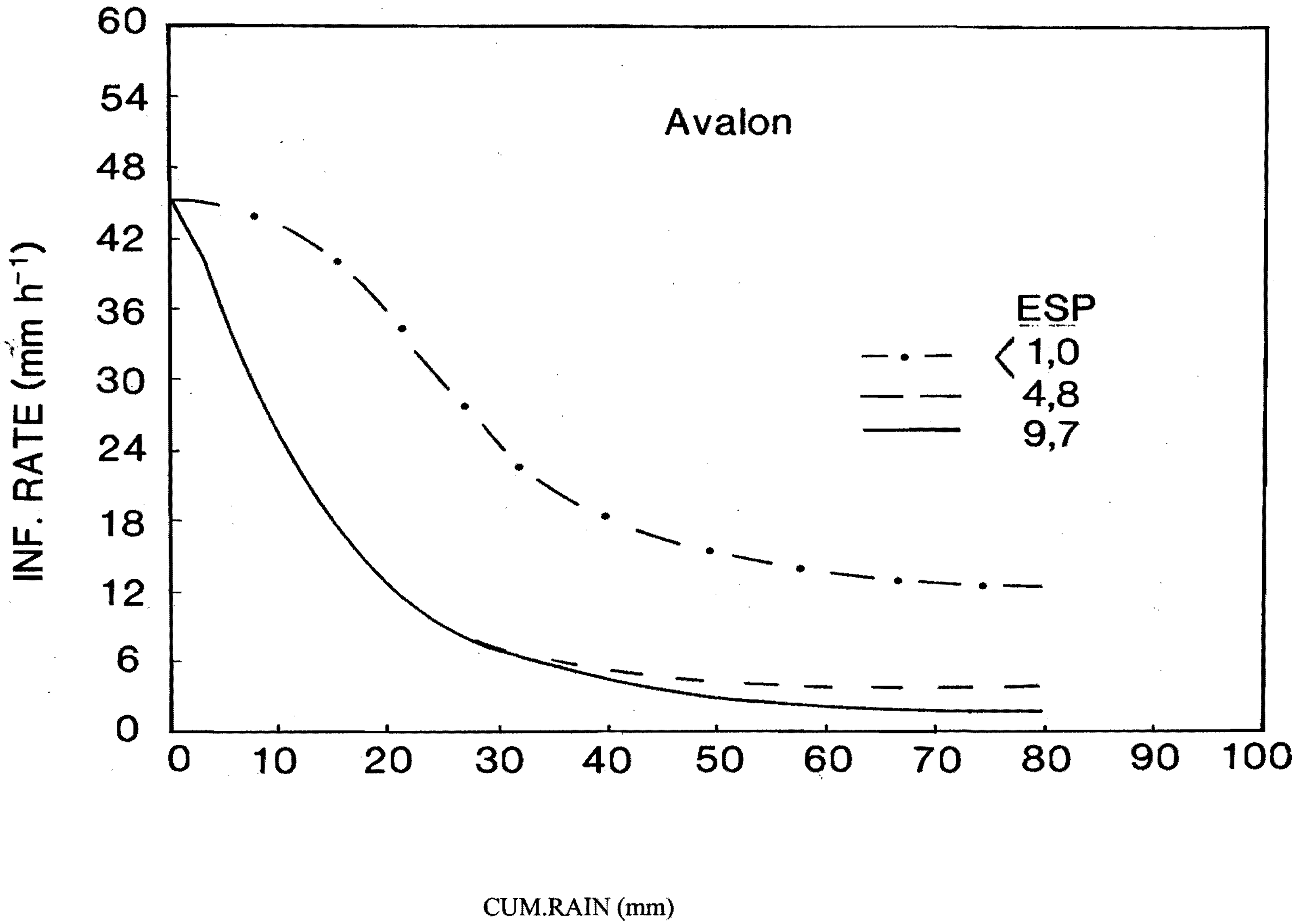

FIG. 3.15 The effect of soil ESP on the infiltration rate of the relatively unstable kaolinitic soils.

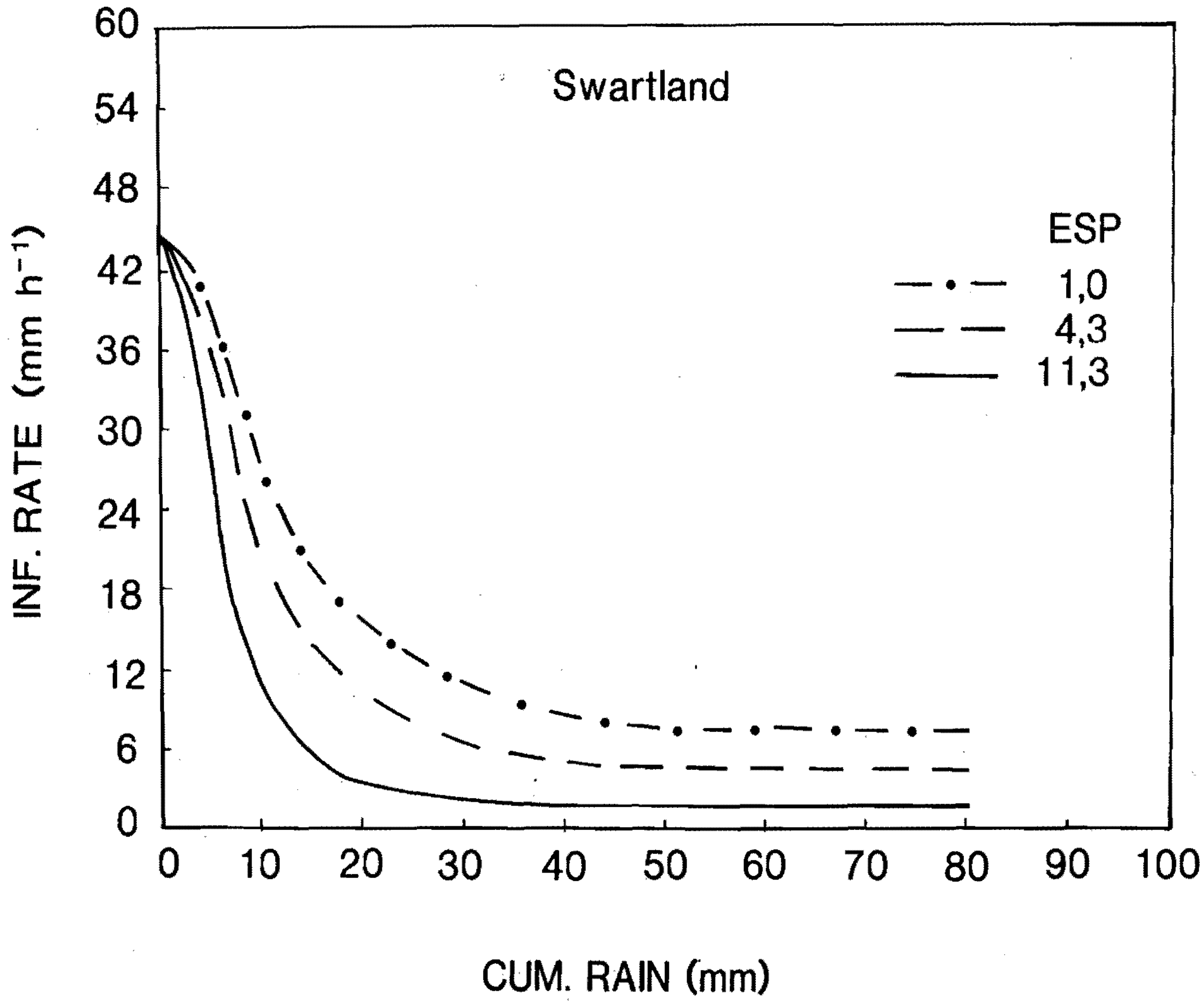

FIG. 3.16 The effect of soil ESP on the infiltration rate of the illitic soil.

Table 3.23 The cumulative infiltration (CIF) after 100 mm of rain and final infiltration rates (FIR) for the four South African soils.

Soil	ESP	CIF[a]		FIR[a]	
		mm		mm h^{-1}	
Hutton (Z)	1,0	85,2	a	21,3	a
	6,3	85,9	a	21,9	a
	13,0	63,4	b	11,0	b
Avalon	<1,0	43,1	a	12,9	a
	4,8	22,5	b	4,1	b
	9,7	21,5	b	2,7	c
Hutton (H)	1,2	38,6	a	6,7	a
	5,3	40,4	a	6,3	a
	8,9	38,7	a	5,4	a
Swartland	1,1	28,1	a	8,4	a
	4,3	21,7	b	5,2	b
	11,3	14,1	c	2,5	c

3.4.3.3 The effects of exchangeable Mg and K on infiltration rate as opposed to their effects on hydraulic conductivity

This study was carried out since some of the waters in South Africa (SA) contain a high concentration of Mg. The molar Ca/Mg ratio in the P.K le Roux and Paul Sauer dams, which supply water to big irrigation schemes in South Africa, is 1:1. The use of such waters for irrigation, coupled with an increase in irrigation efficiency, will result in large increases in exchangeable Mg in the soil. There are many views concerning the effect of adsorbed Mg on hydraulic properties of soil. The U.S salinity laboratory staff (1954) grouped Ca and Mg together as similar ions, beneficial in developing and maintaining soil structure. On the other hand, Van der Merwe and Burger (1969) found that Na-Mg saturated soil was structurally less stable than Na-Ca soil. From these results they concluded that exchangeable Mg can have a **deleterious** effect on the structure and permeability of the soil.

Much of the results presented, describe only the effect of exchangeable Mg on the HC of the soil but not on the infiltration properties of soil exposed to rain. The IR of a soil is much more susceptible to its sodicity and the chemistry of the applied water than to the HC (Oster and Schroer, 1979)

As pointed out earlier, one of Levy's objectives was to study the effects of Mg and K on IR. However, his emphasis was on different ESP levels and the extent to which ESP affected HC of the soil. Levy (1988), indicated that Mg has a negative effect on selected soil physical properties. It is also mentioned that increasing the amount of K in the exchangeable phase resulted in a decrease in the HC as well as in the infiltration rate of the soil. Levy (1988) indicated that the magnitude of these phenomena depended on the clay mineralogy of the soils studied, being the smallest in kaolinitic soils and the largest in an illitic soil. No indications were found that he actually did systematic studies on Mg and K, and the conclusions may be simply due to incidental observation during the study.

Unfortunately the effects of Mg and K were not really evaluated, as the main emphasis was on ESP level.

3.5 SUMNER, ME (1975): THE PHYSICAL AND CHEMICAL PROPERTIES OF TALL GRASSVELD SOILS OF NATAL IN RELATION TO THEIR ERODIBILITY

3.5.1 General background and objectives

The Karoo, former Transkei, parts of the Free State and former Transvaal and a considerable part of Kwazulu-Natal are subject to erosion by water (Sumner, 1957). A large proportion of the erodable soils in South Africa occur on the same geological formation, viz. the Beaufort and Ecca groups of the Karoo supergroup, which form the parent material of most soils in the Karoo, former Transkei, Free State and Western and Northern KwaZulu-Natal.

According to Sumner (1957) erosion in KwaZulu-Natal, the Eastern Cape (including the former Transkei) and the Eastern Free State is most severe in the semi-arid areas, although it is not confined solely to these areas. In KwaZulu-Natal, erosion is most severe in the Thornveld and Tall Grassveld. Soil erosion had been studied by non-soil scientists in these areas and various control measures were introduced, but there was a lack of basic understanding of the role of soil factors in erosion incidence in these areas.

The study was undertaken to:

(1) Conduct laboratory studies of the physical and chemical properties of several soils from the Tall Grassveld of Natal that are known to vary in their erodibilities.

(2) Show that the nature of the soil plays a most significant role in determining its susceptibility to erosion.

3.5.2 Areas studied and research methodologies

3.5.2.1 Areas studied

Sumner conducted studies in the following areas of Northern KwaZulu-Natal:

- Between Estcourt and Bergville.
- Vryheid district.
- North of Utrecht.
- Around Paul Pietersburg

3.5.2.2 Geology of the areas studied

The surface geology of the areas that were studied, consists mainly of sedimentary rocks of the Beaufort and Ecca groups (Sumner, 1957). Intrusions of Jurassic dolerite, a basic igneous rock, occur widespread throughout all the areas.

3.5.2.3 Field survey techniques

The area was transversed by motor transport. At this stage, the soils were studied by inspecting the exposed faces of dongas. On the basis of erodibility, tall grass veld was divided into groups to identify different types of soils (Sumner, 1957). The soils for all the areas studied were divided into two distinct groups, viz. (a) those derived from sedimentary rocks of the Beaufort and Ecca group, and (b) those derived from dolerite.

Six sites were selected, and eleven pits were dug in order to examine typical profiles and to facilitate sampling for laboratory investigations. Soils were classified according to United State Soil Survey Manual (Sumner, 1957), as the study predated the adoption of South African Binomial System (MacVicar and others, 1977) and Soil Classification Working Group (1991).

3.5.2.4 Laboratory analyses

Particle size analyses and determination of exchangeable bases were conducted. The hydrometer method according to Bouyoucos (1935) was employed, using sodium hexametaphosphate as dispersing agent.

Exchangeable Ca, Mg, Na and K were extracted according to the method outlined by Bower, Reitemeyer and Fireman (1952).

3.5.3 Results

3.5.3.1 Particle size distribution

A marked increase in clay content with depth was observed in residual soils derived from Beaufort sediments. The silt content tended to be higher in the upper horizon of the profile. At a depth of about 60 to 120 cm, a layer of ferruginous concretions in varying amounts occurs, indicating a zone of alternating waterlogging and drying. Soil becomes temporarily waterlogged during the wet summer months, leading (according to Sumner) to movement of iron, which is then precipitated in the illuvial horizon during the dry winter months.

The alluvial soils which developed in sediments derived from the sedimentary rocks possess topsoils similar in texture and structure to those of the residual soils. There is an increase in clay content in the subsoil, which probably represents a horizon of illuviation (Sumner, 1957). The silt content is high in the upper layers and gradually decreases with depth. Ferruginous concretions occur at about 120 to 180 cm and the soils are usually permanently wet and waterlogged below 180 cm.

The soil derived from dolerite shows higher clay in the topsoil as compared with that derived from sedimentary rocks. This high clay content is fairly constant down the profile. The silt content is of the same order as that of soil derived from the Beaufort sediments. The decrease in coarse sand with depth can, according to Sumner, be the result of fewer ferriginous concretions of coarse sand size (2.00 – 0.20 mm diameter) occurring in the subsoils of these soils, which are not subject to waterlogging.

3.5.3.2 Exchangeable base contents of Tall Grassveld soils (developed on the Estcourt geological formation) as compared with some other soils

The exchangeable base patterns of the topsoils derived from Beaufort sediments and dolerite are both in accordance with what one would expect for slightly acidic soils. However the pattern in the subsoils from Beaufort sediments is markedly different in that the exchangeable Mg content is high and in the order of the exchangeable Ca content. The high exchangeable Mg content is of importance when compared with the exchangeable Ca content and the total base content. High exchangeable Mg gave rise to very poor structure resulting in very compact soils and highly erodable soils.

The "Natal lateritic yellow earths" (Sumner, 1957), probably refer to Clovelly, Griffin, Hutton and Katspruit form soils (Soil Classification Working Group, 1991) of the higher Tabamhlope Plateau (Turner, personal communication). This area receives more than 950 mm rain per year, and these soils exhibit an extremely low base status as a result of the intense nature of weathering and leaching.

3.5.3.3 Classification of soils according to erodibility

On the basis of erodibility Sumner (1957) divided the soils of the Tall Grassveld into two distinct groups, viz. (a) those derived from Beaufort and Ecca sediments and (b) those derived from dolerite. The soils derived from the Beaufort and Ecca sediments exhibit a marked susceptibility to erosion by water whereas those derived from dolerite are reasonable stable. Since all soils occur under the same environmental conditions, it may be assumed that most of the external factors which might affect the degree of erosion are practically constant. The marked differences in their erodibility must, therefore, be due to variation in their physical and chemical properties (Sumner, 1957).

3.5.3.3.1 Highly erodible soils

a. ***Residual soils on Beaufort sediments***: These soils occur on elevated plateaux and on slopes of hills. According to Sumner (1957) they "are usually" shallow varying from 75 to135 cm in depth – which is actually deep for South African soils. Differentiation within the profile is very marked. The A-horizon is usually a pinkish grey to pinkish white sandy clay loam about 30 cm thick. It is rich in silica and contains abundant grass roots. This horizon merges into a reddish yellow to very pale brown gravelly sandy clay loam to clay, with fewer grass roots and abundant ferruginous concretions mixed with calcareous nodules (Sumner, 1957).

b. ***Alluvial soils on sediments of Beaufort origin***: These soils and sediments occur in the bottom of the large basin of Tabamphlope. In general, the soils and sediments are from 3 to 12 metres in depth. The A-horizon varies from a pinkish grey to light grey loamy sand to sandy loam about 30 cm thick containing abundant grass roots. This is followed by a white to light grey sandy clay to loam, which gradually merges into a pale yellow to very pale brown clay to clay loam with a few scattered ferruginous concentrations and lime nodules. Both the surface and internal drainage of these soils are poor (Sumner, 1957)

There was enough evidence in the field to show that these soils are or were alluvial in nature. This has been indicated by the horizontal and typical sedimentary bedding in profiles when observed on donga faces. Successive layers were no doubt deposited during or after floods. Alluvial soils generally showed more signs of erodibility, as indicated by dongas' width and depth (Sumner,1957).

c. ***Alluvial soils on sediments of Ecca origin***: In general these soils are very similar to the alluvial soils and sediments derived from Beaufort materials.

3.5.3.3.2 Relatively stable residual soils on dolerite

When comparing soils formed from dolerite with soils derived from Beaufort and Ecca sediments, the soils derived from dolerite were found to be less erodable. The soils derived from dolerite can be divided into two groups, viz. (a) red clays and (b) black clays. The latter soils are of less importance and were not studied by Sumner. The red clay soils show very little differentiation within the profile. In general the topsoil varies from a light red to reddish brown granular sandy clay, containing abundant grass roots. This gradually merges into a reddish yellow to light yellowish brown columnar sandy clay loam to clay with fewer grass roots (Sumner, 1957).

3.5.4 Conclusions

The exchangeable Mg content may play an important role in determining the structure of the soil and thereby indirectly affecting its erodibility. It is well-known that high

Mg content is conducive to poor structure. Soils that have high Mg content exhibited very poor structural properties. In general soils derived from Beaufort and Ecca groups are susceptible to erosion by water, whereas soils derived from dolerite are stable.

3.6 RAPP, I (1998): THE EFFECTS OF SOIL PROPERTIES AND EXPERIMENTAL CONDITIONS ON THE RILL ERODIBILITIES OF SELECTED SOILS

3.6.1 General background and objectives

Rapp (1998) studied the effects of soil properties and experimental conditions on the ***rill*** erodibilities of selected soils. Soil erosion by water is commonly divided into two forms: rill and interrill erosion (Mosley, 1974; Foster and Meyer, 1977). These are also termed concentrated flow erosion and sheet erosion, respectively, referring to the hydraulic condition of the runoff water flow.

Rill erosion (RE) depends on flow shear stress and stream's power, the shear strength of the soil and cohesion forces between soil particles and the stream transporting capacity (Rapp, 1998). Soil shear strength depends on soil properties (e.g. texture, clay mineralogy, exchangeable cations, organic matter content) and the experimental conditions (e.g. soil bulk density, moisture content, aging duration, prewetting rate). Rill erodibility (k_r) is often considered as a soil property, related to basic soil physical and chemical characteristics. Attempts to relate k_r to soil properties were of limited success in the past. It was hypothesized (1) that k_r depends in addition to soil properties also on the experimental conditions, and (2) that if both soil properties and experimental conditions are considered, it would be possible to quantify the effects of soil properties on k_r in the laboratory.

Stream power has been advocated as important parameter in soil detachment and transport (Rose, 1985). It is proportional to the product of bed slope and flow rate (Moss *et al.*, 1980; Rose, 1985). However, in most models particle detachment is expressed as a function of the excess of applied shear over the material critical shear stress (Li *et al.*, 1977; Foster *et al.*, 1985; Lane *et al.*, 1988;). Nearing *et al.* (1991a) concluded that neither shear stress nor stream power were appropriate flow parameters for describing detachment rates for their soils, whereas Shainberg *et al.* (1994) found that for the Miami silt loam soil both were adequate.

Soil particles at the rill bed are held against the shear force of flowing water by three complementary forces: gravity, suction and inter-particle cohesive forces. The vector of the gravity forces, which keeps a soil particle from being lifted, depends on the slope of the rill bed, the slopes in the rill and on the mass of the soil particle to be detached and transported. The cohesive forces between soil particles increase with clay content and decrease with increases in organic matter coating (Kemper and Roseman, 1984) and increased ESP. Cohesive forces also increase with increases in soil bulk density and moisture content (Shainberg et al., 1996). There is an optimum moisture content and bulk density for cohesive forces to develop (Shainberg et al., 1996). One should also understand the direct and indirect effects of soil properties such as bulk density, moisture content and hydrology on rill erosion.

For this study, three specific objectives of Rapp (1998) have been selected. These are as follows:

(a) To determine soil erodibility of various South African and Israeli soils using a laboratory miniflume procedure.

(b) To quantify in laboratory studies the effect of inherent soil properties (texture, clay mineralogy, organic matter content, soil pH and sodicity) and water quality (electrolyte concentration) on the cohesive forces between soil particles.

(c) To evaluate the effect of experimental conditions and their interaction with soil properties on the erodibility of a soil.

3.6.2 Materials and methods

3.6.2.1 Description of soils used

Soils from South Africa and Israel were used. The soils were classified according to the South African soil classification system (Soil Classification Working Group, 1991) and USDA's Soil Taxonomy (Soil Survey Staff, 1975). In total 15 soils were used in this study (12 from South Africa and three from Israel). Locations of the soils are presented in Table 3.24. The soils are listed in order of increasing clay content for those from each country.

