

CHAPTER 3

GEOTECHNICAL INVESTIGATION METHODS AND THEIR HYDROGEOLOGICAL INTERPRETATIONS

Engineering geologists and geotechnical engineers are primarily concerned with the strength, potential volumetric change and permeability of soils and rocks. Engineering geologists use a variety of investigation methods to determine relevant soil and rock properties. Many of these methods involve the application of knowledge and experience in the fields of soil and rock mechanics and it is therefore difficult to derive direct correlation with hydrogeological properties. During the last few years, geohydrological investigations have become an important aspect of many engineering investigations, including waste disposal, mining activities (including mine tailings dams) and industrial development. Engineering geologists with training and experience in hydrogeology have become involved in these investigations to an increasing degree. Geotechnical methods and tests are described in detail in numerous textbooks and other publications (Jennings *et al.*, 1973; Brink, 1979; Holtz & Kovacs, 1981; Das, 1990; Bell, 1993)

Engineering geologists have traditionally been involved in the spatial distribution of geological materials with similar geotechnical properties. In particular, engineering geological investigations in assessing the geotechnical suitability for residential development and investigations concerned with the identification of geological materials for road construction purposes have greatly increased the engineering geologist's ability to determine the spatial distribution of geological materials.

Since many engineering geologists are mainly interested in the unconsolidated material overlying the bedrock, they need to consider the following processes in the assessment of geotechnical properties:

- The extent of weathering, which is mainly a function of climate;
- The time the material has been exposed to weathering processes, which implies that knowledge regarding paleo-climate could be necessary;
- Differences in geotechnical properties that could be ascribed to its topographical setting;
- Pedogenic processes

3.1 Geology

A thorough understanding of the geology is of paramount importance in any engineering geology investigation. Geological maps are widely available, and with his knowledge and experience of rock properties and its weathered products, the engineering geologist can predict possible geotechnical

problems even before visiting the site. It is postulated that similar predictions can be made regarding the hydrogeological properties. The hydrogeological characteristics depend on *inter alia* the geological material.

Most pre-Quaternary rocks in South Africa are characterised by two features, namely negligible primary porosity and varying secondary porosity caused by fracturing and weathering. These rocks are generally classified by geohydrologists as “*hard rocks*” (UNESCO, 1984). Hard, competent, impermeable rocks underlie most parts of South Africa. Some authors choose to exclude volcanic and carbonate rocks from the hard rock category (UNESCO, 1984). However, most volcanic and carbonate rocks in South Africa exhibit typical hard rock characteristics and have been included in the hard rock category. In the case of sedimentary rocks, competent impermeable hard rocks occur interbedded between impermeable fine-grained rocks such as shale, mudrocks and siltstone. These fine-grained rocks, although not classified as hard rocks, are also characterised by virtually no primary porosity and may also contain secondary porosity due to fracturing.

Rock masses are seldom homogeneous, but are intersected by joints, tension cracks, bedding planes, geological faults and other discontinuities. This causes more rock surface area to be exposed to weathering. Weathering etches along discontinuities and causes complex weathering profiles. Weathering is usually more intense along geological structures such as faults and often results in deeply weathered soils along these zones. Where three orthogonal joint sets are present in the rock mass, weathering along these joints may result in the development of corestones in the profile.

Flow through hard rocks is generally restricted to an interconnected system of fractures, joints and fissures within the rock mass. These discontinuities are mainly the result of large-scale tectonic events within the earth’s crust. (UNESCO, 1984). Weathering processes have a significant effect on the flow and storage capacity of the discontinuities.

3.1.1 The weathering profile

While rocks are exposed to weathering processes, they go through a series of weathering stages before converted to residual soils. The weathering stages typically manifested within a soil/rock profile from the ground surface to unweathered rock are as follows (Geological Society Engineering Group, 1990):

- Residual soil (Stage VI) – The original structure of the material has been destroyed. A pedological soil profile with characteristic horizons has developed. Large volume changes have occurred, but the soils have not been transported significantly.
- Completely weathered (Stage V) – All rock material has decomposed or disintegrated into soil. The original structure of the rock is still intact and visible.
- Highly weathered (Stage IV) – More than half of the rock material has decomposed or disintegrated into soil. Discoloured rock is present as blocks or rounded core-stones.
- Moderately weathered (Stage III) – Less than half of the rock material has decomposed or disintegrated into soil. Fresh or discoloured rock is present as blocks or core-stones that fit together.
- Slightly weathered (Stage II) – Discoloration on the rock surface indicates weathering, especially along discontinuity surfaces.
- Unweathered (Stage I) – No visible sign of rock material weathering except perhaps slight discoloration along discontinuity surfaces.

The hydrogeological properties of a particular geological material will vary depending on the stage of weathering of the material. Weathering characteristics are a function of both climate and geology and therefore need to be assessed on a site-specific basis. However, the following trends can be identified:

Unweathered rock: Since the South African geology is dominated by “hard rocks”, the unweathered rocks (igneous, sedimentary and metamorphic) are generally characterised by a low primary porosity and very low primary hydraulic conductivity. Preferential weathering frequently develops along joints, fractures and geological boundaries. The depth to unweathered rocks may therefore vary significantly over small distances, and this complicates unsaturated hydrogeological studies significantly.

Slightly weathered rock: Although weathering has not penetrated fractures and joints, discontinuities and rock surfaces are stained by weathering agents. The hydrogeological properties are similar to unweathered rocks, i.e., low primary porosity and low but definite primary hydraulic conductivity as indicated by the stained surfaces.

Moderately weathered rock: Less than half the rock material is decomposed or disintegrated to a soil. The rock could be highly fractured, which could result in very high permeabilities, typically up to 10 000 times higher than unweathered and slightly weathered rock. Moderately weathered rock typically has the highest hydraulic conductivities of all weathering stages in the soil profile resulting it to be the main pathway for contaminants originating from the surface. In particular, moderately weathered dolerite (termed fractured, boulder or gravel dolerite depending on the weathering stage and appearance) is characterised by high hydraulic conductivities. The material has been identified as the major contamination pathway of many pollution sources associated with coal mining in the Mpumalanga coal fields.

Highly weathered rock: More than half the rock has been decomposed or disintegrated into a soil. The rock is characterised by a high porosity, but unlike during the moderately weathered stage, the hydraulic conductivity depends on infilling between joints and fractures. In cases where the infilling has been decomposed to clay, hydraulic conductivity may be very low. In the case of disintegration to sand and gravel, the hydraulic conductivity may be very high. Dolerite generally disintegrates to granular (or sugar) dolerite with high expected hydraulic conductivities.

Completely weathered material: All rock material has been disintegrated or decomposed to a soil. However, the original rock structure is still intact. The hydraulic conductivity will be highly variable and flow of water will mainly occur along preferential pathways such as highly leached preferential weathering zones or disintegrated quartz veins. The field hydraulic conductivity is a function of *inter alia* climate and rock type.

Residual soil: All rock material has been disintegrated or decomposed to soil and the rock structure has been destroyed by pedogenic processes. This weathering stage is characterised by more uniform materials, compared with materials of the other weathering stages. Therefore, flow in residual soils generally conforms to traditional unsaturated flow theory. Hydraulic conductivities may be high in cases where disintegration is dominant, low in the case of decomposition and very low in cases where minerals have been weathered to predominantly clay minerals.

The contact between the weathered material and bedrock can be described as transitional, sharp, regular or irregular (Chorley, Schumm & Sugden, 1984).

Transitional contacts are commonly found in rocks with uniform structure and texture. Weathering susceptibility is limited to mineralogical level and the boundary between soil and bedrock is difficult to define. Arenaceous rocks of the Karoo Sequence generally exhibit transitional contacts in the more humid areas of South Africa.

Sharp contacts are characterised by thin (often only a few millimetres thick) contacts between bedrock and soil. These contacts are frequently referred to as a weathering front. Sharp contacts are associated with dense, mineralogically and structural uniform rocks, rocks susceptible to solution, basic rocks with a

low resistance to weathering and rocks with a low permeability or those with a permanent groundwater level close to surface (Chorley *et al.*, 1984). Basement granitoid rocks typically exhibit clearly defined weathering fronts. However, discontinuities in these rocks could result in preferential weathering along these zones and the weathering contact can then be described as irregular.

Regular contacts refer to contacts characterised by extensive jointed corestones, increasing in size with depth. Weathering stages are more clearly defined than with sharp and transitional contacts. The weathering of dolerite dykes and sills (occurring as intrusive bodies in Karoo rocks) may, to some degree, be described as regular. Extensive fracturing and jointing frequently occurs along the contact between bedrock and soils. The regular contacts of dolerite are probably related to the jointing and low resistance to weathering of the bedrock material.

Irregular contacts refer to conditions where weathering depths may vary considerably. This occurs where the bedrock's susceptibility to weathering and degree of jointing vary spatially. Irregular contacts typically develop in intrusive igneous rocks.

3.1.2 The Weinert N-value

Weinert (1980) developed the N-value to differentiate between regions of similar weathering characteristics. The N-value is based on the climatic situation of a particular area and can be defined as:

$$N_w = \frac{12 E_j}{P_a}$$

where N_w is the Weinert N-value, E_j is the total evaporation for the warmest month (in the case of southern hemisphere countries such as South Africa, this will generally be January) and P_a is the total annual precipitation. A map showing Weinert N-values for southern Africa is contained in **Figure 3.1**.

According to Weinert, physical weathering (disintegration) will predominate in areas where the N-value is larger than 5 and chemical weathering (decomposition) will predominate in areas where the N-values are less than 5. Chemical weathering will result in the formation of secondary minerals such as hydromica, clay minerals and sesquioxides. The type of secondary minerals that will develop will depend on the underlying geology, the time the rock has been exposed to weathering processes and climate and will be discussed in section 3.3. The weathering characteristics for areas of similar N-values are summarised in **Table 3.1**.

Table 3.1: Weathering characteristics as influenced by climate (Weinert, 1980)

N-value	Mode of weathering	Weathering characteristics	Principle secondary minerals
>10	Mainly disintegration	Thin weathering layer, No secondary minerals	Almost none
5-10		Few secondary minerals	Hydromica, illite
2-5	Mainly decomposition	Weathering profile deepens towards N=2, significant secondary minerals	Kaolinite, Montmorillonite
<2		Montmorillonite changes to Kaolinite in top soil layers, deep weathering profile	Kaolinite, Montmorillonite
<1		Montmorillonite and kaolinite change to sesquioxides, very deep weathering profiles	Kaolinite, Sesquioxides

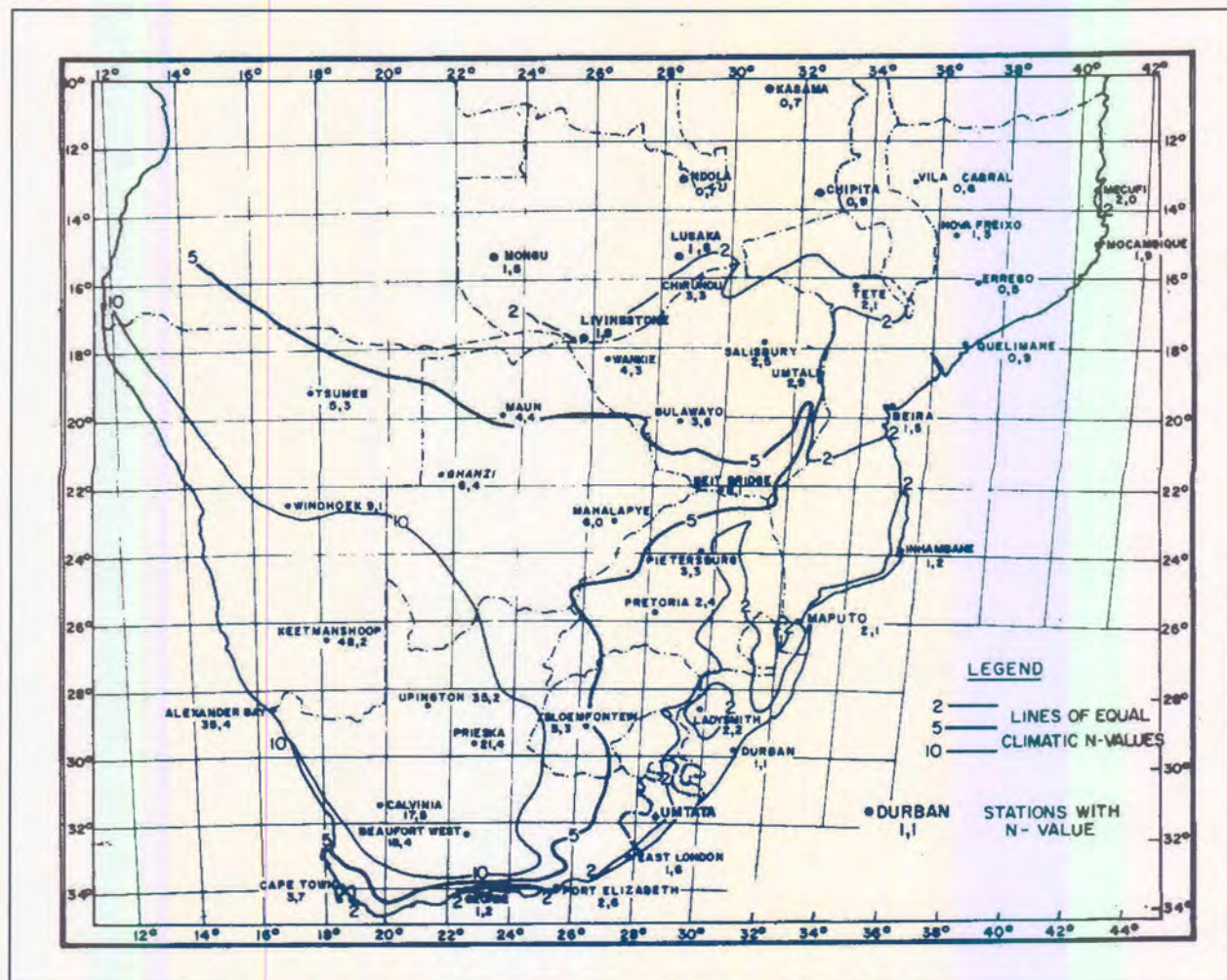


Figure 3.1: Map of South Africa indicating Weinert N-values (after Weinert, 1980)

3.2 Geomorphologic cycles

The present landscape in South Africa is the result of distinct geomorphologic cycles, ranging from the late Jurassic age to the present. Each cycle was initiated by a tectonic episode in which the land surface was uplifted, rifted, tilted or deformed in some way. With a new base level of erosion established after each cycle, renewed denudation proceeded, carving new landforms on top of the older ones (Brink, 1979).

Principal geomorphologic events in South Africa ranging from the Mesozoic age to present are summarised below (Partridge & Maud, 1987).

- Break-up of Gondwanaland by means of rift-faulting occurring in the late Jurassic/early Cretaceous to early Miocene age and manifested in the initiation of the Great Escarpment, owing to high absolute elevation of the southern African portion of Gondwanaland, also leading to the Enon-conglomerate Formation.
- African polycyclic erosion cycle occurring in the late Jurassic/early Cretaceous to early Miocene age, manifested in advanced planation throughout the subcontinent, resulting in erosion surfaces at two levels, above and below the Great Escarpment; also in the development of deep residual soil with extensive ferricrete and silcrete development and leading to the development of the Kalahari basin with the onset of sedimentation towards the end of the Cretaceous period

- Upliftment of 150-300 m comprising a slight westward tilting of the African erosion surface with limited coastal monoclinical warping and subsidence of the Bushveld Basin occurring at the end of the early Miocene age
- Post-African I erosion cycle occurring in the early mid Miocene to late Pliocene age, manifested by the development of the imperfectly planed Post-African I erosion surface and major deposition in the Kalahari basin
- Major upliftment event of up to 900 m, comprising an asymmetrical upliftment of the subcontinent and a major westward tilting of previous land surfaces of the interior, with monoclinical warping along the southern and eastern coastal margins, occurring in the Late Pliocene age.
- Post-African II erosion cycle of major valley incision occurring in the Late Pliocene to Holocene age, manifested in incision of coastal gorges, down-cutting and formation of higher terraces along interior rivers, formation of erosion surfaces and limited planation restricted to the eastern Lowveld region
- Quaternary cycle of climatic oscillations and glacio-eustatic sea-level changes occurring in the Late Pliocene to Holocene age, manifested in low-level marine beaches, coastal dune deposits, river terraces and deposits of the Kalahari sands

The hydrogeological properties of weathered materials and residual soils have mainly been influenced by the African and Post-African erosion cycles. The African polycyclic erosion cycle lasted for more than 100 million years. Soils and rocks have been exposed to weathering processes for very long periods resulting in the development of extensive weathering profiles. In addition, the palaeo-climate during the African erosion cycle was different than the present climate, with many areas in South Africa being wetter for long periods. Much of these soils have been removed by later erosion cycles but soils located on the African erosion surface are characterised by extensive pedocrete development and, particularly in areas underlain by granite, extensive kaolinisation (Partridge & Maud, 1987). In assessing the hydrogeological characteristics of residual soils, the geomorphic situation has to be considered.

