CHAPTER 1

INTRODUCTION

It is well known that South Africa is a water-poor country (Department of Water Affairs and Forestry, 1986; Reynders, 1991; Parsons & Jolly, 1994). Due to a rapidly expanding population, increased consumption of water by the industrial sector and general upliftment in living standards of South African people, it is believed that the demand for water will in the future increase significantly.

The South African government's objectives as regards the management of water resources are "to achieve optimum, long term, environmentally sustainable social and economic benefit for society from their use" (Braune & Dziembowski, 1997). The Department of Water Affairs and Forestry (South Africa) has recognised that a holistic approach should be followed in the management of water resources (Braune & Dziembowski, 1997). This implies an integrated approach to management, with regard to both quantity and quality, for all water resources, i.e. surface and sub-surface water.

The importance of groundwater in South Africa is well-documented (Braune, 1990; Reynders, 1991; Parsons & Jolly, 1994). Approximately 65 per cent of the area in South Africa relies on this water source in one way or another (Braune, 1990). Groundwater is an important source of water for many rural communities, especially those that have been disadvantaged.

While rapid growth puts ever-increasing demands on groundwater as a source of urban, agricultural and rural water supply, it also causes deterioration in groundwater quality, mainly due to poor groundwater management, waste disposal, mining and agricultural activities.

The government's commitment to ensure that every person has access to clean drinking water should include the protection and efficient management of groundwater, to ensure its use as a sustainable resource for both groundwater users, and indirectly for surface water users. Groundwater management should thus strive towards ensuring that a sustainable source of water of an acceptable degree of quality is maintained. This implies that an assessment of groundwater recharge and vulnerability should be an integral part of the management of groundwater resources. Unfortunately, both aspects are as yet poorly understood. This is partly because these processes take place in the vadose zone, an area that has been largely neglected by the geohydrological community.

Much of the theoretical knowledge regarding unsaturated flow and soil-moisture retention is based on investigations conducted largely to determine the moisture retention characteristics for agricultural purposes. However, during the last two decades, this knowledge has primarily been applied to liquid flow and contaminant transport investigations. The theoretical aspects of fluid movement and contaminant transport through the vadose zone are fairly well established and have been proven by numerous vadose zone investigations. However, soil scientists have traditionally focussed on the top 1.2 metres of the soil profile and very little investigations have been conducted in deeper soil profiles, with the result that very little is known about the hydrogeological characteristics for major parts of the vadose zone.

Engineering geologists have also traditionally been involved with investigations regarding the vadose zone primarily for engineering purposes. In most cases, geotechnical investigations comprise an assessment of the weathered geological profile that usually constitutes a portion of the vadose zone. Engineering geologists and geotechnical engineers are mainly interested in soil and/or rock material strength, possible volumetric changes and permeability of materials.

More recently, a general awareness of the environment has compelled engineering geologists to address environmental aspects in many geotechnical investigations. These investigations include developments of waste disposal sites, underground and surface storage facilities as well as pipelines for the storage and transmission of hazardous fluids, water purification works, cemeteries, low cost housing, developments where no sewage services are available and industrial developments.

Depending on the purpose of the investigation, engineering geologists may be required to assess soil conditions for

i. Regional areas (e.g. planning for residential and industrial developments)

ii. Local areas (e.g. residential developments)

iii. Site-specific areas (e.g. foundation investigations).

A large number of proven geotechnical methods and techniques have been developed to assess soil and rock conditions for these situations. These methods may include aerial photo interpretation, description of soil profiles and a number of *in situ* and laboratory tests.

Standard geotechnical tests, such as soil profile descriptions and foundation indicator tests, are conducted at almost every geotechnical investigation. In the case of geotechnical investigations for residential purposes, these geotechnical data may cover large areas. These data are stored for record keeping purposes and are generally available from the relevant engineering and engineering geology institutions. In addition, a number of mainly public institutions have established large geotechnical data sets, mainly for general development and transport planning purposes. This information is generally available from the relevant institutions, in formats varying from hardcopy reports and maps to electronic databases and from Geographical Information Systems (GIS). This information can be used in hydrogeological investigations.

The current groundwater situation can be summarised as follows:

- The importance and vulnerability of groundwater requires that the resource should be effectively managed. This entails an accurate assessment of groundwater recharge and its vulnerability to contamination.

- The geohydrological community has traditionally been involved in the exploration of groundwater resources, groundwater exploitation and the assessment of possible contamination pathways within the groundwater regime. However, due to a lack of knowledge and information regarding the vadose zone, groundwater recharge and vulnerability could often not be accurately assessed.

- The soil science community has contributed significantly to the understanding of unsaturated flow processes. In addition, soil data are available from mainly public sector institutions. However, soil data have traditionally been collected for agricultural purposes, therefore ignoring the deeper soil and rock profile constituting a major portion of the vadose zone.

- Engineering geologists have been traditionally involved in assessing soils and rocks for engineering purposes. Many of these structures, such as waste disposal sites, can adversely affect groundwater. The engineering geologists are therefore required to assess the contamination potential for these structures, often lacking knowledge to accurately assess the situation.

- On the other hand, engineering geologists are in possession of large geotechnical data sets and have the knowledge and expertise to accurately assess geotechnical characteristics for large areas. These data sets and geotechnical techniques can be used to assess groundwater recharge and contamination and will assist geohydrologists to accurately assess the sustainability and vulnerability of groundwater resources.

The research aims to bridge the knowledge gap between practising geohydrologists involved in groundwater recharge and contamination investigations and engineering geologists frequently involved in similar investigations from an engineering point of view. It also aims to incorporate the wealth of existing geotechnical data to estimate important hydrogeological properties. These estimated properties could be used, in conjunction with relevant climatic and geohydrological information, to estimate groundwater recharge and contamination for use in aquifer vulnerability studies and recharge assessments.

1.1 Aims of this study

The hydrogeological properties of soil, climate and vegetation are the main aspects that need to be considered in the assessment of water flow in the vadose zone. Water infiltrates the vadose zone mainly because of precipitation and irrigation. In addition, water is removed from the soil by means of evaporation and transpiration processes. The portion of water not removed by evapotranspiration processes reaches the groundwater and is termed recharge. Rainfall events are erratic and, as such, water flow in the vadose zone is dynamic. This implies high variability in annual recharge that will mainly be a function of rainfall patterns.

The rate at which water infiltrates into, or is removed from the soil will mainly be a function of the hydrogeological properties of the vadose zone. The term "*hydrogeological properties of the vadose zone*" refers to mainly physical properties of soils and rocks that constitute the vadose zone and are affecting the rate at which water is flowing through it. Flow in the vadose zone is a function of *inter alia* moisture content, which implies that the rate of water flow through the vadose zone is determined by complex functions of precipitation and evapotranspiration processes. Considering the highly erratic characteristics of rainfall, it is the author's opinion that these complex relationships can only be described, understood and predicted by means of numerical modelling.

Since infiltration and evapotranspiration theory is well-established (Ward, 1975; Bodman & Colman, 1943; Philip, 1964; Penman, 1963, Horton, 1933), it was decided not to consider climate as part of the project. Rather, it was decided to focus on estimating hydrogeological properties from readily available geotechnical data, thereby providing inputs, which together with climatic inputs, infiltration, redistribution and evapotranspiration theory, could be used in the assessment of groundwater recharge, vulnerability and other processes occurring in the vadose zone.

The aims of this research were:

- To identify those hydrogeological properties important in the assessment of saturated and unsaturated flow occurring in the vadose zone
- To establish relationships between readily available geotechnical data and the hydrogeological properties of residual soils that occur within the vadose zone
- To apply the above-mentioned results in the hydrogeological characterisation of the vadose zone for a specified study area.

The initial approach was to obtain direct relationships between geotechnical and hydrogeological properties as described by established one-dimensional flow equations through the soil matrix of the vadose zone. However, it soon became apparent that preferential flow could have a major impact on flow of fluids and solute transport through the vadose zone. It has also been established that the spatial distribution and variability of soil properties could have a significant impact on recharge within a regional area. Research was conducted on three aspects regarding the hydrogeology of the vadose zone, namely: pedotransfer functions, preferential flow and hydrogeological characterisation on a field scale (**Figure 1.1**).

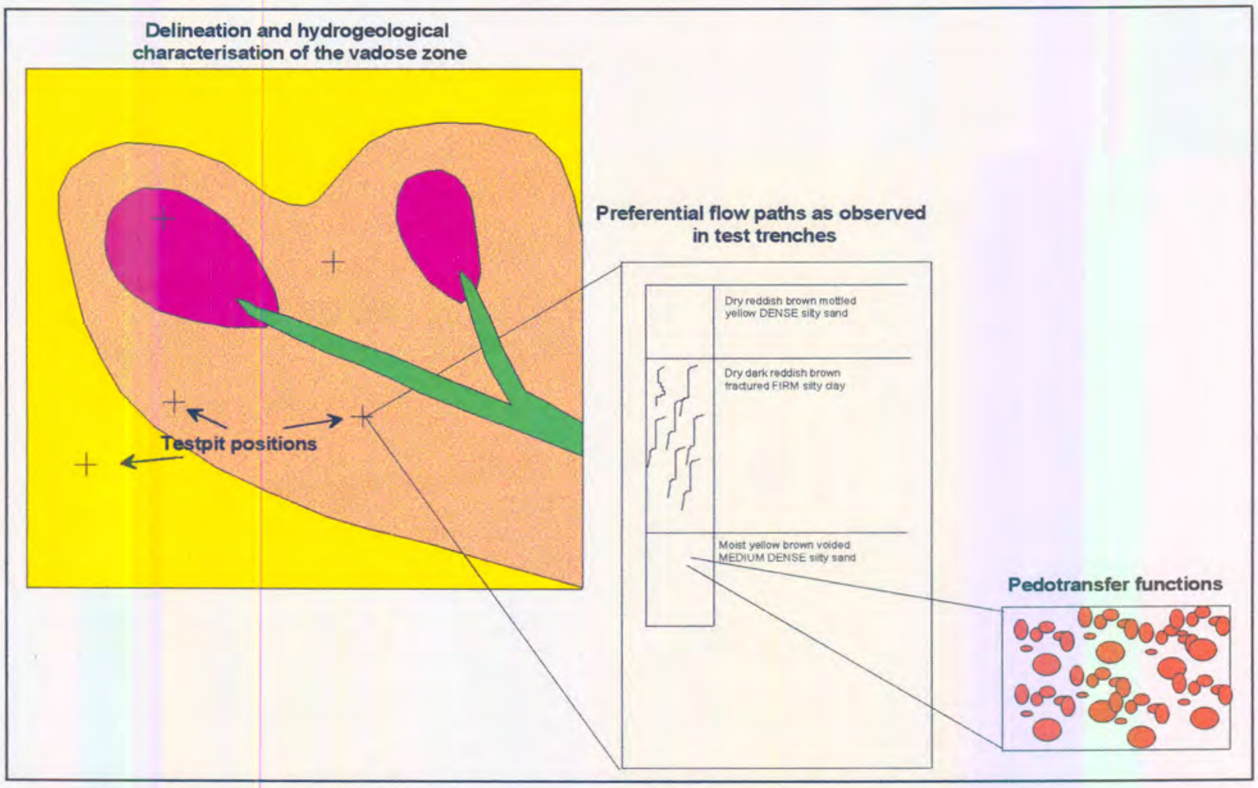

Figure 1.1: Three aspects considered regarding flow in the vadose zone

Pedotransfer functions refer to methods and techniques to estimate important hydrogeological parameters from widely available soil data such as particle-size distribution and Atterberg limits. These functions are based on point source soil samples and generally represent flow characteristics through specific soil layers within the vadose zone. Pedotransfer functions are typically empirical in nature.

Preferential flow represents flow mechanisms where fluids by-pass the soil matrix. In this investigation, the impact of macropore channelling was assessed for three different residual soils.

Both pedotransfer functions and preferential flow were considered in the compilation of a map indicating units of similar hydrogeological characteristics of the vadose zone. This map can be used in groundwater recharge and vulnerability assessments and indicates how geotechnical data can be applied to assist in the assessment of groundwater recharge and vulnerability. Aspects such as attenuation and climate were not considered in the compilation of this map.

It can be seen that the three above-mentioned aspects account for the effect of scaling within the vadose zone. Pedotransfer functions are used to determine hydrogeological parameters for specific soil layers within the vadose zone. Preferential flow accounts for preferred flow through the vadose zone and, in the case of macropore channelling, it can generally be observed in soil profiles. Hydrogeological characterisation implies that units with similar hydrogeological characteristics can be presented by means of either electronic or hardcopy maps.

1.2 Approach

In compiling this thesis, it was taken into consideration that the results could be applicable to a broad spectrum of scientists involved in groundwater research. These scientists include engineering geologists who may not be fully versed in geohydrology and geohydrologists who may not be acquainted with aspects of engineering geology. Likewise, these scientists might not possess knowledge on unsaturated and preferential flow processes occurring in the vadose zone. For this reason, the thesis contains an extensive literature survey, covering aspects ranging from basic textbook information to frontline research findings.

This thesis incorporates the extensive literature study and the application of the research findings to the study area.

Chapters 2, 3 and 4 contain an extensive literature study. Chapter 2 covers basic aspects of the vadose zone, theory of flow through saturated and unsaturated soil, preferential flow, groundwater recharge and vulnerability and numerical unsaturated flow modelling. Chapter 3 describes the general aspects, methods and techniques applied during engineering geology investigations, and also discusses the availability of geotechnical data. Chapter 4 covers relationships between geotechnical data and hydrogeological properties of soils in the vadose zone. Both saturated and unsaturated properties are considered.

Chapter 5 describes the experimental procedures. Those include five field studies conducted at three different areas in South Africa, laboratory experiments and analyses of data sets obtained from the literature. Chapter 6 discusses the results of the experiments.

Chapter 7 describes the application of the above-mentioned relationships in a regional hydrogeological investigation of a study area located in Midrand, South Africa. A map of the study area, depicting zones of similar soil types, has been compiled in GIS and is included as Drawing 1 at the back of the thesis.