The soils from South Africa covered a wide range with respect to climate, parent material and soil properties. The three soils from Israel represented the most important cultivated soil types in Israel. It should be noted that the Israeli Hamra soil, which was classified as typic Rhodexeralf (Dan *et al.*, 1968; Koyumdjisk *et al.*, 1988), has a B horizon with prismatic/columnar structure and is, therefore, much like the similar unstable red prismatic "duplex" soils of South Africa and Australia, e.g. the Natrustalf ANB soil (Table 3.24).

3.6.2.2 Soil analysis methods

The soil samples were air dried, crushed and sieved through a 2 mm sieve. Some of the chemical, physical and mineralogical properties of the soils are summarized in Table 3.25. Discussion refers to the soils by their code names as indicated in Tables 3.24 and 3.25. Sand particle size distribution was determined using a dry sieve analysis while silt and clay were determined using the pipette method (Gee and Bauder, 1986). The clay mineralogies of South African soils were determined using x-ray diffraction (XRD). Semi-quantitative estimates of the amounts of the various clay minerals were obtained from relative peak areas of the diffractoctograms of the clay fractions after removal of organic matter and iron (Hahne and Fitzpatrick, 1985). The clay mineralogical compositions of the Israeli soils were based on the work of Koyumdjisky *et al.* (1988) and Banin and Amiel (1969). Organic carbon (OC) content of the RSA soils was determined with the Walkeley-Black wet oxidation procedure (Nelson and Sommers, 1982) while for the Israeli soils it was determined by multiplying the organic matter value (obtained by ignition at 400 °C) with a factor of 0.581 (Nelson and Sommers, 1982). Exchangeable cations were determined with ammonium acetate extraction. Cation exchange capacity (CEC) was determined by using the ammonia acetate method (US Department of Agriculture, 1972) for the South African soils or the Na acetate method (Richards, 1954) for the IL soils. Soil pH was determined in a 1: 2.5 soil/water ratio suspension on mass basis (McLean, 1982).

3.6.2.3 *Rill erosion (RE) Simulation system*

A small laboratory scale mini-flume apparatus for studying concentrated flow (CF) erosion in rills was used.

Different shear stress/stream power values were induced by means of combinations of different slope angles and flow velocities.

For some soils different ESP levels were used to study the effects of ESP on the rill erodibilities of the soil. The effects of different total electrolyte concentrations of the flowing water on rill erosion were also studied.

3.6.3 RESULTS

3.6.3.1 Effects of soil bulk density (ρ_b) and moisture content (MC) on rill erosion

Bulk density (ρ_b) values of the soils in the flumes are presented in Table 3.26. The ρ_b values decreased with increasing clay content. Based on 13 of the soil samples, a linear regression between ρ_b (y) and the clay content of the soil (x) was obtained ($y = 1.52 - 4.49 \times 10^{-4} \times X$; $r^2 = 0,59$). With increase in clay content the stability of soil aggregates increases and bulk density decreases (Kemper and Kock, 1966).

Soils with high clay content (and low ρ_b) also had high moisture contents (MC) following prewetting (Table 3.24), while soils with low clay content (and high ρ_b) had low MC. A negative correlation ($r^2 = 0,69$) was obtained between moisture contents and bulk densities of the 15 soils, which was even better than the correlation between MC and clay content (Figure 3.17a). The improved correlation with bulk density is due to the fact that bulk density and micro-porosity are also correlated with stable aggregates, which were partially destroyed during the sample preparation. Rill erodibility increases with increase in bulk density, and decreases with increasing moisture content (Figure 3.17).

Kemper and Rosenau (1984) and Shainberg *et al.*, (1996) indicated in their studies that the effect of moisture content is more complicated. General speaking, in unstable soils, when moisture content is high, soil particles are far apart and opportunities for cohesion to develop are small. Conversely, when moisture content is low, soil particles are not free to reorient themselves to a position with low energy and high cohesion between particles (Shainberg *et al.*, 1996) and rill erodibility is high.

It should be noted that in the correlation between rill erodibility and bulk density and rill erodibility and moisture content, four soils with very high erodibilities were not included. The very high erodibilities of these soils were due to their high sodicity and salinity.

TABLE 3.24: Classification and location of soil samples #.

Soil				SA Binomial system ***		USDA ****
Code **	Location	Latitude**	Longitude	Soil form	Soil family	Taxonomy
South African soils						
ANA	Aliwal North	30.38	26.44	Sterkspruit	Smithfield	Natrustalf
TH3	Thabazimbi	24.29	27.08	Hutton	Stella	Typic Camborthid
TH2	Thabazimbi	24.35	27.15	Oakleaf	Richie	Entic Haplumberpt
PB	Piketberg	32.52	18.46	Glenrosa	Williamson	Lithic Haplumberpt
HO	Holfontein	26.33	24.41	Westleigh	Mareetsane	Plinthaqualf
TB	Taaibosbult	26.54	26.55	Westleigh	Mareetsane	Plinthaqualf
CY	Cyres	26.52	27.00	Westleign	Mareetsane	Plinthaqualf
R	Riviersonderend	19.54	34.09	Glenrosa	Robmore	Lithic Xaplumbert
IR	Irene	25.54	28.12	Hutton	Suurbekom	Paleudalf
TH1	Thabazimbi	24.13	26.54	Oakleaf	Ritchie	EnticHaplumbert
ANB	Aliwal North	30.38	26.44	Sterkspruit	Smithfield	Natrstalf
NC	Ncera	32.45	26.55	Shortlands	Bolweni	Typic Kandiustalf
Israel soils##						
H	Bet Dagan	32.00	34.49	Coastal plain	Hamra	Typic Rhodoxeralf
L	Nahal Oz	31.28	34.29	Northern Negev	Loess	Calcic Haploxeral
V	Qedma	31.41	34.46	Pleshet plain	Vertisol	Typic Chromoxerert

\# Soils are arranged according to their texture. Clay percentage increased down the list.

\## The USDA taxonomy of the IL soils are based on Koyumdjisky *et al.*, (1988) and Dan *et al.*, (1968).

* Soil codes are related to site location of the soil samples and to the soil form of the RSA soil and Israeli soils, respectively.

** Latitude and longitudes are given in a format of: degree.minutes

RSA soils are with South latitudes while those of the Israeli soils are North. All longitudes are east.

*** Soil Classification Working Group, 1991.

**** Soil Survey Staff, 1975.

Table 3.25 Some physical and mineralogical properties of the soils

Soil	clay	silt			Sand					CEC	pH	ESP	Organic carbon	Mineralogy Dominant clays (% of clay fraction)				Caco$_3$	EC**
		Fine	Coarse	Total	Very fine	Fine	Medium	Coarse	Total		Soil water Ratio								
	<0.002	0.002-0.02	O.02-0.005	0.0-2-0.05	0.05-0.106	0.106-0.25	0.25-0.5	0.5-2	0.05-2										
	mm																		
code	g kg^{-1}									Cmol(+) (kg soil)$^{-1}$	1:2.5	%	G kg^{-1}	Sm*	M	K	I/Sm	G kg^{-1}	Ds m^{-1}
South African soils																			
ANA	67	30	45	75	144	372	310	2	828	5.33	7.5	3.75	1.8	6	7	10	74		1.25
TH3	85	26	45	71	68	135	424	188	815	4.82	5.6	2.28	10.7	16	14	66			0.55
TH2	113	41	63	104	86	274	387	10	757	15.26	7.9	0.72	13.5	68	3	26		14.3	0.87
PB	173	148	118	266	64	102	305	83	554	5.53	7.3	4.70	7.4	4	55	37			0.68
HO	201	68	76	144	130	194	287	32	643	9.89	5.6	0.90	12.3	10	19	66			0.34
TB	224	67	59	126	124	212	275	16	627	10.18	6.2	1.08	7.1	11	11	71			19.93
CY	234	44	52	96	119	185	306	38	648	12.53	5.8	0.79	10.2	17	8	69		6.76	0.46
R	254	228	169	397	73	94	49	113	329	10.21	6.6	2.25	23.2	6	48	38			0.96
IR	335	106	79	185	58	107	241	60	466	12.51	5.9	0.48	11.7	13	20	61			1.21
TH1	336	175	217	392	74	34	144	1	253	26.14	7.1	0.84	6.3	68	5	24			1.51
ANB	339	74	56	130	89	196	221	1	507	17.32	8.4	12.7 8	2.4	15		10	73		2.45
NC	414	289	126	415	25	14	103	0	142	17.15	6.1	1.05	27.1	9		83			0.40
Israeli soils																			
H	100			50					850	11.86	7.1	1.00	7.0	55**	5	30	10		
L	250			233					517	23.80	8.0	3.50	8.7	50	15	15	20	182	
V	465			225					310	38.70	8.0	1.60	12.0	65	10	15	10	151	

* Kaolinite, M=Mica, Sm=Smectite, I/Sm=intrastified illite and smectite, Cal=Calcite, I=Illite, Qua=Quartz

** EC= electrical conductivity of the saturated soil extract

*** Clay mineralogy of the IL soils is based on Koyumdjisky *et al.*,(1998) and Banin and Amiel, 1969.

Table 3.26 Bulk density (ρb) and moisture content (MC) of the soils in the miniflume

Soil	Soil Bulk density		Capillary Moisture	
Code	**$Kg.m^{-3}$**	**Std. #**	**$g. kg^{-1}$**	**Std**
South African soils				
ANA	1.447	0.04	302.71	12.0
TH3	1.562	0.04	262.66	15.8
TH2	1.281	0.03	410.71	17.8
PB	1.603	0.06	241.04	23.8
HO	1.431	0.04	322.53	7.6
TB	1.485	0.05	305.59	18.4
CY	1.424	0.03	352.83	15.1
R	1.359	0.04	358.84	15.7
IR	1.346	0.03	375.43	14.9
TH1	1.389	0.02	397.70	25.8
ANB	1.436	0.04	369.43	19.7
NC	1.332	0.04	400.11	16.2
Israeli soils				
H	1.407	0.08	420.33	
L	1.391	0.05	448.78	
V	1.287	0.01	553.63	

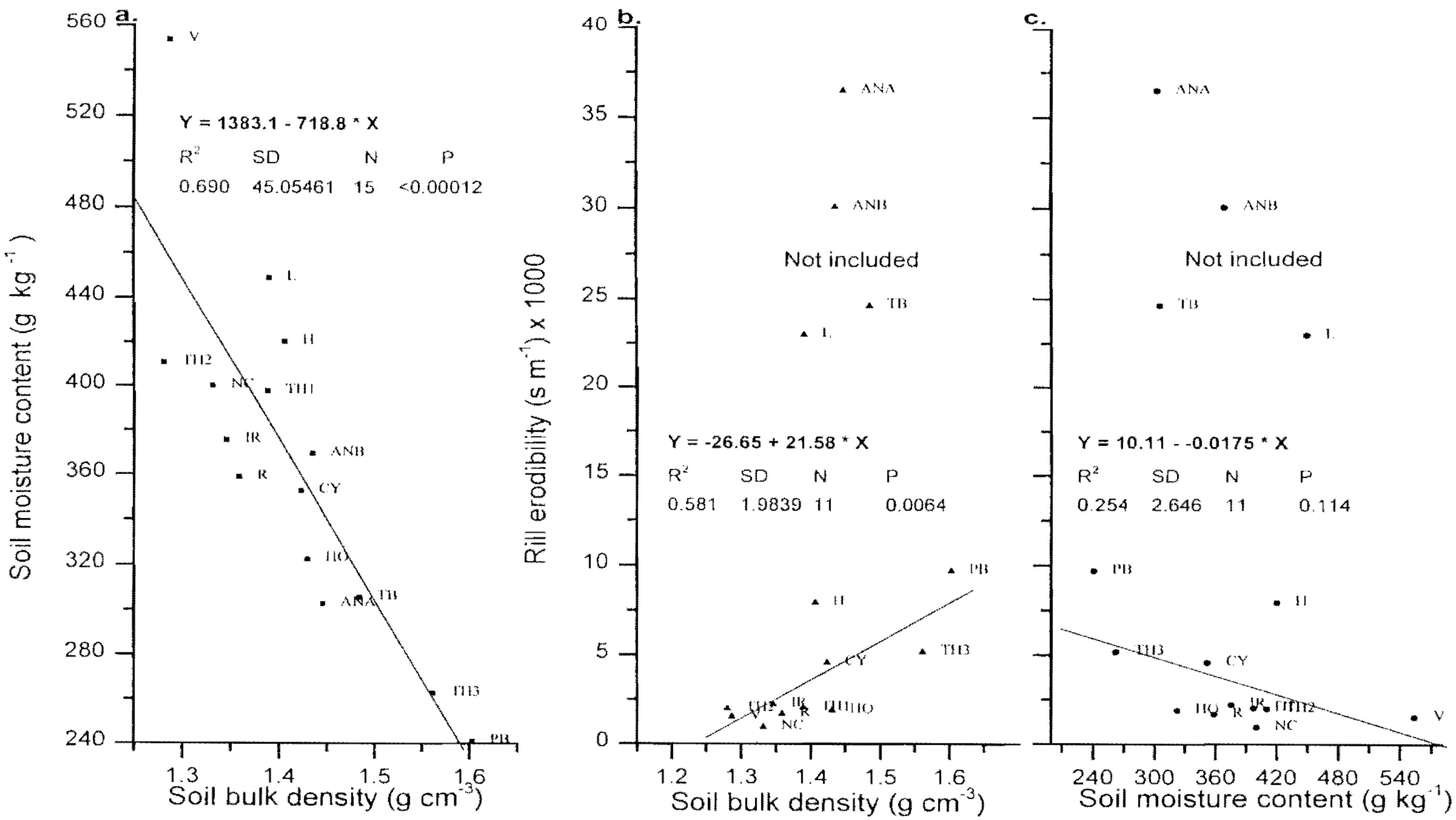

Fig. 3.17: Rill erodibility as a function of: soil bulk density (b), soil moisture content (c), and the relationship between moisture content and soil bulk density (a).

3.6.3.2 Effect of clay mineralogy on rill erosion

Mixed mineralogy characterizes the clay fraction of all the soils used in the study (Table 3.25).

The following trends were observed (Tables 3.25 and 3.27)

a. Soils of which the clay fraction is dominated by interstratified illite/smectite (I/Sm>70%) were associated with high erodibility.

b. With one exception (TB), all soils of which the clay fractions were dominated by kaolinite (TH3, HO, CY, IR and NC) were stable against rill erosion, irrespective of their texture and despite the fact that (except NC) their clay fractions all contained more than 10% smectite.

c. The clay fraction of the very stable NC soil is totally dominated by kaolinite (83%). Most importantly, this soil is stabilized by free ferric oxides which impregnate it, as was shown in thin section studies of this soil. It is well known that organic matter and free iron oxides are the most important agents which stabilize soil structure.

d. The two predominately illitic soils (PB and R) fell in different rill erodibility classes. PB had both a low OC content (7.4 g kg^{-1}) and moderate ESP (4.70%), whereas R had a relatively high OC content (23.2 g kg^{-1}) and lower ESP (2.25%). The latter soil is considerable less erodible

e. With the exception of the L soil, the soils of which the clay fraction were dominated by smectite (TH2, TH1, H and V) were contrary to expectations, stable against rill erosion. As in the case of the PB soil, the L soil is characterized by a relative low OC content (8,4 g.kg^{-1}) and moderate ESP (3.50%).

Table 3.27: Rill erodibility (K_r) and critical shear stress (τ_c) as affected by the salinity of the flowing water and soil sodicity.

soil		Electrolyte concentration of the flowing water											
		De-ionized water				10 mol (+)m^{-3}				50 mol (+) m^{-3}			
code	ESP	**Kr#**	STDs	**Tc##**	STDs	Kr	STDs	**Tc**	STDs	**Kr**	STDs	Tc	STDS
IR	0.48	**2.22**	0.97	**0.51**	0.41	3.01	0.87	022	0.13	**3.85**	1.50	0.16	0.18
TH3	2.28	**5.18**	2.52	**0.48**	0.16	3.77	1.45	**0.30**	0.15	**5.15**	1.54	0.15	0.09
AN A	3.75	**36.50**	14.84	**0.36**	0.27	45.26	4.75	**0.42**	0.02	**43.24**	7.62	0.50	0.03
PB	4.70	**9.70**	2.22	**0.27**	0.09	5.38	0.85	**0.12**	0.10	**4.44**	0.15	0.07	0.06
ANB	12.7	**30.09**	3.02	**0.00**	0.00	22.50	1.52	**0.03**	0.04	**18.81**	2.18	0.00	0.00

\# Rill erodibility (sm^{-1}) X 1000

\## Critical shear stress (Pa)

3.7 STERN, R (1990): EFFECTS OF SOIL PROPERTIES AND CHEMICAL AMELIORANTS ON SEAL FORMATION, RUNOFF AND EROSION

3.7.1 General background and objectives

As mentioned in the introduction, "soil erosion is the biggest environmental problem in South Africa" (Clem Sunter, The Star, 20th August 1990). Although soil lost from one field may be deposited on another, its transport into streams, rivers, or dams often results in quick silting up of dams and pollution by sediments containing nutrients and pesticides. This is of importance because many dams have become uneconomic through sediment accumulation.

Estimation and measurement of runoff erosion have been carried out in SA since 1929, since when runoff plots had been maintained by various organizations, either continuously or for brief periods. Other studies which estimated the rates of erosion using sediment accumulation rates of dams and sediment yield from a number of South African rivers have been carried out by Rooseboom and Harmse (1979) and Rooseboom and Mulke (1982) in the Orange Free State. It should be noted that these studies were conducted by agricultural engineers and pasture scientists. Soil scientists were not part of the teams. Consequently, analysis of mechanisms of runoff and erosion has received scant attention from South African scientists (Garland, 1982).

Seal formation at the soil surface is a very common phenomenon in arid and semi-arid regions all over the world, as well as in SA, and is the predominant factor in promoting water runoff and soil erosion.

The stability of surface soil aggregates under the impact of water drops falling as rain or applied as irrigation will determine surface crusting, water infiltration and soil erodibility.

Forces involved in stability studies (i) impact and shearing forces administered while preparing samples, (ii) abrasive and impact forces during sieving and (iii) forces involved in the entry of water into the aggregates (Kemper and Rosenae, 1986). Since dispersion and crust formation occur under wet conditions, this investigation was focused on disintegration occuring in the liquid phase.

In determining aggregate stability, known amounts of certain some size fractions of aggregates are commonly subjected to disintegration forces designed to simulate some important field phenomenon. There are different types of aggregate stability tests available. One of the tests is described by Emerson (1967) and Rengasamy et al. (1984). In these tests the dispersion of air dried aggregates in water is determined, thus, the measurement of spontaneous dispersion in the absence of any imposed external forces will reflect the behaviour of surface soils during the rainfall events when the soil surface is effective protected by mulch. Imposing different periods of shaking, however, may simulate the extent to which soil microaggregates disperse when bare soil is subjected to raindrop impact (Miller and Beharuddin, 1986).

The study was undertaken to:

1. Study the effect of clay mineral composition on seal formation, runoff and erosion.
2. Evaluate the effect of seal formation on runoff from representative South African soils, dominated by illite and/or kaolinite, under natural rainfall.

3. Study the effect of surface ameliorants (PG, polymer and mulch) on seal, runoff and erosion from bare plots.
4. Study the effect of surface ameliorants on yield and irrigation water use efficiency in irrigated plots.
5. Determine relationships between microaggregate dispersibility and soil susceptibility to runoff and erosion.
6. Develop a laboratory method for particle size analysis of soils from PG treated fields.

The present study focusess mainly on the findings of Stern in regard to 1, 2, 3 and 5

3.7.2 The effect of clay composition on seal formation, runoff and erosion

3.7.2.1 Runoff and erosion from stable and dispersive kaolinitic soils

Most of the investigation on seal formation and the effect of slope and surface amendments on runoff and erosion have been done on soils in which the dominant clay minerals were smectites and illites (Agassi *et al.*, 1981, 1985; Ben-Hur *et al.*, 1985; Ben-Hur *et al.*, 1989). These clay minerals are known to be more dispersive than kaolinitic clays (Goldberg and Glaubic 1987; Frenkel, Goertzen and Rhoades, 1978). The degree of dispersion of the clay fraction of soils is an important factor in determining seal formation, runoff and erosion (Agassi *et al.*, 1981; Miller and Baharuddin, 1986).

Stern studied effects of slope and PG application on seal formation, runoff and erosion from kaolinitic soils which vary in their response to rainfall and proposed mechanisms which account for these variabilities.

3.7.3 Materials and methods

Four kaolinitic soils from the former Transvaal and one from the former Natal were selected. The soils were classified according to the S.A. binomial system (MacVicar, De Villiers, Loxton, Verster, Lambrechts, Merryweather, Le Roux, van Rooyen and Harmse, 1977) and according to the USDA Soil Taxonomy (Soil Survey staff, 1975).

Selected physical and chemical properties are presented in Table 3.28. Cation exchange capacity (CEC) was determined by treatment of the sample with ammonium acetate, buffered at pH 7.0, followed by washing and distillation of ammonia. Organic carbon was determined using the Walkely-Black wet oxidation procedure. A measurement of free iron was obtained using the citratebicarbonate-dithionite (CBD) method. The clay mineralogical composition was determined by x-ray diffraction (XRD), using a C_oK_α

3.7.4. Results and discussion

3.7.4.1 Infiltration rate and runoff

Infiltration curves of the untreated and PG treated Msinga (D) soil from Irene at 5% and 30% slopes are presented in Fig. 3.18. Stern (1990) found that the rapid drop in the IR of the control soil is an indication that the soil is very susceptible to surface sealing. The final IR of the Msinga (D) soil dropped to 2,4 mm h^{-1} at 30 mm of rain (Table 3.29 and Fig. 3.18), which is very low for a kaolinitic soils and is much lower than final values of 8 to 13 mmh^{-1} obtained by Miller (1987), Levy and van der Watt (1988), and Smith (1990) for kaolinitic soils. This value is unexpectedly in fact similar to that obtained by Kazman, Shainberg and Gal (1983) and Ben-Hur et al. (1985) for smectitic soils. Increasing the slope gradient from 5% to 30% reduced the ponding time, but significantly increased the final IR from 2,4 to 5,2 mmh^{-1} respectively and decreased the cumulative runoff from 70,1 to 65,0 mm, respectively (Table 3.29). Stern indicated that decrease in ponding time is probably due to the lower surface water storage on steep slopes for the same surface roughness. The increase in final IR is associated with an increase in soil erosion (Table 3.30). As the slope increased, removal of the seal by erosion took place (Poesen, 1987; Warrington *et al.*, 1989) and the final IR increased.