It is not possible to discuss the effect of geomorphology for all geomorphic cycles since these will depend on the geology, topography, paleo-climate and present climate. These effects should be considered at site-specific and regional (1:50 000) investigations.

3.3 Land pattern classification

The preparation and compilation of any map, indicating the spatial distribution of various elements in the natural environment, imply that these elements can be classified according to the similarity of one or more physical attributes. The physical attributes within each of these classes should not vary significantly in order to allow the deduction of purposeful information. In addition, data collected for a particular land pattern class should be relevant for similar land patterns in other areas. In this way, the physical characteristics of particular land pattern classes can be extrapolated to similar land patterns in the investigation area, thereby making full use of available data and eliminating the need for the duplication of costly investigations on other similar land patterns.

The advantage of this approach is that, with the aid of remote sensing imagery, in particular aerial photographs, the properties of extensive land areas can be inferred at relatively low cost.

Several approaches have been developed in recognising the different land pattern classes. The two main approaches applied in South Africa are:

- The land system classification
- The land type classification

3.3.1 The land system classification approach

The land system classification system originated from soil engineering maps developed in the UK, South Africa and Australia for engineering purposes (Partridge, 1994). The nomenclature employed in this system closely resembles that of Christian and Stewart (1953). The system gained acceptance from approximately 1966 onwards, when it was adapted mainly for engineering purposes.

A land system is a large area with a recurring pattern of land forms, soils and hydrological regimes (Beaven, 1994). Land systems differ from other land systems by their distinctive physical attributes and can be recognised from aerial photos and other remote sensing techniques. Land systems are identified and mapped by their pattern of landforms, streams and vegetation. They are characterised by the following aspects (Beaven, 1994):

- They usually extend over an area of at least 100km² and can be mapped at scales of between 1:250 000 and 1: 1 000 000.
- The climate is uniform.
- Underlying geology is either uniform, or consists of closely related geological units (e.g. alternate layers of sandstone and mudstone) that can be mapped as a single unit.
- Recurrent land patterns can be recognised from aerial photo interpretations. Recurrent land systems may be situated thousands of kilometres from each other and still be named similarly, indicating that they possess similar physical attributes.
- Land systems are contiguous and no gaps of unclassified area occur between them.

Land systems are generally named after a town or village located within such a system. **Figure 3.2.** illustrates the Kyalami and surrounding land systems developed for the Gauteng area.

Land facets are defined as recognisable aspects of the landscape that together comprise a particular land system. These are geomorphologically related to each other. This means that they always occur in a certain relationships to each other. Land facets therefore occur in a particular sequence (e.g. crest, convex slope, concave slope), enabling engineers and geologists to identify land facets by their position in the landscape. Land facets are units of uniform slope, geology, soils, hydrological and hydrogeological conditions. Any deviation in a land facet should be simple and uniform. Land facets are characterised by the following aspects (Beaven, 1994):

- Non-related land facets may occur within a land system. These occur generally due to local geological features (such as intrusion structures) or particular hydrological features (such as pans).
- Land facets can generally be mapped at scales of between 1:10 000 and 1:100 000.
- Hydrological characteristics are consistent within a particular land facet.
- Land facets are named after the particular land form they are comprised of, e.g. hill crest, foot slope and river terrace. These units are not unique for all land systems.
- Land facets are contiguous and no gaps of unclassified land occur between them.

Land elements are the smallest unit in the land system classification methodology. Land elements are subdivisions of land facets and are often too small to be mapped at any practical scale. Yet, these features may have a significant effect on the aspect investigated. Examples include small rock outcrops, small dykes and oxbow lakes.

Chapter 3: Geotechnical investigation methods and their hydrogeological interpretations

Variations within a land facet that cannot always be identified except from field investigations are termed variants. For example, residual soil from different rock types, concealed by transported soil, may have diverse physical properties. Although the area blanketed by the transported soil is described as a single land facet, different residual soils are described as different variants within the land facet.

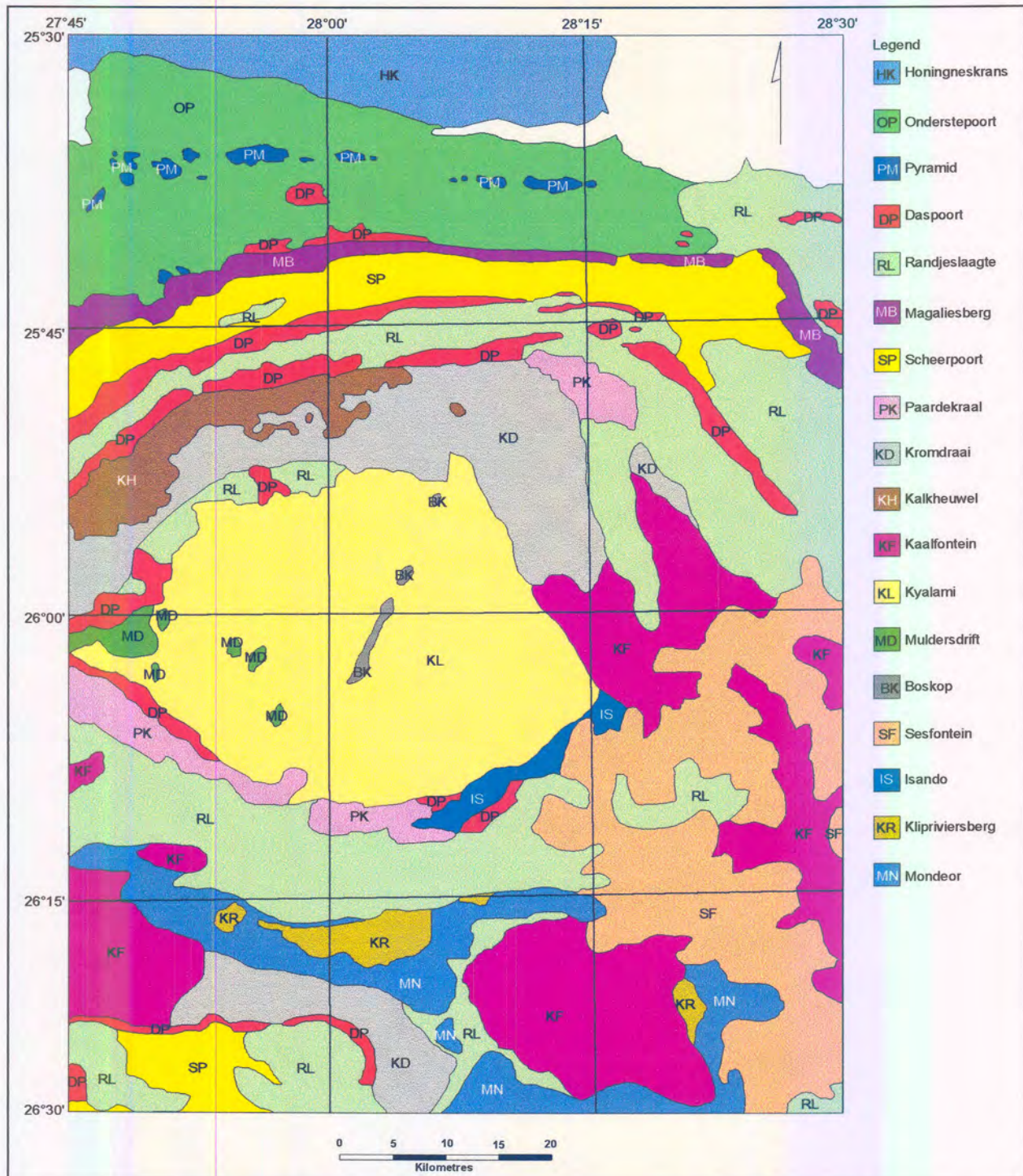


Figure 3.2: Land System map for the Gauteng Province area, South Africa (modified after Partridge, 1994)

The first stage in the assessment of the spatial characteristics of a particular area is to identify and define different mapping units. The scale and extent of the mapping depend on the intended use of the final presentation. In South Africa, national land system maps on a 1:250 000 scale are compiled by the

Institute for Soil, Climate and Water. Large parts of mainly urban areas are mapped for geotechnical purposes, generally at a scale of 1:10 000 or larger.

Land facets are mainly identified from stereoscopic aerial photo interpretations. The scale of the aerial photos chosen will depend on the level of detail required and the availability of the aerial photos. Other sources that can be used to identify land systems and land facets include satellite imagery, aerial colour photos and other remote sensing techniques.

3.3.2 The land type classification system

The land type classification system is similar to the land system classification system generally applied by the engineering fraternity. The land type classification system has been formulated by the Soil and Irrigation Research Institute of the South African Department of Agriculture during the 1970's.

Land type classification denotes an area to be depicted on a scale of approximately 1:250 000 and indicate areas characterised by a high degree of uniformity in respect of climate, terrain form and soil pattern. Land types are analogous to land systems, although differences do exist with regard to delineation of the land type. The chief criteria in land type delineation are pattern and density of the drainage system, relief, slope, profile and extent of every terrain unit.

A terrain unit is defined as any part of the land surface with a homogeneous form and slope, with other land units together comprising a particular land type. Five basic terrain units are used throughout the classification system, namely crest, scarp, midslope, footslope and valley bottom. These units are fixed: unlike the land facet approach, morphology and not genesis determines the delineation and description of terrain units. Variations in the form and sequence of the terrain units are generally indicated on an accompanying cross-section of the particular land type. Terrain units may further be subdivided into phases indicating variations within a specific terrain unit.

3.3.3 Parametric and analogue methods

The parametric and analogue methods differ from the land pattern method in that they identify a large feature in the landscape and subdivide it by weighting and rating separate data points within the feature.

The advantage of the parametric approach is that it reduces the physiographic bias associated with land pattern methods. However, grave disadvantages are associated with this approach. The first problem derives from choosing attributes to define the limits of a particular unit. The different attributes are rated and weighted according to their relevance for the aspect under investigation. In practical terms, the method proves to be less effective and requires much more effort than demarcating land units characterised by similar physical attributes. In addition, the method has proven to be very data intensive, thereby increasing the costs of deriving a parametric model. However, the increasing availability of data and recent progress in Geographical Information Systems may result in the increased application of parametric methods in mainly detailed investigations.

Analogue techniques are similar to parametric methods in that they identify a large feature in the landscape and subdivide it by assigning a rating to the relevant attributes within the land feature. Unlike the parametric method, analogue methods assign a numerical value to the particular attributes, which is entered into a mathematical equation, relating the attributes to the particular aspect investigated. LeGrand (1983) state that these methods may be suited to advanced stages in the investigation and are generally suited to site specific investigations where large amounts of data are available to numerically define the particular physical attribute.

3.4 Data collection and evaluation

After land patterns have been delineated for a specific investigation area, data must be collected to describe typical physical characteristics for a particular land facet. Soil data are usually collected by:

1. Excavating a number of testpits representative of the particular land facet,
2. Describing the soil profiles and
3. Collecting soil samples representative of the different soil horizons for laboratory testing.

Variability in soil properties for a particular land facet can be attributed to a number of reasons and has been discussed in preceding sections. These aspects should be taken into account during the processing and evaluation of data.

3.4.1 Data evaluation by means of GIS

After data collection has been completed, it should be recorded in a database in order to facilitate data evaluation and manipulation. For these purposes, data should preferably be recorded and presented within a spatial database framework. Recent advances in Geographical Information Systems (GIS) make it possible to evaluate point data within a land facet classification context.

GIS provides an invaluable tool for the capturing, processing, evaluation and presentation of spatial data. Digital photogrammetric work stations allow direct digital capturing of aerial photo interpretations. Information from adjoining studies can be incorporated within present studies, assisting in the interpretation (Murphy & Stiff, 1994). A major advantage of GIS is that different data source/layers can be overlain, thereby exposing patterns that may not have been noticed when the data sources were studied individually. Point data sources can be evaluated in terms of its positions within a particular land facet. With point data attributes captured in a database, data can be presented in terms of any of these attributes, thereby exposing patterns and assisting in the interpretation of the data.

Another major advantage of GIS is that data can be presented in any form required by the end user. Maps can be compiled on any scale to present any of the data digitally captured within GIS. However, data should not be presented at a scale larger than the scale in which the data were collected, as this may cause misinterpretation of the data for the particular area.

As is the case with any computerised tool, a major disadvantage of GIS is that the information presented may be based on insufficient data, thereby misleading the end user with regard to the quality and reliability of the end product.

3.5 Standard geotechnical investigation methods

An engineering geological investigation commences with a desk study during which all relevant information is gathered. Geological, topographical, orthophoto-, geohydrological, geophysical and other maps may be consulted to assess site conditions, then to be confirmed by a site visit. Aerial photographic interpretations serve to investigate the area from another angle and to identify geological structures such as faults.

Geophysical investigations are conducted for a number of reasons, but are applicable especially in dolomitic and geohydrological investigations. Gravimetric and Electromagnetic (EM) methods are commonly used in dolomitic investigations, while Vertical Electrical Resistivity (VER), as well as Magnetic and Seismic methods are used in geohydrological investigations.

Field investigations involve the mapping and excavation of a number of testpits on site. The soil profiles are described according to guidelines laid down by Jennings, Brink & Williams (1973). These allow for consistent soil profile descriptions for every type of soil. Percussion or rotary drilling may be conducted and drill cores are described according to guidelines by the Association of Engineering Geologists (1976). Small and large diameter auger drillings may also be conducted to extract samples or to describe soil profiles at greater depths. Samples for laboratory testing can be taken from the testpits, auger holes or drill cores.

In geohydrological investigations, the first step is usually to conduct a borehole census within a radius of a few kilometres from the proposed site. A data sheet containing information on borehole location, depth to groundwater level, borehole yield and groundwater use is completed for each borehole. Water samples are taken from a number of boreholes and tested for pH, electrical conductivity, total dissolved solids, major cations and anions, relevant microconstituents and bacteriological constituents.

During the geohydrological investigation, borehole pumping tests may be conducted to estimate the specific geohydrological characteristics of the aquifer. Step drawdown tests and constant-rate discharge tests are conducted, followed by recovery tests. Important geohydrological properties such as transmissivity, storativity and hydraulic conductivity can be obtained. From these data, estimates regarding the safe yield of the well can be made.

Geotechnical laboratory tests may be conducted to obtain index properties of the soil or rock material, in order to compare them with soils or rocks with similar properties of which the geotechnical behaviour is known. Certain geotechnical properties can be measured directly. Standard laboratory tests for geotechnical investigations are summarised in **Table 3.2**. Standard laboratory testing procedures are generally specified by the American Society for Testing and Materials (ASTM, various dates), the American Association of State Highway and Transportation Officials (AASHTO, 1974) and the British Standards Institution (1990).

The type of laboratory tests conducted depends on the nature and scope of the investigation. In general, index properties are derived for nearly all geotechnical investigations and are most widely available in the literature.

In addition to laboratory tests, *in situ* tests may be conducted to obtain additional information on the typical behaviour of soils and rocks. Standard *in situ* tests conducted during geotechnical investigations are listed in **Table 3.3**. Standard *in situ* tests are generally specified by the American Society for Testing and Materials (ASTM).

Direct and indirect correlation with geohydrological parameters can be made from many of the above-mentioned profiling techniques, laboratory and *in situ* tests. However, not many of results of geotechnical testing are widely available. For purposes of this research, the focus is on those geotechnical data that are readily available. These are the following:

- Soil profile descriptions
- Particle-size distribution
- Atterberg limits

These aspects will be dealt with in more detail.

Table 3.2: Standard geotechnical laboratory tests.

Test	Properties determined	Reference
Soil index properties		BSI 1377 (1990)
Particle-size analysis	Grading, soil fractions, classifications	
Atterberg limits	Plasticity charts, soil classification	
Bulk density	Natural density	
Natural moisture content	With bulk density → dry density	
Specific gravity	With dry density → void ratio, porosity	
pH	PH	
Conductivity	Conductivity	
Dispersivity tests		Elges (1985)
Crumb test	Dispersive/non-dispersive	
Pin hole test	Degree of dispersiveness	
Double hydrometer	Percentage of dispersiveness	
Compaction tests		BSI 1377 (1990)
MOD-AASHTO/PROCTOR	Max. dry density, opt. moisture content	
California Bearing Ratio	Bearing capacity	
Rock strength/durability		AASHTO (1974)
Unconfined compressive strength	Rock strength	
ACV and 10 % FACT	Rock durability	
Water absorption	Degree of weathering	
Consolidation tests		BSI 1377 (1990)
Single oedometer	Degree and rate of consolidation	
Double oedometer	Degree and rate of consolidation	
Collapse potential	Collapse potential	
Swell potential	Swell potential	
Permeability tests		BSI 1377 (1990)
Constant head	Hydraulic conductivity	
Falling head	Hydraulic conductivity	
Triaxial and shear box tests		BSI 1377 (1990)
Unconsolidated undrained	Friction angle, cohesion	
Consolidated undrained	Friction angle, cohesion	
Consolidated drained	Friction angle, cohesion	
Quick shear box	Friction angle	
Slow drained shear box	Friction angle	

Table 3.3: Standard geotechnical *in situ* tests.