CHAPTER 2

GEOHYDROLOGICAL CHARACTERISTICS OF THE VADOSE ZONE

2.1 Behaviour of a fluid in a porous medium

Porous media such as soil and rock are generally three-phase systems, comprising solids, liquids and gasses. Saturated porous media refer to the situation where almost all gases have been replaced by liquid. Unsaturated porous media refer to the situation where both gasses and liquids are present in between the porous medium.

2.1.1 Saturated porous media

Saturated conditions refer to a two-phase system comprising solids and liquids. The liquids are subjected to gravitational force. The behaviour of the liquids is primarily governed by the Fundamental Theorem of Fluid Statics, stating that the difference of pressure in a fluid in a vertical direction can be represented by the following equation:

$$\frac{\delta p}{\delta z} = -\rho_f \cdot g \qquad [2\text{-}1]$$

Equation 2-1 can be applied to calculate any pressure, p, at height, z, at a given pressure p_0, and height, z_0, provided that the density is a constant or a known function of both pressure and height as is shown in the following equation.

$$\int_{p_0}^{p} \frac{\delta p}{\rho_f . g} = \int_{z_0}^{z} \delta z = z - z_0$$

or

$$z = z_0 + \int_{p_0}^{p} \frac{\delta p}{\rho_f \cdot g} \qquad [2\text{-}2]$$

If the fluid is incompressible or the effect of pressure and density is negligible, Equation 2-2 is simplified as follows:

$$z = z_0 + \frac{(p - p_0)}{\rho_f \cdot g} \qquad [2\text{-}3]$$

The pressure at the groundwater surface is equal to atmospheric pressure. In subsurface flow studies, the value of atmospheric pressure, p_0, is by convention considered to be nil. Height, z, is known as the pressure head, piezometric head or Hubert's potential, h_p, and pressure, p, is the difference between pore-water and pore-air pressure. Since pore-air pressure is usually equal to atmospheric pressure (nil), pressure, p, is equal to pore-water pressure, u_w, Equation 2-3 can then be expressed as follows:

$$h_p = \frac{u_w}{\gamma_w} \qquad [2\text{-}4]$$

2.1.2 Unsaturated porous media

Unsaturated conditions refer to a three-phase system comprising solids, liquids and gasses. Under these conditions, the voids are filled mainly by gas, since most of the liquid has been removed because of gravitational force. Forces that counteract the force of gravitation to hold liquid in the porous medium are called *matrix forces*. These forces include *capillary* and *adsorption forces* and *electrical forces on a molecular level.*

In soil-plant environments, the matrix forces may include the effect of osmotic forces. Osmotic forces refer to the attraction of solute ions or molecules to water. If pure water is separated from water containing solutes, by a membrane that is not permeable for solutes, water molecules will move towards the solute water mixture and this will cause a higher pressure in the solute water side of the membrane. Since osmotic suction has little effect on movement of water through a porous medium, osmotic forces have been omitted for purposes of this study.

Capillary forces

A wetting liquid, such as water, will rise in a capillary tube, caused by the difference in pressure between the liquid and gas within the tube. The difference in pressure occurs owing to curvature on the liquid-gas interface, known as the meniscus, in a capillary tube.

The relationship between pressure difference, *dp*, and a double-curved surface element on the liquid-gas interface (such as the curvature that occurs in a capillary tube) is expressed by Laplace's equation:

$$dp = \sigma\left(\frac{1}{r_1} + \frac{1}{r_2}\right) \qquad [2\text{-}5]$$

If the radii are equal in length, Equation 2-5 is simplified to:

$$dp = \frac{2 \cdot \sigma}{r} = \frac{4 \cdot \sigma}{d} \qquad [2\text{-}6]$$

The pressure difference within the capillary tube can be expressed as:

$$dp = \frac{2 \cdot \sigma \cdot \cos(\alpha_w)}{R} \tag{2-7}$$

The height, h_c, to which the liquid rises in the tube is controlled by the downward force caused by the weight of the liquid, and can be calculated as follows:

$$h_c = \frac{2 \cdot \sigma \cdot \cos(\alpha)}{(\rho_l - \rho_g) \cdot g \cdot R} \tag{2-8}$$

The density of the gas, ρ_g, is generally ignored. From Equations 2-7 and 2-8 we obtain the following:

$$h_c = \frac{dp}{\rho_l \cdot g} \tag{2-9}$$

dp has a negative value, which implies that the pressure beneath the meniscus in the capillary tube is lower than atmospheric pressure and that the liquid will rise in the tube.

Equations 2-4 and 2-9 can be combined to yield an equation that is applicable in both saturated and unsaturated conditions:

$$h = \frac{u_w}{\gamma_w} \tag{2-10}$$

Hydraulic head, h, and pore-water pressure, u_w, have negative values in unsaturated conditions. Capillary forces can also be expressed in terms of soil suction, ψ

$$\psi = u_a - u_w \tag{2-11}$$

The pore-air pressure, u_a, is usually equal to atmospheric pressure and is therefore omitted.

A porous medium, such as soil, can be compared to a bundle of capillary tubes, with varying and irregular radii, tied together. A concave meniscus extends from grain to grain across each pore channel. The radius of each curvature reflects the pressure difference between the liquid and the gas (Freeze & Cherry, 1979). Forces that hold liquid in a porous medium owing to capillary action are called capillary forces. Capillary forces are approximately inversely proportional to effective pore diameter, R_{eff} (Hillel, 1980), to be expressed as follows:

$$R_{eff} = \frac{2 \cdot \sigma \cdot \cos(\alpha_w)}{\rho_f \cdot g \cdot h} \tag{2-12}$$

R_{eff} represents the radius of a hypothetical bundle of capillary tubes on a macroscale. On a microscale, however, great variations occur caused by variations in pore size. Ward (1975) states that the concept of soil being compared to a bundle of capillary tubes tied together, is totally inadequate in describing soil-water movement in unsaturated soils. He maintains that water movement in unsaturated soils mainly takes place through films of water in the irregularly shaped inter-particle voids.

Adsorption forces

In addition to capillary forces, the adsorption of liquid molecules onto solid particles is also an important force. Surface tension forces occur on the solid-liquid and solid-gas interfaces. The force that attracts a fluid to a solid surface is known as adhesion. Adsorbed liquids are held very tightly to the solid particles and cannot be removed, except by external forces such as evaporation (Hillel, 1980). Water that is adsorbed onto soil grains is called hygroscopic or adsorbed water.

The volume of water that is adsorbed onto soil grains is directly proportional to the specific surface of the soil, which in turn is inversely proportional to the grain size of the soil. Clay minerals have much higher specific surfaces than silt or sand grains, due to their relatively small sizes. Certain clay minerals, especially smectites, possess large adsorption areas because of their ability to incorporate water into their crystal lattices.

Electrical forces at molecular level

Since water is a bipolar molecule, it is attracted to soil grains resulting from the net electrical charges that may occur on the surfaces of soil grains, especially clay minerals. Permanent negative charges occur on the surfaces of clay minerals, caused by isomorphous substitution. Net electrical charges also occur on the edges of clay minerals and on the surfaces of allophane and hydroxides of iron and aluminium, due to their incomplete crystal lattices (White, 1989). This phenomenon is partly responsible for water being held in the soil matrix, particularly of clay soils. It is also partly responsible for the cohesion and plasticity of clay soils.

The charges on a mineral surface can be calculated by measuring the difference in moles of charge contributed, per unit mass, by cations and anions adsorbed from an electrolyte solution at a known pH. The cations and anions adsorbed, are known as the Cation Exchange Capacity (CEC) and Anion Exchange Capacity (AEC), respectively, and are expressed as cmols charge per kilogram. Typical cation exchange capacities of common clay minerals are shown in **Table 2.1**.

Table 2.1 Typical values of some properties of common clay minerals (White, 1989; Holtz & Kovacs, 1981)

	Kaolinite	Illite	Chlorite	Montmorillonite	Vermiculite
Thickness (nm)	50-2000	30	30	3	NA
Diameter (nm)	300-4000	1000	1000	100-1000	NA
Specific surface (km^2/kg)	0.015	0.08	0.08	0.8	NA
CEC $cmol_{(+)} \cdot kg^{-1}$	3-20	10-40	NA	80-120	100-150
Plasticity	Low	Medium	Medium	High	Medium
Swelling/ Shrinkage	Low	Medium	Low	High	NA

NA = Not available

2.2 The vadose zone

The vadose zone is the portion of the geological profile above the groundwater (phreatic) surface. Voids within the profile are usually, but not always, partially filled by liquid and partially by gas. The terms 'unsaturated zone', 'capillary zone' and 'zone of aeration' are frequently used in literature referring to this portion of the soil profile. To understand the concepts affecting moisture distribution and flow through the vadose zone, knowledge of the processes and forces that govern liquid flow through a porous medium is essential.

Figure 2.1 indicates a homogeneous soil profile through the vadose zone, under static conditions. The effect of evapotranspiration has been omitted. The vadose zone can be divided into three sub-zones namely, the *capillary fringe*, *capillary zone* and *discontinuous zone* (Martin & Koerner, 1984a) as indicated in the figure.

The **capillary fringe** is a zone that occurs directly above the groundwater surface. The zone is completely saturated and subjected to negative pressure. The thickness of the capillary fringe, h_d is analogous to the height of capillary rise. The specific diameter of the capillaries is inversely proportional to the effective pore diameter of the soil (Martin & Koerner, 1984a). The effective pore diameter is dependent on the gradation (soil texture), porosity and other factors. The capillary fringe will be thick in fine-grained soils and thin in coarse-grained soils.

The **capillary zone** consists of soil in which the pores are filled with air and water. Matrix forces hold the water in the soil. Water fills the small pores, while air fills the large pores in the soil. As the pore-water pressure decreases with distance above the groundwater surface, the radius of the curved water surface decreases and the water consequently retreats into smaller pores. This leads to a decrease in water content. Fine-grained soils can retain high water contents for considerable distances above the groundwater surface (Martin & Koerner, 1984a).

In the **discontinuous zone**, water is only retained as adsorbed water, since the pore-water pressure is too low to sustain capillary water. Water is strongly adsorbed on each soil particle. This water can be removed by evaporation.

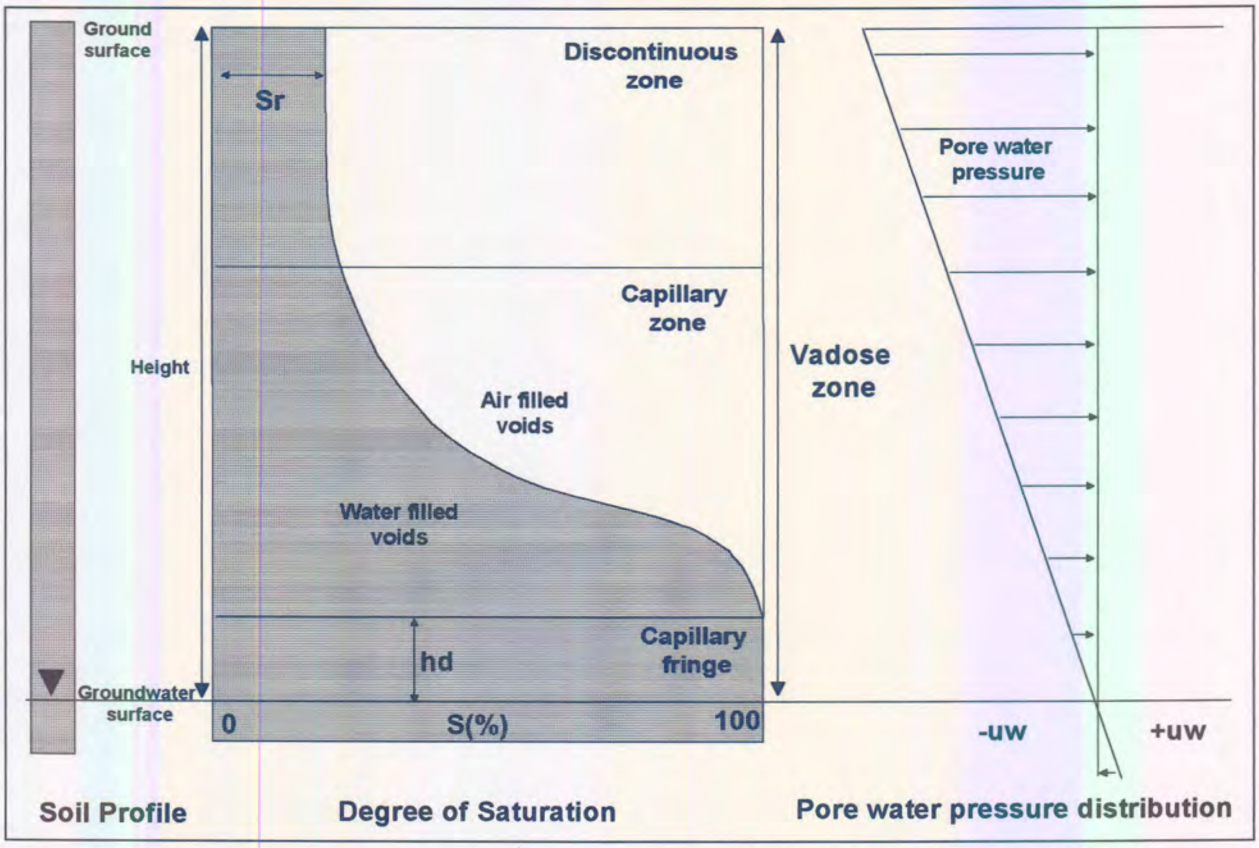

Figure 2.1: The vadose zone indicating soil-water content – pore-water pressure relationship (after Martin and Koerner, 1984a)

Specific retention and storage capacity

Specific retention, θ_r, can be defined as "the volume of water that is retained by a unit volume of soil against the force of gravity during drainage" (Everett, Wilson & Hoylman, 1984). This minimum water content is known as specific retention, field capacity or residual water content and can be determined from soil-water characteristic curves (section 2.3). The concept of specific retention is controversial since it does not exist (Edworthy, 1989). Drainage never really ceases but drainage rates decrease progressively until the drainage rate is practically equal to nil. There is no definite point where the water flow ceases. The extreme variability in unsaturated flow rates, as well as the existence of preferential pathways, considerably complicates the determination of specific retention.