Stern found that PG application was beneficial in increasing the IR of the Msinga (D) soil (Fig. 3.18). PG treatment increased the final IR of this soil from 2,4 mm/h in the control to 9,8 mmh^{-1} and reduced the runoff percentage in an 80 mm rain from 87,6% in the control to 53,0% in the PG treated samples (Fig. 3.18 and Table 3.29). Slope steepness had only a small effect on infiltration curves of the PG treated Msinga (D) soil. The ponding time at 30% slope was shorter due to lower surface storage capacity, but the final IR and cumulative runoff were not affected significantly by the slope increase. By diminishing chemical dispersion at the soil surface, the PG treatment slowed down seal formation and reduced removal of the seal by erosion (Table 3.29).

Infiltration curves of the untreated and PG treated Shorrocks (S) soil from Potchestroom at 5% and 30% slopes are indicated in Fig. 3.19. The moderate decline in the infiltration curve of the untreated Shorrocks (S) soil differed considerably from the sharp drop in the case of the Msinga (D) soil (Fig. 3.18. The final IR of the Shorrocks (S) soil (15,1 mmh^{-1}) was obtained only at 70 mm of cumulative rain. At a slope gradient of 30% the IR curve was slightly lower and the final IR was similar to that at 5% slope. The Shorrocks (S) soil forms stable aggregates which did not disperse when exposed to raindrop impact. Hence, seal formation is limited, infiltration is high and runoff and erosion (Table 3.30) on the steep slope are small and insufficient for removing the surface seal and increasing rain infiltration rate.

Stern also found that PG was beneficial in improving rain infiltration in the Shorrocks (S) soil. The final IR of the PG amended Shorrocks (S) soil was maintained at 40 mmh^{-1} for the 5% slope. Presumably, when clay dispersion is prevented by the dissolved gypsum, the impact of raindrops could not disperse the surface aggregates. Increasing the slope gradient increased the velocity of sheet flow and enhanced soil and PG particle transport downslope, and reduced the PG efficiency in slowing seal formation. At the 30% slope, the final IR of the PG treated soil was maintained at 20,7 mmh^{-1} compared with 15,0 mmh^{-1} for the untreated soil (Table 3.29) and consequently run off from the PG treated soil was 70% than of the untreated soil.

The dominant clay type in both soils was kaolinite (Table 3.28). Clay contents of Shorrocks and Msinga soils were 189 and 319 g kg^{-1}, both in the range where soils are most susceptible to sealing (Ben Hur et al., 1985). The ESP of both soils was low and quite similar (1.4 and 1.5). The pH of the two soils was below 6, and kaolinitic clay at this low pH is not dispersive (Schofield & Sampson, 1954), even in distilled water. A close look at the XRD of the two soils may explain the difference in their susceptibility to sealing. The clay fraction of the Shorrocks (S) soil is predominantly kaolinite (Fig. 3.20) and is not contaminated by small amounts of swelling minerals. Also the amount of illite in this soil is low (Fig. 3.20). The low CEC of 18,9 $cmol_c\,kg^{-1}$ clay supports the observation that Shorrocks (S) soil does not contain smectitic clays. Conversely the Msinga (D) soil contains interstratified material (mostly smectites) and a moderate amount of illite (Fig. 3.20) which accounts for the relatively high CEC of 38.0 $cmol_c\,kg^{-1}$ of clay. Stern suggested that the smectite impurities are responsible for the dramatic differences in the susceptibility of these two kaolinitic soils to seal formation. Addition of small amounts of montmorillonite to kaolinitic soils evidently promotes the dispersion of the kaolinitic flakes. This phenomenon has been ascribed to the breakup of the edge-to-face particle association of kaolinite structure by the adsorption of negatively charged montmorrilonite particles (faces) on the positively charged kaolinite edges (Schofield and Sampson, 1954). A similar conclusion was made by Frekel *et al.* (1978), who observed a reduction in hydraulic conductivity of an acid kaolinitic soil when a small amount of montmorillonitic clay (2% by mass of soil) was mixed with the soil.

The hypothesis that small amounts of smectites in kaolinitic soils increases the dispersivity of the soils and their susceptibility to sealing and erosion needed more supportive data. Thus, three other kaolinitic soils which seemed to produce different amounts of runoff and erosion in the field were sampled and their IR curves were determined in the laboratory. Two of the three soils (Shorrocks (D) and Hutton (D) were found to be dispersive and susceptible to sealing. The final IR of these soils was 9.5 and 4.2 mmh^{-1} and runoff at 5% slope was 62,7 and 76,5% (Table 3.29) respectively. The fifth soil (Msinga soil (Msinga (S)) from Dundee was found to be stable with final IR values similar to those of the stable Shorrocks (S) soil (Table 3.29).

Stern found that when a seal with low IR is formed and high erodibility prevails, as is the case of Msinga (D) and Hutton (D) soils (Table 3.30), the final IR increased with increased slope gradient which is contrary to normal expectations. When less erosive conditions prevail, as is the case in stable soils or when applying PG to the five soils, there is no influence of slope on IR and cumulative runoff. A negative relationship between slope gradient and IR was obtained only in the PG treated Shorrocks (S) soil.

Stern concluded by indicating that in unstable soils which tend to seal, infiltration increases with an increase in slope gradient because the seal is being removed by erosion.

Comparison of the infiltration and runoff data (Table 3.29) and the soil loss data (Table 3.30) of the three additional kaolinitic soils support the hypothesis that kaolinitic soils which contain low concentrations of smectites (Table 3.28) are dispersive. Conversely soils with low CEC of the clay fraction and no interstrafied clay are more stable and less dispersive.

3.28 Some physical, chemical and mineralogical properties of the kaolinitic soils

Soil					Sand[C]		CEC[A]	ESP	organic carbon	Fe	pH (H_2O)	Clay[B] minerals
Series	Location	Taxonomy	Clay	Silt	V.fine	Coarse						
			g kg^{-1}				Cmol$_c$ kg^{-1}	%	g kg^{-1}			
Msinga (D)	Irene	Paleudalf	319	258	66	357	38.0	1,4	17	55	5,6	K(5), I(2), IS(1)
Hutton (D)	Roodeplaat	Rhodudalf	300	142	148	410	38.4	1,3	9	20	6,0	K(5), I(2), Is(2)
Shorrocks (D)	Potchefstroom	Rhodustalf	158	121	187	534	24.7	3,1	5	18	5,0	K(5), I(1), Is(1)
Msinga (S)	Dundee	Paleudalf	238	188	145	429	25.2	2,5	20	42	6,1	K(5), I(2)
Shorrocks (S)	Potchefstroom	Rhodustalf	189	106	156	549	18.9	1,5	14	17	5,4	K(5), I(1)

[A] CEC = Cation exchange capacity cmol$_c$ per kg clay following the subtraction of OM contribution
[B] K = kaolinite , I = illite, Is = interstratified materials (swelling minerals). (1) = Weak… (5) = Strong
[C] V.Fine Sand (<0,10mm); Coarse Sand (<2,00mm)

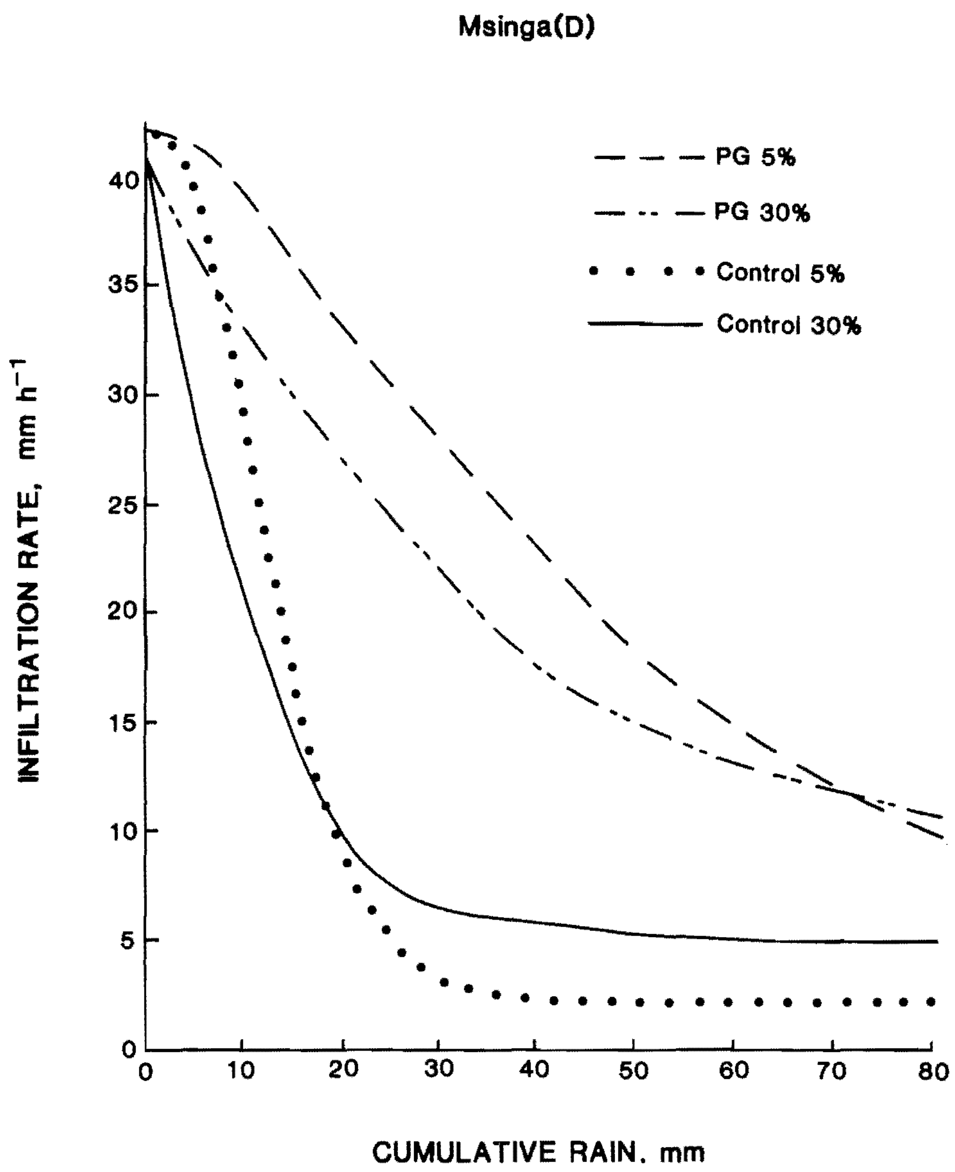

FIG. 3.18 Infiltration curves of untreated and PG treated Msinga (D) soil at 5% and 39% slope gradients

TABLE 3.29 Final infiltration rates and cumulative runoff of untreated and PG treated kaolinitic soils at 5 and 30% slopes.

Soil	Slope	Control			Phosphogypsum			Stat.
		FIR[A]	Runoff		FIR	Runoff		
	%	mm h^{-1}	mm[B]	%[C]	mm h^{-1}	mm	%	Sign.
Msinga (D)	5	2,4	70,1	87,6*	9,8	42,4	53,0	+
Msinga (D)	30	5,2	65,0	81,3	10,8	44,6	55,7	+
Hutton (D)	5	4,2*	61,2	76,5	16,2	36,5	45,6	+
Hutton (D)	30	6,1	58,6	73,2	14,4	37,4	46,7	+
Shorrocks (D)	5	9,5	50,2	62,7	14,8	38,0	47,5	+
Shorrocks (D)	30	8,7	54,2	68,0	15,2	39,9	49,0	+
Msinga (S)	5	14,5	32,6	40,7	22,5	26,4	33,0	+
Msinga (S)	30	15,6	32,2	40,2	21,8	28,0	35,0	+
Shorrocks (S)	5	15,1	35,6	44,5	40,0*	6,8	8,5*	+
Shorrocks (S)	30	15,0	37,9	47,3	20,7	26,3	32,9	+

[A] FIR – IR after 80 mm rain

[B] Cumulative runoff from 80 mm rain

[C] % runoff from 80 mm rain

* Significant differences in slope effect (P=0,05)

\+ Statistical significance between final IR's and cumulative runoff of control and PG treatments for the same soil and level of slope (P=0,05)

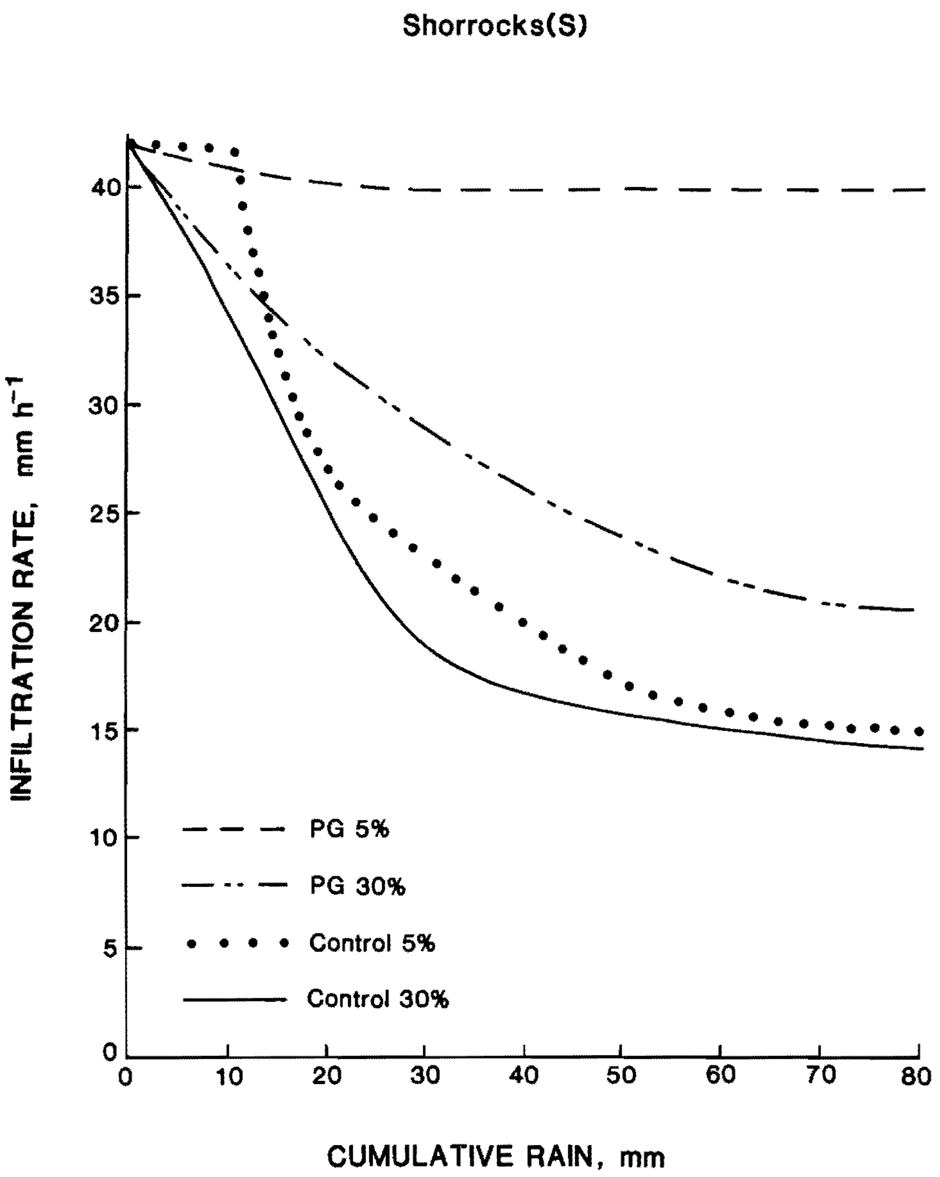

FIG. 3.21 Infiltration curves of untreated and PG treated Shorrocks (D) soil at 5% and 30% slope gradients

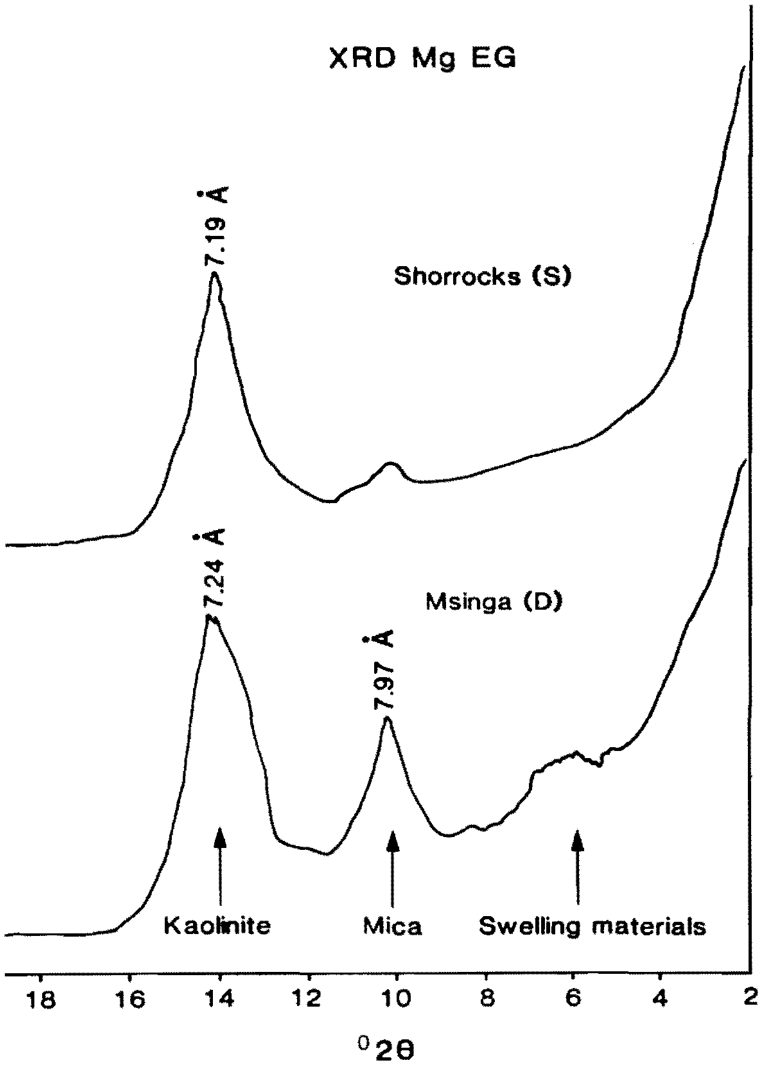

FIG. 3.22 X-ray powder diffraction patterns (CoK_{α}) of Mg saturated Shorrocks (S) and Msinga (D) soils after ethylene glycol treatment

3.7.4.2 Soil Loss

Stern found that soil losses varied markedly between the stable and unstable soils and were affected substantially by both slope gradient and PG amendment (Table 3.30). Soil losses from the untreated stable soils (Shorrocks (S) and Msinga (S) at 5% slope were significantly lower than soil losses from the dispersive ones (Table 3.30). Stern indicated that low soil losses from the stable soils are associated with high IRs and low runoff (Table 3.29). Conversely, a seal with low hydraulic conductivity was formed on the dispersive soils and the high runoff combined with the dispersive nature of the soils caused high soil losses. Increasing slope dramatically increased soil losses from the untreated dispersive soils resulting in slope ratios (Sr = ratio between soil losses at 30% and 5% slope) of 5,2 and 4,5 for the Msinga (D) and Hutton (D) soils respectively. Conversely, only moderate increases in soil loss with slope were obtained for the more stable soils, giving Sr values of 1,1 and 2,7 for the Msinga (S) and Shorrocks (S) soils respectively (Table 3.30). The high Sr values of the dispersive soils explain the increase in IR of these soils with increase in slope (Table 3.29). The intensive erosion on the steep slopes removes the seal that limits rain infiltration.

Stern also explained the effect of slope steepness on soil losses from the stable and dispersive soils (Table 3.30). Stern found that the Shorrocks (S) and Msinga (S) soils form stable aggregates which slake, but do not disperse, during the rainstorm. No clay migration occurred and no clay particles were detected in the infiltrating water. Thus, infiltration rate is high and runoff is low. Since only aggregate slaking take place, the sediment from stable soils are in the form of microaggregates (Meyer et al., 1980) which have low transportability by runoff. Hence, the stable soils are less susceptible to erosion even at steep slope. The Msinga (D), Hutton (D) and to a lesser extent the Shorrocks (D) soils are substantially more dispersive due to the presence of smectite impurities in the clay fraction (Table 3.28 and Figure 3.22). Clay dispersion, being enhanced by raindrop impact, causes low IR (Table 3.29) and high erosivity (Table 3.30). Stern found that high erosivity of the overland flow at 30% slope was due to two complementary mechanisms: (i) the high runoff and flow velocity lifts more particles and "entrains" them downslope, and (ii) the dispersed particles are easily detached by the raindrops and transported downslope by the runoff flow.

Stern found that PG amendment significantly reduced soil losses from all soils at both 5% and 30% slope gradients (Table 3.30). The beneficial effect of PG on soil loss depends on the inherent properties of each soil. The most pronounced effect of PG was obtained in the highly erodable soils, especially at steep slopes (Figure 3.23). Soils losses from the Msinga (D) and Hutton (D) soils at 30% slope were reduced by PG treatment from 6,77 to 2,12 and from 7,15 to 1,18 t ha^{-1} respectively. In the less erodable Shorrocks (S) and Msinga (S) soils and at a gentle slopes the effect of PG was less dramatic but still very important (Table 3.29).