<i>In situ</i> test	Properties determined	References
Penetrometer tests		
Dynamic Cone Penetration (DPL)	Relative density, CBR, depth to bedrock	De Beer (1991)
Standard Penetration Test (SPT)	Effective angle of friction, relative density, compressibility, undrained shear strength (cohesive soils)	ASTM D1586-84
Dynamic Probing (DPM, DPH, DPSH)	Effective angle of friction, relative density, compressibility, shear strength	Melzer (1982)
Cone Penetration Testing (CPT) and Piezocone (CPTU)	Relative density, effective strength parameters, various moduli, Overconsolidation Ratio, permeability, sand type	ASTM 3441-86
<i>In situ</i> field density tests		
Sand replacement	<i>In situ</i> density	ASTM D1556-82
Neutron meter	<i>In situ</i> density, <i>in situ</i> water content	ASTM D2922-81
Borehole tests		
Single and double packer	Water loss	Houlsby (1976)
Permeability tests		
Double ring infiltrometer	Hydraulic conductivity	Daniel (1989)
Constant or falling head borehole permeameters	Hydraulic conductivity	

3.5.1 Soil profile descriptions

The first assessment of soil properties commences with a systematic description of the soil profile. A systematic description of soil profiles is a widely used practice in South Africa, and the soil profile records are readily available. Most engineering geologists and geotechnical engineers use the guidelines developed by Jennings *et al.* (1973) to describe a soil profile.

Observations of a soil profile take place in trenches, test pits or large-diameter auger holes. Geological information of the top three to five metres is usually obtained. Auger holes, however, will provide geological information to depths in excess of 30 m. A soil profile usually consists of several layers/horizons that can be distinguished by changes in moisture content, colour, consistency, structure, soil type and origin. The individual layers are identified and their depths/thicknesses are recorded. The layers are then described by the MCCSSO method (Jennings *et al.*, 1973) on the basis of moisture content, colour, consistency, structure, soil type and origin.

Moisture content

The moisture content varies with time and rainfall events. The moisture content is described as dry, slightly moist, moist, very moist or wet. Wet soils are generally situated below the groundwater or perched groundwater levels. The moisture content must be interpreted in terms of grain-size. Sand with a water content of five to ten per cent may be described as 'wet', while a clay with the same water content may be described as 'slightly moist'. The term 'moist' usually describes a soil with a water content close to optimum. Dry and slightly moist soils require water if they are to be compacted effectively. Likewise, very moist and wet soils require drying before they can be effectively compacted.

Colour

The colour of soil is recorded to allow correlation of the same layer in different holes in the investigation area and it may be used to communicate with other people on the site. The colour of the soil depends on its moisture content. The colour is therefore described 'in profile', i.e. at natural water content, as well as 'wet', i.e. the colour of soil after it has been wetted.

Consistency

Soil consistency is a measure of the hardness or toughness of soil, i.e. the effort required to excavate the soil. It is also a rough measure of the soil's strength and density. The consistency of a soil depends on its moisture content, particularly with regard to cohesive soils. Soil consistency is described in different terms for granular and cohesive soils. The terms relating to consistency and their meanings are presented in **Table 3.4** and **3.5**:

Table 3.4: Consistency of granular soils (Jennings *et al.*, 1973)

Term	Description	Typical density (kg/m ³)
Very Loose	Crumbles very easily when scraped with geological pick	< 1 450
Loose	Slight resistance to penetration with sharp end of geological pick	1 450 - 1 600
Medium dense	Considerable resistance to penetration with sharp end of geological pick	1 600 - 1 750
Dense	High resistance to penetration with sharp end of geological pick; requires many blows of the pick for excavation	1 750 - 1 925
Very dense	High resistance to repeated blows of geological pick; requires power tools for excavation	> 1 925

Table 3.5: Consistency of cohesive soils (Jennings *et al.* 1973)

Term	Description	Unconfined compressive strength (kN/m ²)
Very soft	Pick head can easily be pushed in up to shaft of handle, easily remoulded by fingers	< 35
Soft	Easily penetrated by thumb, sharp end of pick can be pushed in by 30 to 40 mm, moulded by fingers with some pressure	35 – 75
Firm	Indented by thumb with effort, sharp end of pick can be pushed in up to 10 mm	75 – 150
Stiff	Slight indentation with sharp end of pick; requires hand pick for excavation	150 – 300
Very stiff	Slight indentation produced by blow of pick point; requires power tools for excavation	> 300

Structure

The structure of soil refers to the presence or absence of cracks, fissures and other structures not associated with the soil matrix. Granular soils generally exhibit a granular microstructure while the term voided is used for soils with a loosely-packed open microstructure.

Cohesive soils may exhibit a variety of structures:

The term *intact* indicates the absence of macrostructures. If the soil is firm, it may exhibit tension cracks when cut with a geological pick.

The term *fissured* indicates the presence of closed joints. The fissure surfaces are often stained with iron and manganese oxides. In residual soils, the fissures may coincide with relict joints or they may represent planes in which tension or shear has taken place.

The term *slickensided* indicates fissures with highly polished and glossy surfaces that are often striated. This indicates fairly recent shearing in the soil, probably due to heaving conditions.

The term *shattered* indicates the presence of open joints. The fragments usually consist of stiff to very stiff cubical, elongated or granular shapes. Shattered soil is usually associated with shrinkage or heaving conditions.

The term *microshattered* indicates small-scale shattering. The fragments are usually sand-sized. The soil appears to be granular but when wetted and rubbed on the palm of the hand, it breaks down into silt and clay.

Terms such as *stratified*, *laminated*, *foliated* and *warved* are generally used when soil shows relict structures of the parent material; they indicate the origin of the residual soil.

Soil type

The soil type is described in terms of the proportions of the various soil fractions, i.e. gravel, sand, silt or clay. A smaller proportion of soil fractions attains an adjectival function when the main soil fraction type is described, e.g. soil consisting mainly of sand with secondary silt is described as silty sand. A very small proportion of soil fractions is also recorded, e.g. a sandy soil with very little silt is described as sand with silt. Soil with approximately equal parts of soil fractions is described as silt-clay. Sand and gravel fractions are further described as fine, medium or coarse. A well-graded soil with secondary silt and little gravel is therefore be described as silty fine, medium and coarse sand with gravel.

The shapes of gravel, cobbles and boulders are recorded and described as rounded, subrounded or angular. The composition of these fractions is also recorded. The packing of soils with large proportions of gravel or pedogenic material is recorded. These soils are described as, e.g. 'Densely packed, rounded and subrounded quartz, coarse gravel and cobbles in a matrix of silty medium and coarse sand' or 'Loosely packed nodular ferricrete in a matrix of silty fine and medium sand'. Following this convention, consistent descriptions of soil profiles are possible.

Origin

Engineering geologists distinguish between transported and residual soils. The origin of transported soils is described by the mode of transport, i.e. hillwash, fine or coarse colluvium, alluvium, lacustrine, aeolian or beach deposits. A pebble marker usually occurs at the base of transported soil. The origin of residual soils is described in terms of the parent rock. The name of the formation of the parent rock should be included.

Notes

Additional information regarding the general conditions of the soil profile is recorded at the end of the description under the heading; 'notes'. Information regarding seepage, reasons for refusal of the machine, and samples taken, are usually recorded. Whereas descriptions of soil layers may be confirmed by laboratory tests, the information contained in the notes section is in many cases the only indication of shallow or perched groundwater conditions. In the absence of borehole logging data, these notes are the

only indication of weathering depths. Both the laboratory tests and the notes may be important in the characterisation of the vadose zone.

3.5.2 Soil index tests

Particle-size distribution

Several tests have been developed to analyse index properties of soil. These values are compared with the index properties of other soils to enable a prediction of soil behaviour. Mechanical analyses determine the range of grain sizes in a particular soil. Sieve analyses are conducted on particles larger than 0.075 mm, while hydrometer analyses are conducted on particles smaller than 0.075 mm. The methodology of these procedures is described by Das (1990) and several other authors. The results of the mechanical analysis can be presented on semi-logarithmic plots and are known as grain-size distribution curves. Several properties can be derived from particle-size distribution curves. Well-graded soils have a wide range of grain sizes and the curves are therefore smooth and generally concave. Soils with a uniform grain size have a small range of grain sizes and the curves are therefore steep. Gap-graded or skip-graded soils have a deficiency of certain grain sizes that may be caused by leaching or lessivage. Some parameters used for geotechnical purposes are obtained from particle-size distribution curves.

The grading modulus is a function describing the shape of the particle-size distribution curve. It can be defined as follows:

$$Gm = \frac{P_{2.0} + P_{0.425} + P_{0.075}}{100} \quad [7-1]$$

where P is the percentage the material retained on sieve sizes 2 mm, 0.425 mm and 0.075 mm respectively. Soils with a high fine fraction will in general have a grading modulus value of less than 0.8, while soils with a low fine fraction will have a grading modulus value of more than 1.0.

The effective size, D_{10} , corresponds to the sieve size, where 10 per cent of the mass of the soil sample have passed through; 10 per cent of all the grains of the sample therefore have diameters smaller than the effective size. Other sizes frequently used are D_{60} and D_{30} , where 60 per cent and 30 per cent of all grains have diameters smaller than D_{60} and D_{30} , respectively.

The coefficient of uniformity, C_u , is a crudely shape parameter, and is defined as:

$$C_u = \frac{D_{60}}{D_{10}} \quad [7-2]$$

C_u values of 2 to 3 correspond to uniform soils, such as beach sand, while C_u values of more than 15 correspond with well-graded soils (Das, 1990).

The coefficient of curvature, C_c , another shape parameter, is defined by:

$$C_c = \frac{(D_{30})^2}{(D_{10}) \cdot (D_{60})} \quad [7-4]$$

C_c values of between 1 and 3 correspond to well-graded soils, while C_c values are higher than 6 for uniform sands, and higher than 4 for uniform gravel (Das, 1990).

Atterberg limits

The presence of water in the pores of especially clayey soils will significantly change the geotechnical properties of a specific soil. Soil consistency changes as water content increases and can be grouped into four basic states: solid, semi-solid, plastic and liquid, with water content increasing from the solid to the liquid state.

The specific water contents where the nature of a soil changes (e.g. from solid to semisolid), are known as the Atterberg limits. The Shrinkage Limit (SL) refers to transition from a solid to a semisolid state. The Plastic Limit (PL) refers to transition from a semisolid to a plastic state and the Liquid Limit (LL) the transition from a plastic to a liquid state. The Plasticity Index (PI) is the difference in moisture content between the liquid and the plastic limit.

The activity value of soils can be used to predict the heaving potential of clayey soils. The activity of soils is defined as:

$$A = \frac{PI}{C} \quad [7-4]$$

Soil classification systems

The main purpose of soil classification systems is to group together soil types with similar behaviour patterns. Soils are generally classified according to their grain size. Gravel, sand, silt and clay are the main fractions to be distinguished. **Table 3.6** shows the classification system that is generally used in South Africa (Geological Society Engineering Group Working Party Report, 1990).

Table 3.6: Classification according to grain size (Geological Society Engineering Group Working Party Report, 1990)

Grain size (mm)	Classification	Mineralogical composition
< 0.002	Clay	Secondary minerals (clay minerals and sesquioxides)
0.002 – 0.074	Silt	Primary and secondary minerals
0.074 – 0.2	Fine sand	Primary minerals (mainly quartz)
0.2 – 0.6	Medium sand	
0.6 – 2.5	Coarse sand	
2.5 – 12	Fine gravel	Rocks (sometimes vein quartz)
12 – 50	Medium gravel	
50 – 200	Coarse gravel	
200	Boulders	Rocks

Natural soil usually consists of a mixture of various grain-sizes. For this reason, soils are usually classified in terms of their proportions of gravel, sand, silt and clay.

The Unified Soil Classification System (USCS), depicted in **Table 3.8**, is the most widely used classification system for geotechnical purposes. The USCS is based on the premise that the behaviour of a coarse-grained soil depends largely on its grain size, while the behaviour of a fine-grained soil depends on its plasticity. The soil is classified in terms of the results of mechanical analyses, after the Atterberg limits have been determined.

Coarse-grained soils are subdivided into gravel (G) and sand (S). They are further subdivided according to their gradation (well-graded (W) and poorly graded (P)), as well as according to the proportion of fines in the sample. Fine-grained soils are subdivided into silts (M) and clays (C) based on their plasticity

properties, and not on grain size. **Figure 3.3** shows the plasticity values of a fine-grained sample that are plotted on the plasticity chart.

Silts plot below the A-line, while clays plot above this line. Fine-grained soils are further classified according to the value of their liquid limit. Organic soils (O) and peat (Pt) are visually identified.

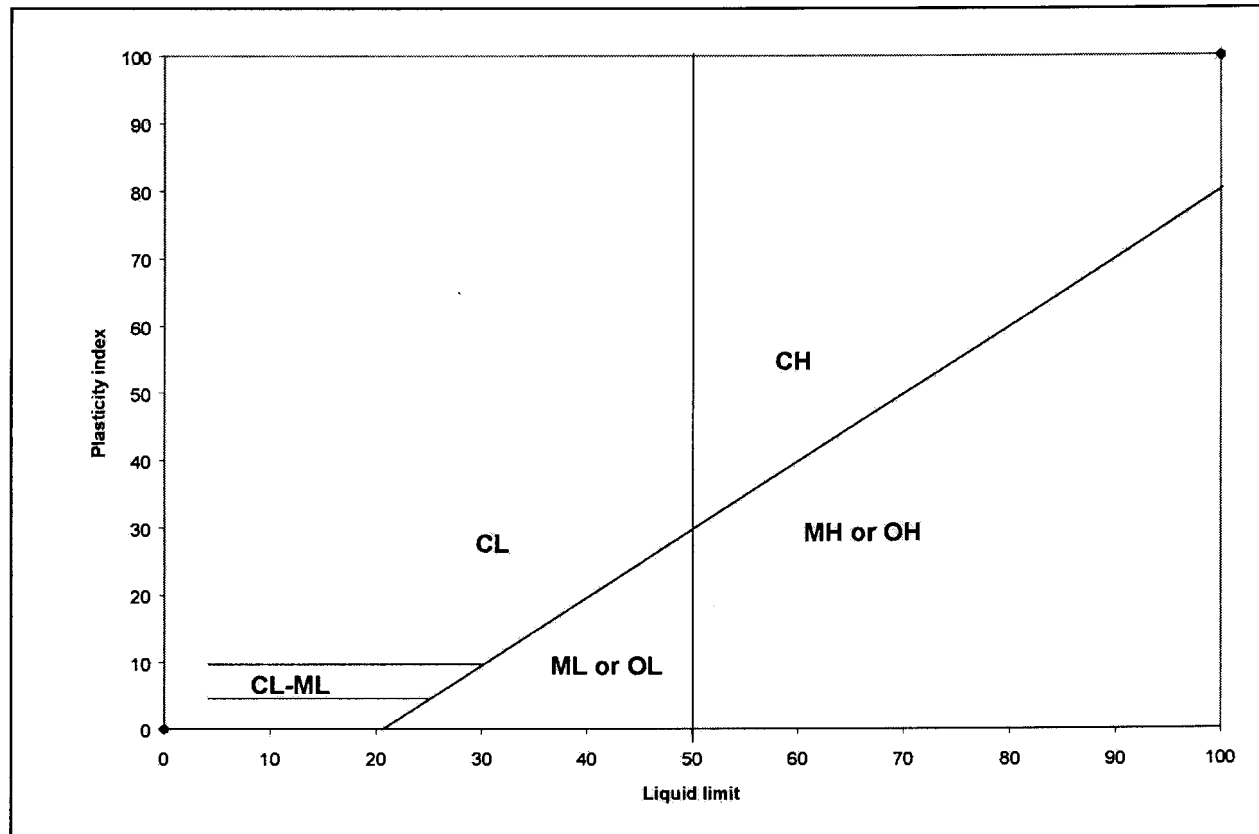


Figure 3.3: The plasticity chart

Table 3.8 contains a summary of the different USCS classes and indicates the criteria for distinguishing between the classes.

Another soil classification system frequently used is the AASHTO system. This system has been developed for use in the construction of roads and is also based on the grain size and plasticity of the soil. Holtz and Kovacs (1981) give a detailed description of the applicability of the AASHTO system in engineering work.

3.5.3 In situ permeability tests

In situ permeability tests provide the best means of obtaining the accurate hydraulic conductivity values of field soils and rock masses. Although laboratory permeability tests may provide accurate hydraulic conductivity values for compacted soils, they may not yield an accurate description of the geohydrological characteristics of field soils. Because of the small size of the laboratory sample and the spatial variability of geohydrological properties of field soils, they cannot be regarded as a representative. In addition, inaccurate measurements may result because of preferential flow along the sides of the ‘undisturbed’ samples. This is especially true for low-permeability soils.