Specific retention is reached in a static situation, i.e. no external factors are considered. However, in field situations evapotranspiration is responsible for a decrease in water content lower than the specific retention value. This zone of water deficiency can reach considerable depths in arid and semiarid environments (Martin & Koerner, 1984a). In the case that water reaches these zones, it will be retained in the soil because of the high sorption of the soil. The maximum volume of water that can be accommodated in the deficiency zone, V_{dz}, also known as the storage capacity of the vadose zone, can be approximated by the following equation (Everett *et al.*, 1984):

$$V_{dz} = (\theta_r - \theta_s) z_{dz} \cdot A \qquad [2\text{-}13]$$

The disposal of hazardous waste in zones of water deficiency seems to be feasible, since leachate will be retained in the soil matrix because of the high sorption of the soil (Martin & Koerner, 1984b; Levin, 1988). However, downward migration of leachate will continue, albeit at a very slow rate. Calculations of storage capacity may be inaccurate because of complications in determining the specific retention value of the soil. The existence of preferential pathways may cause rapid movement of liquid waste and leachate along these pathways.

2.3 Soil-water characteristic curves

When soil suction is slowly increased on a fully saturated non-shrinking soil, either by applying suction on the liquid phase or by exerting pressure on the gas phase, the liquid will begin to retreat below the soil surface. Large pores will be emptied and a curved liquid surface, initially with a large radius, will develop. At low suctions, only large pores are partially filled with gas, while smaller pores are filled with liquid. With increasing suction, the radius of the curved liquid surface becomes smaller in accordance with Laplace's equation, and the liquid retreats into smaller pores. This results in a decrease in water content. Water in large pores is held less tightly and drains more easily.

A similar process is observed in shrinking soils such as clay with a high smectite content. However, under low suction the material will shrink, resulting in a reduction in pore size. Although water will drain from the soil at these pressures, the material will still be completely saturated. Under increasing suction, pores will begin to fill with gas and the process described above will develop. However, in field soils, cycles of shrinkage and heaving will cause cracks to develop. At very low suctions, a portion of the water will drain through cracks, resulting in a decrease in water content. After the water has drained through the cracks, the drainage rate becomes much slower, with a progressively higher suction.

The unique relationship between water content and soil suction for a specific soil is presented by soil-water characteristic curves. **Figure 2.2** shows typical adsorption and desorption soil-water characteristic curves for silty sand. A number of parameters can be defined from these curves, namely the saturated water content, θ_s, residual water content, θ_r, the air-entry or bubbling pressure, ψ_a, and the residual air content, θ_a.

The difference between adsorption (wetting) and desorption (drying) cycles may be attributed to hysteresis. This phenomenon is caused by the 'ink bottle' effect where many pores have narrow connections or entry channels to adjacent pores. This means that, when drying, pores will not drain until suctions are large enough to drain water from the entry channel. Likewise, with wetting, water will not enter a pore until equal suctions are reached. The water content will therefore be lower for wetting cycles at equal suction values. The difference between wetting and drying cycles at saturation can be attributed to air entrapment during the wetting cycle. With time, air bubbles trapped in pores will be released and the water content will be equal to the saturated water content. The drying curve is usually employed in experimental studies. Many researchers feel that the wetting cycle gives a better indication of pore-size distribution than the drying cycle, since the former is governed by the size of the pore while the latter is governed by the size of the entry channel (Ward, 1975). However, because of experimental difficulties involving the wetting cycle, many researchers prefer to use the drying curve in their studies.

Figure 2.3 indicates typical drying soil-water retention curves for sand, silty sand and clay respectively. Soil-water characteristic curves generally have an inverse-S shape and three sections can be identified, namely a gradual slope at low soil suctions, a steeper slope at increasing suctions and again a gradual slope at high soil suctions. The differences between the soil-water characteristic curves of different soils are attributed mainly to differences in pore-size distribution in each soil. The curves are sensitive to changes in soil density and disturbances of soil structures (Miyazaki, 1993).

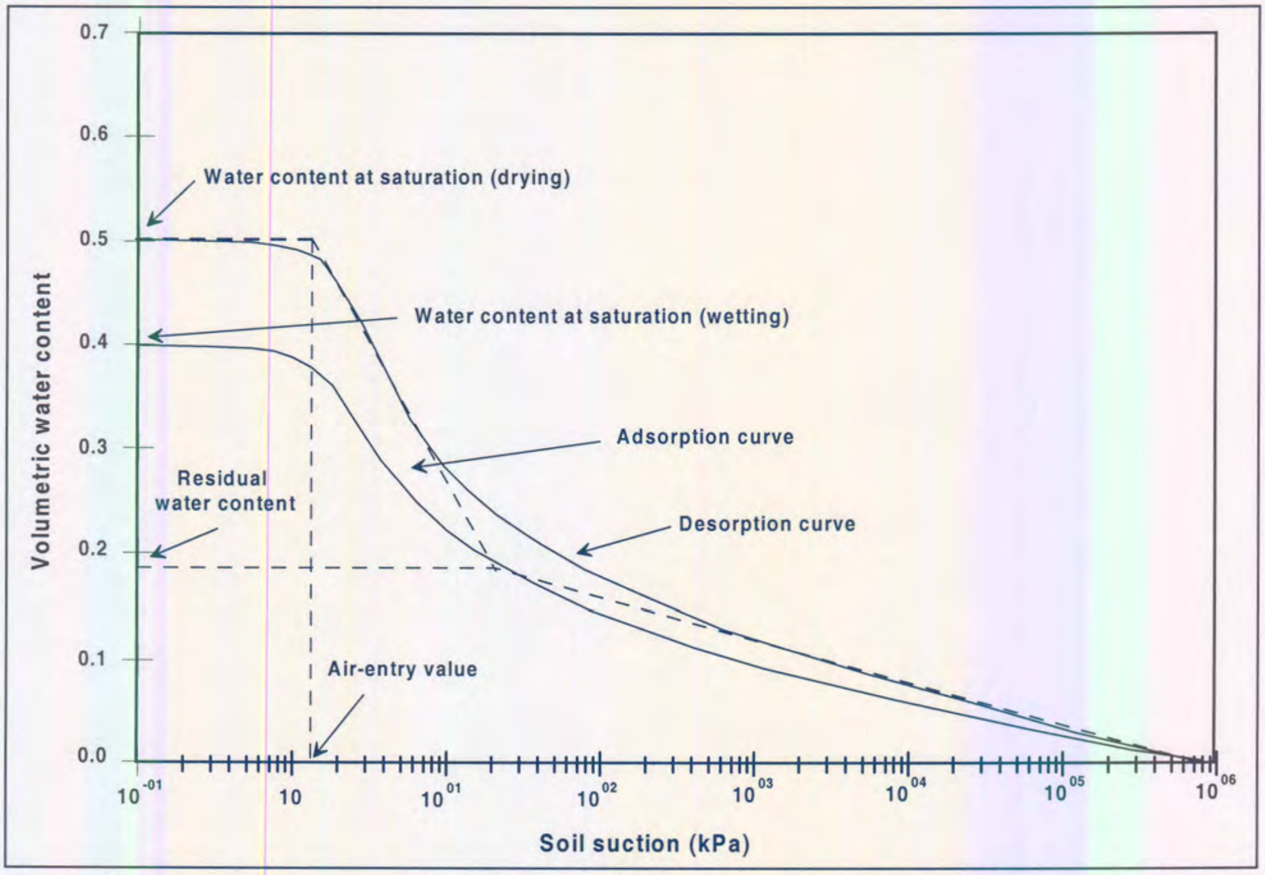

Figure 2.2: Typical soil-water characteristic curve for a silty sand (after Fredlund & Xing, 1994)

Since the soil suction, ψ, is inversely proportional to pore-water pressure, u_w, and hydraulic head, h, and proportional to effective pore-size, R, water content can be related to either of the above-mentioned parameters.

Soil-water characteristic curves are frequently used by the agricultural soil science and geotechnical communities in vadose zone studies (Martin & Koerner, 1984a; Everett *et al.*, 1984; Fredlund & Xing, 1994, Leong & Rahardjo, 1997). Soil-water retention data may be available from agricultural soil science institutions and many soils laboratories are equipped to determine soil-water retention characteristics.

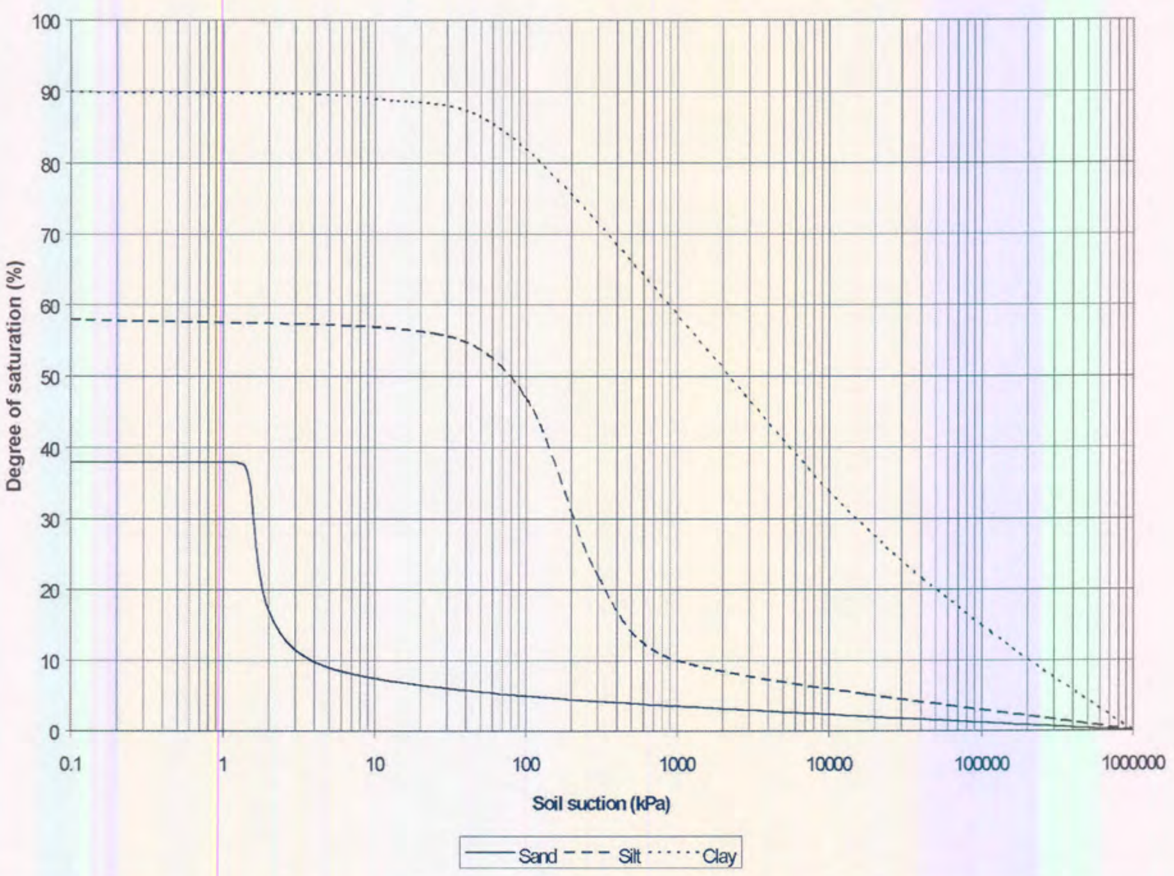

Figure 2.3: Typical soil-water characteristic curves for a sand silt and clay (Modified after Brady, 1974)

2.4 Flow of a liquid in a porous medium

Liquid in a porous medium possesses different forms of energy, largely due to the effects of gravitational and matrix forces. Energy in liquid is generally expressed in terms of a difference in liquid pressure and therefore a difference in the hydraulic head.

The hydraulic gradient, *dh/dl*, represents the loss of hydraulic head between two points, and can be expressed as:

$$\frac{dh}{dl} \text{ or } \frac{dh}{dz} \qquad [2\text{-}14]$$

In this context, the equation represents a one-dimensional vertical situation in which z is measured positive downwards.

If, in a continuous area of liquid, there is a difference in the amount of energy between two points, the flow of liquid will occur in the direction of lesser energy (Bell, 1993). Other things being equal, and on condition that laminar flow exists, the velocity of flow between two points is directly proportional to the difference in hydraulic head between them (Das, 1990).

2.4.1 Steady-state saturated flow

Flow through soil is a complicated process, inhibited by numerous restrictions, bottlenecks, and occasional "dead end" spaces. It is therefore too complicated to describe in microscopic detail, and is rather described in terms of a macroscopic flow vector. The detailed flow pattern is ignored and the conducting body is treated as though it were a uniform medium, with the flow spread out over the entire cross-section, solids and pore space alike (Hillel, 1980).

Darcy's empirical expressions describe steady-state macroscopic flow of fluids through a porous medium. Darcy (1856), while observing the flow of water through sand filters, related the macroscopic velocity (also known as the Darcian velocity), v, to the hydraulic gradient (Hazen, 1930).

$$v = -K\frac{dh}{dz} \qquad [2\text{-}15]$$

The flow of liquid is therefore directly proportional to the driving force acting on the liquid (i.e., the hydraulic gradient) and also to the ability of the conducting medium to transmit the liquid (i.e., the hydraulic conductivity) (Hillel, 1980). The minus or negative symbol is used by convention and indicates that flow occurs in the direction of decreasing hydraulic head.