TABLE 3.30 Effect of PG and slope gradient on soil loss (t ha^{-1}) from kaolinitic soils (48 mm rain)

Soils	Control slope 5%	Control slope 30%	Control slope ratio[A]	PG slope 5%	PG slope 30%	Treatment ratio[B] slope 5%	Treatment ratio[B] slope 30%	K[C]	Erodibility order[E] K	Erodibility order[E] Data
Msinga (D)	1,31bD	6,77c	5,2	0,47b	2,12c	2,8	3,2	0,10	1	4
Hutton (D)	1,60b	7,15c	4,5	0,58b	1,18b	2,7	6,1	0,10	1	5
Shorrocks (D)	1,18b	3,39b	2,9	0,84c	1,16b	1,4	2,9	0,16	3	3
Msinga (S)	0,38a	0,42a	1,1	0,22a	0,25a	1,7	1,7	0,11	2	1
Shorrocks (S)	0,32a	0,84a	2,7	0,14a	0,45a	2,3	1,9	0,10	1	2

[A] Slope ratio – The ratio between soil losses at 30% slope and that at 5% slope for the control

[B] Ratio between soil losses of untreated to PG treated soils at 5% and 30% slopes respectively

[C] K – Soil erodibility factor. Values obtained from the nomograph (Wischmeier and Smith, 1978)

D Different letters indicate significant differences between soils of each treatment (P=0,05)

[E] Erodibility order (1 – least erodable, 5-most erosdible). K- according to calculation in C. Data – according to soil losses of untreated soils at 30% slope.

3.7.4.3 Particle concentration in the runoff

The changes in sediment concentration in runoff from the Msinga (D) and Shorrocks (S) soils at two slope angles during a 48 mm simulated rainstorm are demonstrated in Figure 3.24. The sediment concentration of the Msinga (D) soil at 30% slope rapidly increased over the first 20 to 30 mm, followed by a slight decrease to steady state concentration at 40 mm cumulative rain. At 40 mm both runoff volume and amount of sediment transported become constant and steady state particle concentration was maintained. Stern found that changes in sediment concentration in the Msinga (D) soil at 5% slope are small (Figure 3.24). Similarly in the Shorrocks (S) soil, the concentration of sediment in runoff is low, with minor fluctuations during storm events.

Stern found that the concentration of sediment in runoff varies during a rainstorm. However, in order to compare soils and treatments (Slope and PG) the average concentration of sediment in runoff was calculated (Table 3.31). The sediment concentration in runoff of the untreated unstable soils at 30% slope was found to be between 2 to 4 times that at 5% slope. Stern established that the increase in sediment concentration was responsible for most of the increase in soil losses. The sediment concentration of the PG amended soils, except the Msinga (D), increased more moderately with slope increase and was associated with a moderate increase in soil loss with slope increase (Table 3.30).

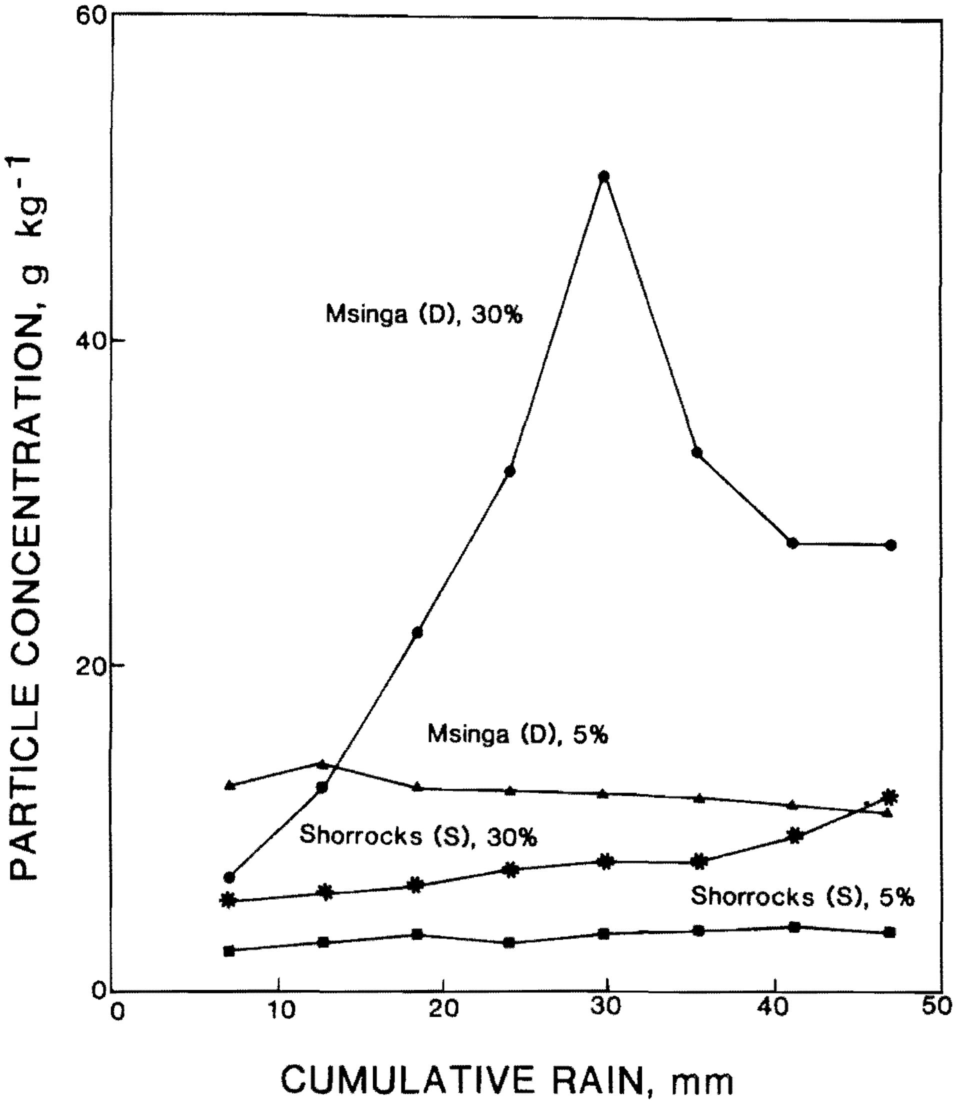

FIG. 3.24 Particle concentration in the runoff of Msinga (D) and Shorrocks soils at 5% and 30% slope gradients

TABLE 3.31 Particle concentration in runoff as affected by slope and PG teament

Soil	Particle concentration in runoff			
	Control		PG	
	slope			
	5%	30%	5%	30%
	g kg^{-1}			
Msinga (D)	11,5	29,6	4,5	14,9
Hutton (D)	12,1	51,8	4,8	10,6
Shorrocks (S)	12,0	22,7	13,2	15,0
Msinga (S)	6,2	7,6	4,5	4,8
Shorrocks (S)	3,8	8,5	3,0	6,4

3.7.5 Conclusions

Stern (1990) found that the soils which contain pure kaolinite form stable aggregates, maintain high IR and have low erosion. However, kaolinitic soils which contain small amounts of smectites are dispersive even at pH below 6. They form seals with low hydraulic conductivity and the high runoff, due to low IR, causes high soil losses. Soil loss from the dispersive soils increased markedly with an increase in slope gradient. The intensive erosion on the steep slope removes the seal which limits rain infiltration. Thus the final IR of the dispersive soils increases with increased slope gradient. PG amendment reduces runoff and soil loss from all soils.

3.7.6 RUNOFF AND EROSION FROM ILLITIC SOILS

3.7.6.1 Introduction

Most studies on soil crust were carried out with soils containing montmorrilonite as the dominant clay mineral (Kazman *et al.*, 1983; Ben Hur *et al.*, 1985). Miller (1987) studied the phenomenon of crusting in some kaolinitic soils from Georgia. His results showed that the soils formed crusts when exposed to rain, but maintained relatively high final infiltration rates (5-10 mm h^{-1}).

The results presented by Arora and Coleman (1979) imply that soils with illitic clay are more dispersive than soils dominated by montmorillonite clays, especially at low levels of ESP.

Stern's objective in this study was to investigate the susceptibility of illitic soils to seal formation, runoff and erosion.

3.7.6.2 Materials and methods

Three illitic soils were selected, one from Riviersonderend (Rosehill soil) in the South Cape Rûens area, one from Aliwal North (Jozini soil) in the Northern Cape and a sodic soil from Piketberg (Trevanian soil) in the Swartland area. Some properties of the soils are presented in Table 3.32

Fog rain was applied to the Trevanian soil, using the rain simulator. The fog rain was applied via emitters. The diameter of the drops was < 0,1 mm, with a maximum drop velocity of 0,1 ms^{-1}, kinetic energy of < 0,001kj m^{-3}, producing fog intensity similar to high energy rainfall (42 mm h^{-1}).

TABLE 3.32 Some physical, chemical and mineralogical properties of the illitic soils studied

SA Soil series	Jozini	Rosehill	Trevanian
Location	Aliwal North	Riviersonderend	Piketberg
Soil taxonomy	Haplustalf	haplargid	Haplargid
Clay g kg^{-1}	166		
Silt -"-	78	233	189
V F Sand -"-	651	317	323
Sand -"-	105	165	88
CEC[B]cmol (+) kg^{-1}	35,4	285	400
ESP %	1,6	43,8	20,8
C g kg^{-1}	11	2,5	8,4
Fe g kg^{-1}	7	20	9
pH (H_2O)	6,7	5,7	6,8
Clay	I(5), K(1)	I(4), K(3)	I(5),K(4)
Minerals[C]	Is(1)	Cl(2), Is(1)	Cl(2)

A Very fine sand <0,10 mm

[B] CEC=Cation exchange capacity $cmol_c$ per kg clay following the subtraction of organic carbon contribution

[C] k = kaolinite, I = illite, Cl = Chlorite, Is = interstratified material (swelling material) (1) = weak… (5) = strong

3.7.6.3 Interpretation of results and discussion by Stern

3.7.6.3.1 Infiltraton rates of Jozini and Rosehill soils

The infiltration rates of the untreated Jozini and Rosehill soils declined rapidly, to reach final infiltration values after application of less than 25 mm of simulated rain (Figure 3.25 and Figure 3.26). The rapid decline to low values is associated with unstable aggregates at the soil surface and clay dispersion. At 5% slope the final IRs of both soils stabilized at 4,2 mm h^{-1} (Table 3.33). Stern also found that cumulative runoff from the Rosehill soil at 5% slope was somewhat higher than from the Jozini soil (Table 3.33). This was due to the a lower initial IR on the strongly crusting Rosehill soil, which is caused by the low saturated hydraulic conductivity with application of distilled water.

Stern found that the slope gradient affected IR's of the untreated Jozini and Rosehill soils in opposite ways (Table 3.33). An increase in slope gradient from 5 to 30% increased the final IR of the Jozini soil from 4,2 to 9,0 mm h^{-1}, but decreased that of the Rosehill soil from 4,2 to 3,0 mm h^{-1}. The positive effect of increased slope gradient on final IR of the Jozini soil is associated with high erosivity (Table 3.34) and rapid removal of the surface, probably dispersed, layer by the overland flow at steep slope. In contrast the Rosehill soil developed a highly resistant crust at the soil surface which could not be removed by flow on the steep slope, as was indicated by the relatively small amounts of soil loss recorded from this soil at both moderate and steep slopes (Table 3.34).

PG beneficially affected water infiltration into both the Jozini and Rosehill soils (Table 3.33). Stern found that the PG moderated the rate of decline in IR and brought about a large increase in final IRs, leading to a significant reduction in cumulative runoff. The increase in final IR was particularly large for the Rosehill soil at both 5 and 30% slope and Jozini soil at 5% slope. In the Jozini soil at 30% slope the difference between the untreated and PG treated soil was smaller due to the relatively high final IR of the untreated soil. The latter was caused by continuous removal of the surface layer and consequent prevention of the formation of a dense seal.

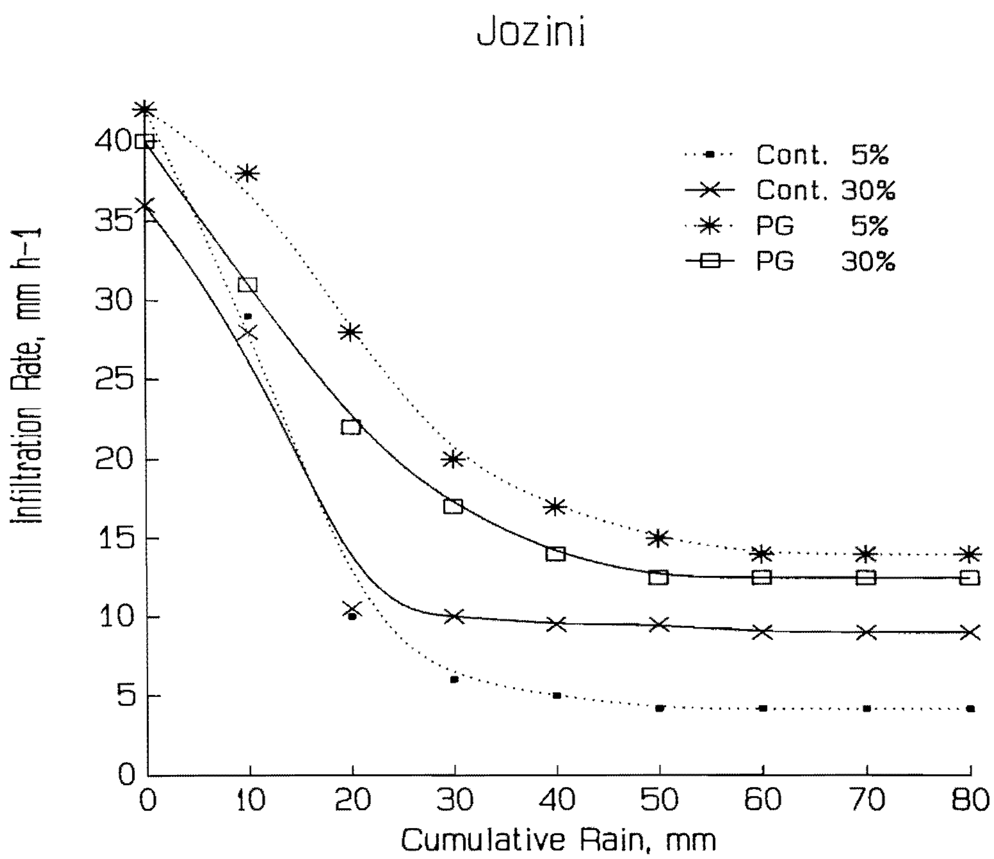

FIG. 3.25 Effect of PG and slope on the IR of the Jozini soil

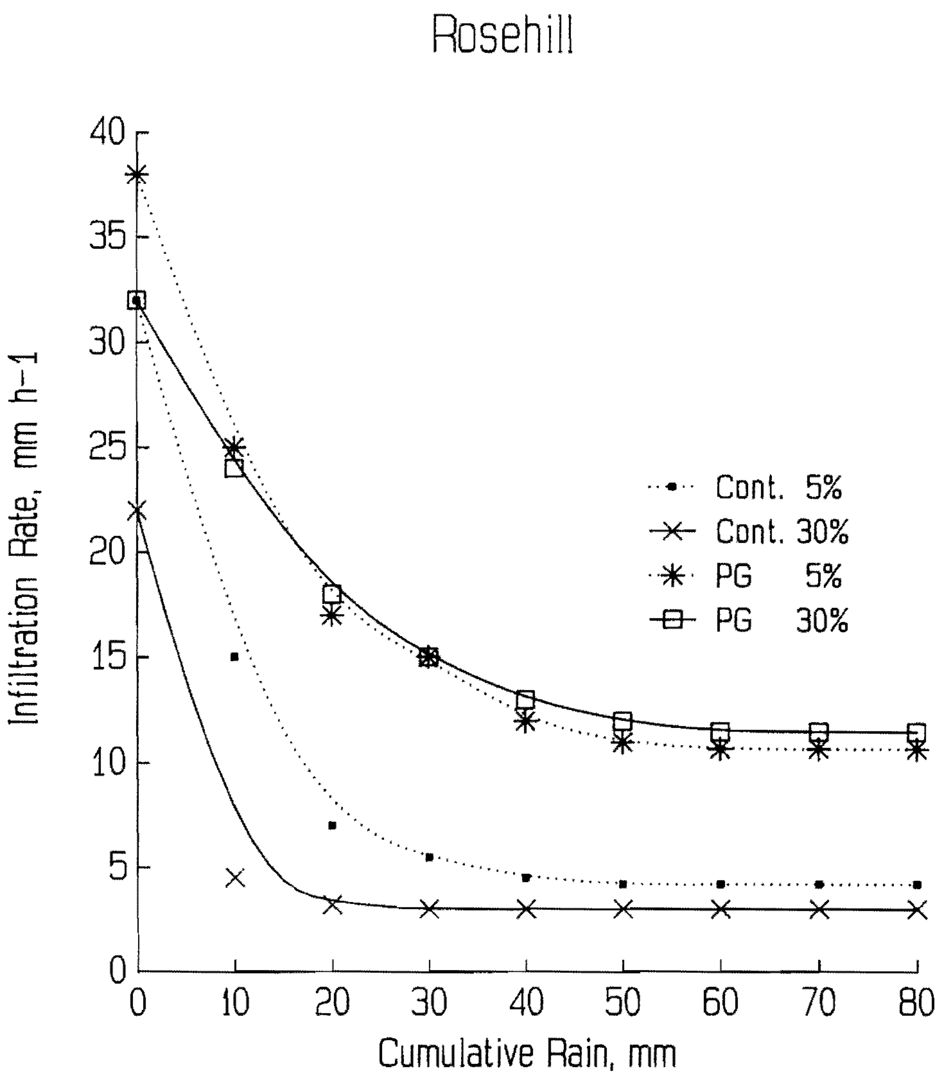

FIG. 3.26 Effect of PG and slope on the IR of the Rosehill soil

TABLE 3.33 Infiltration rates and cumulative runoff of untreated and PG treated soils at different slope gradients

Soil	Slope %	Control			Phosphogypsum			
		FIR[A] mm h^{-1}	Cum. mm[B]	Runoff %[C]	FIR mm h^{-1}	Cum. mm	Runoff %	Stat. Sign.
Jozini	5	4,2	58,5	73,1*	14,2	40,1	50,0	+
Jozini	30	9,0	50,4	63,0	12,5	44,4	55,6	+
Rosehill	5	4,2*	64,8	81,0*	10,7	52,3	65,3	+
Rosehill	30	3,0	71,6	89,5	11,5	54,8	68,5	+
Trevanian	5	1,8	73,8	92,7	2,0	73,2	91,5	+
Trevanian	30	2,1	74,6	93,2	2,0	74,0	92,5	+

[A] FIR – IR after 80 mm rain

[B] Cumulative runoff from 80 mm rain

[C] % runoff from 80 mm rain

* Significant differences in slope effect (P = 0,05)

+ Statistical significance between IR's and cumulative runoff of control and PG treatments for the same soil and level of slope (P=0,05)

TABLE 3.34 Effect of slope gradient and PG on soil loss from the illitic soils

Soils	Soil loss t ha^{-1} (48 mm rain) Untreated ---slope------- 5%	30%	Slope ratio[A]	PG ---------slope------- 5%	30%	Treatment ratio[B] 5%	30%	K[C]
Jozini	1,62	9,28	5,7	0,55	1,85	2,9	5,0	0,36
Rosehill	0,84	2,12	2,5	0,51	1,34	1,6	1,6	0,27

[A] Slope ratio – the rate between soil loss of 30% and 5% slope for the untreated soils

[B] Ratio between soil losses of untreated to PG treated soils upon 5% and 30% slope

[C] K – Soil erodibility factor. Values obtained from the nomograph (Wischmeier and Smith, 1978)

3.7.6.3.2 Soil loss from Jozini and Rosehill soils

Stern found that soil loss differed widely between the Jozini and Rosehill soils and was also affected substantially by slope gradient and PG amendment (Table 3.34). At 5% slope soil loss from the Rosehill soil was significantly lower than from the Jozini soil, being 0,84 and 1,62 t ha^{-1} respectively. This low soil loss was unexpected since runoff from it was higher than from Jozini soil (Table 3.33). Stern found that a crust with low hydraulic conductivity was rapidly formed on the Rosehill soil, causing high total runoff.

The extremely severe crusting nature of the Rosehill soil relative to the Jozini soil was somewhat unexpected in view of the fact that the clay fraction of the Rosehill soil contains significant quantities of kaolinite. On the other hand, the presence of significant quantities of chlorite, which have been implicated in aggravating dispersion and reduction of hydraulic conductivities of soils (Nel, 1989), and the high silt probably contributed greatly to this phenomenon.

Stern indicated that increasing the slope gradient dramatically increased soil loss from the Jozini soil but only moderately affected soil loss from the Rosehill soil. Slope ratios were 5,7 and 2,5 for the Jozini and Rosehill soils respectively. This difference further supports the assumption that seal formation plays an important role in controlling soil loss. The role of seal formation in dominating soil loss is further demonstrated by the changes in particle concentration with rain depth for the Jozini and Rosehill soils (Figure 3.27). The sediment concentration of the Jozini soil at 30% slope rapidly increased over the first 15 to 25 mm followed by a slight decrease to level out at 40 mm cumulative rain. As the loose particles were removed and a compacted seal developed, the concentration of sediment in runoff decreased. The pattern of the Jozini soil curves (Figure 3.27) is comparable to that of the Msinga (D) soil (Figure 3.24). Stern found that the increase in soil loss curve at 30% is faster and reaches a higher maximum value than the Msinga (D) soil curve. The particle concentration of the Rosehill soil was nearly constant during the rain storm, both at 5 and 30% slope (Figure 3.27). A low permeability seal was rapidly developed at the beginning of the rainstorm as was reflected by the rapid decrease in infiltration rate (Figure 3.26).

Once formed, the seal is rigid and is not susceptible to particle detachment or entrainment by the runoff. After only 10 mm of cumulative rain the Rosehill soil maintained the runoff and particle concentration values which remained constant during the rainfall event. The low particle concentration during the rainstorm is associated with the relatively low soil loss at 30% slope (Table 3.34).

The high erodibility of the Jozini soil could be derived from its clay mineralogy. Since its clay fraction is dominated by illite it could be expected to be highly susceptible to crusting (Levy and van der Watt, 1988). Stern found that the presence of small amounts of smectites (which in the case of kaolinite aggravated dispersion) could also enhance seal formation. The shear strength of the crust in this soil is evidently low and could not prevent the intensive entrainment of sediment and the development of rills by the high flow velocity on steep slope.