Table 3.7: The Unified Soil Classification System (Holtz & Kovacs, 1981)

Major Division			Symbol	Description
Coarse-grained soil	Gravel	No fines	GW	Well-graded gravel, gravel-sand mixture, little fines
			GP	Poorly graded gravel, little fines
		With fines	GM	Silty gravel, gravel-sand-silt mixture
			GC	Clayey gravel, gravel-sand-clay mixture
	Sand	No fines	SW	Well-graded sand, gravelly sand with little fines
			SP	Poorly graded sand with little fines
		With fines	SM	Silty sand, sand-silt mixture
			SC	Clayey sand, sand-clay mixture
Fine-grained soil	Silt and clay LL<50	ML	Inorganic silt and very fine sand, silty or clayey fine sand, clayey silt with slight plasticity	
		CL	Inorganic clay with low to medium plasticity, gravelly or sandy clay, lean clay	
		OL	Organic silt or clay with low plasticity	
	Silt and clay LL>50	MH	Inorganic silt, micaceous or diatomaceous fine sandy or silty soil, elastic silt	
		CH	Inorganic clay with high plasticity, fat clay	
		OH	Organic clay with medium to high plasticity, organic silt	
		Pt	Peat, highly organic soils	
Coarse-grained soil:		less than 50% of material passes through No. 200 (0.075mm) sieve size		
Fine-grained soil:		more than 50% of material passes No. 200 (0.075mm) sieve size		
Gravel:		more than 50% of coarse fraction larger than No. 4 (4.75mm) sieve size		
Sand:		more than 50% of coarse fraction smaller than No.4 (4.75mm) sieve size		
Less than 5% fines:		GW, GP, SW, SP		
More than 12% fines:		GM, GC, SM, SW		
Between 5 and 12% fines:		Dual symbols e.g. GW-GM		
GW:		Cu > 41 < Cc < 3		
SW:		Cu > 61 < Cc < 3		
Fine-grained fraction:		Plot on plasticity chart		
Silt:		Beneath A-line or PI < 4		
Clay:		Above A-line or PI >7		
CL-ML:		Above A-line, 4 < PI < 7		

In situ permeability tests have the advantage that a large area of soil, and therefore a more representative sample, can be tested. The effect of preferential flow paths may be assessed. *In situ* permeability tests are not often conducted in geotechnical investigations and data are seldom available. Several methods are available to determine the hydraulic conductivity *in situ*. Daniel (1989) identifies four categories of *in situ* permeability tests, namely:

- Borehole tests
- Porous probe tests
- Infiltrometers
- Underdrains

He compares nine different in situ permeability tests. The methodologies are described by Daniel (1989). His findings are summarised in **Table 3.8**. The methodology for conducting slug tests is described in a number of geohydrological textbooks such as Fetter (1994).

Table 3.8: In situ permeability tests

Type	Device	Equipment cost	Duration	Direction	Accuracy
Borehole	Slug test	Low	Hours	Horizontal	Medium
	Packer tests	High	Hours	Horizontal	Medium
	Boutwell permaemeter	Low	Days to weeks	Vertical Horizontal	Medium
	Constant head permaemeter	Low	Hours to days	Horizontal	High to Medium
Porous probe	BAT permaemeter	High	Minutes to hours	Horizontal	Medium
Infil-trometer	Open, single- ring	Low	Weeks to months	Vertical	Medium
	Open double-ring	Low	Days to months	Vertical	High
	Closed, single-ring	Low	Weeks to months	Vertical	Medium
	Sealed double-ring	Medium	Weeks to months	Vertical	Very high
	Air-entry	Medium	Hours to days	Vertical	Medium
Under-drain	Lysimeter pan	Low	Weeks to months	Vertical	High

3.6 Availability of geotechnical data

The success with which geotechnical data can be applied to estimate hydrogeological properties depends on the availability and reliability of the geotechnical data. Geotechnical data must be available over large parts of the country and have to be easily accessible to be of use in future hydrogeological studies. In addition, geotechnical data must be available at a reasonable cost.

Standard geotechnical techniques and tests such as soil profile descriptions and soil index tests are conducted during almost every geotechnical investigation. These investigation methods are applied *inter alia* for residential development and may cover large areas. In contrast, the more sophisticated triaxial and double oedometer tests are only conducted during the design stages of specific structures and therefore represent small site-specific areas. Some types of geotechnical data are more accessible than others. The reliability of estimating geohydrological properties from geotechnical data, depends on the type and quality of the geotechnical data

The availability of geotechnical data, especially data from investigations for residential development, has been researched and published by Milford (1994). Much of the following discussion is based on his information.

Geotechnical data are available at various organisations, which include:

- The Council for Geoscience
- CSIR (Transportek Division)
- Local authorities, regional and metropolitan councils

- Provincial administrations
- Department of Transport
- Private sector

The Council for Geoscience

Over the years, the Council has accumulated substantial amounts of geotechnical data that cover large parts of mainly urban areas in South Africa. The data are presented in different formats, but mostly as engineering geology maps with accompanying reports. Raw data such as soil profile plots and laboratory tests results are included, usually as appendices. All geotechnical reports conducted by the Council are available to the public.

The Council for Geoscience has initiated a digital geotechnical database, ENGEODE, whereby geotechnical data sources are indexed and all soil profiles, together with laboratory test results, are digitally captured (Du Plessis, 1998). This database was set up because of the need for geotechnical data experienced by engineering geologists and other persons operating in the civil engineering field

By March 1998, ENGEODE contained an index of approximately 85 000 soil profiles, mainly reports on housing developments (Croukamp, 1998). Approximately 80 per cent of these soil profiles are from Gauteng. The soil profile descriptions are available in both digital and paper format. Results of laboratory tests (mainly particle-size distribution curves, Unified Soil Classification and Atterberg limits) are usually included with the soil profiles. Very few permeability test results are available, but estimations of saturated hydraulic conductivity values according to the Hazen equation (see Chapter 9) are included. Data on ENGEODE are stored on Oracle databases. Data can be made available on ASCII format and on third party software (notably dot.PLOT) to provide the graphical representation of soil profiles.

The Council of Geoscience has access and control over large quantities of geotechnical data. Unfortunately, these data cover mainly urban areas of South Africa. Geotechnical data are readily available, but there is a cost attached.

CSIR – Transportek

Significant geotechnical data sets are currently in the possession of the Transportek Division of the CSIR. The CSIR has been undertaking the engineering geology mapping of central Gauteng, with emphasis on the dolomite girdle, on a scale of 1:10 000. The mapping is based on the land system and land facet approach. The CSIR has established a GIS database, based on land facet parameters. The database includes generalised soil profiles for each facet, as well as approximately 1 200 additional soil profile descriptions. Most profile descriptions include laboratory results, mainly particle-size distribution curves, Atterberg limits and Unified Soil Classifications. All soil profile descriptions are spatially referenced and many of these can be displayed on and questioned by means of GIS software. Very few permeability studies have been carried out by the CSIR. Estimations of permeability of soil are made by means of tables relating Unified Soil Classification symbols to saturated hydraulic conductivity.

In addition, the CSIR is in possession of data, mainly engineering geology maps indicating construction materials for road construction purposes. These data have previously been collected and maintained by the Department of Transport but this database was closed down in the mid-1980's.

Local authorities, regional and metropolitan councils

Geotechnical data, mainly information regarding geotechnical suitability for residential development, are available from several local authorities and from regional and metropolitan councils. The Greater Johannesburg Metropolitan Council, East Rand Metropolitan Council and the Johannesburg City Council maintain the most extensive geotechnical data sets.

Geotechnical information at the Greater Johannesburg Metropolitan Council comprises several spatial layers comprising speciality geotechnical maps such a dolomite stability map, a landform map and a land use potential/land allocation map.

At least two data sets maintained at the East Rand Metropolitan Council are relevant to this study, namely geotechnical suitability maps for residential development and an undermined land study comprising maps (1:10 000 and 1:5 000) indicating areas of shallow undermining.

The Johannesburg City Council Geotechnical Data Bank contains approximately 2 000 geotechnical reports of varying detail. These cover the greater Johannesburg area. Many reports include soil profiles and laboratory tests. The data are indexed on a cover sheet as well as on a bibliographical reference list. The data bank does not contain digital data. The data are available to the public at minimal reproduction costs.

The Durban Metropolitan Council does not maintain any geotechnical data sets. However, it does have 1:15 000 geological maps, covering the Durban - Westville area. These maps are available from the Surveying Department.

The Cape Town Metropolitan Council also does not maintain any geotechnical data.

Private sector

The private sector has over the years, accumulated substantial quantities of geotechnical data. The private sector (with the exception of mining houses whose data are typical propriety information) is mostly willing to make their information available to other organisations. Copies of all geotechnical investigations are usually indexed and can be accessed by means of the project name, number and location. Although the geotechnical investigations have been conducted in accordance with clients' needs, general geotechnical data, such as soil profile descriptions and index test results, are usually included in the appendices. In addition, the general geological and geotechnical conditions of the specific investigation area are discussed in the report. The investigation areas have usually already been developed by the time the report is accessed, but estimates on the geotechnical characteristics in similar adjacent areas can be made by extrapolation.

3.6.1 Information from other data sources

The assessment of the geotechnical and geohydrological conditions of a specific area requires a thorough understanding of the general conditions of that area. Information on the topography, geology, hydrology, climatology, land use and existing infrastructure has to be available. Many institutions can assist in supplying the relevant information.

A large variety of data collected during geotechnical and hydrogeological investigations that may be used in estimating hydrogeological properties, are available from a number of institutions. These are listed in Table 8.1.

Government Printing Works

A series of topographical sheets on various scales covering the whole of South Africa can be obtained from the Government Printing Works. These include the 1:50 000, 1: 250 000 and 1:500 000 topographical maps and a range of other less relevant maps.

Surveyor-General

A series of aerial photos of various scales covering the whole of South Africa can be obtained from the Surveyor-General. In addition, 1: 10 000 scale orthophoto maps, covering most parts of South Africa can also be obtained.

The Council for Geoscience

A 1:1 000 000 regional geological series is available in paper and digital (either CAD or ArcView/ArcInfo) formats. Twenty-seven sheets of the 1:250 000 regional geological series are available in paper format and are in the process of being captured in digital form. An accompanying information booklet on the geology of the specific map can also be obtained. Several 1:50 000 geological sheets of selected areas (mainly the central Witwatersrand area) are in the process of being captured in digital format.

Weather Bureau

There are weather stations at many towns and cities around the country. Meteorological data are recorded, including average monthly minimum and maximum temperatures, average monthly evaporation, average monthly rainfall and cloud cover. The climate statistics are available from the Weather Bureau at minimal cost.

Department of Water Affairs and Forestry (DWAF)

The Directorate of Geohydrology of the DWAF is responsible for maintaining the National Groundwater Database (NGDB). General information regarding boreholes such as their location, longitude and latitude is recorded. Groundwater data such as the depth to groundwater level, groundwater level fluctuations, borehole yield and depth of water strike are also on record. Data can be accessed by co-ordinates and are available free of charge. Digital data can be made available in popular formats such as dBase, Excel, Quattro Pro and Access.

Institute for Soil, Climate and Water (ISCW)

Three data sets of interest are maintained:

- A national land type database
- A national soil profile database
- A soil mapping database

The ISCW land type database consists of land type units surveyed and presented on maps at a scale of 1:250 000. It represents areas with a high degree of uniformity with regard to terrain form, soil pattern and climate (Milford, 1994). It covers 80 per cent of the country. Gaps occur mainly in the Eastern Cape and in the winter rainfall regions.

The soil profiles are classified according to the South African Binomial or Taxonomic Soil Classification Systems. The relevant soil profiles typically extend to a depth of about 1.2 m.

Agricultural soil mapping at a scale of 1:50 000 is available in digital form for the Gauteng area.

ISCW soil profile descriptions and soil classification systems are different from geotechnical descriptions and extend only to 1.2 m in depth, probably describing mostly transported soil, which further limit the geotechnical use of the data.

Soil-water retention data can be used to establish soil-water characteristic curves for different soils, which can assist with the estimated unsaturated geohydrological properties. However, such data are very scarce.

The ISCW frequently uses the Soil Texture Chart to classify soil according to the sand, silt and clay proportions. It is used to differentiate between the different agricultural soils.

Chapter 3: Geotechnical investigation methods and their hydrogeological interpretations

Local authorities

Maps indicating the layout of towns and cities are available from most local authorities.

Educational and research institutions

Data on water retention characteristics can be obtained from a number of educational and research institutions which have conducted research in this regard. The University of Natal, University of Fort Hare and the Soil and Irrigation Research Institute have water retention data from research conducted in many parts of the country (Hutson, 1984).

Available geotechnical data and sources are summarised in **Table 3.9** while **Table 3.10** contains information on other data sources.

Table 3.9: Availability of geotechnical and other relevant information (Modified from Keyter; 1994)

Source	Information	Coverage	Format	Accessibility
Council for Geoscience	Engineering geology	Central Wits	Digital	Interpreted information
	Residential development reports	All provinces with the exception of Western, Northern and Eastern Cape	Paper & digital	Interpreted information and soil profiles with lab tests
	ENGEODE	Mainly Gauteng and Port Elizabeth	Paper & digital 60 000 soil profiles 25 000 digital	R40 per soil profile some with lab tests
	Dolomite stability	Central Gauteng	Digital	Interpreted information
CSIR (TRANS-ORTEK)	Engineering geology Problem soils Development suitability	Central Gauteng	Digital	Yes
	Soils database	Central Gauteng	GIS: 1 200 soil profiles	Yes
Gauteng Province	TPA Engineering soils	Gauteng	Digital	Unknown
Greater Johannesburg Metropolitan Council	Geology 1:50 000 Construction materials Landform Dolomite stability Problem soils Land use potential Undermined land	Central Witwatersrand area	Digital	Unknown
East Rand Metropolitan Council	Development suitability Undermined land	East Rand	Digital	
Johannesburg City Council	Geotechnical database	Greater Johannesburg	Paper	Reproduction cost
Department of Transport	Soils engineering maps	Gauteng	Paper	At CSIR

Table 3.10: Other available relevant information

Source	Information	Coverage	Format	Accessibility
General soil data				
ISCW	Land type maps	80% of South Africa	Paper & digital	Interpreted information
	Soil profile database	80% of RSA 200 000 profiles	Paper	Interpreted information
	Soil types	Gauteng 6000 profiles	Paper & digital	Interpreted information
Educational and research institutions	Soil-water retention data	Selected parts in South Africa	Paper	On request
General data				
Government Printing Works	Topographical sheets 1:50 000 1:250 000 1:500 000	Whole of South Africa	Paper	Moderate costs
Surveyor-General	Aerial photos Various scales	Whole of South Africa	Paper & film	Moderate costs
	Orthophotos 1:10 000	Most of South Africa	Paper	Moderate costs
Local authorities	Town layout plans	Most of South African towns	Paper	Most towns free or at reproduction cost
Council for Geoscience	Geology 1:1000 000	100% of RSA	Paper & digital	Moderate costs
	Geology 1:250 000	74% of RSA	Paper	Moderate costs
	Geology 1: 50 000	Gauteng	Paper	Moderate costs
Weather Bureau	Climate statistics	Whole of South Africa	Booklet	Minimal costs
Department of Water Affairs and Forestry	Groundwater data	Whole of South Africa	Digital	Free

CHAPTER 4

RELATIONSHIPS BETWEEN GEOTECHNICAL AND HYDROGEOLOGICAL PROPERTIES

4.1 Estimations of porosity

Porosity data are not collected as frequently as other geotechnical data such as indicator tests. Porosity is an important parameter in the estimation of both saturated and unsaturated hydraulic conductivity. In addition, the soil-water retention characteristics are highly sensitive to porosity. It may be necessary to estimate porosity from other geotechnical data such as soil profile descriptions and particle-size distribution curves. In field soils the porosity depends on a number of factors, the most important being particle-size distribution, bulk density, shape of the grains and mineralogical content.

In coarse-grained soils, soil profile descriptions may provide an indication of the density of the soil. The porosity and void ratio can be calculated if the dry density value is known. However, soil consistency is determined in a very subjective manner, often resulting in inaccurate estimations of primary porosity.

Table 4.1: Typical dry density, porosity and void ratio values for granular soils (Modified from Jennings *et al.*; 1973)

Description	Typical dry density (kg/m ³)	Typical porosity	Typical void ratio
		(Specific weight of grains = 2.65)	
Very loose	< 1450	> 0.45	> 0.83
Loose	1450 – 1600	0.45 – 0.40	0.83 – 0.66
Medium dense	1600 – 1750	0.40 – 0.34	0.66 – 0.51
Dense	1750 – 1925	0.34 – 0.27	0.51 – 0.38
Very dense	> 1925	< 0.27	< 0.38

In fine-grained soils, soil consistency is a rough indication of the shear strength of the soil. The shear strength is strongly influenced by moisture content and density. The relationship between soil density and consistency is not reliable and the moisture content has to be considered in order to interpret the soil consistency. Dry, fine-grained soils have a much higher apparent shear strength than wet, fine-grained soils.