The total quantity of flow, Q_{tot}, is the volume of water that flows through an area within a given time. It can be expressed as:

$$Q_{tot} = -K\frac{dh}{dz}A \cdot t \qquad [2\text{-}16]$$

The steady-state rate of flow, Q, through an area can be expressed as:

$$Q = -K\frac{dh}{dz}A \qquad [2\text{-}17]$$

Macroscopic velocity, v, is a macroscopic quantity. The conducting body is treated as though it were a uniform medium, with the flow spread out over the entire cross-section, solids and pore space alike (Hillel, 1980). The cross-sectional area of flow, however, includes both solids and voids (Everett *et al.*, 1984). The real area through which flow takes place is therefore smaller than the entire cross-sectional area, suggesting greater actual velocity values. In addition, the length of the liquid body flowing through the soil is much longer because of the tortuous nature of the pore passages, implying even higher actual velocity values. The seepage rate or true velocity of flow, v_s, is the microscopic tortuous flow of the liquid and can be approximated by means of the following equation:

$$v_s = -\frac{K}{\theta_s}\frac{dh}{dz} \qquad [2\text{-}18]$$

Since the water content at saturation is approximately equal to the porosity of the soil, ε, Equation 4-5 can be expressed as:

$$v_s \approx -\frac{K}{\varepsilon}\frac{dh}{dz} \qquad [2\text{-}19]$$

Tortuosity is the average ratio of the actual flow of liquid along pore passages to the physical straight length of the flow path. The value of tortuosity will always be greater than one.

Specific discharge, q, also known as the flux or Darcian flux, is the volume of fluid discharge per unit area per unit time. It has the same value and units as Darcian velocity:

$$q = -K\frac{dh}{dz} \qquad [2\text{-}19]$$

Specific discharge differs from macroscopic velocity in that it does not represent the velocity of the liquid, but rather the discharge rate of the liquid. However, the value of specific discharge is numerically equal to macroscopic velocity.

Limitations of Darcian expressions

Darcian expressions describe the one-dimensional, steady-state flow of an incompressible fluid through a homogeneous porous medium. The flow of fluids in field soils is much more complicated. Certain assumptions have to be made when Darcian expressions are applied (Daniel, 1989):

- The porous medium is homogeneous and isotropic.
- One-dimensional flow takes place.
- The fluid is incompressible.
- A steady-state of flow exists.
- Flow in the porous medium is laminar.
- There is no interaction between the fluid and the porous medium.
- The porous medium remains physically stable.
- Atmospheric pressure and fluid temperature are constant.

In addition to the above-mentioned constraints, the relationship between hydraulic gradient and quantity of flow is not always linear, as many authors have found (Scheidegger, 1957; Childs, 1969; Swartzendruber, 1962; Miller & Low, 1963; Mabula, 1997). At high velocities, the flow may become turbulent and the hydraulic potential may become less effective in inducing flow. At low velocities water in close vicinity to particles becomes rigid and resists flow, probably due to adsorption forces.

Darcian expressions have been validated in both sandy soils and clayey soils by laboratory and *in situ* experiments (Tavenas, Jean, LeBlond & Leroueil, 1983b; Daniel, 1989), indicating that although not all the assumed criteria are met, a reasonably accurate value of hydraulic conductivity can be obtained.

2.4.2 Unsaturated flow

Flow in unsaturated media is more complex than in saturated media. Complex relationships exist between hydraulic conductivity, moisture content and soil suction, and these are further complicated by hysteresis. The basic Darcian principles also apply to unsaturated conditions, i.e. flow takes place in the direction of decreasing potential, the rate of flow is proportional to the hydraulic gradient and is affected by the geometric properties of the pore channels. However, soil moisture in unsaturated soil is subject to pressures lower than atmospheric pressure and the flow is governed by both matrix and gravitational forces.

An important difference between the flow of liquid in saturated and in unsaturated conditions is reflected by hydraulic conductivity values. Under saturated conditions almost all the pores are filled with liquid, except for approximately 5 per cent of the pore volume occupied by entrapped air (Hillel, 1980), and are therefore conductive. As the soil desaturates, some pores are filled with air and the soil becomes less conductive. The larger, more conductive pores drain first, with the result that the liquid flows only in the smaller pores. The water has to circumvent empty pores, resulting in an increase in tortuosity. Transition from saturated to unsaturated conditions usually entails an initial sharp drop in hydraulic conductivity, especially in sand (Hillel, 1980).

The hydraulic conductivity in unsaturated porous media is a non-linear function of volumetric water content, $K(\theta)$. Since the volumetric water content is a function of hydraulic head, $\theta(h)$, (as shown in soil-water characteristic curves), the unsaturated hydraulic conductivity is also a function of hydraulic head. Unsaturated hydraulic conductivity can also be expressed as a function of soil suction, $K(\psi)$, pore-water pressure, $K(u_w)$, and effective pore radius $K(R_{eff})$.

Darcian expressions are valid in unsaturated porous media. Both liquid and gas fill the voids but liquid flow is confined to spaces occupied by liquid.

The rate of flow, Q, in unsaturated porous media can be expressed as:

$$Q = -K(\theta)\frac{dh}{dz}A \qquad [2\text{-}20]$$

where volumetric water content can be substituted by hydraulic head, soil suction, pore-water pressure or effective pore radius.

The cross-sectional area of flow includes solids, liquid and gas. The seepage rate, v_s, in unsaturated conditions can therefore be expressed as:

$$v_s = \frac{K(\theta)}{\theta}\frac{dh}{dz} \qquad [2\text{-}21]$$

(Everett *et al.*, 1984)

2.4.3 Transient flow of water through the vadose zone

Rainfall events are erratic, not only in frequency but also in duration and intensity. These factors have a great effect on the infiltration and redistribution of water flowing through the vadose zone. On the other hand, evaporation and transpiration are less erratic, especially in areas where the climate is characterised by relatively small climatic variations. While rainfall events are time-specific, evaporation and transpiration take place continuously.

Soil-water movement during infiltration

While precipitation or irrigation events take place, water infiltrates the soil. Soil layers close to ground surface will become saturated and downward movement of a zone with higher volumetric water content occurs. The downward movement is caused by both gravitational forces and a suction gradient between the top saturated soil layers and the bottom drier soil layers. Bodman & Colman (1943) identified a number of zones that can be distinguished during infiltration.

The saturated zone at ground surface is approximately one centimetre thick. Immediately below this is the transition zone of a few centimetres in thickness and characterised by a fairly sharp decrease of volumetric water content with depth. The transmission zone occurs below the transition zone and is characterised by little variation in moisture content with depth. The transmission zone, as the name implies, transmits water from the transition zone to the underlying wetting zone. The wetting zone is characterised by a sharp decrease in volumetric water content with depth. The wetting front marks the limit between the wetted soil profile and the underlying dry soil (Ward, 1975). The sharp decrease in volumetric water content implies that a high suction gradient exists in the wetting zone.

With continuing infiltration, the length of the transmission zone increases and little changes in water content take place in the saturated, transition and transmission zones. However, within the wetting zone, the high suction gradient causes an initial rapid movement of water. With time (that implies deeper soil profiles) the suction gradient decreases and the sharp boundary between the wetted and dry soil of the wetting zone becomes less well defined. Ward (1975) argues that as volumetric water content approaches values similar to the transmission zone, the effect of the suction gradient on downward movement of water becomes of lesser importance and the gravitational force becomes the main downward driving force. Since the gravitational gradient has a value of one, downward flux will approximate the value of the saturated hydraulic conductivity.

Infiltration in field soils is more complex than the above-mentioned conceptual processes. Reasons for this include non-uniform soil and water content properties, the effect of hysteresis, boundary conditions that vary with time and the effect of preferential flow paths.

Experiments have shown that the final infiltration rate in field soils is one-half to two-thirds of the saturated hydraulic conductivity (Miyazaki, 1993). This discrepancy can probably be explained by the fact that large volumes of entrapped air occur within the field soil as opposed to laboratory determined saturated hydraulic conductivity, where the sample is usually saturated from the bottom up, thereby minimising entrapped air (Miyazaki, 1993). The difference between the saturated hydraulic conductivity and the final infiltration rate can thus be attributed to entrapped air obstructing the flow of water in field soils

Various equations were derived to describe infiltration of water into soil. These can be classified as empirically and physically based methods. Empirical equations describe the infiltration curve applying two or more fitting parameters. These equations are discussed by Miyazaki (1993).

The Green-Ampt (1911) equation was the first to describe infiltration into soil on the basis of physical methods. The model assumes a piston-like moisture profile and the wetting front is represented by a sharp transition between the saturated soil and the underlying unsaturated soil. The reliability of the Green-Ampt equation relies on the accurate description of the hydraulic head at the wetting front. The vagueness in definition of this parameter has lowered the theoretical reliability of this equation (Miyazaki, 1993). Although the Green-Ampt equation has been derived from the oversimplified assumption of piston-like movement of water through soil, it is still used because of its ability to predict the infiltration rate with no less validity than equations based on the more realistic movement of water through soil.

The equation derived by Phillip (1969) is sometimes used to predict infiltration rates. The method is restricted to particular initial and boundary conditions and to uniform soils. It is very cumbersome and requires knowledge of unsaturated hydraulic conductivity relations in terms of both volumetric water content and soil suction. (Miyazaki, 1993).

Soil-water movement during redistribution

After the infiltration process has ended, infiltrated water continues to move downwards through the soil profile. The downward flow is termed redistribution or the drainage stage. Water, and therefore also the wetting front, continues to move downwards through the soil profile under gravitational and suction forces. However, as water moves downwards, water in the upper zones begins to drain and the transmission zone subsequently becomes a drainage zone (Ward, 1975). The hydraulic conductivity decreases with lower volumetric water content and subsequently, the water (and therefore also the wetting front) moves downwards at a lower rate. The hydraulic gradient in this zone is slightly higher than one, while the hydraulic gradient at the wetting front is much higher (Miyazaki, 1993). A large quantity of downward flow thus takes place around the wetting front, while less downward flow occurs in the wetted front. At volumetric water content values close to field capacity, very little downward movement of water is taking place. The permeability of the soil and the initial volumetric water content are two important factors which affect redistribution. Wierenga (1995) has shown that soils with a higher initial water content allow recharge to take place much more quickly than drier soils in similar conditions.

Soil water movement during evapotranspiration

After infiltration has ended, evaporation takes place and results in the removal of water from the surface layers. Evaporation and redistribution processes occur simultaneously, and while redistribution will result in a lower rate of downward movement of water, evaporation causes an upward movement of water due to a suction gradient between the drier surface layers and the wetter deeper soil layers. Evaporation is affected by temperature, relative humidity and wind velocity, as well as hydrogeological properties such as the unsaturated hydraulic conductivity and the hydraulic gradient. Three distinct patterns for rate of evaporation can be identified, depending on the evaporation potential of the soil. These include:

i A constant-rate stage where the evaporation rate does not change with time,

ii The first falling-rate stage where evaporation rate decreases linearly with time and

iii A second falling-rate stage where the evaporation rate decreases exponentially with time

(Miyazaki, 1993).

It is difficult to estimate rates of evaporation and transpiration in the vadose zone due to the many factors affecting evaporation rate and the difficulties in measuring these factors. Water movement during redistribution and evapotranspiration is graphically represented in **Figure 2.4**.

The value of the hydraulic gradient in the vadose zone

Flow of water within the vadose zone is caused mainly by rainfall events and evapotranspiration. These events cause a positive (downward flow) and negative (upward flow) hydraulic gradient respectively within the vadose zone. Miyazaki (1993) states that large positive hydraulic gradients only occur at the wetting front while the hydraulic gradient in the wetted zone is almost equal to one. Miyazaki (1993) maintains that in the case of gravity-dominant flow in deep soils (not affected by evapotranspiration), hydraulic gradients tend to equal one. According to Wierenga (1995) hydraulic gradients in moist soils is almost equal one and do not change much. Based on approximations over long periods of time, Unlu, Kemblowski, Parker, Stevens, Chong & Kamil (1992) suggest that the value of the hydraulic gradient can be assumed to be one, since conditions of vertical flow prevail.

In contrast, Miyazaki (1993) states that these conditions may only prevail in regions where the annual precipitation exceeds annual evaporation. In semi-arid and arid regions or regions characterised by distinct wet and dry seasons, evaporation can cause the surface soil layers to dry out, causing a negative hydraulic gradient. In addition, the effect of ponding must be considered. Ponding can result in relatively large positive hydraulic gradients. Perched groundwater levels may also result in relatively large hydraulic gradients. Further investigations should be conducted on the effect of hydraulic gradients within the vadose zone under South African conditions.

