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.

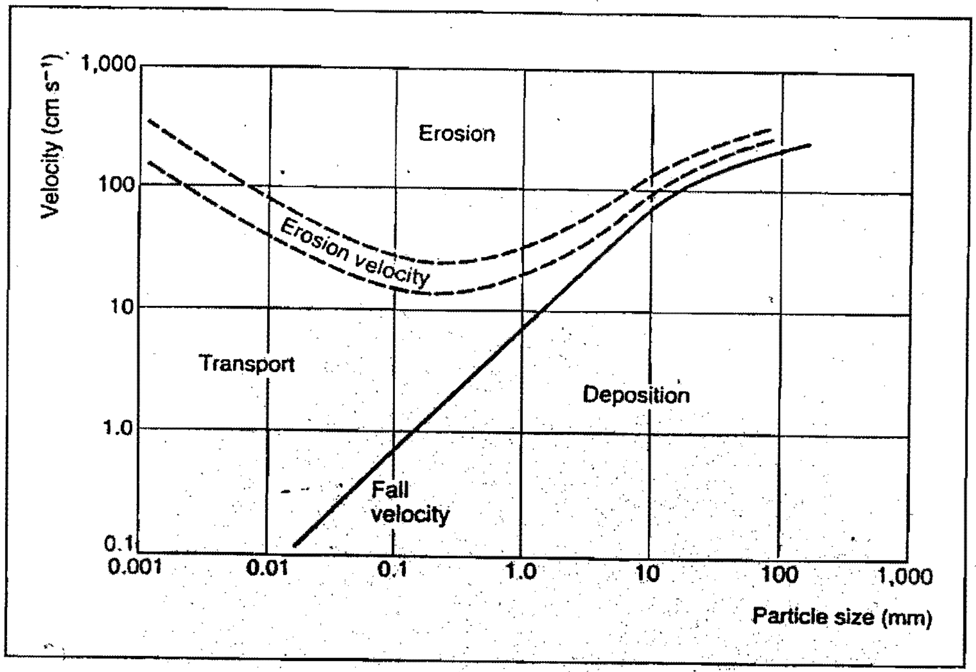

Fig 2.1 Critical water velocities for erosion, transport and deposition as a function of particle size (From: Hjulstrom, 1935).

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.