Table 4.2: Typical dry density, porosity and void ratio values for cohesive soils (Modified from Mathewson, 1981)

Description	Typical dry density (kg/m ³)	Typical porosity	Typical void ratio
		(SG of grains = 2.7)	
Very soft	< 1000	> 0.62	> 1.65
Soft to firm	1000 – 1700	0.36 – 0.62	0.56 – 1.65
Stiff	1700 – 1800	0.32 – 0.36	0.47 – 0.56
Very stiff	> 1800	< 0.32	< 0.47

In the case of an ideal soil consisting of spherically shaped grains of equal diameter, the porosity will be a factor only of the packing of the grains. Porosity values may vary between a maximum of 0.476, in the case of unstable packing to 0.260 in the case of fully compact packing (Davis & De Wiest, 1966). In the case of grains consisting of plate-like structures of identical size, porosity may vary from 0.60 for a box-like structure to 0.00 for ideally stacked plates (Davis & De Wiest, 1966). However, grains in field soils are neither ideally spherical nor plate-like and considerations involving these ideal shapes cannot be directly applied to field soils. The grain shape has a significant effect on porosity, since angular grains may create vault-like structures in the soil, resulting in a 2 to 5 per cent higher porosity value than expected. These vault-like structures may have an even more significant effect on saturated water flow through the soil. The Poiseuille equation indicates that the rate of flow through narrow tubes is very sensitive to the radius of the tubes and therefore also to the radius of pores.

Grain-size distribution has a major effect on porosity, since smaller grains tend to be situated within large pores. The grain-size distribution can be expressed by the coefficient of uniformity. Laboratory results conducted by Hazen (1930) show a definite correlation between porosity and the coefficient of uniformity, as indicated in **Table 4.3**.

Table 4.3: Coefficient of uniformity and corresponding porosity values (presented by Vukovic & Soro, 1992)

C _u	1.8	2.0	2.3	2.3	2.4	7.8	9.0
ε	0.45	0.42	0.44	0.42	0.40	0.32	0.36

Research by Istomina (1957) shows that porosity can be approximated from the coefficient of uniformity by means of the following equation:

$$\varepsilon = 0.255(1 + 0.83^{C_u}) \quad [4-1]$$

However, Istomina has found that significant deviations occur in soils comprising high clay percentages. Equation 4-1 can therefore not be applied if a high percentage of clay is present in the soil.

With the development of the ACRU agrohydrological modelling system, Schulze (1995) indicated that a number of soil properties, including porosity, could be estimated from soil texture classes. These statistically determined relationships have been developed by a number of authors (Rawls; Brakensiek & Saxton, 1982; Schulze George & Angus, 1987; Buitendag, 1990; Everett, 1990). The porosity value range for the twelve agricultural soil types is indicated in **Table 4.4**.

Table 4.4: Estimated porosity values for twelve agricultural soil types (from Schulze 1995)

Soil type	Bulk density (kg/m ³)	Total porosity
Clay	1 220 – 1 370	0.470 – 0.482
Clay loam	1 220 – 1 410	0.474 – 0.456
Loam	1 260 – 1 420	0.512 – 0.480
Loamy sand	1 310 – 1 510	0.452 – 0.477
Silt	N.A.	N.A.
Silt loam	1 130 – 1 340	0.530 – 0.500
Silty clay	1 230 – 1 380	0.476 – 0.480
Silty clay loam	1 250 – 1 400	0.489 – 0.473
Sand	1 320 – 1 500	0.446 – 0.440
Sandy clay	1 350 – 1 530	0.393 – 0.428
Sandy clay loam	1 350 – 1 580	0.435 – 0.405
Sandy loam	1 260 – 1 460	0.486 – 0.466

4.2 Estimations of saturated hydraulic conductivity

Various authors have tried to relate the saturated hydraulic conductivity to descriptions of soil type (Das, 1990; Mathewson, 1981) and these relationships are presented in **Table 4.5**. Since hydraulic conductivity is not entirely dependent on soil type, the predicted saturated hydraulic conductivities are not very accurate. In addition, the percentages of clay fractions present in the profile greatly affect saturated hydraulic conductivity. For example, two soils consisting of sand with clay contents of 10 per cent and 30 per cent respectively can both be described as clayey sand. However, the soil with 30 per cent clay may have a saturated hydraulic conductivity of a few orders of magnitude lower than the soil with 10 per cent clay. In addition, soil types are described subjectively. Saturated hydraulic conductivity values derived from soil-type descriptions may not be very reliable. However, in the absence of more reliable data, saturated hydraulic conductivity values of reasonable reliability can be obtained.

Table 4.5: Estimated hydraulic conductivity values from soil type (modified from Mathewson, 1981)

Soil type	Saturated hydraulic conductivity (m/s)
Gravel	$10^{-1} - 10^{-4}$
Gravel (Well-sorted)	$10^{-1} - 10^{-3}$
Gravel (Poorly sorted)	$10^{-1} - 10^{-4}$
Silty gravel	$10^{-5} - 10^{-8}$
Clayey gravel	$10^{-6} - 10^{-9}$
Sand	$10^{-3} - 10^{-7}$
Coarse sand (Well-sorted)	$10^{-3} - 10^{-5}$
Coarse sand (Poorly sorted)	$10^{-3} - 10^{-5}$
Medium sand (Well-sorted)	$10^{-4} - 10^{-6}$
Medium sand (Poorly sorted)	$10^{-3} - 10^{-5}$
Fine sand (Well-sorted)	$10^{-5} - 10^{-7}$
Fine sand (Poorly sorted)	$10^{-4} - 10^{-6}$
Silty sand	$10^{-6} - 10^{-8}$
Clayey sand	$10^{-7} - 10^{-9}$
Silt	$10^{-6} - 10^{-9}$
Clay (Low plasticity)	$10^{-8} - 10^{-10}$
Clay (High plasticity)	$10^{-9} - 10^{-11}$

A number of authors have attempted to relate saturated hydraulic conductivity to USCS soil groups (Mathewson, 1981; Badenhorst, 1988) and these are summarised in **Table 4.6**.

Table 4.6: Estimation of hydraulic conductivity from USCS soil groups (modified from Mathewson, 1981 and Badenhorst, 1988)

USCS soil Groups	Hydraulic conductivity (m/s)	
	Mathewson	Badenhorst
GP	$10^{-5} - 10^{-2}$	$10^{-5} - 10^{-1}$
GW	$10^{-4} - 10^{-2}$	$10^{-5} - 10^{-3}$
GM	$10^{-8} - 10^{-5}$	$10^{-9} - 10^{-6}$
GC	$10^{-9} - 10^{-6}$	$10^{-10} - 10^{-7}$
SP	$10^{-6} - 10^{-3}$	$10^{-6} - 10^{-3}$
SW	$10^{-7} - 10^{-3}$	$10^{-5} - 10^{-3}$
SM	$10^{-8} - 10^{-6}$	$10^{-9} - 10^{-5}$
SC	$10^{-9} - 10^{-7}$	$10^{-10} - 10^{-6}$
ML	$10^{-9} - 10^{-6}$	$10^{-10} - 10^{-6}$
MH	$10^{-9} - 10^{-7}$	$10^{-11} - 10^{-9}$
CL	$10^{-10} - 10^{-8}$	$10^{-10} - 10^{-8}$
CH	$10^{-11} - 10^{-9}$	$10^{-13} - 10^{-6}$

Although soil classification using the USCS system is much more objective as a means of obtaining geotechnical properties, it does not necessarily follow that precise estimations of hydraulic conductivity can be obtained. Estimated hydraulic conductivities from systematically described soil profiles may be more accurate and precise than estimations from USCS soil groups because of the large variations in hydrogeological properties within many USCS soil groups. The USCS has been developed for engineering purposes and is less applicable to hydrogeological situations.

Carsel and Parrish (1988) conducted normal distribution and joint probability distribution tests for saturated hydraulic conductivity and other hydrogeological properties. These were based on the regression equation developed by Rawls, Brankensiek & Saxton (1982) (Equation 4-7). Joint probability distributions can be used to produce a multivariate normal distribution model for Monte Carlo modelling and other simulation studies. Estimations of the saturated hydraulic conductivity for 12 agricultural texture classes, as determined by Carsel and Parrish (1988) by means of Equation 4-7, are indicated in **Table 4.7**.

Table 4.7: Estimated saturated hydraulic conductivity for twelve agricultural soil types (modified from Carsel & Parrish, 1988 and Schulze, 1995)

Soil type	Estimated saturated hydraulic conductivity (m/s)	
	Carsel & Parrish	Schulze
Clay	5.56×10^{-7}	1.67×10^{-7}
Clay loam	7.22×10^{-7}	6.38×10^{-7}
Loam	2.89×10^{-6}	3.61×10^{-6}
Loamy sand	4.05×10^{-5}	1.69×10^{-5}
Silt	6.94×10^{-7}	N.A.
Silt loam	1.25×10^{-6}	1.89×10^{-6}
Silty clay	5.56×10^{-8}	2.50×10^{-7}
Silty clay loam	1.94×10^{-7}	4.17×10^{-7}
Sand	8.25×10^{-5}	5.83×10^{-5}
Sandy clay	3.33×10^{-7}	3.33×10^{-7}
Sandy clay loam	3.64×10^{-6}	1.19×10^{-6}
Sandy loam	1.23×10^{-5}	7.22×10^{-6}

With the development of the ACRU agrohydrological modelling system, Schulze (1995) has shown that the saturated hydraulic conductivity can be estimated from soil textural classes. Estimations of saturated hydraulic conductivity are also indicated in **Table 4.7**

4.2.1 Estimation of saturated hydraulic conductivity based on empirical relationships

Soil fractions (Soil texture)

In general, the following empirical equation is applicable to predict saturated hydraulic conductivity:

$$K_s = a \cdot f(S, M, C, \varepsilon) \quad [4-2]$$

The parameter a is a constant and reflects factors with an effect on saturated hydraulic conductivity other than grain sizes, such as the shape of the grains and packing.

Many authors have related saturated hydraulic conductivity to physical soil properties based on fractions, sand, silt and clay and, in some cases, bulk density (Campbell, 1974; Campbell, 1985; Rawls *et al.*, 1982; Campbell & Shiozawa, 1992). These relationships are expressed as empirical models, many of which have been generated by multiple regression analysis. The more popular equations are the following:

The Campbell equation (1974) is widely used to predict saturated hydraulic conductivity:

$$K_s = a \exp(-6.9C - 3.7M) \quad [4-3]$$

This equation is valid for a bulk density value of 1 300 kg/m³, with grain sizes of 1.025, 0.026 and 0.001 mm for sand, silt and clay respectively.

After extensive research to relate physical soil properties to soil-water retention characteristics, Campbell (1985) points out the relationship between saturated hydraulic conductivity and bulk density, air-entry value and the geometric mean grain diameter. Based on this research, Campbell and Shiozawa (1992) suggested that saturated hydraulic conductivity could merely be expressed as a function of percentages of silt and clay and of the bulk density of the soil. The saturated hydraulic conductivity can be expressed as:

$$K_s = a \left(\frac{1.3}{\rho_b} \right)^{1.3b} \exp(-0.025 - 0.363M - 0.0688C) \quad [4-4]$$

where

$$b = \exp(-0.025 - 0.0363M - 0.0688C)^{-0.5} + 0.2 \exp(0.1332M + 0.47C - \ln^2[\exp(-0.025 - 0.0363M - 0.0688C)])^{0.5} \quad [4-5]$$

and a is a constant to be experimentally determined.

Campbell and Shiozawa (1992) also suggest a more simple equation as a means to derive saturated hydraulic conductivity:

$$K_s = 1.5 \times 10^{-5} \exp(-0.07M - 0.167C) \quad [4-6]$$

Based on soil-water retention and saturated hydraulic conductivity data collected by Luxmoore and

Sharma (1980), Rawls, Ahuja and Brakensiek (1992) developed the following regression model to predict saturated hydraulic conductivity.

$$K_s = 2.7778 \times 10^{-6} \exp(19.523\varepsilon - 8.968 - 0.028C + 0.0002S^2 - 0.009C^2 - 8.395\varepsilon^2 + 0.078S\varepsilon - 0.003S^2\varepsilon^2 - 0.019C^2\varepsilon^2 + 0.00002S^2C + 0.027C^2\varepsilon + 0.001S^2\varepsilon - 0.000004C^2S) \quad [4-7]$$

where the fractions sand and clay and porosity are expressed as percentages. Equation 4-7 has been validated for soils with clay contents ranging from 5 to 60 per cent, sands contents ranging from 5 to 70 per cent and all porosity values representative of field soils.

Particle-size distribution: Coarse grained soils

It has been shown in Chapter 2 that the permeability or intrinsic hydraulic conductivity for water flowing through a narrow tube can be expressed in terms of Poiseuille's law:

$$k = 0.125 \cdot R^2 \quad [4-8]$$

In the case of fluids flowing through a porous medium, the effective pore diameter, R_e , can be related to the effective grain diameter, d_e , by means of the following equation (Vukovic & Soro, 1992):

$$R_e = c f(\varepsilon) d_e \quad [4-9]$$

The parameter, c , is a dimensionless constant depending on a number of porous medium properties (e.g. structure and grain shape), $f(\varepsilon)$ is the function defining the relationship between modelled and actual porous media and d_e is the effective grain diameter related to the tube diameter. Large volumes of the porous media consist of solids through which fluids will not be able to flow. The permeability can be expressed as:

$$k = 0.125 (c f(\varepsilon) d_e)^2 \varepsilon \quad [4-10]$$

The intrinsic hydraulic conductivity is a function of the physical properties of the porous medium. However, Equation 4-10 does not address the effect of the retention forces of the porous media or the physiochemical properties of liquid and porous media. These aspects may significantly influence the hydraulic conductivity in fine-grained soils. However, for homogeneous coarse-grained soils, these aspects may be neglected.

By redefining the constant, c , and the function, $f(\varepsilon)$, Equation 4-10 can be expressed as:

$$k = c_1 g(\varepsilon) d_e^2 \quad [4-11]$$

where

$$c_1 = 0.125 \cdot c^2 \quad [4-12]$$

and

$$g(\varepsilon) = \varepsilon \cdot (f(\varepsilon))^2 \quad [4-13]$$

Saturated hydraulic conductivity can be expressed as

$$K_s = c_1 \cdot g(\varepsilon) \cdot d_e^2 \frac{\rho g}{\eta} \quad [4-14]$$

The effective grain diameter represents the diameter of the grain corresponding to the effective pore diameter in the Poiseuille equation. The effective pore diameter has a significant effect on the saturated hydraulic conductivity of soils. Many researchers have regarded the grain diameter of the smaller part of the soil fraction as the effective pore diameter. Since smaller grains tend to be situated within larger pores, this assumption could be justified. In the case of significant silt and clay contents, saturated hydraulic conductivity will mainly be governed by the fine material fraction.

Saturated hydraulic conductivity varies to a large extent with change in porosity. Vukovic and Soro (1992) show that in the case of sand, saturated hydraulic conductivity may increase by a factor of about 3 for a porosity value increase from 30 to 40 per cent. Since temperature does affect fluid, viscosity, it also has an effect on saturated hydraulic conductivity. In the case of water, the hydraulic conductivity may increase by three per cent for every one degree Celsius increase in temperature (Vukovic & Soro, 1992).

Fluidity, as discussed in Chapter 2, represents the properties of the fluid that effect flow through soil pores. The most important factors with an effect on fluidity are fluid density and viscosity. The viscosity is a function of temperature. The fluidity at 20°C has a value of:

$$f = \frac{\rho_w g}{\eta_{w 20^\circ C}} = 975124 m^{-1} s^{-1} \quad [4-15]$$

The dimensionless constant, c_1 , represents all factors not addressed, the most important probably the shape of the grains. Empirical models frequently assume grains to be spherical and differences in the shape of the grains may have an effect on saturated hydraulic conductivity.

The equation does not consider the effect of electrochemical forces. These forces may have a significant effect on clayey soils. However, in the case of sandy soils with little or no fines, electrochemical forces may be negligible.

Adsorption forces may have a significant effect on saturated hydraulic conductivity, especially if large silt and clay fractions are present. Only the Kozeny-Carman and Zamarin equations take the effect of adsorption into account, by including the specific surface of the soils. Since the saturated hydraulic conductivity basically represents friction forces between water and soil grains, specific surfaces may be a more representative parameter for estimating saturated hydraulic conductivity.