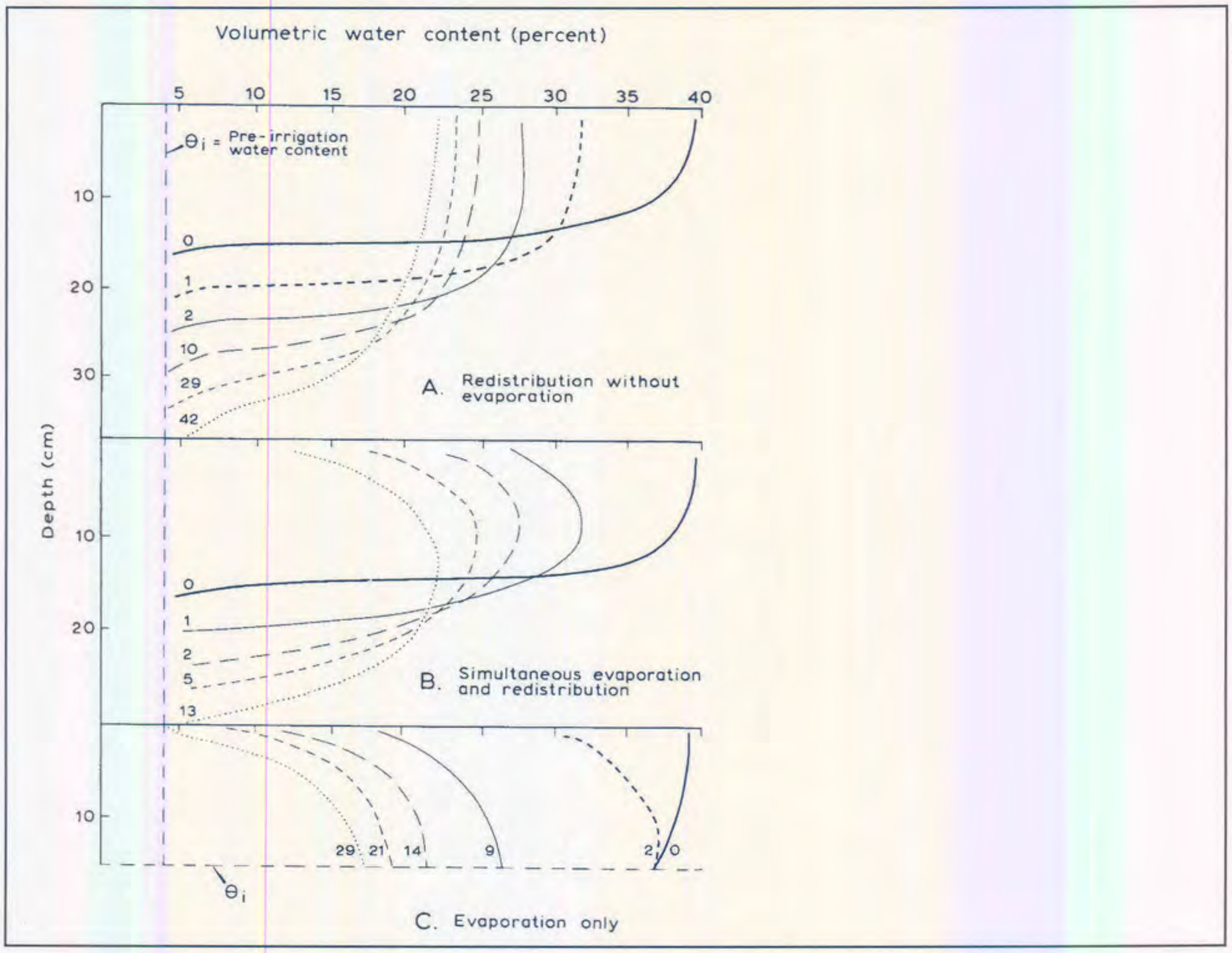

Figure 2.4: Soil-water content profiles during redistribution and evapotranspiration (numbers indicate days after wetting) (after Ward, 1975)

Upward movement of soil-water

The effect of evaporation and transpiration causes a suction gradient to develop between the lower wetter zones and the upper dried-up zones. The suction forces cause upward movement of soil-water up to the root zone from where it is released to the atmosphere by evaporation and transpiration processes. The rate at which upward movement of soil-water occurs, is controlled by various factors, the most important being the permeability of soil, the rate of drying of the surface layers, the density of the plant root system and the depth to the groundwater level. Penman (1948) has shown that in the case of slowly drying soil surfaces, the rate of upward movement of soil-water will keep pace with the rate at which soil-water is removed from soil surfaces by evaporation processes. However, in the case where drying occurs rapidly, upward movement of soil-water will be constrained by the reduced hydraulic conductivity at the dry (hence less permeable) surface layers. In extreme cases, evaporation losses from soils may become negligible even though soil-water contents at a few centimetres depth are still high. The depth to the groundwater surface is a major factor in the upward movement of soil-water. Wind (1961) has shown that the rate of upward movement of soil water is mainly a factor of groundwater level rather than the suction imposed at the soil surface. High rates of water movement will therefore occur where the groundwater is shallow (<60cm) This aspect is more pronounced in coarse-grained soils than in fine-grained soils.

Wind (1961) also shows that the rate of upward water movement is sensitive to the juxtaposition of different horizons in the soil profile. In wet conditions, the hydraulic conductivity of clay soils is low while in drier conditions, the hydraulic conductivity of clay soils can be higher than sandy soils. As such, sandy soil overlying a clayey soil will result in low rates of upward soil-water movement. According to Wind (1961), the highest rate of upward movement of soil water will occur in a soil profile comprising of a lower sandy soil horizon overlain by a silty soil horizon with a clay horizon at the surface.

In conclusion, it can be stated that in humid areas, where rainfall exceeds evaporation, downward movement of soil-water will dominate. This results in the leaching of soluble minerals from the top soil layers and the accumulation of these minerals in lower soil profiles. These processes can eventually result in the development of iron and aluminium-rich horizons and even the development of hardpan ferricrete.

In dry areas, where evaporation exceeds rainfall, upward movement of water will dominate. This can cause upward transport of soluble minerals and the eventual development of typically lime-rich soil horizons. These processes can lead to the accumulation of salts on the surface layers.

2.4.4 Numerical simulation modelling of unsaturated flow

Simulation can be defined as the application of models as a substitute for the study of real or hypothesised systems. These models may comprise either physical or numerical models. With advances in the processing capabilities of personal computers, most researchers have focussed on the development of easy to construct and low-cost numerical models. Simulation modelling is a powerful tool for understanding a particular problem and in some instances to understand what the problem is. Time can either be stretched or compressed for processes that happen too fast for observation or too slowly for practical observation. Ultimately, numerical models can be used as a prediction and decision-making tools. A number of "what if…" scenarios can be simulated and the relevant designs can be optimised for the particular problem.

Since flow in the vadose zone is dynamic and transient in nature, numerical modelling by means of computer models offers the best tool to understand, simulate and ultimately predict flow of water and solutes through the vadose zone. Since flow in the vadose zone is predominantly downward (or upward), it can be presented by a one-dimensional model. Transient flow in the vadose zone can be described by the Richard's equation, which is based on Darcy's law (modified from Feddes, Kowalik & Zaradny, 1978):

$$\frac{\partial \psi}{\partial t} = \frac{1}{\theta'(\psi)} \cdot \frac{\partial}{\partial z}\left[K(\psi)\left(\frac{\partial \psi}{\partial z}\right)+1\right] - \frac{S(\psi)}{\theta'(\psi)} \qquad [2\text{-}22]$$

where $S(\psi)$ is a sink term and represents the volume of water taken up by plant roots (volume water per volume soil per unit time). Water uptake by plant roots does not constitute part of the research and as such, readers are referred to Feddes *et al* (1978).

A boundary condition is defined at the top of the system where water infiltrates the system by means of precipitation or irrigation and where water leaves the system by means of evapotranspiration. Evapotranspiration can be directly measured or can be calculated by means of a number of equations of which the Penman (1948) equation is the most popular. Rainfall is measured directly. Another boundary condition is defined at the bottom of the system and this will depend on the particular situation.

A number of numerical solution schemes can be used to simulate flow of water in the vadose zone. The finite-difference technique is the most frequently used technique to solve the extended Richard's equation. Other techniques, including the finite element method, are also used to simulate flow of water through the vadose zone. Wates, Rykaart, Vermaak & Bezuidenhout (1999) identified forty-five models capable of handling unsaturated flow conditions. The most popular models include HELP, FEMWATER, LANDFIL, SEEP/W and ACRU. However, many of these models are analytical models and can not simulate dynamic conditions. The conceptual unsaturated flow model is shown in **Figure 2.5**.

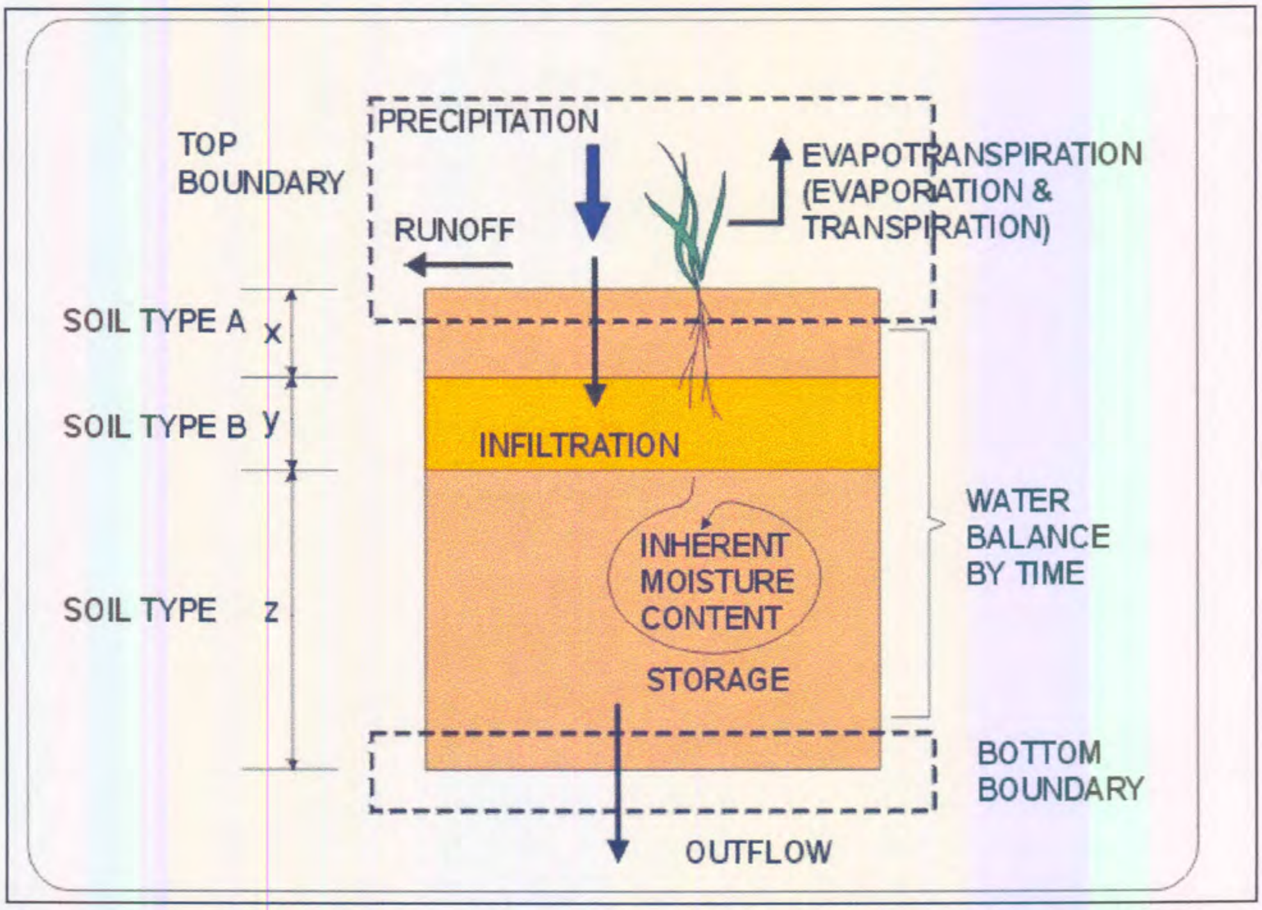

Figure 2.5 Conceptual unsaturated flow model for a layered soil (after Wates *et al*, 1999)

2.4.5 Saturated hydraulic conductivity

Saturated hydraulic conductivity represents the factors, other than the hydraulic gradient, which affect the flow of liquid through a porous medium. Three important factors have to be considered:

- Properties of the fluid
- Properties of the porous medium (soil or rock)
- Interaction between the fluid and the porous medium

Interaction between the fluid and the porous medium generally refers to electric forces at molecular level. These forces may be an important factor in saturated hydraulic conductivity, especially in clayey soils. However, in sandy soils, these forces do not generally have a significant effect on saturated hydraulic conductivity. In these instances, saturated hydraulic conductivity, K, can be described in terms of the properties of the fluid medium, known as fluidity, f, and in terms of properties of the porous medium, known as intrinsic hydraulic conductivity or permeability, k:

$$K = kf \qquad [2\text{-}23]$$

Fluidity is inversely proportional to viscosity:

$$f = \frac{\rho_f g}{\eta} \qquad [2\text{-}24]$$

The density of most liquids is nearly constant, although it may vary somewhat due to the effect of temperature and solute concentration, but the viscosity value varies due to the effect of temperature.

Equation 2-24 indicates that saturated hydraulic conductivity will vary in accordance with the type of liquid. The petroleum industry uses Equation 2-24 to differentiate between the velocity of flow of oil, water and methane gas.

Intrinsic permeability represents all factors associated with the porous medium which affects saturated hydraulic conductivity. Intrinsic permeability is a factor of grain-size distribution, structure, density and other properties of the soil. It depends primarily on the size of the conducting pores as well as on the porosity of the soil.

In a case where the porous medium consists of a bundle of straight tubes with equal and constant diameters, the intrinsic permeability can be calculated from Equation 2-25 that is based on Poiseuille's law:

$$k = 0.125 \cdot R^2 \qquad [2\text{-}25]$$

It is much more difficult to determine the intrinsic permeability for soils and rocks because of the irregular geometry of the pores and because pore-size distributions depend on grain-size distribution, packing, the shape of the grains and various other factors.

The effect of electro-chemical forces at molecular level

Electro-chemical forces have already been discussed (Section 2.1.2) with regard to their effect on retaining fluids in unsaturated soil. These forces play an important role in flow through soils with active particle surfaces, such as clays. The flow of fluids through these soil types will strongly depend on the electro-chemical properties of the fluid.

Fernandez and Quigley (1985) found that the hydraulic conductivity of clayey soils was strongly influenced by the physico-chemical properties of liquid hydrocarbons. They ascribed this to the electro-chemical reactions between the liquid and the soil media, especially clay minerals.

Fernandez and Quigley (1985) found that liquids with a relatively high dielectric constant, such as water, inhibited flow through the clay. The hydraulic conductivity of water (with a dielectric constant of 80) flowing through a clayey soil (with a void ratio of 0.8) was measured at 5×10^{-9} $m \cdot s^{-1}$. The hydraulic conductivity of liquid hydrocarbons (with a dielectric constant of 2) flowing through the same soil was measured at 1×10^{-4} $m \cdot s^{-1}$.

Equation 2-23 does not apply to these conditions, because the effect of the dielectric constant on the double clay layers completely swamps the effects of fluid viscosity and density.

Fernandez and Quigley (1985) found no changes in hydraulic conductivity when water-insoluble liquids with a low dielectric constant were used in a water-wet compacted clayey soil sample. These hydrophobic liquids were probably forced through micro-channels or macropores, displacing only 10% of the pore water from the samples.