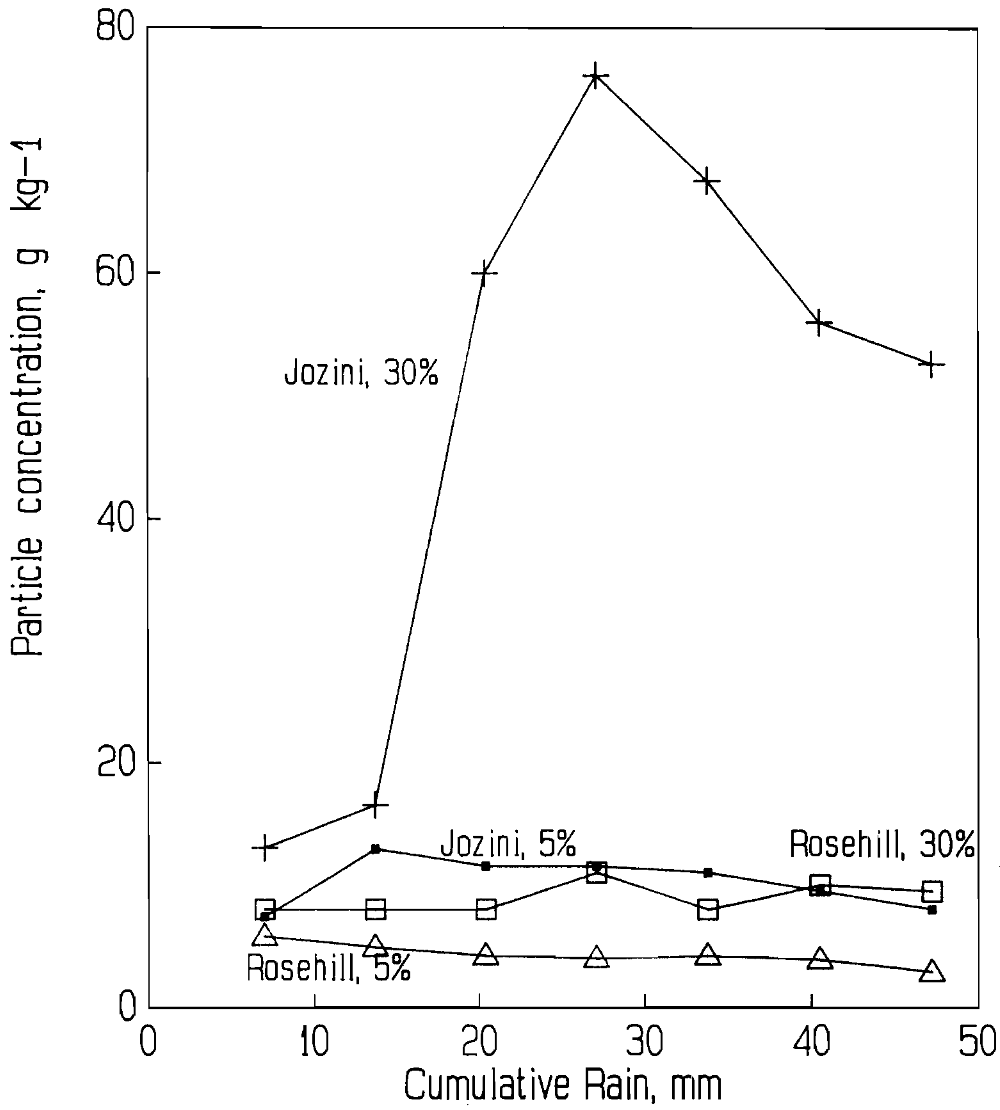

FIG. 3.27 Particle concentration in the runoff of Jozini and Rosehill Soils at 5% and 30% slope gradients

3.7.6.4 Trevanian soil from Piketberg

Infiltration curves of the Trevanian soil are presented in Figure 3.28. The initial infiltration rates of the different treatments when rain with energy was applied were below 15 mm h^{-1} and the IR's dropped rapidly to reach constant rates of about 2 mm h^{-1}. The slope effect was detectable at the beginning of the storm when due to lower surface capacity, the initial infiltration rates of the steeper slope treatments were lower. Nevertheless, after 20 to 30 mm of rain all the curves appeared similar.

Stern found that a severe deconsolidation of the soil structure, when subjected to rainfall, took place virtually immediately. The raindrop energy predominantly governed the dispersion process which promoted the formation of a very low permeability seal at the soil surface.

Applying fog rain (low energy rain) maintained the IR of the soil at 14,2 mm h^{-1} during the storm (Figure 3.28). This value is low when compared with the other soils studied which maintained the IR at 42 mm h^{-1} (the application rate) under fog rain i.e. their saturated hydraulic conductivity was above 42 mm h^{-1}. Stern found that when fog rain is applied no seal is formed at the soil surface and evidently the saturated hydraulic conductivity of the profile is determined. PG application beneficially affected the IR of the Trevanian soil when subjected to fog rain, maintaining a value of 21 mm h^{-1}. The effect of PG was more pronounced when using fog rain, since the disruptive effect of the raindrops, which represent the physical effect, was less dominant.

It is important to note that the Trevanian soil is sodic and therefore does not fit the main objective of the study, which were to study the nature of seals in non-sodic soils. However, this soil represents considerable tracts of arable land in the Swartland area, one of the biggest wheat production regions in SA, which suffers from high runoff.

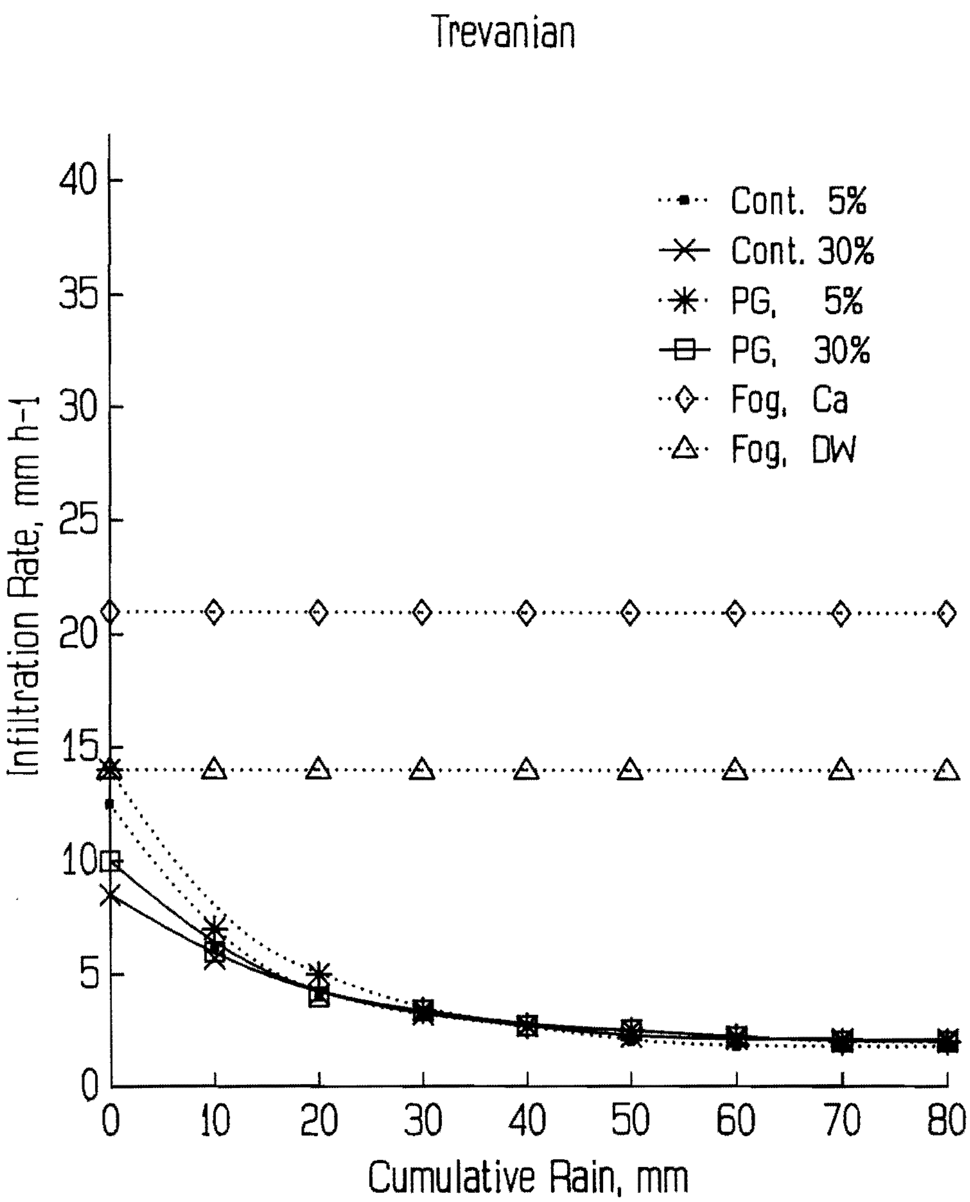

FIG. 3.28 Effect of drop energy, water quality and slope gradient on the IR of the Travanian soil

3.7.6.5 Conclusions

Stern found that soil structure and clay mineralogy largely determine the fundamental physical and chemical properties of the soil surface and therefore have large effects on the degree of sealing and the nature of the seal. From the experimental results it was evident that, depending on soil texture, soil clay mineralogy and slope gradient, the intensity of seal development and the nature of seals imposes different relationships between IR and soil erosion.

3.7.7 FIELD STUDIES WITH RUNOFF PLOTS UNDER NATURAL RAINFALL

3.7.7.1 Introduction

Runoff is particularly critical in the arid and semi-arid zones where it influences the amount of plant available water during the growing season. When rainfall rate exceeds the soil infiltration rate, water will flow over the soil, transporting soil particles downslope.

Prevention of the physio-chemical process leading to seal formation can be achieved by one or a combination of the following methods: (i) protecting the soil surface against the impact of raindrops by mulch; (ii) introducing electrolytes at the soil surface to control the chemical dispersion of clay particles, e.g. phosphogypsum (PG); (iii) stabilization of the aggregates at the soil surface by soil conditioners such as polyacrylamide (PAM).

Preventing the impact of raindrops by mulching maintains a relatively high hydraulic conductivity at the soil surface (Aggassi et al., 1985). Plant residue cover was used as treatment in Stern's research in order to distinguish between the effects of seal formation and the hydraulic properties of the profile on water infiltration into the soil. PG was included in the field studies and its efficiency under natural rainfall was studied.

Synthetic chemical polymers improve soil physical properties by stabilizing soil aggregates. When properly applied, polymers impart desirable stability to soil materials that are structurally unstable in their natural condition. Rainfall simulator studies conducted by Shainberg *et al.*, (1990) on soils inclined to seal, showed that the application of PAM together with PG increased rain infiltration by three to four fold. The combined application of a polymer and gypsum had a synergistic effect in improving infiltration (Shainberg *et al.*, 1990).

Stern studied the role of seal formation in enhancing runoff under natural rainfall conditions. He obtained information from field studies on (i) runoff from representative soils, throughout the country, which are susceptible to surface sealing, (ii) the effect of PG and PAM application and mulching on runoff from non-sodic kaolinitic and illitic soil and (iii) association of results obtained in the field with those of the laboratory studies.

3.7.7.2 Materials and methods

3.7.7.2.1 Soil and plot preparation

Soil were selected from four areas where crusting, low rain intake, runoff and erosion problems were encountered. The soils contained low organic carbon and varied in clay mineralogy. Some properties of the soils from Irene and Potchefstroom are presented in Table 3.28 (Msinga (D) and Shorrocks (D) respectively) and the soil from Aliwal North and Piketberg) in Table 3.32 (Jozini and Trevanian respectively). The former two soils represent crusting kaolinitic soil and the latter two strongly crusting illitic soils studied under the rainfall simulator.

The experimental areas were cultivated for wheat seed-bed preparation and then sprayed with a long acting herbicide. No further cultivation was undertaken during the rainy season at the Irene, Potchefstroom and Aliwal North sites. During April 1989 these areas were recultivated and surface treatments re-introduced.

Short (1,5 m) as well as long (10 m) runoff plots were constructed at the Irene, Potchefstroom and Aliwal North sites. Each plot was bounded along the top and side by metal sheets driven 22 cm into the soil. Two sheets were used as legs of a triangle to channel the runoff and sediment into a flume installed at the bottom end of each plot.

The treatments of the small plots, during the first year included a control, PG spread on the soil surface at 5 t ha^{-1}. A PAM (polyacrylamide) treatment was introduced during the successive rainy season. Dissolved PAM at a rate of 29 kg ha^{-1} (Irene, Potchefstroom) and rates of 5 and 20 kg ha^{-1} (Aliwal North) was sprayed onto the soil surface. One set of PG plots at Irene was left untreated from the previous year in order to study the long-term effectiveness of the PG. The small plots included 5% and 30% slope gradients. Six long plots (10 m × 1,8 m) with 5% slope gradient were also constructed at each site. The treatments of the long plots included PG and control. The treatments were carried out in triplicate and randomized block design was used. The treatments at the Piketberg site, which is a winter rainfall area, were applied once before the rainy season and were tested during two consecutive rainy seasons (1988,1989). Only small plots were used at this site. Treatments included a control, PG and straw mulch.

3.7.7.2.2 Measurement of runoff

A 200 L drum with a 20 L bucket inside was buried at each plot. Runoff and sediment were collected in the bucket. Runoff was measured after each effective rain (usually >10 mm). A total of 10 –20 rain events were recorded for each site during the first year and about 5-10 during the second year.

A recording rain gauge was installed to measure the amount and intensity of rain at each site. Standard error of the mean, obtained when using ANOVA to analyze all storm events for each sites, is presented in Tables 3.35 to 3.37, 3.38 and 3.39.

TABLE 3.35 Effects of surface treatments on runoff from several rainstorms (1988/1989)- Irene (Msinga (D) soil)

No	Rainstorm Depth (mm)	Surface runoff, % of rainfall Interval (days)	Control	PG	Mulch	Stat. Sign.[A]	PG/Ct
1	44	-	11,8	3,8	0,1	+	0,32
2	14	24	9,8	2,5	0,6	+	0,25
3	20	6	17,5	7,2	0,7	+	0,41
4	21	2	58,4	37,9	2,9	*	0,65
5	8	18	28,8	15,7	0,0	*	0,54
6	30	22	9,9	2,4	0,5	+	0,24
7	38	4	61,5	38,6	12,3	*	0,63
8	19	9	21,2	5,6	0,6	+	0,26
9	17	5	11,7	2,1	0,1	+	0,18
10	21	10	66,5	48,2	8,4	*	0,72
11	7	4	29,3	16,4	1,5	*	0,56
12	8	7	9,5	1,3	0,0	+	0,14
12	39	14	40,4	31,0	2,4	*	0,77
14	54	8	33,4	25,7	6,2	*	0,77
Total 340							
Percent of annual rain			32,9	18,3	3,5	*	0,56

S.E of Mean – 4,1

[A] Statistical significance

+ Significant differences between control and other treatments

* Significant differnces between the different treatments

TABLE 3.36 Effects of surface treatments on runoff from several rainstorms (1988/1989) – Potchefstroom (Shorrocks D) soil).

No	Rainstorm Depth (mm)	Intervals (days)	Surface runoff, % of rainfall Control	PG	Mulch	Stat. Sign.[A]	PG/Ct
1	66	-	64,0	26,4	4,6	*	0,41
2	66	9	70,1	32,8	3,5	*	0,47
3	20	4	43,3	14,5	2,7	*	0,33
5	28	15	80,5	56,7	0,5	*	0,70
6	17	16	23,9	4,4	0,0	*	0,18
7	30	6	45,1	22,0	0,1	*	0,49
8	10	2	27,6	5,9	1,4	+	0,21
9	60	12	91,4	73,5	14,2	*	0,80
10	25	11	51,9	30,0	1,2	*	0,58
Total 316							
Percent of annual rain			63,6	35,9	4,6	*	0,56

S.E of Mean- 1,5

[A] Statistical significance

+ Significant differences between control and other treatments

* Significant differences between the different treatments

TABLE 3.37 Effect of PG application on runoff per cent from several rainstorms (1988/1989) – Aliwal North (Jozini soil)

No	Rainstorm Depth (mm)	Intervals (days)	Surface runoff, % of rainfall Control	PG	Mulch	Stat. Sign.[A]	PG/Ct
1	60	-	45,5	20,0	4,7	*	0,44
2	18	6	43,3	20,8	0,1	*	0,48
3	21	2	44,8	25,0	0,1	*	0,56
4	46	11	51,0	18,9	3,0	*	0,37
5	11	10	18,5	6,4	0,0	+	0,34
6	8	11	4,0	2,0	0,0	N.S.	0,50
7	140	14	94,6	75,5	25,8	*	0,80
8	19	5	27,5	10,1	0,0	*	0,37
9	9	2	74,6	49,9	2,1	*	0,67
10	10	4	31,9	20,5	0,0	*	0,64
11	13	20	49,5	39,0	1,0	*	0,79
12	8	5	68,2	51,4	1,8	*	0,75
13	19	8	79,5	65,6	1,5	*	0,82
14	13	1	89,7	70,6	3,1	*	0,79
15	12	1	80,7	66,0	3,1	*	0,82
16	11	22	36,1	20,8	0,1	*	0,58
17	8	5	37,0	6,5	0,1	+	0,18
18	46	6	94,3	78,3	2,6	*	0,83
19	17	7	42,4	7,9	0,1	*	0,19
20	12	18	53,9	13,4	0,1	*	0,25
Total	501						
Percent of annual rain			66,9	45,3	8,1	*	0,68
Excluding storm no 7			56,0	33,8	1,4	*	0,60

S.E of Mean – 3,8

[A] Statistical significance

+ Significant differences between control and other treatments

* Significant differences between the different treatments

[B] PG at rate of 6 t ha^{-1} was re-applied to the PG plots

3.7.7.3 Interpretation of results and discussion by Stern

The results presented below are for short plots with 5% slope only.

3.7.7.3.1 *Runoff from untreated (control) plots*

Runoff data collected during the 1988/1989 season from three summer rainfall sites, viz. Irene, Potchefstroom and Aliwal North are presented in Table 3.35 to 3.37. The runoff from control plots ranged from less than 10% during some storms to over 94% of the rainfall during some high intensity rainstorms. Final infiltration rates (IR) obtained for some of the soils in the rain simulator studies ranged between 2,4 to 9,5 mm h^{-1}. Most of the rainstorms at the Potchefstroom and Aliwal North sites had intensities much higher than 9,5 mm h^{-1} which resulted in runoff exceeding 63% and 67% of the annual rainfall (Table 3.36 and 3.37). Stern found that, at the Irene site, the intensities of rainstorms were lower and despite the high tendency of the soil to form a low permeability seal (Figure 3.20) the annual runoff was only 32,9%. In his studies, Stern indicated that rainstorms, with more cumulative rain but low intensities, produced less runoff than rainstorms with less cumulative rain but higher intensities. Runoff from two different storm types at the Irene site reflected the role of rain intensity in producing runoff.

Stern found that the clay fraction of the soil from Aliwal North is dominated by illite (Table 3.32) and is regarded as prone to seal formation. As a result, low water intake rates and high runoff were expected (Levy and van der Watt, 1988). The clay fraction of soils from Irene and Potchefstroom on the other hand were dominated by kaolinite, and were expected to be less dispersive. However, the high runoff during the season (Table 3.35 and 3.36) showed that the kaolinitic soils also formed seals with very low permeability when subjected to rainstorms. Stern indicated that there was a presence of small amounts of smectites in the clay fraction which induced dispersive behaviour in the kaolinite.

TABLE 3.38 Particle size distribution (<100 µm fraction) of sediments and original Soil from the Potchefstroom site (Shorrocks (D) soil)

Treatment	Particle size distribution (%)		
	V.F. Sand (100-20 µm)	Silt (20-2 µm)	Clay (<2 µm)
Mulch	25,5	73,2	1,3
PG	22,4	68,5	9,1
Natural soil	40,1	26,0	33,9

3.7.7.3.2 Effect of mulch on seal formation

Runoff from mulch covered plots was significantly lower than runoff from other treatments (Table 3.35 to 3.37). It was often less than 10% of the runoff from control plots, and ranged between 0 to 25,8% during the various storm events (Table 3.37). The percentages of annual runoff from mulch covered plots were low and ranged between 3,5% (Table 3.35) and 8,8% (Table 3.37).

When raindrops impact mulch the energy of the drops is dissipated and physical disintegration of soil aggregates is prevented. Since the stirring effect of raindrops is prevented, chemical dispersion processes are also inhibited. Particle size analysis of the sediments showed that those from mulched plots contained relatively more silt and much less clay and very fine sand particles than the natural soil (Table 3.38). Stern found that the reduction in very fine sand particle size in the sediments was due to lower preferability in transportation by the overland flow. However, the enrichment in silt size particles implies that minimal dispersion took place during the rainfall event and that the sediment from the mulched plots was transported as flocculated aggregates.

3.7.7.3.3 Effected of PG on runoff

Percentages of runoff obtained from PG amended plots are presented in Table 3.35 to 3.37. PG reduced runoff to 0,15 to 0,82 of that from the control. Final infiltration rates obtained for PG treated soils when using the laboratory rainfall simulator ranged between 6 to 15 mm h^{-1} (i.e. more than double that of the control plots). Stern found that most of the sediments from PG treated plots were transported by the runoff as silt size rather than clay size particles with are normally preferentially transported. This indicates that PG promoted flocculation and thus limited the amount of dispersive clay available for erosion. Stern indicated that the PG efficiency on reducing runoff under natural rainfall, markedly depends on the rainstorm intensity.

The interrelationship between rain intensity and PG efficiencies was described by multiple correlation between runoff ratio PG/control (Ct) and the Ln of runoff percentage from PG plots from all sites (Figure 3.29). Runoff percentage reflected the role of rainfall pattern in determining runoff and was used as an index for describing the intensities of the rainstorms. The percentage runoff from PG treated plots was preferred to percentage runoff from the control plots as an index since it correlated better with rainfall intensity. The runoff ratio PG/Ct ranged from 0,18 for low runoff (i.e. low intensity rainfall) to 0,80 where runoff from PG plots exceeded 70% of rainfall. PG amendment was most beneficial during storm events producing low runoff from PG treated plots which resulted in low PG/Ct ratios (Figure 3.29). The beneficial effect of PG was more pronounced at the Irene site (Msinga (D) soil) where most storm events were moderate and 50% of the storm events provided a PG/Ct ratio below 0,41 (Table 3.35).