Fourteen frequently used empirical equations were analysed by Vukovic and Soro (1992). The equations have been expressed in the format of Equation 4-14. Predicted hydraulic conductivity was compared to measured hydraulic conductivity and the reliability of the equations was quantified. All fourteen equations discussed can be expressed in the general empirical equation as expressed by Equation 4-14 and are summarised in **Table 4.8**. Detailed descriptions of the fourteen empirical equations are discussed by Vukovic and Soro, (1992) and van Schalkwyk and Vermaak (1999)

Chapter 4: Relationship between geotechnical and hydrogeological properties

Table 4.8: Empirical equations expressed in terms of the general empirical equation (Vukovic & Soro, 1992)

	Author	Coefficient, c_1	Function of porosity, $g(\epsilon)$	Effective grain diameter, D_e
1	Hazen (1930)	4.717×10^{-9}	$4.6 \times 10^{-2} (\epsilon - 0.2)$	d_{10}
2	Amer & Awad (1974)	$3.63 \times 10^{-8} C_u^{0.6} d_{10}^{0.32}$	$\frac{\epsilon^3}{(1-\epsilon)^2}$	d_{10}
3	Shahabi, Das & Tarquin. (1984)	$1.23 \times 10^{-6} C_u^{0.735} d_{10}^{-0.11}$	$\frac{\epsilon^3}{(1-\epsilon)^2}$	d_{10}
4	Kenney, Lau & Ofoegbu. (1984)	5.128×10^{-8} to 1.026×10^{-6}	1	d_5
5	Slichter (Vukovic & Soro, 1992)	1.038×10^{-7}	$\epsilon^{3.287}$	d_{10}
6	Terzaghi (Vukovic & Soro, 1992)	2.082×10^{-2} to 8.204×10^{-2}	$\left[\frac{\epsilon - 0.13}{\sqrt[3]{1-\epsilon}} \right]^2$	d_{10}
7	Beyer (Vukovic & Soro, 1992)	$4.615 \times 10^{-9} \log \frac{500}{C_u}$	1	d_{10}
8	Sauerbrei (Vukovic & Soro, 1992)	3.579×10^{-8}	$\frac{\epsilon^3}{(1-\epsilon)^2}$	d_{17}
9	Pavchich (Pravednyi, 1966)	$3.579 \times 10^{-8} C_u^{0.333}$	$\frac{\epsilon^3}{(1-\epsilon)^2}$	d_{17}
10	USBR (Vukovic & Soro, 1992)	$4.923 \times 10^{-10} d_{20}^{0.3}$	1	d_{20}
11	Krüger (Vukovic & Soro, 1992)	5.076×10^{-9}	$\frac{\epsilon}{(1-\epsilon)^2}$	$\frac{1}{\sum_{i=1}^{i=n} \frac{2\Delta w_i}{d_i^{\max} + d_i^{\min}}}$
12	Kozeny (Vukovic & Soro, 1992)	6.409×10^{-8}	$\frac{\epsilon^3}{(1-\epsilon)^2}$	$\frac{1}{\frac{3\Delta w_1}{2d_1} + \sum_{i=2}^{i=n} \frac{\Delta w_i (d_i^{\max} + d_i^{\min})}{2d_i^{\max} d_i^{\min}}}$
13	Zunker (Vukovic & Soro, 1992)	1.377×10^{-8} to 4.041×10^{-9}	$\frac{\epsilon}{1-\epsilon}$	$\frac{1}{\frac{3\Delta w_1}{2d_1} + \sum_{i=2}^{i=n} \frac{\Delta w_i (d_i^{\max} - d_i^{\min})}{d_i^{\max} d_i^{\min} \ln \frac{d_i^{\max}}{d_i^{\min}}}}$
14	Zamarin (Vukovic & Soro, 1992)	8.276×10^{-8}	$\frac{\epsilon^3}{(1-\epsilon)^2}$	$\frac{1}{\frac{3\Delta w_1}{2d_1} + \sum_{i=2}^{i=n} \frac{\Delta w_i \left(\ln \frac{d_i^{\max}}{d_i^{\min}} \right)}{d_i^{\max} - d_i^{\min}}}$

Skabalanovich (1961) investigated the reliability of the Hazen (1930), Slichter, Sauerbrei, Kruger, Zunker and Zamarin (Vukovic & Soro, 1992) equations in relation to a sandy aquifer. Saturated hydraulic conductivity values were derived from pumping test evaluations and these were compared to the

predicted saturated hydraulic conductivity for the above-mentioned equations. He found that all equations consistently predicted, on average, lower hydraulic conductivity values than were derived from the pumping tests.

Skabalanovich (1961) showed that in 80 per cent of the cases, calculated hydraulic conductivity values were two times higher or lower than the actual saturated hydraulic conductivity of the water-bearing strata. The Sauerbrei (Vukovic & Soro, 1992) equation yielded the most reliable estimation of saturated hydraulic conductivity, but about 60 per cent of the relevant calculations showed significant deviation. Skabalanovich (1961) concluded that the equations under investigation may not yield reliable estimations of saturated hydraulic conductivity and recommended that the equations should not be used for the site-specific estimation of saturated hydraulic conductivity.

Particle-size distribution, clay fraction and Atterberg limits: Fine-grained soil

Research in deriving empirical equations for saturated hydraulic conductivity in coarse-grained soil is relatively well-established, due largely to numerous investigations into sandy aquifers. In the case of clayey soils, researchers tend to calculate saturated hydraulic conductivity on the basis of permeability tests and research tends to focus in obtaining representative measured values of saturated hydraulic conductivity for use in clay liners and other engineering structures. Little research has been done on deriving saturated hydraulic conductivity from empirical equations. In addition, it is much more difficult to obtain empirical expressions for estimating saturated hydraulic conductivity in clays than in coarse-grained soils (Tavenas, LeBlond, Jean & Leroueil, 1983b).

Taylor (1948) has shown that general equations, which are valid for coarse-grained soils, are not valid for clayey soils. Samarasinghe, Huang and Drnevich (1982) suggest the following empirical equation:

$$K_s = c \frac{e^b}{e + 1} \quad [4-16]$$

where b and c are experimentally determined constants. Equation 4-16 is applicable to normally consolidated clays, with the value of b typically between 4 and 5 and c a constant indicating soil characteristics. Taylor suggested an empirical linear relationship between the logarithm of saturated hydraulic conductivity and the void ratio:

$$\lg(K_s) = \lg(K_0) \frac{e_0 - e}{C_k} \quad [4-17]$$

where C_k is the permeability change index and K_0 and e_0 respectively are the *in situ* hydraulic conductivity and void ratio values. This equation can be used in consolidation studies where the void ratio and therefore the hydraulic conductivity change as a result of the load applied to the soils. Mesri and Olson (1971) suggest a linear relationship over a very wide range of void ratios:

$$\lg(K_s) = a \lg(e) + b \quad [4-18]$$

where a and b are experimentally determined constants. According to Tavenas *et al.* (1983a) equations 4-13, 4-14 and 4-15, all derived from experiments on remoulded clay, may be valid for some clays or certain void ratio ranges but may not be valid in other circumstances.

Hydraulic conductivity in clayey soils is a function of *inter alia* the void ratio of the clay. Tavenas *et al.* (1983a) points out that clay fraction and plasticity index are also significant parameters in determining the hydraulic conductivity. These parameters are related to the *hydraulic conductivity – void ratio* relation by the empirical parameter I_c :

$$I_c = PI + C$$

[4-19]

where PI is the plasticity index and C is the clay fraction. **Figure 4.1** indicates void ratio versus logarithmic saturated hydraulic conductivity relations as a function of the empirical parameter I_c . **Figure 4.1** can be used as an indication of the saturated hydraulic conductivity of clay soils.

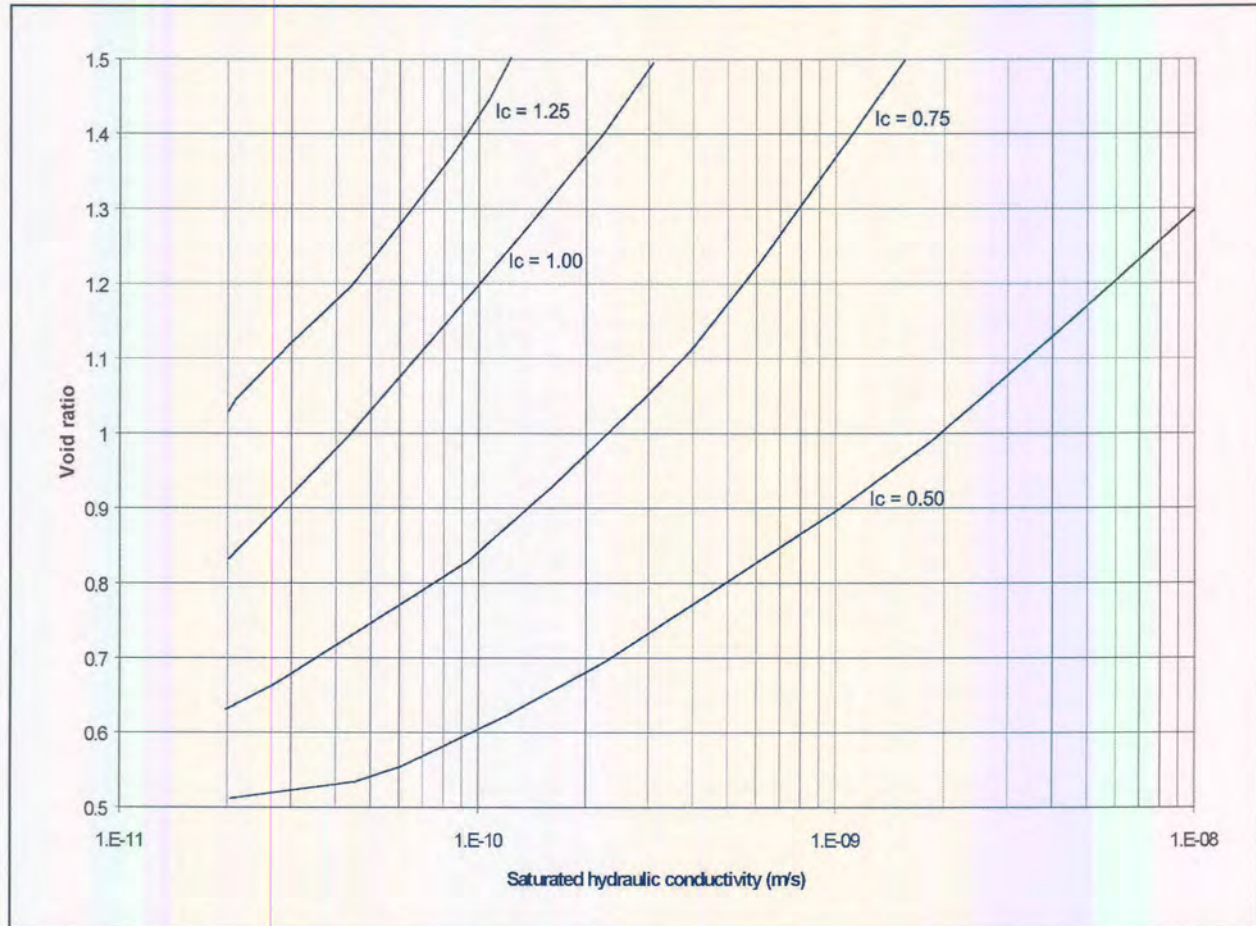


Figure 4.1: Estimation of saturated hydraulic conductivity for fine-grained soils (after Tavenas *et al.*, 1983a)

However, the relationship as shown in **Figure 4.1** may not very reliable, since many factors affect saturated hydraulic conductivity in clayey soils.

The rate of consolidation of clay soils under applied stress is partly governed by their permeabilities (Tavenas *et al.*, 1983a). Consolidation tests (in particular conventional step-loaded tests) are routinely conducted during geotechnical investigations. The hydraulic conductivity of soft clays can be determined by observing the initial consolidation rate; the *in situ* void ratio at zero applied pressure under increasing pressure and hence the decreasing void ratio. According to the Terzhagi theory, hydraulic conductivity can be calculated by the equation (Tavenas *et al.*, 1983b):

$$K_s = \frac{C_v(1+e)\sigma_1'}{0.434 \cdot \gamma_w C_c} \quad [4-20]$$

where C_v is the coefficient of consolidation, assumed to be a constant, and C_c is the compression index of the clay. The Terzhagi theory makes use of a series of assumptions that do not fit the behaviour of natural clay (Tavenas *et al.*, 1983b). As a result, significant errors of saturated hydraulic conductivity values may occur. Tavenas *et al.* (1983b) proposes that hydraulic conductivity values, obtained by indirect methods,

should not be used in engineering applications. However, these values can be used to get an indication of hydraulic conductivity.

4.2.2 Soil-water retention characteristics

Brutsaert (1967) derives the following relationship with regard to saturated hydraulic conductivity:

$$K_s = c \frac{(\varepsilon - \theta_r)^2}{\psi_a^2} \frac{\lambda^2}{(\lambda + 1)(\lambda + 2)} \quad [4-21]$$

where c is a constant representing the effects of the fluid. According to Brutsaert, the theoretically derived constant, c , equals 2.70 in the case of water. The parameter, λ , known as the Brooks and Corey pore size distribution index, is important in describing the soil-water characteristic curve and will be discussed in Section 4.3. Rawls *et al.* (1992) investigated the relationship between soil texture, bulk density and water retention parameters and found that the empirically derived value of the constant c is 0.21, when geometric mean Brooks and Corey pore-size distribution index values are applied. They attained a correlation coefficient of 0.96.

Campbell (1985) suggests that saturated hydraulic conductivity can be calculated by means of the following equation:

$$K_s = \frac{\sigma^2 \theta_s^2}{2\rho_w \psi_a^2 (2b + 1)(2b + 2)} \quad [4-22]$$

where b is a pore-size distribution index similar in definition to the Brooks and Corey parameter, but with a different value.

Ahuja, Naney & Williams (1985) show that saturated hydraulic conductivity can be related to water content at -33kPa in terms of the following equation:

$$K_s = 0.00282 \cdot (\varepsilon - \theta_{33})^4 \quad [4-23]$$

where θ_{33} is the water content at -33 kPa. Experimental studies have indicated that equation 4.23 is suitable for soils with less than 65 percent sand fraction and less than 40 percent clay fraction. According to Rawls *et al.* (1992), equation 4-23 is the best model to use for the study of a wide range of soil properties.

Hutson (1984) developed an empirical equation relating the soil-water characteristic curve to physical soil properties. According to Hutson, the saturated hydraulic conductivity can be estimated from the capillary model developed by Childs & Collis-George (1950):

$$K_s = 5167.36 \cdot \frac{\theta_s}{a_H^2} \left[\frac{Si^{2b_H} + 1}{2b_H + 1} - \frac{Si^{2b_H} + 2}{2b_H + 2} + \frac{(1 - Si)^2}{Si^{-2b_H}} \right] \quad [4-24]$$

where

$$Si = \frac{2b_H}{1 + 2b_H} \quad [4-25]$$

where a_H and b_H are parameters to be determined experimentally.

4.3 Estimation of soil-water retention characteristics

The theoretical framework for unsaturated flow has been well-established (Chapter 2). However, the direct measurement of unsaturated soil hydraulic properties requires demanding laboratory or field tests. Since it has been found that the indirect determination of unsaturated soil hydraulic properties is adequate for most practical cases (Papagiannakis & Fredlund, 1984), the indirect application of methods to determine unsaturated hydraulic conductivity has become acceptable.

Various models have been developed to determine the unsaturated hydraulic conductivity indirectly from saturated hydraulic conductivity and the soil-water characteristic curve (Burdine, 1953; Millington & Quirk, 1961; Brooks & Corey, 1964; Campbell, 1974; Mualem, 1976; Fredlund Xing & Huang, 1994 and others). These models will be discussed in the following sections. Since the soil-water characteristic curve is used as basis for predicting unsaturated hydraulic conductivity, it is important to describe it accurately.

Soil-water characteristic curves are not readily available in South Africa and many other countries. Laboratory soil-water retention tests can easily be conducted for site- specific investigations but in the case of regional aquifer vulnerability and recharge investigations, soil-water retention data can only be acquired at great cost. In these cases, it may be viable to estimate soil-water retention characteristics from available soil data such as soil fractions and particle-size distribution curves.

4.3.1 Soil fractions

Much research has been conducted to relate soil-water retention data to basic soil properties applying regression analysis (Bartelli & Peters, 1959; Salter, Berry & Williams, 1966; Peterson, Cunningham & Matelski, 1968; Gupta & Larson, 1979; Brakensiek, Engleman & Rawls, 1981; McCeun, Rawls & Brakensiek, 1985; Rawls *et al.*, 1982 and others). Similar regression models have been established in South Africa, especially with regard to soils in certain geographical crop production and irrigation areas. (Van der Merwe, 1973; Hutson, 1984 and others). Much of the early research was conducted at selected soil suctions, most notably at -33 and -1500 kPa. Little statistical evidence has been provided to support the regression models. Gupta and Larson (1979) propose that the regression coefficients can be calculated in terms of the following general regression equation:

$$\theta(\psi) = a_1 S + a_2 M + a_3 C + a_4 OM + a_5 \rho_b \quad [4-26]$$

where the regression coefficients at the various soil suction values are indicated in **Table 4.9**:

Table 4.9: Regression coefficients for volumetric water content at various soil suction values (Gupta & Larson, 1979)

Soil suction (kPa)	Regression coefficients					R
	a_1	a_2	a_3	a_4	a_5	
-4	7.053	10.242	10.070	6.333	-321.2	0.950
-7	5.678	9.228	9.135	6.103	-269.6	0.959
-10	5.018	8.548	8.833	4.966	-242.3	0.961
-20	3.890	7.066	8.408	2.817	-187.8	0.962
-33	3.075	5.886	8.039	2.208	-143.4	0.962
-60	2.181	4.557	7.557	2.191	-92.8	0.964
-100	1.563	3.620	7.154	2.388	-57.6	0.966
-200	0.932	2.643	6.636	2.717	-22.1	0.967
-400	0.483	1.943	6.128	2.925	-2.0	0.962
-700	0.214	1.538	5.908	2.855	15.3	0.954
-1000	0.074	1.334	5.802	2.653	21.5	0.951
-1500	0.059	1.142	5.766	2.228	26.7	0.947

Chapter 4: Relationship between geotechnical and hydrogeological properties

Hutson (1984) has established the following general regression model for eight South African soils:

$$\theta(\psi) = b_1 + b_2 C + b_3 M + b_4 \rho_b \quad [4-27]$$

where the regression coefficients at the various soil suction values are indicated in **Table 4.10**.