The use of water-soluble hydrocarbons in water-wet compacted clayey soil samples resulted in an up to ten-fold increase in hydraulic conductivity and the extensive removal of pore water, while permeation with liquid aromatics with a very low dielectric constant resulted in a thousand-fold increase in hydraulic conductivity.

Interaction between the liquid and porous medium can have a significant effect in cases where leachate is moving through clay.

2.5 Heterogeneity and the effect of scale

Heterogeneity of soils is defined in terms of the spatial variability in physical soil properties, such as bulk density, water content, grain-size distribution, pore-size distribution, consistency and other properties. Since all these properties are defined and measured with respect to an elementary volume, the definition of soil heterogeneity is expressed in terms of an elementary volume.

In general, the larger the elementary volume, the more physical properties are averaged in each elementary volume. Therefore, more homogeneous properties of a specific field soil are defined. On the other hand, the smaller the elementary volumes in the same field, the larger the differences in physical properties between the elementary volumes, resulting in defining more heterogeneous properties of a specific soil.

Figure 2.6 shows the value of a specific soil property within a number of elementary volumes with increasing sizes. It can be seen that the variability of the soil property decreases as elementary volumes increase. The elementary volume in which little variability exists, is known as the representative elementary volume (REV) (Bear, 1979). The larger the REV, the larger the heterogeneity of the soil within a specific field soil type.

Although the general theory for establishing REV has not yet been established, various authors have determined empirical REVs for various purposes. These tentatively acceptable empirical REV values (designated by representative lengths) are expressed in **Table 2.2**:

Table 2.2: Empirical REV designated for various purposes

REV (m)	Subject	Reference
REV > 1000	Hydrological modelling in a river basin	Wood, Sivapalan, Beven & Band (1988)
REV > 5	Water balance in a field with cracks	Inoue, Hasegawa & Miyazaki (1988)
REV > 0.5	Saturated hydraulic conductivity of a soil with macropores	Lauren, Wagnent, Bouma & Wosten (1988)
REV > 0.05	Bulk density, water content and solute concentration	Tokunaga and Sato (1975)
REV > 0.01	Microstructure	Cogels (1983)

The REV concept assumes statistical correlation of an intrinsic property independent of scale. As such, it assumes the porous medium to be composed of homogeneous material above some critical scale (Tyler and Wheatcraft, 1990). Although the REV concepts can be and have been applied at small to medium scale (millimetres to metres) in hydrogeological investigations, such correlations are limited at large field scale. Many intrinsic properties can be correlated in space due to differences in geology and soil formation processes. Serious errors could result in the averaging process if the REV approach is applied without consideration of the geology and soil formation processes. The spatial distribution of soils with similar physical and hydraulic properties can be identified, presented and quantified by means of their geological origin and topographical setting. These aspects are discussed under **Section 2.8**.

Miller & Miller (1956) applied the theory of similitude analyses to unsaturated hydraulic conductivity and hydraulic head. Two porous media can be described as Miller-similar if a defined scale factor can describe these media exactly relative to each other (Warrick, 1990, Sposito & Jury, 1990). Similar media exhibit microscopic structures that are similar in shape, but occur on different scales. According to Miller & Miller (1956), the unsaturated hydraulic conductivity – hydraulic head relationship of two similar media can be related by the equation:

$$\frac{K'(\lambda' h')}{\lambda'^2} = \frac{K(\lambda h)}{\lambda^2} \quad [2\text{-}26]$$

where $K'(h')$ and $K(h)$ denotes the unsaturated hydraulic conductivity – hydraulic head relationship of two points within a specific field soil and λ is the characteristic length that relates the unsaturated hydraulic conductivities.

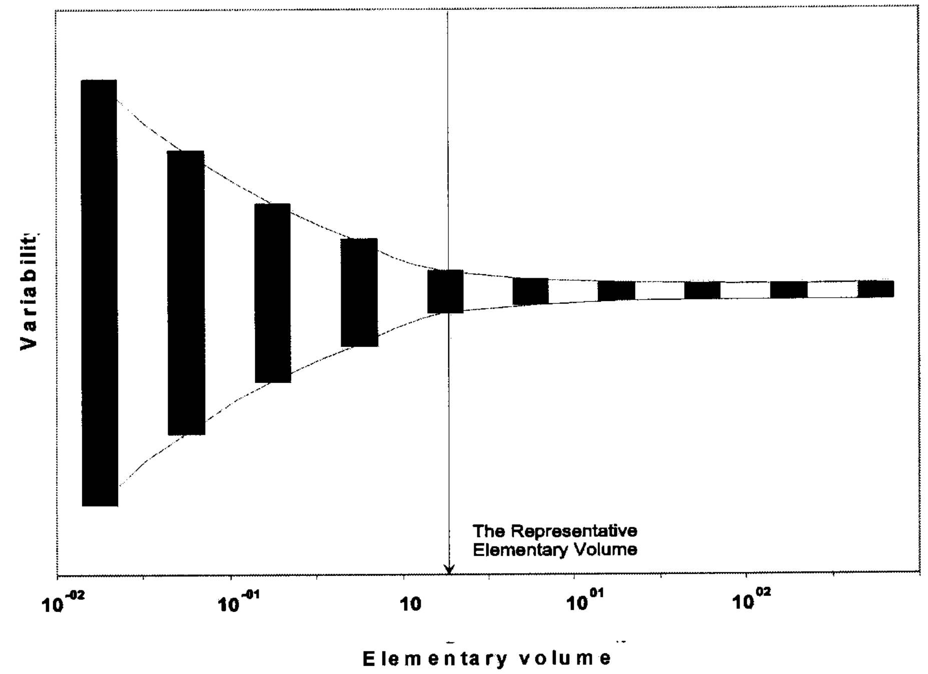

Figure 2.6: The Representative Elementary Volume (modified after Bear, 1979)

Similitude analyses have successfully been applied in describing unsaturated flow through heterogeneous field soils (Sposito & Jury, 1990; Youngs, 1990) using the modified Richard's equation. It has also been successfully applied to characterise the spatial variability of field soils. Warrick (1990) found that by applying scaling techniques in heterogeneous soils, the best fit could be applied to the soils with considerably more accuracy. Warrick recorded up to 80 per cent reduction in the Sum of Squares for scaled fitted data points compared to the unscaled fitted data. The results are indicated in **Figure 2.7**.

The basic principles and applications of scaling techniques are described by Hillel & Elrick (1990).

2.6 Hydrogeological properties important in the assessment of saturated and unsaturated flow through the vadose zone

The flow situation in the vadose zone varies according to climatic events, i.e. rainfall and evapotranspiration, as well as irrigation and other human activities. Since these events are generally erratic, flow in the vadose zone will be transient in nature and vary accordingly. Climatic events result in either the addition or removal of soil-water from the soil. All properties that are functions of soil-water, i.e., soil-suction gradient, soil-water retention characteristics and the unsaturated hydraulic conductivity, will vary accordingly.

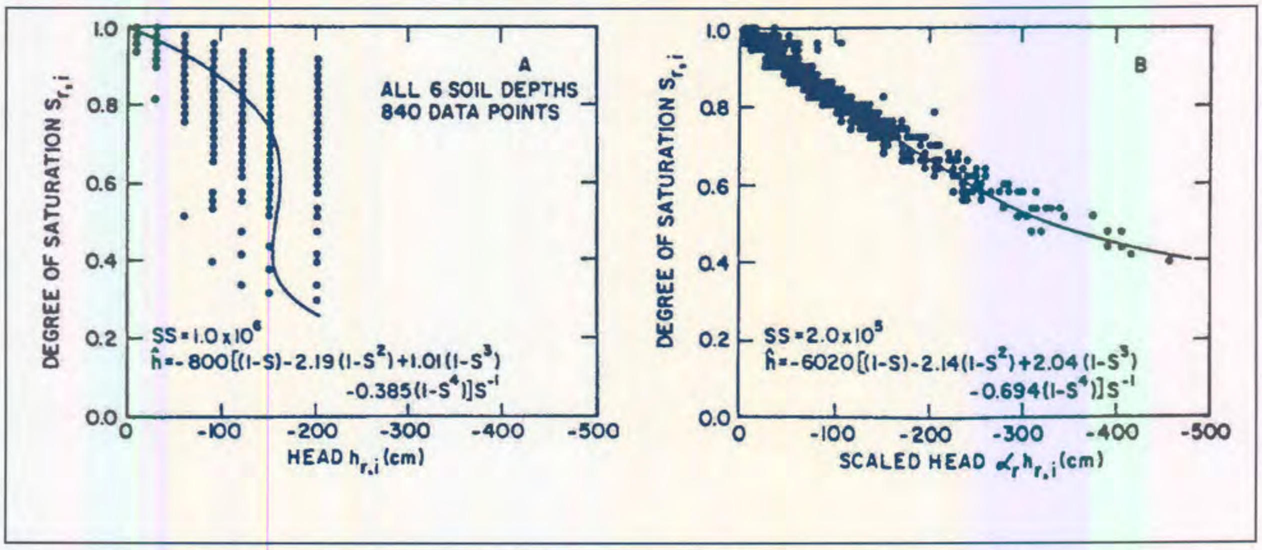

Figure 2.7: Soil-water retention data for fine loam: unscaled (A) and scaled (B) (after Warrick *et al.*, 1977)

The hydrogeological properties of importance in the assessment of saturated and unsaturated flow through the vadose zone, are summarised in **Table 2.3**. The properties described in **Table 2.3** refer to Darcian type flow mechanisms as described by the Richard's equation. These properties may not be relevant in the case of other flow mechanisms such as preferential flow.

Table 2.3: Hydrogeological properties affecting groundwater recharge and contamination

Property	Symbol	Application
Saturated hydraulic conductivity	K_s	Estimates of saturated and unsaturated flow
Hydraulic gradient	$\frac{dh}{dz}$	Estimates of saturated and unsaturated flow
Soil suction gradient	$\frac{d\psi}{dz}, \frac{d\theta}{dz}$	Estimates of saturated and unsaturated flow
Depth to groundwater level	D	Estimates of time before reaching the groundwater level
Volumetric water content	θ	Estimates of unsaturated flow
Soil suction/hydraulic head/ effective pore size	$\psi/h/R_{eff}$	Estimates of unsaturated flow
Soil-water /soil suction relationship	$\theta(\psi)$	Estimates of unsaturated flow and storage capacity
Specific retention	θ_r	Estimates of unsaturated flow and storage capacity
Saturated volumetric water content	θ_s	Estimates of unsaturated flow
Unsaturated hydraulic conductivity	$K(\theta)$, $K(\psi)$	Estimates of unsaturated flow
Dry density, porosity and void ratio	ρ_d, ε, e	Estimates of saturated and unsaturated flow

2.6.1 Characteristics of water movement through geological units

The geological framework will dictate the flow of water and contaminants through the vadose zone. According to Kramer and Keller (1995), the geological profile containing an aquifer may comprise materials hosting primary flow and materials hosting secondary flow. Primary flow can be compared to matrix flow where water and contaminants flow through the soil matrix in accordance with Darcy's law. Secondary flow refers to water and contaminants preferentially flowing along discontinuities in the soil or rock mass.

Matrix flow will be hosted mainly in unconsolidated materials. These include deeply weathered residual material, alluvium and other transported material and many Quaternary deposits. Soils containing high, active clays where preferential flow along cracks may take place, can be exceptions. Other exceptions may be well-cemented geological formations and pedogenic materials that may be impermeable, except for occasional discontinuities in which preferential flow will take place.

Secondary flow is mainly hosted in rock masses. These include most igneous, sedimentary and metamorphic rocks of pre-Quaternary age. Discontinuities such as joints, fault zones, geological contacts, bedding planes, solution cavities and lava tubes occur naturally within the rock masses. Underground mining, oil and water extraction and other processes can cause other discontinuities to develop within the rock mass. Secondary flow may be more rapid in comparison to matrix flow and, in the case of solution cavities and lava tubes, may reach flow velocities of several metres per second.

2.7 Preferential flow

Preferential flow is the process by which water and solutes move along preferred pathways through a porous medium (Helling & Gish, 1991). In the case of flow through the vadose zone, water and solutes by-pass large parts of the soil matrix, conventional convection equations may not be valid and water flow may be faster (or slower) than anticipated. In addition, monitoring devices, such as suction lysimeters, may be located outside (or inside) preferential flow paths and may not yield representative readings.

In many cases preferential flow plays an important role in groundwater recharge and contamination. This may be the predominant factor in groundwater recharge, as shown by Van Tonder and Kirchner (1990) and Kirchner, Van Tonder & Lukas (1991).

Flury, Flürer, Jury and Leuenberger (1994) conducted field experiments on various soils in Switzerland to study the effect of preferential flow. A dye tracer was irrigated on the soil and the pathways of the tracer were described. Flury *et al.* stated that preferential flow was "the rule rather than the exception" and that in most soils, water by-passed a portion of the soil matrix. The extent of by-passing, differed but Flury *et al.* (1994) found that it was likely that during heavy rainstorms, water by-passed the soil matrix in most arable Swiss soils. Flury *et al.* (1994) also found that the (coloured) water penetrated much deeper into structured clayey soils than into non-structured soils. They warned that well-structured clayey soils might be susceptible to pesticide leaching.