The effectiveness of PG treatment was maintained throughout the first rainy season at the Irene and Potchefstroom sites. However, during the following season at Irene, after the accumulated rainfall exceeded 550 mm and the electrical conductivity (EC) in the runoff dropped to 0,2 dS m^{-1} (Figure 3.30), The effectiveness of PG decreased rapidly.

The reduction of EC rates from above 1 dS m^{-1} at the beginning of the rainy season to low rates of 0,3 and 0,4 dS m^{-1} (Figure 3.30) towards the end of the season did not reduce effectiveness of PG. However, further reduction in the EC during the successive season resulted in a sharp decrease in the PG effectiveness. The long-term effectiveness of PG is dependent on the pattern of rainfall and on the characteristics of the seal, both of which determine infiltration rate and runoff volume. Depending on the ESP and clay mineralogy, a minimum electrolyte concentration is required to enhance flocculation. This is known as the threshold electrolyte concentration. Stern suggested that the EC of 0,3 to 0,4 dS m^{-1} is the threshold for the Irene soil.

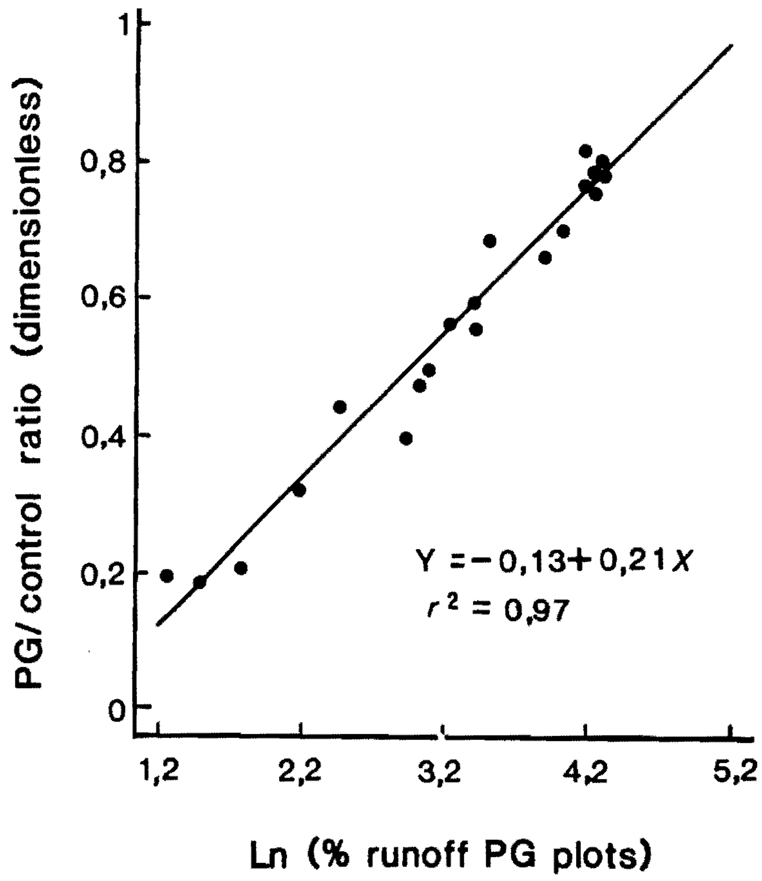

FIG. 3.29 Correlation between PG/Ct runoff ratio and the natural logarithm of runoff percent from PG plots

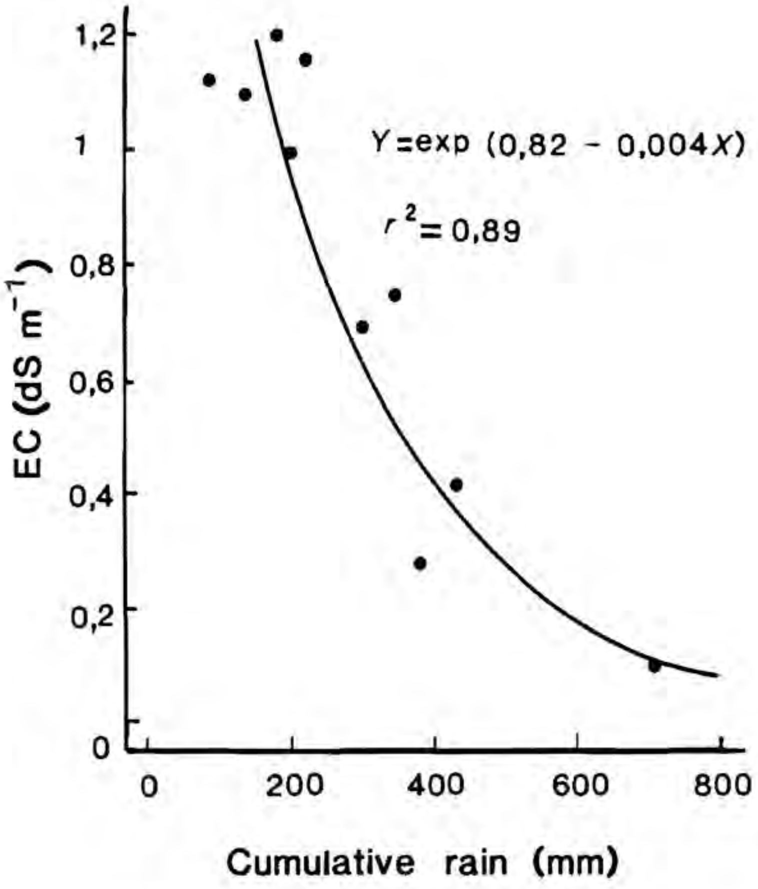

FIG. 3.30 Electrical Conductivity (EC) of runoff with cumulative rain

3.7.7.4 Runoff from Piketberg site

The Piketberg site was treated separately from the other sites, since it contained about 6,7% Na on the exchange complex. It was therefore expected to respond differently to raindrop impact and PG amelioration.

The effects of surface treatments on runoff at the Piketberg site are presented in Table 3.39. Mulch cover reduced runoff significantly during all storm events, indicating the role that seal formation played in enhancing runoff from bare plots. Nevertheless, the runoff percentages from mulched plots in several storms were higher than the values obtained at the other sites (Table 3.35 to 3.37) despite the moderate winter rains at Piketberg. The low initial infiltration rate of the Piketberg soil and rapid decline in the infiltration to reach low final rates as measured in the rainfall simulator (Figure 3.25), illustrated the rapid development of a low permeability seal on the soil surface when subjected to rain drops. Accordingly, runoff commenced even during very low depth rainstorms such as No's 2, 5, 7 and 14 of 5 mm or less (Table 3.39).

The PG treatment showed no significant effect on runoff compared with untreated soil during the 92 mm of rain accumulated from the first 7 storm events (PG/Ct = 0,96). Yet a significant influence on reducing runoff through PG amendment was observed during the subsequent storms, 8 to 15 (PG/Ct = 0,64), as well as during the following season (PG/Ct = 0,60) (Table 3.39).

The permeability of a soil to water depends both on its exchangeable sodium percentage (ESP) and on the electrolyte concentration (Quirk and Schofield, 1955; McNeal, Layfield, Norvell and Rhoades, 1968). In sodic soils, a relatively high electrolyte concentration is needed to reduce runoff and the beneficial effect of PG is somewhat limited (Keren et al., 1983). Stern found that at Piketberg, during the first 7 storms events some of the exchangeable Na^{+} was replaced by Ca^{++} in the surface soil. Thereafter, the beneficial effect of PG in increasing the infiltrability of the soil by supplying electrolytes to the percolating water was more pronounced. Stern also found that the rain depths during storm events at Piketberg, situated in a winter rainfall region, were lower than at Aliwal north and Potchefstroom, which are situated in thunderstorm regions, and did not exceed 28 mm. The intensities of the rain were also lower. However, the annual percentages of runoff in Piketberg ranged between 61,8 and 57,9% for the first and second season respectively and were as high as at Aliwal North and Potchestroom.

3.7.7.5 Effect of PAM on runoff

The effect of PAM on runoff at the Irene, Potchefstroom and Aliwal North sites is presented in Table 3.40 to 3.42. During all the storms, except No 7 (Table 3.42), the PAM treatments resulted in significantly lower runoff than the control. Furthermore, the annual runoff from the PAM treated plots did not exceed 50%, 38% and 35% of the runoff from the control at Irene, Potchefstroom and Aliwal North respectively. At Potchefstroom the PAM provided significantly lower runoff than PG (Table 3.41). Stern found that the surfaces of the PAM treated plots were rough and aggregated compared with the untreated soil which had a smooth surface on which no solid aggregates could be detected.

Stern found that PAM was less efficient in reducing runoff at the Irene site under field conditions (Table 3.40) than in the rainfall simulation study. This could be due to either (i) more efficient application of PAM in the rainfall experiment or (ii) degradation of PAM under field conditions.

3.7.7.6 Conclusions

Field studies were conducted to determine the effect of surface treatments on runoff. It was found that local rainstorms, some having high intensities, resulted in high percentages of runoff from the control plots. High runoff from the control plots of the kaolinitic soils was also associated with the susceptibility of these soils to dispersion and seal formation as observed during rainfall simulation studies. Stern indicated that mulch cover inhibited the disintegration of aggregates and the dispersion of clay particles and consequently markedly reduced runoff. PG was found to be beneficial in reducing runoff. PAM treatment was more effective than the PG treatment in controlling runoff.

TABLE 3.39 Effects of surface treatments on runoff from several rainstorms during 1988 and 1989 at the Piketberg site (Trevanian soil)

No	Rainstorm depth	Interval (days)	Surface runoff, % of rainfall Control	PG	Mulch	Stat. Sign.[A]	PG/CT
1	13	-	21,4	25,5	5,4	+	1,19
2	5	2	14,2	5,9	0,6	*	0,41
3	26	14	61,3	61,3	20,3	+	0,01
4	12	4	43,3	43,3	2,2	+	0,88
5	5	22	38,6	38,6	2,9	*	0,49
6	28	19	66,5	66,5	10,2	+	0,99
7	3	1	64,5	64,5	6,8	+	0,92
8	15	6	78,0	78,0	7,2	*	0,73
9	9	14	29,5	29,5	1,5	*	0,41
10	20	15	80,3	80,3	10,0	*	0,64
11	9	11	47,8	47,8	2,4	*	0,74
12	15	32	83,4	83,4	11,3	*	0,65
13	5	1	72,2	72,2	15,9	*	0,82
14	11	11	55,3	55,3	6,9	*	0,49
15	9	3	43,7	43,7	3,0	*	0,50
1988	185						
Percent of annual rain			61,8	48,5	9,5	*	0,78
16	10	-	59,8	33,1	1,7	*	0,55
17	5	3	24,9	9,5	0,0	*	0,38
18	9	4	44,6	14,8	0,1	*	0,33
19	16	12	68,5	47,8	2,0	*	0,70
20	7	1	63,1	48,2	2,0	*	0,76
21	20	19	60,8	36,0	1,2	*	0,59
1989	67						
Percent of annual rain			57,9	34,8	1,4	*	0,60

S.E of mean – 3,7

[A] Statistical Significances

+ Significant differences between mulch and other treatments

* Significant differences between all treatments

TABLE 3.40 Effects of PAM amendment on runoff from several rainstorms during 1989 and 1990 at the Irene site (Msinga (D) soil)

No	Rainstorm Depth (mm)	Interval (days)	Surface runoff, % of rainfall Control	PAM	Stat. Sign.[A]
1	66	-	44,2	24,3	*
2	41	3	39,2	14,9	*
3	16	2	49,6	35,2	*
4	44	22	45,0	25,5	*
5	16	3	49,6	21,7	*
6	31	10	19,5	7,9	*
Total	214				
Percent of annual rain			40,2	20,7	*

[A] * Significant differences between the control and PAM treatments

TABLE 3.41 Effect of PAM and PG treatments on runoff from several rainstorms during 1989 and 1990 at the Potchefstroom site (Shorrocks (D) soil

No	Rainstorm Depth (mm)	Interval (days)	Surface runoff, % of rainfall Control	PAM	PG	Stat. Sign.[A]
1	20	-	78,0	36,0	42,7	+
2	20	15	68,0	25,3	35,7	*
3	21	4	65,1	15,4	24,7	+
4	12	1	82,8	33,3	49,4	*
5	20	4	72,7	38,7	45,7	*
6	14	2	69,1	16,7	29,4	*
Total						
Percent of total rain			72,5	27,9	37,7	*

[A] Statistical Significance

+ Significant differences between control and other treatments

* Significant differences between the different treatments

TABLE 3.42 Effect of PAM treatments on runoff from several rainstorms during 1989 and 1990 at the Aliwal North site (Jozini)

No	Rainstorm Depth (mm)	Interval (days)	Surface runoff, % of rainfall Control	PAMA[B]	PAMB[B]	Stat. Sign.[A]
1	27	-	39,7	7,3	1,2	+
2	6	15	59,3	20,9	20,2	+
3	6	6	35,7	5,7	2,1	+
4	14	11	37,0	21,9	19,8	+
5	8	9	39,7	8,3	3,0	+
6	6	9	26,4	10,0	6,4	+
7	5	6	4,2	0,0	0,0	N.S
8	9	2	54,5	27,2	22,7	+
9	18	6	39,1	5,9	2,3	+
10	8	6	52,2	30,7	29,1	+
11	15	22	50,0	31,3	28,2	+
Total						
Percent of annual rain			40,9	15,3	11,6	+

[A] Statistical significance

+ Significant differences between control and other treatments

* Significant differences between the different treatments

[B] PAMA: PAM at a rate of 20 kg ha^{-1}. PAM at a rate of 5 kg ha^{-1}

3.7.8. RELATIONSHIP BETWEEN MICROAGGREGATE STABILITY AND SOIL SURFACE DISPERSIBILITY

3.7.8.1 Introduction

The stability of surface soil aggregates under the impact of water drops falling as rain or applied as irrigation will determine surface crusting, water infiltration and soil erodibility.

Forces involved in stability studies include (i) impact and shearing forces administered while preparing samples, (ii) abrasive and impact forces during sieving and (iii) forces involved in the entry of water into the aggregates (Kemper and Rosenae, 1986). Since dispersion and crust formation occur under wet conditions, this investigation was focused on disintegration occuring in the liquid phase.

In determining aggregate stability, known amounts of certain size fractions of aggregates are commonly subjected to disintegration forces designed to simulate some important field phenomenon. There are different types of aggregate stability test that are available. One of the tests is described by Emerson (1967) and Rengasamy *et al.*, (1984). In these tests the dispersion of air dried aggregates in water is determined. The measurement of spontaneous dispersion in the absence of any imposed external forces will reflect the behaviour of surface soils during the rainfall events when the soil surface is effectively protected by mulch. Imposing different periods of shaking, however, may stimulate the extent to which soil microaggregates disperse when bare soil is subjected to raindrop impact (Miller and Beharuddin, 1986).

3.7.8.2 MATERIALS AND METHODS

All seven non-sodic soils were studied, particulars of which are given in Table 3.43 and 3.44. In addition a soil of the Williamson series (Macvicar *et al.*, 1977) from the East London pineapple producing area was included. This was a representative soil from an area where severe erosion is prevalent. This lithitic inceptisol (USDA, Soil

Taxonomy) contains 28% clay, 51% silt and 2,1% organic carbon and has a CEC of 17,3 $cmol_c\ kg^{-1}$ and a pH (water) of 5,3.

3.7.8.2.1 ***Procedure and Apparatus***

A two gram sample of air dried aggregates (<105 µm) was weighed into a 50 ml beaker and 20 ml distilled water was added. The soil-water suspension was vibrated by placing a sonifier probe to a depth of 50 mm in the suspension. The soil suspension was subjected to a range of ultrasonic energy levels with three replicates per energy level (Gregorich *et al.*, 1988). The particle size distribution of the soil suspension was determined using the light scattering technique.

A high intensity probe type ultrasonic generating unit (Brason B-30) was used in this study. The output control was set to operate at 40 W on 50% duty cycle. These settings were chosen to a best represent the range of water drop energy which normally occurs under field conditions.

A malvern model 2600 particle size dector unit including a laser source unit, with interchangeable lenses coupled to a microcomputer was used to measure particle size distribution.

Table 3.43 Some physical, chemical and mineralogical properties of the kaolinitic soils used

Soil			Sand[C]				CEC[A]	ESP	Organic carbon	Fe	pH	Clay[B]
Series	Location	Taxonomy	Clay	Silt	V.Fine	Coarse					(H_2O)	Minerals
			g kg^{-1} %				Cmol$_c$ kg^{-1}	%	g kg^{-1}			
Msinga (D)	Irene	Paleudalf	319	258	66	357	38,0	1,4	17	55	5,6	K(5), I(2), Is(1)
Hutton (D)	Roodeplaat	Rhodudalf	300	142	148	410	38,4	1,3	9	20	6,0	K(5), I(2), Is(2)
Shorrocks (D)	Potchestroom	Rhodustalf	158	121	187	534	24,7	3,1	5	18	5,0	K(5), I(1), Is(1)
Msinga (S)	Dundee	Paleudalf	238	188	145	429	25,2	2,5	20	42	6,1	K(5), I(2)
Shorrocks (S)	Potchestroom	Rhodustalf	189	106	156	549	18,9	1,5	14	17	5,5	K(5), I(I)

[A] CEC = Cation exchange capacity cmol$_c$ per kg clay following the subtraction of OM contribution

[B] k = kaolinite, I = illite, Is = interstratified materials (swelling minerals). (1) = weak... (5) = strong

[C] V.Fine sand (<0,10mm); Coarse Sand (<2,00mm)

Table 3.44 Some physical, chemical and mineralogical properties of the illitic soils

S.A. Soil series	Jozini	Rosehill	Trevanian
Location	Aliwal North	Riviersonderend	Piketberg
Soil taxonomy	Haplustalf	Haplargid	Haplargid
Clay $g\ kg^{-1}$	166	233	189
Silt "	78	317	323
V.Fine Sand[A] "	651	165	88
Sand "	105	285	400
CEC[B] cmol(+) kg-1	35,4	43,8	20,8
ESP %	1,6	2,5	8,4
C $g\ kg^{-1}$	11	17	9
Fe $g\ kg^{-1}$	7	20	19
pH (H_20)	6,7	5,7	6,8
Clay	I(5), K(1)	I(4), K(3)	I(5), K(4)
Minerals[C]	Is(I)	CI(2), IS(1)	CI(2)

[A] very fine sand <0,10 mm

[B] CEC = Cation exchange capacity $cmol_c$ per kg clay following the subtraction of organic carbon contribution

[C] k = kaolinitr, I = illite, cl = cchlorite, Is = interstratified material (swelling minerals). (1) = weak …(5) = strong

3.7.8.3 Discussion and Interpretation of Results by Stern

Dispersion of microaggregates accelerated by external forces imposed through ultrasound waves reflected the percentage dispersible clay (PDC). The PDC of eight South African soils as affected by prolonged sonication time is presented in Figure 3.29. The dispersion rate can in general be described by multiple regressions (Figure 3.29a and 3.29c to 3.29h). It was found that the amount of dispersed clay increases rapidly with sonication time to level off latter on. However, for simulating natural conditions of high intensity rainfall, the energy input during 180 seconds was sufficient (North, 1979). The PDC of the Shorrocks (S) was found to be very low and ranged between 0,1 to 1,6% for 0 and 180 s sonication time. The disintegration rate of the stable microaggregates is nearly constant and has not been completed. This indicates that only a small portion of microaggregates from this stable soil was dispersed and maximum PDC had not been achieved after 180 s of sonication (Figure 3.29a. Stern (1990) found that the same pattern applies for the Shorrocks (D) soil but the PDC of this soil is higher and ranged between 0,2 to 9% for 0 and 180 s of sonication (Figure 3.29c). The Msinga(s) is the exception in which the correlation between the PDC and sonication time is an exponential one (Fig. 3.29b).

Maximum values of PDC were achieved in the Rosehill, Jozini, Williamson and Hutton (D) soils. A rapid increase in PDC to reach a maximum dispersion after only 60 and 120 s of sonication was obtained in the Jozini, Rosehill and Hutton (D) soils respectively, reflecting a high tendency of the microaggregates to disperse. Msinga (D) also give a high PDC, ranging between 2 and 16,4% for 0 and 180 s sonication. However a maximum value was not achieved. Stern (1990) found that the rapid disintegration of microaggregates with sonication time and the high PDC obtained as is shown in Figures. 3.29d to 3.29h are associated with rapid development of seal formation and decrease in IR.

Stern (1990) found that the ultrasound waves affect the disruption of microaggregates of the various soils since the dispersibility ratio (DR) differs according to the soil. For each soil the DR was calculated by dividing the percentage dispersible clay by the percentage of < 2μm particles found by conventional pipette particle size analysis (particle <100 μm) using sodium hexametaphosphate dispersion (Day, 1965). The DR of the eight soils as affected by sonication time is presented in Figure 3.30. Stern (1990) found out that the Shorrocks (S) and Msinga (S) soils have stable aggregates which highly resist dispersion by ultrasound. They provided low spontaneous dispersion (SD) (Rengasamy *et al.*, 1984) obtained by 0 sonication time and maintained comparatively low disintegration rates with sonication time. The rates of dispersion among Msing (D) and, Williamson, Hutton (D)and Rosehill soils are more susceptible to dispersion. It was found that Jozini soil provided high SD, high disintegration rate with sonication time between 0 to 60 s and reached considerably higher maximum DR than the other unstable soils.

Stern (1990) found that adequate description of the dispersion curve is required if one wants to relate the aggregate stability of the soils to their susceptibility to runoff and erosion.