Table 4.10: Regression coefficients for volumetric water content at various soil suction values for South African soils (Hutson, 1984)

Soil suction (kPa)	Regression coefficients				R
	b_1	b_2	b_3	b_4	
10	55.8	3.65	5.54	30.3	0.681
30	-15.5	3.84	5.72	46.3	0.764
100	29.0	3.61	4.41	4.9	0.769
500	158.8	3.47	1.70	-83.8	0.823
1500	602.0	3.22	3.08	-26.0	0.785

Soil-water retention regression models in South Africa have in general been developed for certain geographical regions and may not be applicable to all soils in South Africa. Turner (Hutson, 1984) developed a regression model with correlation coefficients higher than 0.90. However, the regression equation was developed for KwaZulu-Natal clay and clay loam, characterised by high clay contents, low bulk densities and high organic matter, hardly representative of typical South African soils.

4.3.2 Description of the soil-water characteristic curve

Several mathematical functions have been developed to describe and characterise the soil-water characteristic curve. Soil-water retention tests are usually conducted at a series of suction points. The accuracy with which the soil-water characteristic curve is described, will depend on the number of suction points used. The various mathematical functions are fitted through the points, and the values of the fitting parameters can be determined.

The residual water content specifies the upper limit in volumetric water content with a negligible effect on flow in the soil. As such, many researchers have described the soil-water characteristic curve as a function of the reduced water content, also known as the effective degree of saturation:

$$\Theta = \frac{\theta - \theta_r}{\theta_s - \theta_r} \quad [4-28]$$

Since it is very difficult to determine the value of the residual water content, many authors regard the residual water content as another fitting parameter rather than to determine it experimentally.

Gardner (1958) proposes that the soil-water characteristic curve can be derived from the following equation:

$$\Theta = \frac{1}{1 + a\psi^n} \quad [4-29]$$

The fitting parameter, a , is related to the inverse air-entry value of the soil and denotes the suction value at which air will enter the soil pores. The parameter, n , is related to the slope of the soil-water characteristic curve.

Brooks and Corey (1964) have developed the following function to describe the soil-water characteristic curve:

$$\Theta = \left(\frac{\psi_a}{\psi} \right)^\lambda \quad [4-30]$$

where the parameter, λ , is known as the Brooks and Corey pore-size distribution index and is a function of the physical soil properties. This parameter is related to the slope of the soil-water characteristic curve.

Farrel and Larson (1972) proposed the following function to describe the soil-water characteristic curve:

$$\psi = \psi_a \exp[\alpha(1 - \Theta)] \quad [4-31]$$

Campbell (1974) proposed a function similar to the Brooks and Corey model:

$$\psi = \psi_a \left(\frac{\theta}{\theta_s} \right)^{-b} \quad [4-32]$$

where b is a parameter related to the Brooks and Corey pore-size index.

The Brooks & Corey (1964) function does not describe the soil-water characteristic curve accurately in the lower suction (or near-saturated) ranges of the soil-water characteristic curve, Van Genuchten (1980) proposes the following equation, closely related to the Gardner equation:

$$\Theta = \left[\frac{1}{1 + (\alpha\psi)^n} \right]^m \quad [4-33]$$

The fitting parameter, α , is related to the inverse air-entry value, the parameter, n , is related to the slope of the soil-water characteristic curve and the parameter, m , is related to the slope of the soil-water characteristic curve at higher suction values.

Based on the Campbell function, Williams, Ross & Bristow (1992) developed the following function to describe the soil-water characteristic curve for a large range of Australian soils:

$$\ln(\Theta) = a + b \ln(\psi) \quad [4-34]$$

McKee and Bumb (1984) have developed the following function to describe the soil-water characteristic curve:

$$\Theta = a \exp(c\psi - b) \quad [4-35]$$

This equation is also known as the Boltzmann distribution. In order to improve the fit at lower suction values, McKee and Bumb suggest the following modification:

$$\Theta = \frac{1}{1 + a \exp(c\psi - b)} \quad [4-36]$$

Hutson (1984) proposes that the soil-water characteristic curve can be described by a function comprising of two parts, i.e., an exponential function (based on the Campbell's function, 1974) and a parabolic function. The parabolic function describes the soil-water characteristic curve at near saturated conditions while the exponential function describes the less saturated conditions. Hutson (1984) identifies a point of inflection, θ_m , which occurs on the soil-water characteristic curve and indicates the point between the parabolic and exponential functions.

The soil-water characteristic curve can be described as:

$$\psi = a_H \left(\frac{\theta}{\theta_s} \right)^{-b_H} \quad \text{for } \psi \geq \psi(\theta_{in}) \quad [4-37]$$

and

$$\psi = a_H \left(1 - \frac{\theta}{\theta_s} \right)^{0.5} \left(\frac{\theta_{in}}{\theta_s} \right)^{-b_H} \left(1 - \frac{\theta_{in}}{\theta_s} \right)^{-0.5} \quad \text{for } \psi(\theta_{in}) \geq \psi \geq 0 \quad [4-38]$$

The parameters, a_H , b_H and θ_{in} are determined by curve fitting.

Fredlund and Xing (1994) related the fitting parameters to physical soil properties. They suggest that the following equation can describe the soil-water characteristic curve over the entire soil suction range, up to 1 000 000 kPa, that, according to some researchers (Crony & Coleman, 1961), is the range where the volumetric water content approaches nil. This assumption is supported by thermodynamic considerations.

$$\theta = C(\psi) \frac{\theta_s}{\left\{ \ln \left[e + \left(\frac{\psi}{a} \right)^n \right] \right\}^m} \quad [4-39]$$

where e is the natural number 2.71828..., a is the fitting parameter related to the air-entry value, n and m are related to the pore-size function in the lower and higher soil suction ranges respectively. Fredlund and Xing (1994) introduced a correction function, $C(\psi)$, that corresponds to the residual water content. Leong and Rehardjo (1997a) show that a good fit can be obtained in instances where the correction factor is equal to one. The correction factor can be expressed as:

$$C(\psi) = \frac{\ln \left(1 + \frac{\psi}{C_r} \right)}{\ln \left(1 + \frac{1\,000\,000}{C_r} \right)} \quad [4-40]$$

According to Leong and Rahardjo (1997), these functions can be grouped into functions that describe typical exponential curves (such as those of Brooks & Corey, 1964; Farrel & Larson, 1972; Williams, Pebble, Williams & Hignett, 1983 and McKee & Bumb, 1984) and functions that describe typical sigmoidal curves (Gardner, 1958; Van Genuchten, 1980; McKee & Bumb, 1984 and Fredlund & Xing, 1994). Leong and Rehardjo (1997a) show that all these equations can be derived from the following generic function:

$$a_1 \Theta^{b_1} + a_2 \exp(a_3 \Theta^{b_1}) = a_4 \psi^{b_2} + a_5 \exp(a_6 \psi^{b_2}) + a_7 \quad [4-41]$$

They show that functions pertaining to typical sigmoidal curves describe the soil-water characteristic curve better than non-sigmoidal functions. They also show that functions with four fitting parameters describe the soil-water characteristic curve better than functions with two or three fitting parameters. The Fredlund and Xing function (Equation 4-39) produces better results than the Van Genuchten function (Equation 4-33). However, in the case of sandy soils, the Van Genuchten function produces better results.

4.3.3 Estimation of parameters that describes the soil-water characteristic curve

Many researchers have developed empirical models describing typical soil-water hydraulic parameters from basic soil data such as soil fractions, bulk density and particle-size distribution curves. In addition, many researchers have developed empirical models describing fitting parameters for various functions of soil-water characteristic curves. They argue that although the fitting parameters for the various soil-water characteristic curve functions are obtained through non-linear regression and other curve-fitting methods, the fitting parameters do have a physical meaning. Leong and Rehardjo (1997a) show that the parameters a and α used in the Gardner (1958), Farrel and Larson (1972), Van Genuchten (1980) and Fredlund and Xing (1994) functions are either directly or inversely related to the air-entry value. Similar relationships of fitting parameters with pore-size distribution and other soil-water hydraulic parameters have been noted.

Rawls *et al.* (1992) propose that the air-entry pressure, ψ_a , the pore-size distribution index, λ , and the residual water content, θ_r (the three unknown parameters in the Brooks and Corey (1964) equation) can be estimated by means of the following regression equations:

$$\begin{aligned} \psi_a = & 0.001\rho_w g \exp(5.340 + 0.185C - 2.484E - 0.002C^2 - 0.044SE - \\ & 0.617CE + 0.001S^2E^2 - 0.009C^2E^2 - 0.00001S^2C + \\ & 0.009C^2S - 0.0007S^2E + 0.000005C^2S - 0.5E^2C) \end{aligned} \quad [4-42]$$

$$\begin{aligned} \lambda = & \exp(-0.784 + 0.018S - 1.062E - 0.00005S^2 - 0.003C^2 + 1.111E^2 - \\ & 0.031SE + 0.0003S^2E^2 - 0.006C^2E^2 - 0.000002S^2C + 0.008C^2E - \\ & 0.007E^2C) \end{aligned} \quad [4-43]$$

$$\begin{aligned} \theta_r = & -0.018 + 0.0009S + 0.005C + 0.029E - 0.0002C^2 - 0.001SE - \\ & 0.0002C^2E^2 + 0.0003C^2E - 0.002E^2C \end{aligned} \quad [4-44]$$

where C is the clay content expressed as a percentage value and is valid for soils with between 5 and 60 per cent clay. S is the sand content expressed as a percentage value and is valid for soils with between 5 and 70 per cent sand.

Rawls *et al.* (1992) propose that the Campbell (1974) constant, b , (Equation 4-32) is related to the Brooks and Corey pore-size distribution index by:

$$b = 1/\lambda \quad [4-45]$$

They show that the Van Genuchten parameters n and m (Equation 4-33) are also related to the Brooks and Corey pore-size distribution index by:

$$n = \lambda + 1 \quad [4-46]$$

$$m = \frac{\lambda}{\lambda + 1} \quad [4-47]$$

Williams *et al.* (1992) have developed eight functions describing the parameters a and b of the Williams *et al.* (1983) equation (Equation 4-34). Williams *et al.* (1992) employed eight frequently measured properties, including a percentage of coarse sand, fine sand, silt and clay, bulk density, organic matter and texture groups, as defined by Northcote (1971) and a structure index, where different values were attributed to structured and intact soils.

Vereecken, Maes, Feyen & Darius (1989) developed multiple regression models to estimate the residual water content, saturated volumetric water content and the Van Genuchten parameters, α and n (Equation

4-33) from basic soil properties. After investigating several closed-form modifications of the Van Genuchten equation, Vereecken *et al.* (1989) have concluded that the van Genuchten equation can be expressed with the restriction, $m = 1$ without significant loss of flexibility.

Vereecken *et al.* maintain that the soil-water hydraulic parameters can be expressed as:

$$\theta_s = 0.81 - 0.283\rho_b - 0.001C \quad [4-48]$$

$$\theta_r = 0.015 + 0.005C + 0.014C_m \quad [4-49]$$

$$\log(\alpha) = -2.486 + 0.025S - 0.351C_m - 2.617\rho_b - 0.023C \quad [4-50]$$

$$\log(n) = 0.053 - 0.009S - 0.013C + 0.00015S^2 \quad [4-51]$$

Vereecken *et al.* (1989) did extensive statistical analysis on the relationships, including a sensitivity analysis. Correlation ranged from 0.84 for saturated volumetric water content to 0.56 for the Van Genuchten fitting parameter, n .

The sensitivity analysis indicated that the saturated volumetric water content was the most sensitive parameter, with a relative sensitivity of 200 for both over- and under-estimations of 30 per cent. The residual water content was the least sensitive parameter, with a relative sensitivity lower than 10 for over- and under-estimations of 90 per cent. The Van Genuchten parameters, α and n , were more sensitive for under-estimations than for over-estimations of the parameter value. The parameter, α , was found to be more sensitive than the parameter, n .

Tinjum, Bensen & Blotz (1997) investigated the relationship between soil-water characteristic curves and compacted clay. During the investigation, optimum water contents were determined by means of the Standard Proctor and Modified Proctor Compactive tests. In addition, Atterberg limits were determined. Tinjum *et al.* then determined the relationship between these values and the Van Genuchten parameters n and α (Equation 4-33) by means of stepwise regression. They suggested that the Van Genuchten parameters could be expressed as:

$$\log(\alpha) = -1.127 - 0.017PI - 0.092(w - w_{opt}) - 0.263c_{proc} \quad [4-52]$$

$$n = -1.060 + 0.0002PI - 0.0005(w - w_{opt}) \quad [4-53]$$

where c_{proc} is a constant indicating the different compaction tests applied. For the Standard Proctor tests the value of the constant is equal to 1, while for the Modified Proctor, the value is equal to -1. Tinjum *et al.* (1997) found a correlation coefficient of 0.76 and 0.65 for α and n respectively. Better correlation could have been obtained if the parameters were correlated with the liquid limit rather than with the plasticity index. Measurements of the plasticity limit are notoriously subjective, resulting in large discrepancies.

Peterson, Moldrup, Jacobsen & Rolsten (1996) found strong relationships between the specific surface area of the soil and water retention at high suction values (1 500 kPa). However, weak relationships at lower suction values were observed. According to Peterson *et al.* (1996), specific surface areas play a dominant role in water retention at high suction values. This is mainly due to the permanent negative charge on clay particles and the polar nature of water. At lower suction values, physical attributes such as structure and packing are likely to overrule the effect of specific surface areas. In contrast to the findings of Peterson *et al.*, Call (1957), Campbell and Shiozawa (1992) and Banin and Amiel (1970) all found strong relationships between specific surface areas and water retention at low suction values. The apparent discrepancy may be due to the higher clay contents of the soils tested by said authors. The clay particles are situated between larger pores of the soil and override the effect of packing in the soil. Peterson *et al.* (1996) found a higher correlation between soil-water retention and specific surface area expressed in volumetric units ($m^2 \cdot m^{-3}$) than in weight units ($m^2 \cdot g^{-1}$). Peterson *et al.* (1996) found a correlation coefficient of 0.70 for 29 soils tested. The correlation coefficient is higher for deeper soils, probably because of less preferential pathways in these zones.

Many authors infer soil-water hydraulic parameters from grain-size distribution data (Jonasson, 1992; Paydar & Cresswell, 1996; Tyler & Wheatcraft, 1992; Smettem, Bristow, Ross, Haverkamp, Cook & Johnson, 1994). It was found that the slope of the soil-water characteristic curve could be related to the slope of the particle-size distribution curve, assuming that both curves conformed to the power law (Chang & Uehara, 1992; Tyler & Wheatcraft, 1992; Smettem *et al.*, 1994). Tyler and Wheatcraft (1992) show that by assuming that the fractal increment obtained from particle-size distribution curves can be used to estimate the fractal dimension of the pore space, the relationship between the slope of the particle-size distribution curve and the slope of the soil-water characteristic curve can be determined by means of:

$$\ln b = -\ln(n_d + 1) \quad [4-54]$$

where b is the Campbell parameter (Equation 4-32) related to the slope of the soil-water characteristic curve and n_d represents the slope of the particle-size distribution curve. Chang and Uehara (1992) suggest a linear relationship between the inverse Campbell parameter and the slope of the particle-size distribution curve.

$$\ln b = 0.4 - \ln n_d \quad [4-55]$$

Smetten *et al.* (1994) suggest the following linear relationship:

$$\ln b = -0.02 - \ln n_d \quad [4-56]$$

If the air-entry value and saturated volumetric water content are known, the Campbell parameter can be estimated and the soil-water characteristic curve can be derived by means of the Campbell equation (Equation 4-32).