2.7.1 Classification of preferential flow

The following types of preferential flow have been identified:

- Macropore channelling that includes
  - Channelling through bio-pores such as relict root structures and burrowing animals
  - Channelling through fractured soil
  - Channelling through relict rock structures
- Other preferential flow types such as
  - Fingering flow

- Funnelled flow
- Preferential flow due to water repellent soil
- Preferential flow along protruding structures in the ground

Preferential flow mechanisms are presented in **Figure 2.8**

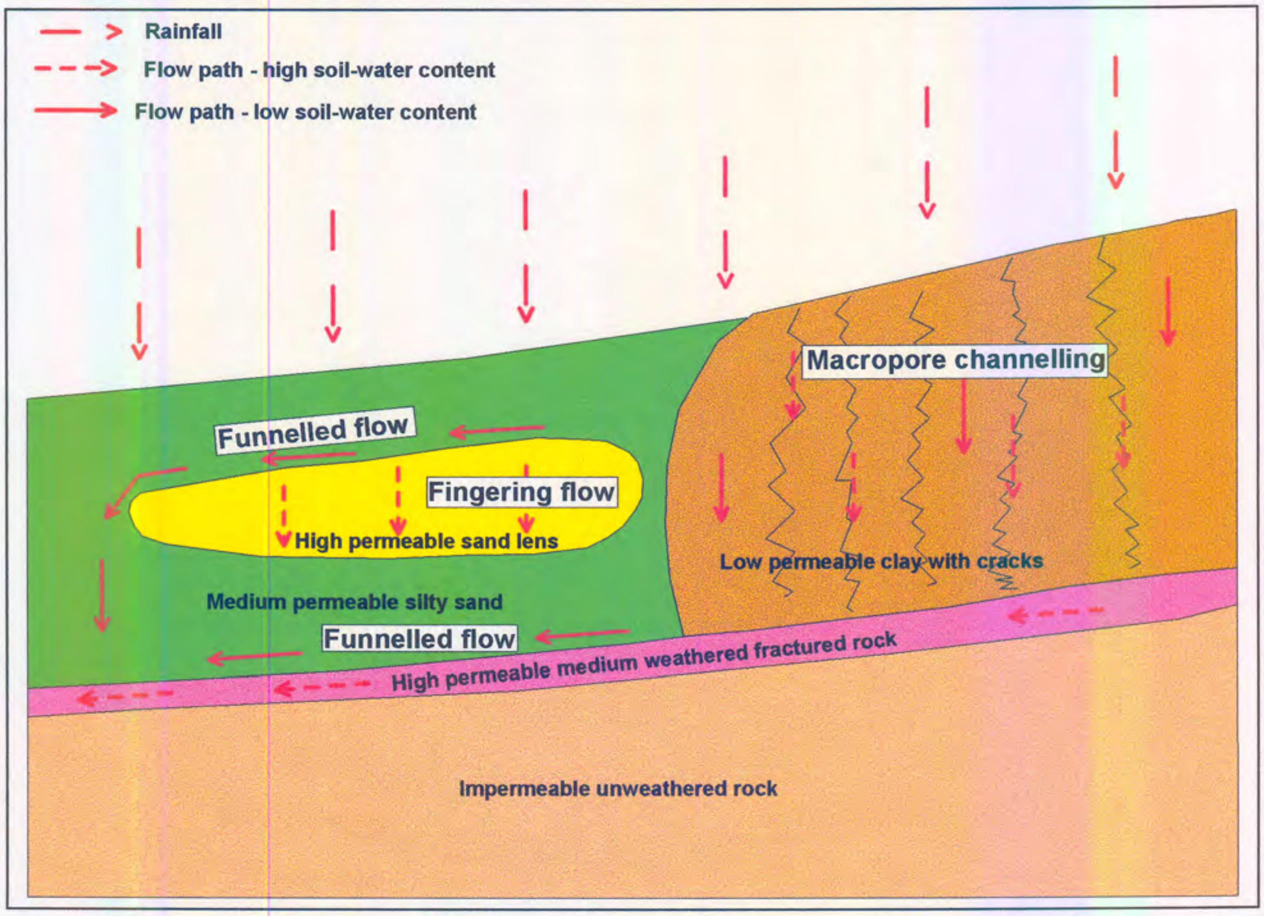

Figure 2.8: Conceptual preferential mechanisms indicating macropore channelling, funnelled flow and fingering flow

2.7.2 Macropore channelling

Macropore channelling refers to a liquid that by-passes the soil matrix via macropores. It can cause rapid water movement through the soil, with the result that the water by-passes the soil matrix and increases the flux value, resulting in a higher recharge rate. Pollutants may be transported through macropores in a very short travel time before the aquifer is reached. This may decrease the attenuation capacity of the vadose zone. Very little interaction between the fluid and soil matrix can also result in a decrease in attenuation, since attenuation processes such as dispersion and cation exchange are by-passed. Macropore channelling may be a very important factor in the flow process of water through soil.

Van Tonder and Kirchner (1990) conducted a recharge study of Karoo aquifers in South Africa. Data on both the phreatic and vadose zones were collected. A major flood occurred during the end of the study period and provided the opportunity to study recharge mechanisms in this semi-arid environment. No measurable matrix flow occurred in the flat-lying areas. However, much recharge took place, as was evident from higher groundwater levels. This lead to the conclusion that recharge occurred through preferential pathways, probably through cracks in clay layers. Van Tonder and Kirchner stated that water could move downwards until the heavy clay layer is reached, after which water movement is predominantly horizontal. Water then moves down natural cracks occurring at specific locations. Recharge estimates, applying common models of water flow through unsaturated soils, do not apply in this situation.

2.7.3 Types of macropores

Macropores refer to openings in the soil which are larger than the pores occurring in the soil matrix. These voids are readily visible and may be continuous for several metres, both in vertical and horizontal directions. Macropores are generally classified according to their morphology and origin of the pores.

Fissures and fractures are generally caused by shrinking/heaving clays, chemical weathering and freeze/thaw cycles (Beven & Germann, 1982). The extent of shrinkage/heave can be predicted by determination of the activity of clay soils. A high plasticity index and clay fraction increases the activity of the soil. Cracks formed in shrinking and heaving soils depend on moisture content changes and may vary seasonally.

Processes causing the formation of discontinuities in clay are dynamic. In the case of desiccation, cracks appear during dehydration of the soils caused by evaporation processes. These cracks will close during hydration processes caused by precipitation events.

Formation of macropores by soil fauna and flora

Pores caused by soil fauna are normally tubular in shape with diameters of 1 – 500 mm (Beven & Germann, 1982). Various animal species, including insects (ants and termites), earthworms, moles, rodents and aardvarks, are responsible for burrowing holes. Most pores caused by soil fauna are close to the ground surface (up to one metre down), except for termite holes, which are known to extend down to the groundwater level. The types of species that occur in the soil depend on the properties of the soil and climate. Insects are found in acidic soils, while earthworms prefer low-acidic to neutral soils. Pores caused by soil flora are also generally tubular in form. These pores are generally caused by plant roots, alive or decayed (Beven & Germann, 1982). The extent and depth of the macropore network caused by soil flora are related to the plant species contingent to the climate of the area.

2.7.4 Other preferential flow mechanisms

Fingering flow

Fingering flow refers to the process where water or solutes move down the soil matrix in columnar structures, approaching velocities of saturated flow. Fingering is caused where the wetting front becomes unstable. This may occur for a number of reasons, including a change in hydraulic conductivity with depth and the compression of air ahead of the wetting front (Steenhuis & Perlange, 1990).

Fingering causes an increase in flow velocity within the more permeable zone. The lower travel time results in less contact time for attenuation processes to take place. In addition, large parts of the soil matrix are by-passed. Since fingering flow tends to be confined to certain areas in the soil profile, attenuation processes are also confined to these areas and the attenuation capacity of the soil will therefore be lower compared to soil where no fingering flow occurs.

Water flowing through a less permeable layer into a more permeable layer may cause wetting front instability. Water in the less permeable zone cannot enter the larger pores of the more permeable zone, due to a difference in pore water pressure. Water will not enter the larger pore as long as the suction at the bottom of the upper pore is greater than the suction of the lower pore. When the pore suction of the top layers finally exceeds the suction of the pore layers at the bottom, water moves down into the coarse-grained layer. Because of the reduced flux in the coarse-grained layer (the flux cannot exceed that in the less permeable layer on top), the effect of gravity offsets the effect of the surface tension of the liquid. This causes wetting front instability. The wetting front splits and fingering takes place.

Wetting front instabilities may also occur due to low intensity rainfall which is much lower than is necessary for ponding to develop. Fingering flow can occur when soils are very permeable (coarse sand or gravel) or if soils are water-repellent.

Funnelled flow

Funnelled flow refers to the preferential flow of the water being 'funnelled' to flow laterally on top of coarse-grained soil layers (Kung, 1990). This phenomenon occurs because water in the less permeable zone cannot enter the larger pores of the more permeable zone because of a difference in pore water pressure. Water will not enter the larger pore as long as the pore suction at the bottom of the upper layer is greater than the pore suction at the top of the lower layer. Since vertical flow is impossible, the water will move laterally.

Funnelled flow also refers to the funnelling of water laterally on top of a less permeable zone. In this instance, the hydraulic conductivity of the underlying less permeable layer is too low to accept incoming water.

The funnelled water eventually moves downward in concentrated columnar flows when it reaches the edges of the coarse-grained (or fine-grained) layer. If the water content within the funnelled area is high enough, fingering within the coarse-grained soil may result.

Funnelling causes the decrease of the total flow area. Since most of the flow is concentrated in columnar structures, very little or no flow occurs in adjacent areas. Kung found that funnelled flow might be restricted to one-tenth of the total flow area. The subsequent increase in water content also causes an increase in flow velocity. Kung found that the steady-state flow velocity in the columnar structures increased 100 times.

2.8 Important aspects regarding unsaturated flow within field soils

The vadose zone represents the top portion of the geological profile. This zone is subjected to weathering, erosion, pedogenic and other processes, often resulting in a complex geological setting. The soils and rocks in the vadose zone are very rarely homogeneous and the situation presented in **Figure 2.1** is therefore rarely a reflection of the situation in the field.

The top portion of the geological profile can consist of thick layers of transported material, unweathered to completely weathered *in situ* material, poorly to well developed pedogenic soils with clearly defined soil layers, poorly to well developed pedocrete layers or a combination of the above-mentioned and other materials. A range of materials with a variety of physical properties may occur within this portion and significantly complicate studies of the vadose zone. The geological setting of the vadose zone has to be thoroughly understood before hydrogeological studies can be conducted.

The hydrogeological characteristics of specific geological materials are discussed in **Chapter 3**.

2.8.1 Shallow weathered and perched aquifers

Shallow weathered aquifers generally occur in a weathering profile overlying hard rock types. The groundwater is mainly stored within primary voids in the weathered zone. The depth of the shallow weathered aquifer may vary considerably depending on the climate, geological conditions and rock type.

Weathered zones in more humid areas could be more than 100m in thickness while thicknesses of less than 10m could be expected in dryer parts of South Africa. Shallow groundwater is recharged mainly by rainfall and influent (losing) streams. Groundwater stored within the weathered zone is, in most cases, hydraulically connected to deeper fractured aquifer systems. Shallow groundwater generally acts as storage for deeper fractured aquifer systems and provides base flow for streams and rivers.

Perched aquifers are defined as independent and isolated areas of groundwater situated above the groundwater level and separated from it by unsaturated soils/rocks; i.e. they occur in the vadose zone (Monkhouse, Steyn & Boshoff, 1983). Perched aquifer systems generally occur on top of clay lenses with low permeability. In South Africa, perched aquifers often occur on top of pedogenic layers such as hardpan ferricrete and calcrete.

The geotechnical fraternity in South Africa often uses the term "*perched aquifer*" incorrectly to describe shallow, (less than 5m), often seasonal, groundwater conditions. The engineering geologist records shallow seepage during the description of soil profiles since this could impact significantly on the development of the site and is generally not concerned with the hydraulic setting of the so-called, *perched aquifer*, relevant to deeper aquifer and surface water systems. Although this body of groundwater is frequently (but not always) located above the groundwater level, it is rarely independent and isolated and is normally hydraulically connected to surface water systems. These groundwater bodies are therefore not perched aquifers *sensu stricto*.

2.8.2 Hypothetical flow paths of contaminants seeping from a pollution source

Hypothetical flow paths of contaminants seeping from a pollution source are schematically represented in **Figure 2.9.** This figure represents a typical geological profile which comprises residual material overlying unweathered fractured bedrock. A fractured aquifer, developed in a highly weathered fractured fault zone, is located upstream of the site. Half the disposal area is underlain by impermeable hardpan ferricrete on which a perched water table has developed. The contaminants flow along the slope of the hardpan ferricrete (A). Funnelling flow occurs in more permeable weakly developed ferruginised material (B) and contaminants flow downwards until they reach impermeable hard rock. Contaminants flow preferentially through highly permeable materials to highly weathered and highly fractured materials (C) and enter the fractured rock aquifer (D). Contaminants may also flow along highly permeable leached relict rock structures (G) resulting from preferential weathering along these structures. On the other hand, in the event of lower water contents, contaminants may flow preferentially along the residual soil – highly leached zone interface. Fingering flow (I) may occur within the highly leached zone.

Figure 2.9 illustrates the complexity of aquifer recharge and contamination processes. Although it is not feasible to identify all these aspects on a regional scale, it may be possible to identify major recharge and contamination mechanisms.