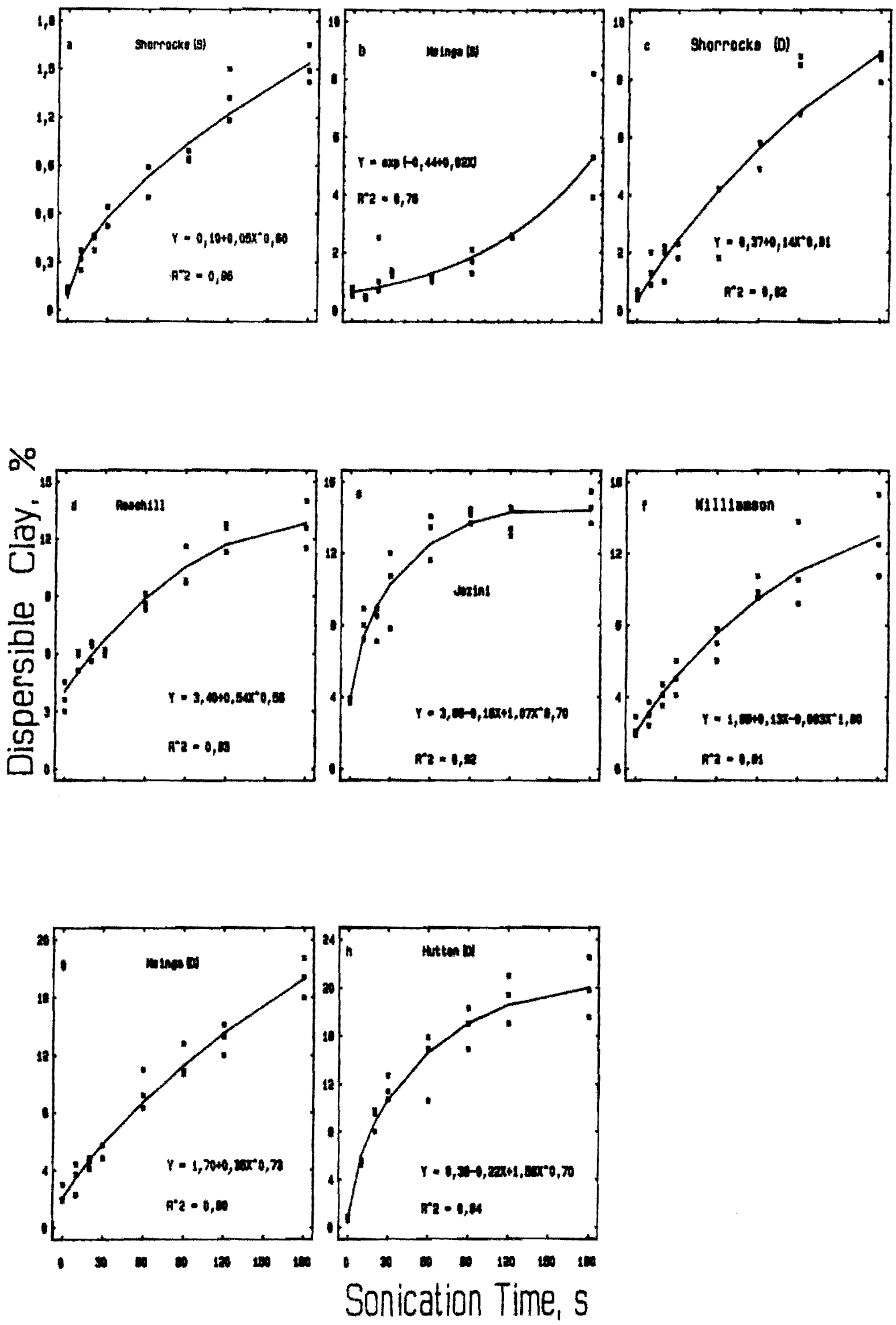

FIG. 3.29 Percentage dispersible clay with sonication time for eight soils

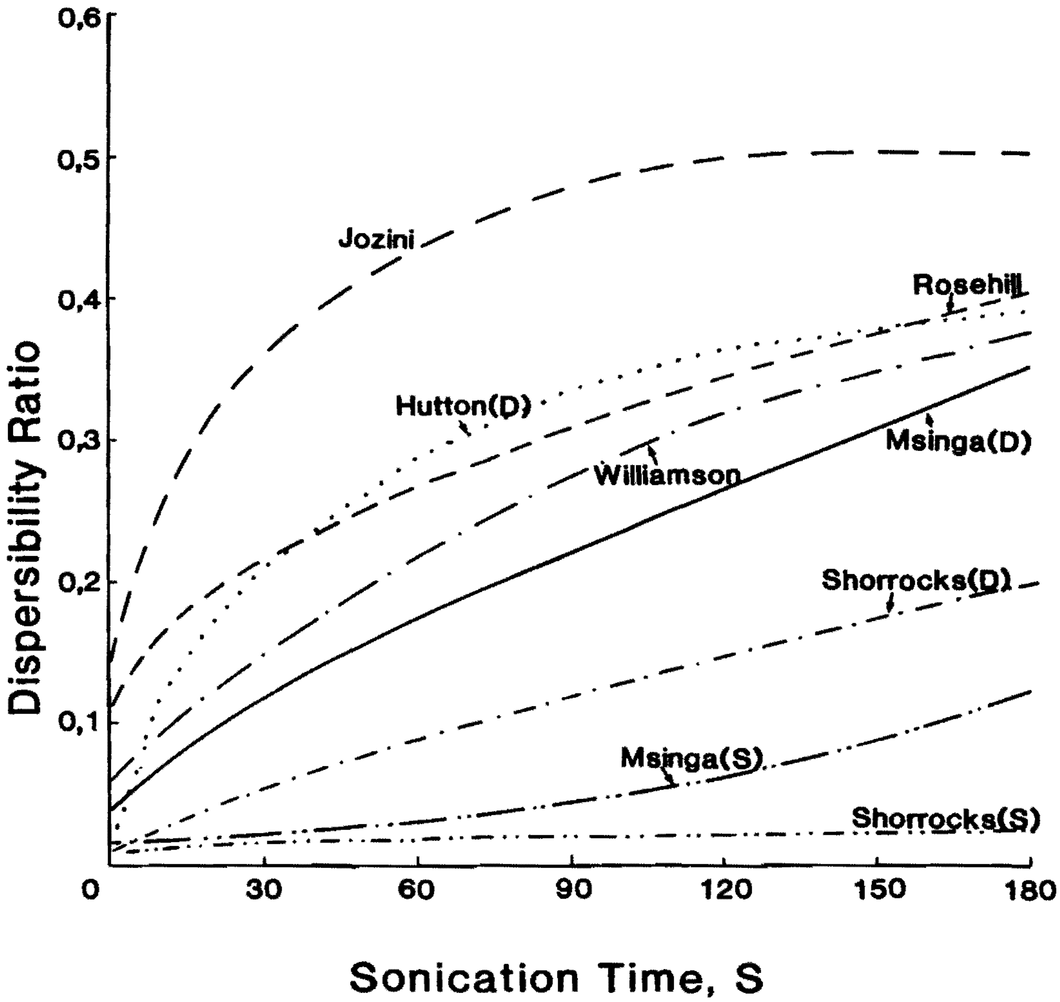

FIG. 3.32 Dispersibility Ratio (DR) vs. sonication time for the eight soils studied

3.8 BLOEM, AA (1992): CRITERIA FOR ADAPTION OF THE DESIGN AND MANAGEMENT OF OVERHEAD IRRIGATION SYSTEMS TO THE INFILTRABILITY OF SOILS.

3.8.1 General background and objective of the study

Due to the dramatic rise in energy costs and increasing competition for water resources by urban users, it has become essential to ensure maximum water-use efficiency in irrigation system design. In order to achieve this, irrigation systems should be designed and managed so that the application rate of water does not exceed the infiltrability of the soil. If this is not accomplished it will cause the following adverse effects:

- Excessive evaporation of ponded water
- Inefficient water use and sub-optimal crop production (Stern *at al.,* 1992)
- Soil erosion
- Pollution of dams and rivers with agricultural chemicals

The infiltrability of soils decreases with progressive infiltration due to a decrease in matric suction gradient (Hillel, 1980), but may also decrease due to the formation of a surface seal (McIntre, 1958). In cultivated soils the infiltrability is often controlled by the formation of a surface seal (Moore and Larson, 1979; Morin et al., 1989, Bloem, 1992).

The objective of this study was to determine how irrigation systems should be designed and managed within the constraints of soil infiltrability.

3.8.2 Materials and methods

Samples collected from the upper layer (0 to 200 mm) of 45 soils throughout SA were used in this study. The soil samples were air-dried and sieved through a 4 mm sieve. The physical and chemical properties of these soils are presented in Table 3.45. A modified model of the drip type laboratory-scale irrigation simulator described by

Bubenzer and Jones (1971) was used in this study. The sieved soil samples were packed (20 mm deep) into 50 mm deep, 272 × 221 mm soil trays. The trays were oscillated horizontally during the irrigation application to achieve an even distribution of droplets over the soil surface. The packed soil was saturated with water with an electrical conductivity (EC) of 70 mSm^{-1}. Water percolating through the soil was collected after each 5 mm application so that the IR could be calculated. The runoff rate was computed likewise from runoff collected from the soil trays. From these data the cumulative application that could be made without ponding (CAWP) was determined. Because of the relatively steep slope (9%) and smooth soil surface (aggregates were smaller than 4 mm), it was assumed that the surface storage was negligible. The CAWP values for each soil were correlated with the kinetic energy (KE) of the applied water, and from these regression curves the KE for CAWP values of 15, 20 and 25 mm were determined. These KE values represent the maximum allowable kinetic energy (MAKE) at which an irrigation system should be designed to make a certain application without ponding or runoff.

3.8.3 Results and discussion

It is evident from Table 3.46 that the CAWP decreases with an increase in falling height, i.e. with an increase in kinetic energy. This can be ascribed to a more rapid formation of a surface seal at higher levels of kinetic energy. From the data presented in Table 3.46 it is clear that there are vast differences between the reactions of different soils to surface sealing. Because the physical and chemical properties of a soil determine its susceptibility to surface sealing, the MAKE values (Table 3.47) were correlated with soil properties.

After inspection of a plot of MAKE values against clay content (Figure 3.31) Bloem found that samples having MAKE values lower than the general trend, referred to as dispersive soils in Figure 3.31, have at least one of following properties:

- ESP values higher than 2.0
- Clay mineralogy dominated by smectite (more than 50% smectite and less than 28% kaolinite)

- Ca:Mg ratio smaller than 1.0
- Organic matter content lower than 0.2%

It is well known that these 4 factors are involved in soil dispersion, and consequently in surface sealing and lowering of the infiltrabilities of soils. Higher dispersivity of soils dominated by smectite than those dominated by illite or kaolinite has been well-documented internationally.

Several researchers, e.g. Rengasamy *et al*., (1986), have found that chemical dispersion of soils increased when the Ca:Mg ration is less than 1.0. Soils with Ca:Mg ratios of less than 1.0 are widespread in SA. Bloem et al., (1994), indicated that the instability of these Mg-rich soils and their susceptibility to erosion are well-known, though not well documented in scientific literature.

Further investigation of Figure 3.31 by Bloem showed that the "stable" soils containing small amounts of smectite and could be separated, due to a lower stability, from those without any smectite. These soils are in fact not very stable, just relatively more so than the highly dispersive group.

Bloem grouped the soils into the following 3 groups:

- Dispersive
- Stable with minor presence of smectite
- Stable without any smectite

It is clear from Figure 3.31 that clay content is the soil parameter with the biggest effect on surface sealing in each of the three groups. An increase in clay content caused an increase in surface sealing (decrease in MAKE) in all three cases.

Bloem found that the MAKE correlated well with clay content but the three groups of soils showed different relationships. Some statistical data of the curves fitted for three design applications of 15, 20 and 25 mm are presented in Table 3.48. The correlation coefficients and significance could not be improved by including clay content, ESP, Ca:Mg ratio, organic matter content or clay mineralogy in multiple regressions. For

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the dispersive soils a sharp decrease in MAKE occurred with a rise in clay content until about 20%, above which it remained very low.

Table 3.45
Some physical and chemical properties of the soil studied

Sample no.	Clay Content %	Silt+ Content %	CEC* $cmol_ckg^{-1}$	ESP	pH H_2O	Organic Matter %	Ca:Mg**	Clay mineralogy **
1	36.2	32.1	51.8	9.7	8.5	2.08	2.43	St 33 Mi55 Kt12
2	16.1	11.1	33.5	5.7	8.0	0.19	2.24	
3	34.0	14.5	40.8	13.3	7.8	0.95	0.11	
5	37.8	39.1	94.3	0.6	8.0	1.33	1.45	St18 Mi29 Kt49 Go4
6	6.1	33.2	52.1	4.1	6.6	0.19	1.94	St10 Mi50 Kt40
7	10.0	5.4	39.4	4.1	5.9	0.38	1.91	
8	6.1	5.4	29.9	4.9	6.5	0.01	2.00	Mi19 Kt81
9	6.2	1.4	38.6	4.6	6.2	0.19	2.05	Mi19 Kt81
11	21.4	37.4	124.5	2.5	7.7	1.51	10.59	
12	16.1	8.3	84.7	6.7	8.1	0.95	2.25	St17 Mi29 Kt50 Go4
13	14.8	4.1	118.8	1.8	7.2	0.57	1.27	
15	43.6	28.7	80.4	6.7	7.8	2.46	1.12	
17	17.0	11.7	52.9	0.1	7.9	0.76	1.84	St11 Mi12 Kt55 Go4
18	13.7	5.0	40.2	3.4	6.4	0.19	1.68	Mi17 Kt50 Kt/Is3
19	13.5	7.1	38.8	3.6	6.4	0.38	1.50	Mi24 Ktg1 Vm15
21	11.6	1.0	44.1	1.0	7.7	0.57	1.83	Mi50 Kt25 Kt/Is25
22	9..5	1.0	45.7	0.9	7.0	0.38	2.38	St13 Mi37 Kt30 Kt/Is20
23	14.9	3.1	19.3	1.0	6.5	0.38	0.92	Mi22 Kt62 Go9 Py17
24	12.6	4.3	7.7	2.5	4.9	0.35	1.16	Mi22 Kt54 G014 Py10
25	11.0	4.2	52.0	2.4	5.9	0.49	2.40	Mi20 Kt75 Vm5
28	17.7	2.0	22.7	0.8	7.8	0.45	1.44	St16 Mi30 Kt37 Tc17
29	14.8	10.0	39.5	2.1	5.3	0.45	2.07	St41 Mi37 Kt22
30	29.7	18.7	79.1	0.3	7.3	2.27	2.08	St63 Mi6 Kt28 G03
33	32.2	38.4	57.2	0.9	8.2	2.08	2.03	St19 Mi69 Kt12
35	25.4	15.9	56.1	0.6	8.0	1.14	10.72	St31 Mi14 Kt55 Kt/Is6
36	31.9	22.7	56.3	1.3	7.9	1.14	1.47	St56 Mi8 Kt27 Tc7 Go2
38	25.3	15.1	60.3	0.6	7.4	1.14	2.07	St66 Mi19 Kt7 Kt/Ist8
40	24.9	8.2	40.4	0.1	6.5	3.98	1.49	Mi21 Kt79
41	32.6	18.4	12.9	1.5	5.7	1.82	-	St25 Mi25 Kt50
42	22.6	10.4	14.2	2.4	6.4	0.96	-	Mi11 Kt48 Go15 Tc17 Vm9
43	29.0	18.0	26.0	0.8	7.3	0.86	123	St6 Mi38 Kt56
45	72.0	14.9	69.1	2.1	6.9	3.80	1.92	Mi23 Kt44 Kt/Is10
47	37.3	41.9	67.1	0.4	7.8	2.46	1.27	St20 M16 Kt60 Go4
48	57.5	36.8	67..1	1.3	6.9	2.65	1.67	St53 Mi30 Kt17
50	38.6	14.4	51.8	2.1	6.1	2.44	-	St59 Mi3 Kt28 Kt/Is10
51	38.6	18.4	22.5	0.1	6.7	1.77	-	St14 M23 Kt45 Go18
52	55.0	19.0	43.0	1.7	6.1	2.36	1.57	Mi19 Kt81
54	39.0	42.0	54.5	0.1	4.8	4.58	-	Kt10 Vm45 C145
55	30.4	5.1	20.5	0.6	6.5	1.14	1.13	Kt89 Vm11
56	20.0	8.9	84.9	0.4	7.4	1.70	2.26	St46 Mi20 Kt34
57	27.4	8.3	27.4	0.3	7.4	1.50	1.65	Mi 19 Kt71 Go10
58	20.9	11.4	36.5	1.6	7.8	1.10	1.97	Mi10 K77 Go13
59	46.4	15.8	34.8	2.0	6.6	1.33	1.76	Mi27 Kt73
60	42.0	29.0	45.0	0.2	6.9	1.77	1.13	Mi13 Kt62 Go4 Tc13 Vm7
61	8.7	1.3	41.0	1.1	4.8	0.19	2.61	Mi6 Kt83 Go7 Py4

* Cation exchange capacity of the clay fraction
** Relationship between exchangeable cations ($Cmol_c.Kg^{-1}$)
*** A semi-quantitative estimation of the clay mineralogy

Where St = Smectite
Mi = Mica
Kt = Kaolinite
Vm = Vermiculite
Tc = Talc
Py = Pyrophylite
Go = Goethite
Cl = Chlorite
Kt/Is = Kaolinite

\+ Size of 2 to 50 μm

Table 3.46

THE CUMULATIVE APPLICATION (MM) THAT CAN BE MADE BEFORE PONDING OCCURS

Sample no.	Falling height			
	3.0 m	**2.0m**	**1.0 m**	**0.3 m**
1	0.3	6.1	9.5	13.5
2	9.0	13.1	18.5	60.2
3	4.4	10.2	10.8	12.1
5	0.1	0.7	13.5	14.9
6	-	12.2	20.3	33.8
7	7.6	16.0	12.1	20.0
8	4.1	6.1	-	42.7
9	16.9	27.1	45.3	54.1
11	4.4	8.3	9.1	22.0
12	0.7	8.8	15.6	31.8
13	0.1	1.8	-	10.0
15	0.1	1.0	2.2	3.1
17	10.2	14.2	33.8	-
18	6.1	12.2	15.6	25.0
19	0.1	14.2	18.3	-
21	34.5	40.7	47.6	60.7
22	90.0	90.0	90.0	90.0
23	-	4.1	7.4	14.9
24	-	0.7	3.4	22.3
25	10.2	29.1	34.5	75.8
28	4.7	13.5	18.9	35.2
29	0.1	3.4	8.8	16.2
30	0.7	6.8	12.2	41.3
33	12.9	15.6	21.0	27.1
35	4.1	18.3	23.0	-
36	4.1	10.8	28.4	-
38	0.1	0.1	1.4	2.7
40	10.8	18.9	27.1	48.7
41	-	0.1	3.4	20.3
42	-	8.8	15.6	33.3
43	-	0.1	7.4	15.5
45	0.1	3.4	10.2	23.7
47	3.4	12.2	18.3	28.4
48	0.1	0.7	-	14.2
50	4.1	8.1	13.5	27.1
51	-	3.4	7.4	44.0
52	12.9	16.2	23.0	44.0
54	16.2	25.7	45.3	-
55	3.4	10.2	-	45.3
56	9.5	13.5	33.8	-
57	10.2	23.1	29.8	-
58	19.6	27.1	36.5	50.1
59	6.8	12.2	16.9	35.2
60	6.1	12.9	20.3	33.0
61	2.5	11.1	17.9	26.7

TABLE 3.47

THE MAXIMUM ALLOWABLE KINETIC ENERGY
($J.mm^{-1}m^{-2}$) FOR A CERTAIN DESIGN APPLICATION

Sample	Design application		
No.	25 mm	20 mm	15 mm
1	0.1	0.1	0.7
2	6.5	8.1	11.2
3	0.1	0.1	0.1
5	0.1	0.1	3.5
6	7.1	8.6	12.2
7	1.7	3.0	9.1
8	2.2	11.2	13.3
9	15.6	18.1	20.3
11	2.6	2.8	4.2
12	5.0	7.7	10.7
13	0.1	0.1	0.1
15	0.1	0.1	0.1
17	11.5	11.6	14.4
18	1.7	5.9	10.9
19	5.5	7.9	11.2
21	24.0	24.0	24.0
22	24.0	24.0	24.0
23	1.3	1.5	1.8
24	2.3	2.3	3.3
25	10.8	14.9	17.7
28	7.2	8.0	10.7
29	0.1	0.1	2.6
30	6.0	6.2	7.6
33	4.4	10.1	16.1
35	8.3	10.9	14.1
36	9.1	11.5	13.8
38	0.1	0.1	0.1
40	11.7	12.7	22.1
41	2.7	2.9	3.5
42	1.0	6.6	9.4
43	0.1	0.1	2.5
45	0.2	3.2	6.0
47	4.7	7.9	11.5
48	1.6	2.1	2.6
50	3.5	5.0	7.7
51	3.8	4.1	5.0
52	7.0	10.0	16.0
54	14.9	17.2	20.5
55	9.9	12.1	14.0
56	10.1	12.2	14.8
57	11.8	14.8	17.8
58	15.5	19.6	24.0
59	6.7	7.8	11.0
60	6.8	9.9	13.1
61	6.6	7.1	10.9

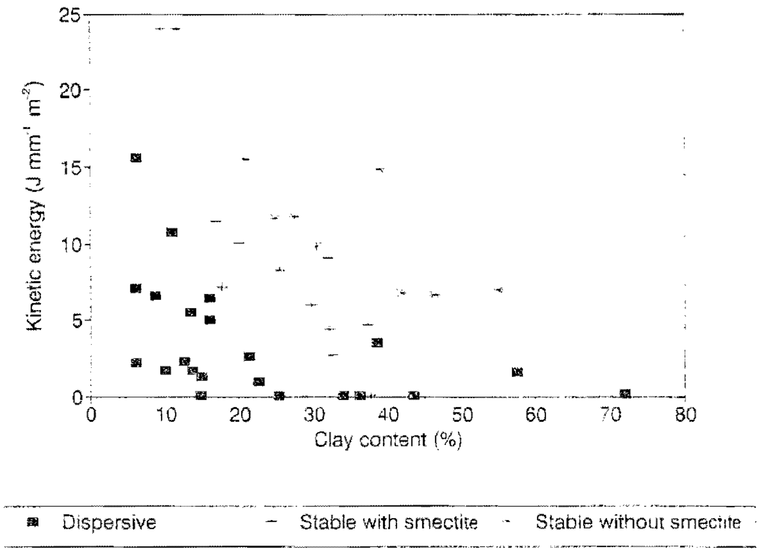

FIG. 3.31 Relationship between maximum allowable kinetic energy and clay content for the samples used in this study.