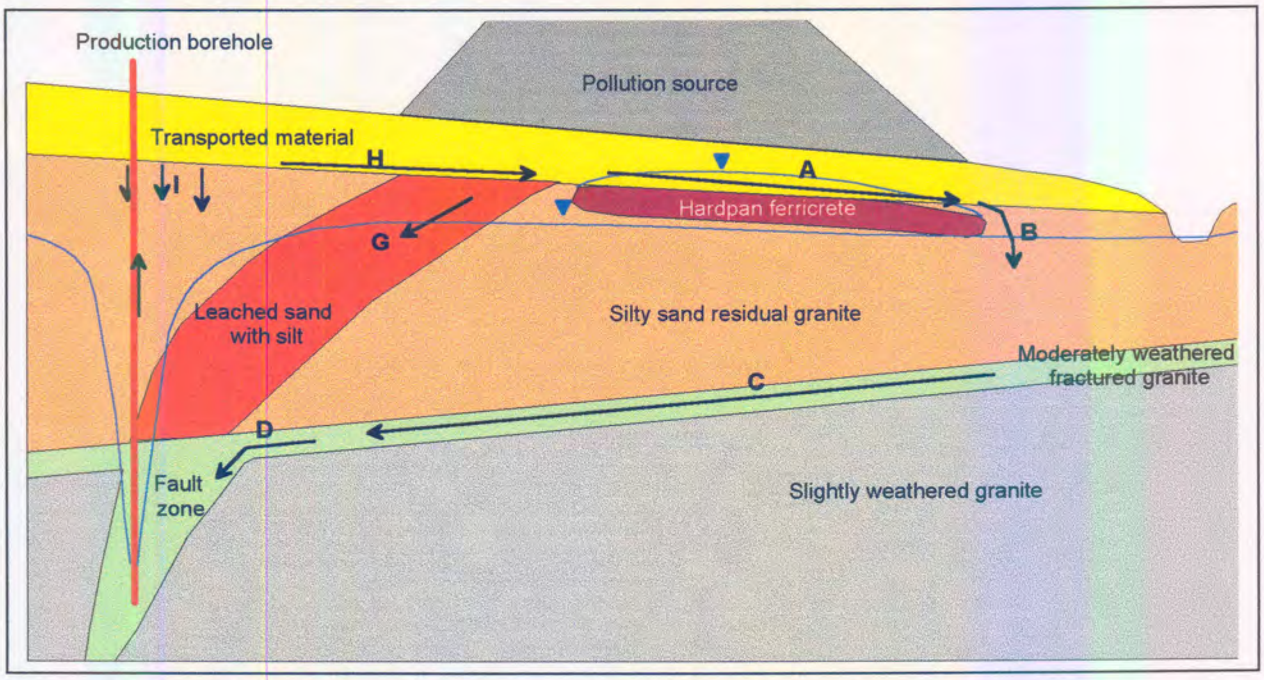

Figure 2.9: Hypothetical flow paths of contaminated water flowing through the vadose zone

2.9 Quantification of groundwater recharge

Groundwater recharge can be defined as the quantity of water flowing through the groundwater surface per unit time (Everett *et al.*, 1984), and is generally measured over a long period of time. It represents the portion of rainfall that reaches the groundwater surface and is frequently expressed as a percentage of mean annual rainfall (Bredenkamp, Botha, Van Tonder & Van Rensburg, 1995).

Groundwater recharge estimations are of great importance in groundwater management. In many parts of South Africa and other semi-arid regions of the world, future development and sustainable growth depend to a large extent on the availability of groundwater. In many rural areas groundwater is the sole or bulk water resource and, during droughts, often the only exploitable resource available (Bredenkamp *et al.*, 1995).

Many well-established methods for the quantitative estimation of groundwater recharge have been developed (Lerner *et al.*, 1990), but few can be applied with success to all the different climatic regions, especially to the semi-arid environments. Simmers (1988) states that no single comprehensive technique for the estimation of groundwater recharge can be identified from the available methods. Bredenkamp *et al.* (1995) suggest that it is preferable to average the results of the different methods.

The methodologies suitable for the estimation of groundwater recharge have been classified by Bredenkamp *et al.* (1995) in a number of categories and are discussed in detail by them and in many other hydrogeological textbooks.

2.9.1 Application of direct methods in the quantitative estimation of recharge

A flow model based on Darcy's law can be used to estimate groundwater recharge. However, conditions in the vadose zone are very complex and knowledge of a number of parameters is essential when recharge is estimated. These parameters include porosity, soil suction as a function of volumetric water content (i.e. the soil-water characteristic curve), saturated hydraulic conductivity and the residual water content.

Recharge can be expressed as the rate of flow per square metre and in the case that matrix flow is the dominant flow type, can be expressed as:

$$RE = q_{z_0} = -K(\theta)\frac{dh}{dz} \qquad [2\text{-}27]$$

where q_{z_0} is the specific discharge at groundwater surface depth.

The depth to the groundwater surface, volumetric water content, and hydraulic head are site-specific parameters and vary with time due to changing climatic conditions.

Bredenkamp *et al.* (1995) state that the unsaturated Darcy flow model fails as a practical method for estimating recharge because:

- Boundary conditions are difficult to determine and have to be approximated by sub-models. These models may apply only to part of the aquifer or for only part of the time.
- The lack of relevant precipitation and evaporation data is a practical restraint.
- The method relies on point measurements to derive values that cannot generally be extrapolated because of considerable variation in moisture content and hydraulic properties.

Freeze (1969) states that rainfall, with its influence on soil moisture conditions, is the main variable in recharge control. He also states that estimates of recharge based on saturated hydraulic conductivity and textural classification can be misleading.

Another drawback of the method is that preferential pathways and soil layering, frequently occurring in the vadose zone, have a major effect on fluid flow in the vadose zone and may render Darcy's law invalid.

Notwithstanding the above-mentioned criticism of direct methods in estimating recharge, recent advances in the processing capabilities of PCs and advances in unsaturated flow numerical modelling enables hydrogeologists to simulate and predict recharge processes.

2.10 Techniques for assessing aquifer vulnerability

Groundwater is polluted by many human activities, such as waste disposal and mining, as well as industrial and agricultural activities. The cleaning of polluted aquifers is technically very difficult and extremely expensive (Parsons & Jolly, 1994). It is therefore necessary to adopt a proactive approach to groundwater protection.

The concept of aquifer vulnerability originated from the assumption that the physical environment might provide some degree of protection against contaminants entering the ground surface (Vrba & Zaporozec, 1994). Contaminants enter the ground and are attenuated by a number of processes that are active in vadose and phreatic zones. Attenuation processes are discussed in section 2.10.1. Different physical environments have different capacities for the attenuation of contaminants.

Aquifer vulnerability maps indicating areas more (or less) sensitive to contamination will provide local, national and water authorities, as well as anybody involved in the planning and development, with knowledge concerning ways to site potential groundwater degrading activities away from vulnerable areas, or enable sites to be engineered in such a way that contamination is avoided. In South Africa and other developing countries, aquifer vulnerability maps can also promote the assessment of vulnerability in the case of informal settlements where services are lacking.

A number of methods have been developed to assess aquifer vulnerability. These include popular methods such as DRASTIC (Aller, Bennet, Lehr, Petty & Hacket, 1987), SINTACS (Civita, 1990) and GOD (Foster, 1987). The assessment of aquifer vulnerability does not constitute part of this research and readers are referred to Vrba and Zaporozec (1994) and Hearne, Wireman & Campbell (1991) for detailed discussions on aquifer vulnerability assessment and mapping.

Parsons and Jolly (1994) developed the Waste-Aquifer Separation Principle (WASP) to assess the effect of waste disposal sites on aquifers. This concept is based on the principle that waste and groundwater ought to be separated. Potential groundwater-degrading activities have to be situated away from vulnerable aquifers and such sites must be engineered to prevent infiltration of contaminants into the ground (Parsons & Jolly, 1994).

Three components have to be considered when pollution potential is assessed:

- Source of contamination (Threat factor)
- Pollution pathways (Barrier factor)
- Pollution receptor (Resource factor)

Figure 2.10 indicates the factors affecting groundwater resource pollution. The vadose zone acts as a barrier against pollution of the groundwater resource.

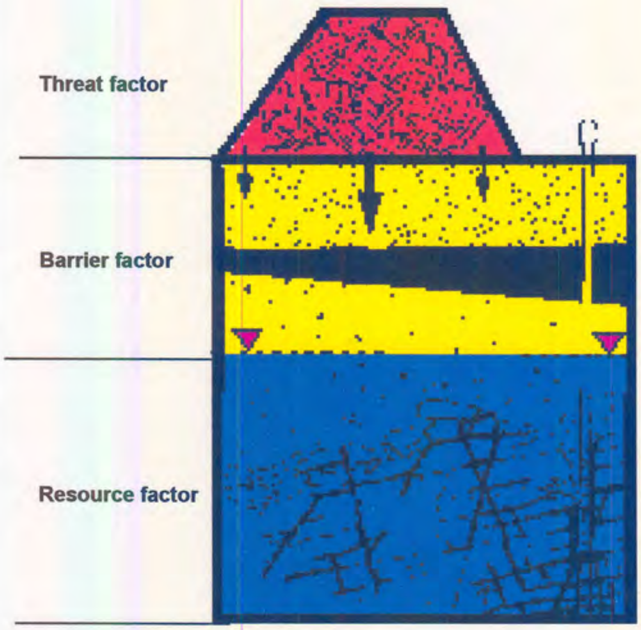

Figure 2.10: Factors that could have an effect on contamination of the groundwater resource (after Parsons and Jolly, 1994)

Parsons and Jolly (1994) use travel time to quantify the barrier factor which prevents the pollutants from a waste disposal site from reaching groundwater level. The travel time, also known as residence time, is the time it takes for the dissolved contaminant species to move to the groundwater surface. A longer travel time suggests that the contaminants are exposed to attenuation processes for a longer period, thus increasing the probability of complete attenuation before the contaminants reach the groundwater. The travel time can therefore be used to quantify the attenuation capacity of the vadose zone.

Travel time can be defined as:

$$Tt = \frac{d}{v_s} \qquad [2\text{-}28]$$

In a case where matrix flow is the predominant flow type of a specific area, Equation 2-28 can be written in Darcian terms:

$$Tt = \frac{z_{gs} \cdot \theta}{K(\theta)\dfrac{dh}{dz}} \qquad [2\text{-}29]$$

The depth to groundwater surface, volumetric water content and hydraulic gradient are site-specific parameters that vary with time due to changing climatic conditions. In addition, the saturated hydraulic conductivity may also vary, depending on the type of leachate moving through the soil. This is especially true in the case of hydrocarbon-based leachate moving through clay.

This simplistic approach may be criticised, since the most basic attenuation properties of the soil are not considered. Attenuation properties can either be an advantage or a disadvantage, depending on the type of contaminant that is involved. In the absence of attenuation processes, contamination is delayed and will resume after reaching the groundwater surface. In the case of a long travel time, specific discharge will be very low and dilution processes may be effective in preventing serious groundwater pollution. However, it has long been recognised in the hydrogeological fraternity that "*dilution is not the solution for pollution*". Attenuation processes have to be addressed in groundwater vulnerability studies. However, in the absence of relevant data, the travel time can be used to quantify the attenuation capacity of the vadose zone since long travel times are generally the result of a high clay content, which is a desirable property in attenuation.

2.10.1 Attenuation

When contaminants enter the ground, the dissolved contaminant species or leachate migrates downwards because of the water flow in the vadose zone. This process is known as advection. Several physical, chemical and biological processes that can improve the quality of the leachate take place in both the vadose and phreatic zones. These processes are known as attenuation processes. **Table 2.4** lists the types of attenuation processes that are operative in the vadose and phreatic zones.

Table 2.4 Types of attenuation processes (Sililo, Conrad, Murphy, Tredoux, Eigenhuis, Ferguson & Moolman, 1997)

Physical	Chemical	Biological
Dispersion	Hydrolysis	Aerobic biodegradation
(Ad)sorption	Dehydrohalogenation	Anaerobic biodegradation
Volatilisation	Precipitation	Hypoxic biodegradation
Filtration	Cation exchange	Nitrification
Dilution	Oxidation/reduction	Denitrification
Advection		Recarbonation (of high pH effluent)
		Cell synthesis

The effectiveness of these attenuation processes depends on:

- The properties of the contaminant species
- The properties of the soil medium
- Environmental factors (e.g. temperature)
- The travel time of the contaminants (the longer this period, the longer the contaminants are exposed to attenuation processes).

Because of the numerous interactions between different kinds of contaminants and the different kinds of soil media present, the evaluation of aquifer vulnerability is complicated. Some researchers have therefore suggested that a list of soil properties in the vadose zone may contribute either favourably or unfavourably to attenuation (Thorton, Lerner, Bright & Tellman, 1993):

Favourable properties

- A high clay and silt content increases the sorption capacity of organic and inorganic solutes.
- Particulate organic matter possesses a much higher cation exchange capacity (CEC) per unit weight, than clay. This can contribute favourably to the retention of leachate cations, especially of heavy metals.
- An alkaline pH enhances the immobilisation of heavy metals by hydroxides and/or carbonates. It also controls the effective CEC of the system.
- High levels of ferric oxide may contribute favourably to the attenuation of heavy metals in aerobic conditions.
- High levels of lime may also contribute favourably to attenuation.

Unfavourable properties

- High levels of soluble salts in the soil medium may mobilise toxic constituents.
- Adsorbed metals in contact with a reducing leachate may, in the presence of ferric oxide in the soil, be released through dissolution.

Attenuation processes are complex and are not yet fully comprehended. It is difficult to apply quantitative values to the above-mentioned properties, since attenuation is affected by the properties of both the soil and the contaminant.

It is generally accepted that pH is the primary variable in control of attenuation processes (Sililo *et al.*, 1997). Metal cations are adsorbed and precipitate as oxides, hydroxides and carbonates in high pH environments. Their mobility increases as pH decreases. The opposite is true for metal anions such as chromium, selenium and arsenic in some valence states. Some soils and rocks can resist pH changes where acidity or alkalinity is introduced. Important mechanisms include carbonate mineral buffering, exchangeable base cation buffering and buffering by alumina-silicate mineral decomposition in strongly acidic soils (McBride, 1994).

Sililo *et al.* (1998) are of the opinion that several attenuation processes operate in the soil/aquifer system and that these systems occur simultaneously and in certain cases compete with each other. The attenuation potential of a soil is not a fixed parameter, but a variable. However, Sililo *et al.* maintain that a soil with a high level of adsorption surfaces, pH well-buffered in the neutral or higher range and a low level of dissolved organic carbon will successfully immobilise metals.

Sililo *et al.* verified said propositions with an investigation of the attenuation potential of six different soils with regard to an array of heavy metals and organic compounds. The strongest attenuation occurred in soils with high pH and CEC values. However, chromium - anionic in groundwater - was most strongly attenuated in acidic soils, while other metals were also attenuated in these soils due to a high sesquioxide content with a subsequent high adsorption potential.

Attenuation processes continue in the phreatic zone, mainly by means of dilution. The effectiveness of attenuation in the phreatic zone will be dependent on the rate of flow at the groundwater surface, i.e. the specific discharge or flux.