TABLE 3.48

STATISTICAL PARAMETERS OF THE CURVES FITTED BETWEEN KINETIC ENERGY (MAKE. THE DEPENDENT VARIABLE) AND CLAY CONTENT OF THE SOIL CATEGORIES

Soil class	**Design Application mm**	**Equation**	**Correlation R^2-values**	**Significance level**
Disperse	25 20 15	$Y=61.92X^{-1}-0.849$ $Y=74.81X^{-1}-0.185$ $Y=90.00^{-1}-0.905$	0.449 0.471 0.479	$P<0.001$ $P<0.001$ $P<0.001$
Stable With smectite	25 20 15	$Y=1.152X-16.643^{0.5}+61.830$ $Y=0.779X-12.371X^{0.5}+51.763$ $Y=0.314X-7.167X^{0.5}+44$	0.851 0.779 0.648	$P<0.001$ $P<0.001$ $P<0.01$
Stable Without Smectite	25 20 15	$Y=131.749X^{-0.732}$ $Y=0.313X+24.529$ $Y=0.229X+24.983$	0.733 0.659 0.460	$P<0.01$ $P<0.01$ $P<0.05$

CHAPTER 4

SYNTHESIS OF THE RESULTS OF THE DISSERTATIONS AND THESES SUMMARIZED IN CHAPTER 3

4.1 INTRODUCTION

In this chapter data regarding all the important factors that play significant roles in the erodibility of soil that have been presented in the previous chapter are synthesized. It is important to note that in some cases the findings of the researcher(s) per soil property/factor may be related to published data from elsewhere.

4.2 PARENT MATERIAL

The results presented in the previous chapter by Sumner (1957, Section 3.5 and D'Huvetter 1985, Section 3.2) showed that soils derived from basic igneous rocks, especially dolerite, had higher stability against erosion as compared with the majority of other soils, mainly those from sedimentary rocks of the Beaufort and Ecca groups. As a result of high base content, soil that form from dolerite are dominated by smectite (2:1 swelling clays) except in the humid tropics and subtropical regions.

The Beaufort (i.e. shales and mudstone) and Ecca groups are associated with substantial amounts of magnesium, sodium and the clay mineral illite and as a result they produce soils with silt percentages. Furthermore they produce unstable duplex soils that are incredible erodable. It is well known that high Mg give rise to very poor structure resulting in very compact soils and high erodable soils. Generally it is accepted that soils derived from Beaufort and Ecca series are more subject to water erosion. It is important to note that Sumner (1957, Section 3.5) found that soils associated with Beaufort and Ecca group have a marked ability to swell, especially in the subsoils.

Soils derived from dolerite such as Shortlands, Hutton and Arcadia soil forms in the Mavuso pedosystem, Amatola basin and Middledrift were characterised by a significantly higher stability against erosion compared with the majority of other soils in these areas, D'Huyvetter (1985, Section, 3.3).

In Comparison of dolerite and granite as parent material, Smith (1990, Section 3.2), found that soils developed on dolerite have higher amounts of cations, free iron and aluminium, higher amounts of clay (higher amounts of clay forming minerals present) and also generally higher pH values (higher amounts of basic cations) as compared to soil developed on granite. It was found that soils that form from granite usually contain large amounts of course sand that originate from the quartz. Smith (1990, Section 3.2, Table 3.1) also found that granite soils are rich in potasium but very poor in calcium and magnesium. In granite soils, the weathering of the clay minerals progresses easily to the 1:1 phase.

4.3 CLIMATE

The combination of rainfall and temperature, especially, dominates soil formation. In general South Africa is a dry country. At least 65% of the country has a mean annual rainfall of less than 500 mm.

In all of the areas studied by Sumner (1957, Section 3.3) and D'Huyvetter (1985, Section 3.5) rainfall was found to be more than 500 mm per annum. In South Africa this can be regarded as good rainfall when one considers the other parts of the country. All areas studied by Sumner (1957) and D'huyvetter (1985) can be regarded as warm and humid regions. It is well known that warm and humid conditions promote weathering and therefore soils in the humid tropics and sub-tropics are in a very advanced stage of weathering.

Sumner (1957, Section 3.5) found that Northern Kwazulu Natal receives rainfall in the form of aggressive thunderstorms. The three main pedosystems of the former Ciskei, namely, Mavuso, Keiskammahoek and Middledrift also receive rainfall in the form of aggressive thunderstorms (D'Huyvetter, 1985) – reducing their efficiency and increasing their erosivity. As a result of low rainfall, poor quality soils develop. In areas where there was a combination of low rainfall and unfavourable parent material, shallow soils were found, E.g. Valsrivier, Vilafontes, Kroonstad and shallow Glenrosa in the Middledrift pedosystem were formed on colluvial mudstone.

Large differences between soils were observed over short distances in the higher Tabamhlophe as a result of sharp differences in climatic conditions (Turner, personal communication). Sumner (1957, Section, 3.5) found that soils developed in high rainfall areas in the higher Tabamhlope were stable against erosion when compared with those from low rainfall areas. This clearly indicates the role of climatic conditions on the stability of soils

4.4 SOIL FACTORS

4.4.1 Clay mineralogy

The erodibility of soils can largely be derived from their clay mineralogy. From the results presented by Levy (1988, Section 3.4), Stern (1990, Section 3.7) and Rapp (1998, Section 3.8) it is clear that clay mineral type is an important factor influencing the stability of soil aggregates and hence erodibility. Clay mineralogy also determines the fundamental physical and chemical properties of the soil surface and therefore has large effects on the degree of sealing and the nature of the seal (Stern, 1990).

The results presented by Levy (1988) and Smith (1990) showed that most of South African soils were dominated by either kaolinite or illite with smectite and interstratified material as the secondary minerals. Levy and van der Watt (1988) found that if soil is dominated by illite it could be expected to be susceptible to crust formation. Soil from Aliwal North was dominated by illite and, as expected, was prone to crusting (Stern, 1990).

Soils which contain pure kaolinite form stable aggregates, maintain high IR and have low erosion. Conversely, kaolinitic soils which contain small amounts of smectites are dispersive. The high erodibility of the Westleigh soil form in Thabazimbi is attributed to its kaolinitic clay mineralogy "contaminated" by smectite, low organic carbon content and very high soil salinity Rapp (1998, Section 3.6). Soils which do not contain smectite are more stable, less erodible, and are less susceptible to seal formation.

4.4.2 Texture

Sumner (1957) indicated that the amount of water absorbed by the soil is influenced by its mechanical composition and structure. In sand and other light textured soils, the coarseness of the particles is probably the most important factor contributing to rapid water absorption and permeability and decreased erodibility. In heavy soils, there is apparently no direct relationship between the clay content and the amount of erosion. Degree of aggregation and size and stability of the aggregate, therefore might have a pronounced effect upon rate of percolation and, consequently, upon amount of runoff. Bloem (1992, Section 3.8) showed that clay content is the soil parameter with the biggest effect on surface sealing of the soil.

D'Huyvetter (1985) found that fine sand content of the A horizon was statistically correlated with the degree of erosion. High fine sand levels partly explains the high degree of erodibility of the Kroonstad and Glenrosa soils of the Middledrift pedosystem. In all three main pedosystems of the Eastern Cape studied, namely, Mavuso, Keiskammahoek and Middledrift, topsoils with less than 20 per cent clay content were found to be highly erodable while those which had clay contents of more than 20 per cent were found to be less erodable.

Rapp (1998, Section 3.8) found that as clay content in soil increases, cohesion forces between soil particles increased as a direct effect, while at the same time the interparticle spacing increased (lower bulk density and higher saturated moisture content), leading to lower attraction between soil particles as an indirect effect. Also, indirectly with increased clay content, the hydraulic conductivity of the soils and the infiltration rates decreased, leading to high shear stress and transportability.

The results presented by Sumner (1957, Section 3.5) showed that clay content affected crusting. He indicated that crust formation could develop with any type of texture except sand with very low silt and clay. Although sandy soils may not develop crusts, they can exhibit low intakes after several irrigations due to surface sealing. The crusts formed in medium textured soils of South African and Israeli soils were found to be the most dense and maintained the lowest final infiltration value (Levy, 1988, Section 3.4).

4.4.3 Chemical factors

The results regarding the effect of exchangeable K and exchangeable Ca on the hydraulic properties of soil (i.e. hydraulic conductivity and infiltration rate) have been presented. Levy (1988, Section 3.4) found that K had different effects on the soils than did exchangeable Ca. Increasing the amount of K in the exchangeable phase resulted in a decrease in the HC as well as in the infiltration rate of the soil.

The study on the effects of exchangeable Mg on hydraulic properties of soils shows that exchangeable Mg is not as efficient as exchangeable Ca in maintaining HC under sodic conditions. However, from the IR experiments, it is evident that Mg treated soils are similar in their behaviour when exposed to rain. The presence of Mg in the exchangeable phase probably enhances clay dispersion in comparison to Ca and therefore the HC obtained with Mg treated soils was found to be lower (Levy, 1988).

The exchangeable sodium percentage (ESP) in soil was also studied. Normally an ESP of 15 is considered to be the threshold above which problems occur. In South Africa and Australia, sodium induced dispersion, crusting, and erosion start at much lower ESPs (in some soils low as 2 or 3 per cent). Magnesium is another problem cation, and high magnesium: calcium ratios (usually above 1.7:1) are often the cause of dispersion and enhanced erosion in many parts of SA and elsewhere in the world, such as Australia and Russia. Calcium is a flocculating cation thus counteracting dispersion, but it cannot stabilise aggregates against disintegration (Laker, 2000).

Rapp (1998, Section 3.6) found that the presence of exchangeable sodium in kaolinitic soils influences crust formation only when the soils contain some smectite. Furthemore, in soils with moderate to high ESP values, where clay mineralogy is "contaminated" with smectite, dispersive conditions prevail and the efficiency of clay particles as cementing agents between soil particles is very small.

A highly significant correlation exists between the exchangeable sodium percentage (ESP) of the soil and the degree of erosion. In general this correlation is stronger for the A horizon than for the B horizon. The highest ESP values were found in soils of the Valsrivier, Vilafontes, Westleigh, and Kroonstad forms of the Mavuso and Middledrift pedosystem, all of which have high erodibilities. Soils with low A and B horizon ESP values, such Oakleaf and Shortlands soils, in Middledrift were associated with weak to moderate degrees of erosion. The Glenrosa form, having low ESP values in both A and B horizon, yet showing severe erosion, was an exception. A relatively high ESP in the A horizon induces colloidal dispersion, resulting in the formation of dense surface crusts. Such crusts strongly reduce infiltration of water into the soil. The resulting increase in surface runoff is conducive to erosion (D'Huyvetter 1985, Section 3.3).

Soils showed different responses to ESP levels. Some soils were affected by exchangeable Na at low levels, others at moderate and high levels. It was found that FIR decreases with an increase in ESP, and it is important to note that some soils were hardly affected by sodicity up to an ESP level of 9 (Fig. 3.15).

4.4.4 Degree of weathering

The results presented by Smith (1990) showed that many soils have features which relate them to the parent rock from which they were formed. Some minerals like quartz, muscovite mica and some feldspars were found to be more resistant to weathering than other minerals.

Conditions in the soils play a significant role in determining the types of clay minerals that formed from weathered products. In soil conditions where the Si and basic cations are not removed by the leaching, 2:1 type clay minerals, such as smectites and micas, will dominate in the soil environment. Highly weathered soils in humid tropical and subtropical areas, where intensive weathering and leaching of bases and silica has occurred, are dominated by minerals that represent advanced stages of weathering, Examples of this are Clovelly and Griffin soils of Tabamhlope in KwaZulu Natal (Turner, personal communication, 2001).

It was found in the mineralogical data (Table 3.5), that soils from high rainfall areas contain mainly some of end-products of weathering and resistant clay-size mineral. The amounts of 2:1 minerals was found to decrease with an increase in rainfall in soil samples representing both acidic and basic parent material.

In general the results presented indicated that fine crystalline structured rocks (basic rocks) are much more resistant to weathering than coarse crystalline rocks (acidic rocks).

4.4.5 Types of soils

Different types of soils have been studied from different areas in South Africa. Soil forms and the stabilities of these soils against erosion differed from region to region. One should note that major part of South Africa is covered by very poor quality soils. Most of the country is dominated by shallow soils.

It was found that different soils have different slope-erosion relationships, different threshold slopes and they respond differently from treatment e.g. application of phosphogypsum (PG). Ultimately they vary in regard to their erodibility.

Soil losses varied markedly between the stable and unstable soils and were also affected substantially by slope gradient and PG amendment (Table 3.30). Stern (1990) found that low soil losses from the stable soils is associated with high infiltration rate and low runoff (Table 4.2).

Generally it has been found that red soils are much more stable against erosion than those of other colours, probably because the hydraulic properties of these red soils are better than that of non-red soils. Example of this is the very stable red soils of the Shortlands and Hutton forms that have studied as compared to other soils, such as the Sterkspruit, Valsrivier, Vilafontes and Estcourt forms. These soils occurred on footslopes and valley bottoms.

In most of the developing areas in South Africa, and especially in semi-arid and sub-humid regions, the lower landscape positions are dominated by very unstable duplex and pseudo-duplex soils which are extremely vulnerable to water erosion. Due to unfavourable textural properties of most of the deep soils on the river terraces these soils are also very vulnerable to erosion.

4.5 SLOPE

From the soil-slope-erosion data, D'Huyvetter (1985) attempted to derive models which could be used to predict threshold slopes, above which erosion would be a danger under normal cropping practices, for unknown sites where land use planning has to be done. The topographical factors which received attention were: slope length, slope gradient, slope form (convex, concave or plane) and slope feature above the point of study.

D'Huyvetter (1985) found out that for a few soil forms the length of the slope above the point of observation, especially very long slopes, resulted in increased erosion as a result of increased surface runoff.

The soil-slope-erosion relationships depicted in Figures 3.3 to 3.14 revealed clear differences between different soils in regard to inherent erodibility. D'Huyvetter (1985, Section, 3.3) and Stern (1990, Section 3.7) showed that slope percentages have the greatest influence on the degree of erosion.

The threshold slope criterion of 12 per cent used in the past was far too steep for the vast majority of soils in the Mavuso, Keikammahoek and Middledrift pedosystem. The criteria were found to be more-or-less valid for only the small areas of Hutton and Shortlands soils.

The relationhip between slope gradient, IR and soil loss in soils which are susceptible to crusting depends on the properties of the seal (Stern, 1990). Increase in slope gradient from 5 to 30% resulted in soil losses increasing by 520 and 570% for the untreated Msinga (D) and Jozini soils respectively (Stern, 1990, Section 3.7). The rapid increase in soil loss from the untreated Jozini soil with increasing slope gradient started from the flattest slope, with a higher increase rate between 20 and 30%. Stern (1990) found that treatment with phosphogypsum (PG) reduced soil loss.

4.6 THE IMPORTANCE OF PLANNING

Planning involves anticipation of the need for change as well as reactions to it. Its objectives are set by social or political imperatives and must take account of the existing situation. In many places, the existing situation cannot continue because the land itself is being degraded (FAO., Guidelines for Land Use-Planning, 1993).

Environmental and/or social disasters caused by incorrect land use planning in the former homelands are numerous. D'Huyvetter (1988) gave an example of poor planning, which lead to disastrous erosion, as a result of lack of understanding of the resource base.

In the former Ciskei, large parts of the arable areas which were "scientifically" selected and planned since the mid-1950's have suffered devastating erosion, despite good countouring and other measures. This has been the result of lack of basic soil chemical and soil physical knowledge (Laker, 1990).

In the Mavuso, Keiskammahoek and Middledrift pedosystems, for example, a slope of 12% was used as the cut-off point above which soils were deemed to be non-erodable due to erosion hazard, irrespective of the inherent stability and erodibility of the specific soil. It was found that even at low slope gradients many soils suffered serious erosion. This clearly indicates that generalized criteria cannot be used in a situation where there are different types of soils (D'Huyvetter, 1985, Section 3.3).

A large fraction of the areas in the former Ciskei covered by the Glenrosa, Mispah, Sterkspruit, Valsrivier, Swartland, Vilafontes and Westleigh soils should never have been cultivated. The land use planner did not have the necessary data regarding the different degrees of stability and appropriate critical slope criteria for the different soils, and as a result this led to the severe erosion (D'Huyvetter, 1985).

CHAPTER 5

CONCLUSIONS AND RECOMMENDATIONS

5.1 CONCLUSIONS

The major part of South Africa is dominated by very poor quality soils. These soils are inherently unstable and extremely vulnerable to erosion. Most of the country is, in fact, dominated by shallow soils. This is caused by low and inefficient rainfall, which limits soil formation, and the hard geology (Laker, 1990). Runoff, besides depriving crops of essential moisture supplies, is also the cause of soil erosion. South Africa, with its poorly distributed, relatively unreliable rainfall and frequently shallow soils can ill afford moisture and top soil losses resulting from often avoidable runoff. One should note that soil erodibility varies seasonally and the magnitude of seasonal variation varies with soil texture.

It is well known that soils formed on basic parent material have higher amounts of clay, iron, aluminium and basic ions when compared to acidic parent material. This is the result of the higher amounts of primary minerals in basic rocks than in their acidic counterparts (Smith, 1990).

The data presented in the previous Chapters have conclusively shown that Beaufort and Ecca soils are highly erodible and their subsoils are extremely impermeable to water and are almost structureless. The erodibility of the Beaufort and Ecca soils is due, firstly, to the low infiltration capacity which gives rise to a large amount of runoff. Secondly, the instability of aggregation and the dispersibility of these soils are additional factors contributing to erodibility (Sumner, 1957). The actions of raindrops are additional factors contributing to erodibility. When comparing the Ecca and Beaufort soil with dolerite soils on the basis of infitration alone, it has been found that the Beaufort and Ecca soils are more erodible than the doleritic soils. The impermeability of the Ecca and Beaufort soils is due primarily to the compact nature of the subsoil and the highly dispersed nature of the colloidal complex. The relatively low infiltration capacities recorded for the Beaufort and Ecca topsoils which have a reasonably good structure and are not readily dispersible, can only be explained on the basis of crust formation.

The doleritic soils in all areas studied were found to be stable due to a good water stable structure, low dispersibility and high infiltration capacity. Such properties are conducive to low runoff and consequently very little erosion results. It was also found that soils like Shortlands and Hutton forms, both derived from dolerite, are characterised by a significantly higher stability against erosion compared with the majority of other soils, mainly formed on mudstone.

Vast differences in erodibility have been found between different soil forms, resulting in vastly different threshold criteria (D'Huyvetter, 1985).

The clay mineralogy affects the swelling and shrinkage potential of a soil, which in turn affects soil structure.The mineralogical composition of the clay in the soil accounts for the differences in seal formation, infiltration rate and soil loss rates of soils. In most areas where studies have been carried out, it was found that the kaolinitic and illitic soils have a greater resistance to the impact of raindrops than smectitic soils and the surface conditions of these soils maintain higher infiltration rates, which in turn increase the amount of water taken up by during rain. Furthermore, in a soil in which kaolinite is virtually the only clay mineral, crust formation is not affected by the presence of exchangeable sodium up to an ESP of 10. The presence of exchangeable sodium in kaolinitic soils influences crust formation only when the soil contains some smectite.

In a comparative study of South African and Israeli soils it was found that crusts formed in South Africa soils were less dense and more permeable than those formed in Israeli soils. The differences are ascribed to differences in clay mineralogy between the two groups of soils (Levy, 1988).

From the results presented by Rapp (1998) it is evident that rill erodibility decreases with an increase in clay and organic carbon content, conversely rill erodibility increases with an increase in total electrolyte concentration and increases with an increase in soil exchangeable sodium percentage. The effect of organic carbon content on rill erodibility was found to be high in kaolinitic soils, low in smectite soils and intermediate in the soils dominated by illite. It was also found that high rill erodibility values are associated with soils with interstratified illite/smectite mineralogy and coarse soils with kaolinitic mineralogy. The relative order of rill erodibility of soils composed of different clay mineralogy depends mostly on the organic content of the soil and its exchangeable sodium percentage.

5.2 RECOMMENDATIONS

In considering possibilities for the utilization of land, a prediction should be made on how a certain land use or technology is going to affect the land, before any damage is done. The range of technical approaches must be broad, and specific techniques matched carefully with environment- both physical and socio-economic.

Every piece of land should be used optimally i.e. to the best advantage of the people, without causing soil erosion. Erosion can be horrific on steep slopes when land is cleared

Cultivation should be avoided in any area where Beaufort soils (shales and mudstone), Ecca soils and unstable duplex soils are dominant. The main reason for this is that ploughing leads to exposure of subsoils, which is highly undesirable. In a situation where there are animals in the field, good pasture management is needed on these soils. There should be no overgrazing nor overstocking as these practices lead to erosion. A judicious system of veld burning should also be introduced. It is imperative to maintain maximum vegetal cover at all times in area(s) where the soils mentioned above are prominant.

Soil analysis should be undertaken prior to the establishment or rehabilitation of any irrigation project, or for that matter for cultivation of any suitable land.

A single slope criterion cannot be used for all soils, simply because the differences between different types of soil are too great. Appropriate threshold slope criteria should be identified for the different soils in an area, or according to soil properties determining the erodibility of the soils (D'Huyvetter, 1985).

People involved in the planning of farming areas, residential areas, etc. should have adequate knowledge of how to correctly evaluate the qualities of the resources. One would not want a situation whereby people use generalized norms, in view of the vast variation in resources, as this would simply aggravate the situation.

Mapping and classification of eroded areas must be regarded as a priority research area, since regular updating of erosion maps can give a clear indication of the rate of soil erosion and the efficiency of soil conservation measures. Remote sensing and GIS can be employed and play a vital role in identifying and monitoring soil erosion at various levels of management.

Efforts should be made to prevent recurrence of the mistakes of the past. This may be achieved by involving a multi-disciplinary task team in a project.

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