4.3.4 Physico-empirical equations relating the soil-water characteristic curve with particle-size distribution curves

Arya and Paris (1981) and Haverkamp and Parlange (1986) propose that the soil-water characteristic curve could be estimated from the particle-size distribution curve. This approach is based on the similarity in shape of the two curves. The similarity in shape can be explained by the LaPlace equation that relates soil suction to pore radii. As water drains from the soil, larger pores are drained first and water retention is dominated by water occurring in smaller pores. Since pore-size distribution is related to grain-size distribution, models based on the LaPlace equation can be applied to estimate the soil-water characteristic curve from the particle-size distribution curve. According to the LaPlace equation and when said fluid is water, soil suction can be expressed as:

$$\psi = \frac{0.146}{R} \quad [4-57]$$

At a given soil suction, all pores with a radius size equal to or smaller than the corresponding pore radius are filled with water. Higher soil suction values will result from drainage of pores with radii larger than the corresponding soil suction value. Haverkamp and Parlange (1986) assumed that the relationship between grain diameter and pore radius is expressed by constant, λ_p , representing the packing characteristics of the soil.

$$d = \lambda_p R \quad [4-58]$$

Equation 4-58 is only true for uniform soils. In these cases, the packing constant may range from 4.8309 to 8.8889 for pyramidal and tetrahedral arrangements respectively (Gupta & Larson, 1979). However, the arrangements in field soils are much more complicated and the packing parameter is a function mainly of grain-size distribution and bulk density.

According to Arya and Dierolf (1992), the volumetric water content, θ_i , corresponding to the upper limit of the i th grain-size range, can be computed from:

$$\theta_i = \sum_{j=1}^{j=i} V_j \cdot \rho_b \quad i = 1, 2, \dots, n \quad [4-59]$$

where V_j is the volume of soil filled with water. For small particle-size intervals, the average volumetric water content, θ_i^* , corresponding to the midpoint of a given particle size class is:

$$\theta_i^* = (\theta_i + \theta_{i+1}) / 2 \quad [4-60]$$

The corresponding soil suction value, ψ_i , can be obtained by means of the LaPlace equation:

$$\psi_i = \frac{0.146}{R_i} \quad i = 1, 2, \dots, n \quad [4-61]$$

where R_i , is the corresponding pore radius.

Arya and Dierolf (1992) propose that the pore radius can be estimated from the grain diameter by means of the following equation:

$$R_i = \left(\frac{(0.5d_i)^3 4e}{3\alpha} \right)^{0.5} \quad i = 1, 2, \dots, n \quad [4-62]$$

where R_i is pore radius at the i th grain-size range, d_i is the grain diameter and α is a parameter representing the effective pore length associated with each grain, to be determined empirically. The parameter α has been derived from a previous model proposed by Arya and Paris (1981):

$$R_i = 0.5d \left[\frac{4en_i^{(1-\alpha_o)}}{6} \right]^{0.5} \quad [4-63]$$

Equation 4.62 suggests that the pore-size is a function of the number on particles, n_i . No physical explanation can be provided for this situation and Arya and Dierolf suggested that the parameter can be replaced by a single empirical parameter, α .

Arya and Dierolf (1992) found that the value for α , varied between 3 and 15 mm while the average value was found to be 9.38 mm for all soils tested.

The subsequent implementation of the model with an effective pore length of 9.38 mm indicates a high correlation with measured data. However, the authors do not provide any statistical evidence for the relationship. Arya and Dierolf (1992) are not sure if the effective pore length is related to grain size and, if so, how it varies from soil to soil. In a subsequent sensitivity analysis where the value of the effective pore length is estimated at between 5 and 13 mm., the predicted water content varies only slightly and the curves remain within the range of measured data. Variation of the effective pore length compared to pore radius is therefore negligible.

4.3.5 Application of fractal geometry in soil-water characteristic curves

In the last two decades, many scientists have recognised the scale-invariable or self-similar behaviour of objects and processes. The concepts of fractals, as discussed at length by Mandelbrot (1983) have

revolutionised the way scientists view many natural systems. Fractal geometry views the world as a multitude of scales, each with levels of detail and intricacy. Examples of fractal objects and processes are a rugged coastline, atmospheric turbulence, the fracturing of polycrystalline materials and natural porous media. In each case, the object appears the same, or the process appears to repeat itself in a similar manner regardless of the scale at which the object or process is observed. The understanding of fractal concepts has led to the quantification of many disordered systems.

Tyler and Wheatcraft (1990) show that the soil-water retention function, as expressed by Brooks and Corey (1964), can be developed for the Sierpinski carpet. The fractal function can be expressed as:

$$\frac{\theta}{\theta_s} = \left(\frac{\psi}{\psi_a} \right)^{D-2} \quad [4-64]$$

where D represents the fractal dimension of the carpet. In contrast to the exponent function of the Brooks and Corey equation, the Sierpinski carpet exponent, $D-2$ is physically meaningful and is defined by the recursion algorithm chosen (Tyler & Wheatcraft, 1990). By choosing the Sierpinski model as a model for fractal soil pore space, Tyler and Wheatcraft, in effect, map the three-dimensional soil pore network to a plane (Perrier Rieu, Sposito & De Marsily, 1996).

Rieu and Sposito (1991) have developed a lacunarity model of an aggregated soil based on the space partition of the solid interior into a specific number of parts that are then reduced by a specified factor. The resulting model can be expressed as:

$$\theta(\psi) = \theta_s - 1 + \left(\frac{\psi_{\min}}{\psi} \right)^{3-D} \quad 0 < D < 3 \quad [4-65]$$

Rieu and Sposito (1991) found excellent fits to data of six soils varying from sand to clay, with D ranging from 2.758 to 2.986 for sand and clay respectively. Perrier *et al.* (1996) maintain that, although Equations 4-64 and 4-65 do have identical fractal porosities, the two models do not portray soil-water properties in the same way, with the result that significant differences occur in the fractal dimension, D , even though excellent correlation has been achieved for both models.

Tyler and Wheatcraft (1992) show that the Arya and Paris (1981) model (Equation 4-62) can be represented in terms of fractal geometry. According to the original Arya and Paris model the capillary tube length, h_i , associated with particle diameter is calculated by:

$$h_i = d_i N_i \quad [4-66]$$

Tyler and Wheatcraft (1992) suggest that the pore space can be scaled by the power law:

$$h_i = F \cdot d_i^{1-D} \quad [4-67]$$

where F can be evaluated if the “ruler” length, N_i , is assumed to be the straight line length of the pore trace. Equation 4-67 can therefore be represented as:

$$h_i = d_i N_i^D \quad [4-68]$$

Turcott (1986) shows that the particle sizes of many geological materials can be expressed by the following fractal power function in the form:

$$N(0.5d_i)^D = c \quad [4-69]$$

where c is a constant.

The applications of fractal geometry to soil-water characteristic curves may result in a better understanding and, ultimately, a more accurate estimation of soil-water characteristic curves from readily available soil data. However, much research has yet to be conducted regarding the type of model to be used in these cases. Tyler and Wheatcraft (1992) acknowledge that the theoretical development is not yet adequate. Perrier *et al.* (1996) state that fractal analysis of the water retention curve cannot be carried out without also analysing the underlying fractal object with regard to its geometrical interpretation. Simulations regarding random fractal soil structures show that connectivity is an important aspect in soil hydraulic properties (Perrier, Mullen, Rieu & De Marsily, 1995). The effect of hysteresis should also be considered.

4.4 Estimation of unsaturated hydraulic conductivity

As already stated, the vadose zone is characterised mainly by unsaturated flow. The unsaturated hydraulic conductivity is not a constant, but a function of the volumetric water content that in turn varies in accordance with precipitation events and evapotranspiration. No other soil property shows greater variability than unsaturated hydraulic conductivity. It may vary by up to ten orders of magnitude within a *single* soil horizon, depending on the volumetric water content. Since unsaturated hydraulic conductivity is an important aspect of fluid flow analysis in the vadose zone, it is necessary to be able to determine this property. A continuous record of volumetric water content with time is required to determine travel time and recharge. Unsaturated hydraulic conductivity as a function of volumetric water content can be determined by direct measurements or calculated by indirect procedures and prediction models. Although direct measurement is by far preferable, it is for a number of reasons highly impractical in most unsaturated flow problem situations:

- Direct measurements are costly and time-consuming.
- Hydraulic properties of soil are hysteretic in nature, i.e. different relationships between unsaturated hydraulic conductivity and volumetric water content exist depending on the wetting or drying process.
- Soil hydraulic properties are highly variable in nature and large amounts of data are required to accurately represent the field value of unsaturated hydraulic conductivity.
- Unsaturated hydraulic conductivity may vary by a few orders of magnitude within the water content range of interest. Most measurement systems cannot efficiently cover the entire range.

(Mualem, 1992)

These restrictions are especially relevant in the case of regional or catchment studies where only basic soil data are available. Several approaches have been followed to estimate the unsaturated hydraulic conductivity by means of indirect methods.

Unsaturated hydraulic conductivity can be expressed as a function of volumetric water content, soil suction or pore-water pressure. Many researchers prefer to express the unsaturated hydraulic conductivity in terms of the relative hydraulic conductivity, by means of the following equation:

$$K_r(\psi) = \frac{K(\psi)}{K_s} \quad [4-70]$$

4.4.1 Empirical expressions

Empirical equations are often applied to determine hydraulic properties. No single equation with constant parameters is valid for all types of soil and the parameters have to be adjusted for each particular soil. The parameters of empirical equations are determined by curve-fitting methods. The empirical expressions most frequently applied are listed in **Table 4.11**.

Table 4.11: Empirical equations for determining unsaturated hydraulic conductivity

Equation	Reference	
$K(\psi) = a \psi + b$	Richards (1931)	[4-71]
$K_r(\Theta) = \Theta^n$ where $\Theta = \frac{(\theta - \theta_r)}{(\theta_s - \theta_r)}$	Averjanov (1950)	[4-72]
$K(\psi) = \alpha \psi ^{-n}$	Wind (1955)	[4-73]
$K_r(\psi) = \exp(-\alpha\psi)$	Gardner (1958)	[4-75]
$K(\psi) = \frac{K_s}{a\psi^n + 1}$	Gardner (1958)	[4-76]
$K(\psi) = K_s$ for $\psi \leq \psi_a$	Brooks and Corey (1964)	[4-77]
$K_r(\psi) = \frac{\psi^{-n}}{\psi_a^{-n}}$ for $\psi \geq \psi_a$	Brooks and Corey (1964)	[4-78]
$K(\psi) = K_s$ for $\psi \leq \psi_a$	Rijtema (1965)	[4-79]
$K_r(\psi) = \exp[-\alpha(\psi - \psi_a)]$ for $\psi_a \leq \psi \leq \psi_1$	Rijtema (1965)	[4-80]
$K = K_1 \left(\frac{\psi}{\psi_1} \right)$ for $\psi > \psi_1$	Rijtema (1965)	[4-81]
$K = K_s \exp[\alpha(\theta - \theta_s)]$	Davidson, Stone, Nielson & Larue (1969)	[4-82]
$K = K_s \left(\frac{\theta}{\theta_s} \right)^n$	Campbell (1974)	[4-83]

Mualem (1986) observed that the value of n in Equation 4-72 could be estimated as follows:

$n = 3.0$ to 3.5 for sands

$n = 3.5$ to 5.0 for clay, silt and sand mixtures

$n = 5.0$ to 8.0 for clays

$n = 3.5$ to 4.0 for well-structured (micro-shattered) soils

It has been shown that many of the fitting parameters used in empirical equations can be derived from readily available soil data such as soil fraction and grain-size distribution data. Likewise, many authors encounter relationships between the fitting parameters used in empirical equations to derive unsaturated hydraulic conductivity and the fitting parameters used in functions to describe the soil-water characteristic curve. Rawls *et al.* (1992) maintain that the Brooks and Corey equation for unsaturated hydraulic conductivity (Equation 4-30) can be expressed as:

$$K_r(\theta) = \left(\frac{\theta - \theta_r}{\varepsilon - \theta_r} \right)^{3+2\lambda} \quad [4-84]$$

where λ is the Brooks and Corey pore size index parameter, as defined by Equation 4-30. Rawls *et al.* (1992) found that the Brooks and Corey pore size index parameter, the residual water content and saturated hydraulic conductivity can be estimated from soil fraction data as indicated in Equations 4.30, 4-42, 4-43 and 4-44 respectively. The unsaturated hydraulic conductivity at any volumetric water content can then be determined, provided that porosity and soil fraction data are available.

Rawls *et al.* (1992) found a similar relationship for the Campbell equation (Equation 4-32). According to Rawls *et al.*, the Campbell parameter, b , (Equation 4-32) is related to the Brooks and Corey pore size index parameter, λ , by $b = 1/\lambda$. Likewise, the Campbell parameter, b , is related to the Campbell parameter, n , as defined by Equation 4-83. The Campbell empirical equation for unsaturated hydraulic conductivity (Equation 4-83) can be expressed as:

$$K_r(\theta) = \left(\frac{\theta}{\varepsilon} \right)^{3+2/\lambda} \quad [4-85]$$

As with Equation 4-84, unsaturated hydraulic conductivity as a function of volumetric water content, as described by the Campbell equation, can be estimated, provided that porosity and soil fraction data are available.

The main advantages of empirical equations are that relatively simple mathematical expressions are employed to represent soil hydraulic properties, allowing for a closed-form mathematical solution and thereby simplify analysis. The main disadvantage of empirical expressions in this context is that measured unsaturated hydraulic conductivity at some soil suction values has to be known.

4.4.2 Statistical models

Many researchers have expressed unsaturated hydraulic conductivity in terms of statistical pore-water distribution functions (Purcell, 1949; Gates & Templaar-Lietz, 1950; Childs & Collis-George, 1950; Fatt & Dykstra, 1951; Burdine, 1953; Wyllie & Gardner, 1958; Marshall, 1958; Nielson, Kirkham & Perrier, 1960; Millington & Quirk, 1961; Brutsaert, 1967; Farrel & Larson, 1972; Jackson, 1972; Mualem, 1976; Mualem & Dagan, 1978; Van Genuchten, 1980; Van Genuchten & Nielson, 1985; Fredlund *et al.*, 1994). These models are based on the Hagan-Poiseuille equation and visualise the porous medium as a set of randomly distributed pores. The pores are characterised by their length scale (pore diameter) and the density distribution function, $f(d)$. The total hydraulic conductivity, in accordance with any volumetric water content value is determined by integration along the entire range of liquid-filled pores. Mualem (1992) and Leong and Rehadjo (1997) present an excellent review and theoretical background on the different statistical models.

Childs and Collis-George (1950) investigated the effect of the random distribution of pores on unsaturated hydraulic conductivity. They considered the probability of two sections of a porous medium that were randomly connected in such a way that the larger pores with radius R_1 in one section were connected to smaller pores with radius R_2 . The probability, $prob$, is calculated as follows:

$$prob(R_1, R_2) = f(R_1)f(R_2)dR_1 dR_2 \quad [4-86]$$

Computations have been simplified in terms of two assumptions:

- The resistance to flow is from the smaller pore radius, R_2 .

- There is only one connection between the pores.

By applying the Hagen-Poiseuille equation and Equation 4-86, the discharge flow contributed by the pair of pores under consideration is:

$$dq = M \text{prob}(R_1, R_2) R_2 \nabla \psi \quad [4-87]$$

where M is a constant representing fluid properties and pore geometry. By integrating equation 4-87 with regard to filled pores at any volumetric water content value and applying Darcy's law, the unsaturated hydraulic conductivity can be expressed by means of the following equation:

$$K(\theta) = M \left[\int_{R_2=r_{\min}}^{R_2=r(\theta)} \int_{R_1=R_2}^{R_1=r(\theta)} R_2^2 f(R_2) f(R_1) dR_1 dR_2 + \int_{R_2=r(\theta)}^{R_2=r_{\min}} \int_{R_1=R_2}^{R_1=r(\theta)} R_1^2 f(R_1) f(R_2) dR_2 dR_1 \right] \quad [4-88]$$

Childs and Collis-George (1950) proposed transforming the soil-water characteristic curve to a function relating volumetric water content to pore radii based on Kelvin's capillary law and then carrying out the integration. The model proposed by Childs and Collis-George (1950) was subsequently improved by Marshall (1958) and Kunze, Uehara & Graham (1968) to yield:

$$K(\theta_i) = \frac{K_s}{K_{sc}} \frac{\sigma^2 \rho_w g}{2\eta_w} \frac{\theta_s^\zeta}{n^2} \sum_{j=1}^m [(2j+1-2i)\psi_j^{-2}] \quad [4-89]$$

where K_{sc} is the calculated saturated hydraulic conductivity and, i is the interval number that increases with decreasing water content. The first interval corresponds to the saturated volumetric water content and the last interval, where i is equal to m to the lowest water content, j is a counter from i to m , n is the total number of intervals between saturated volumetric water content and zero water content and ζ is a constant that accounts for the interaction between pores of various sizes.

Nielson *et al.* (1960) maintain that the computation of unsaturated hydraulic conductivity is significantly improved if an adjusting factor is used to match the computed values with measured saturated hydraulic conductivity. The unsaturated hydraulic conductivity can then be expressed in its analytical form by means of:

$$K(\theta) = \frac{\int_0^\theta \frac{(\theta - y)}{\psi^2} dy}{\int_0^{\theta_s} \frac{(\theta - y)}{\psi^2} dy} \quad [4-90]$$

where y is a dummy variable (Maulem, 1974; Mualem, 1976).

Burdine (1953) proposes the following equation to calculate relative hydraulic conductivity:

$$K_r(\theta) = \frac{K(\theta)}{K_s} = \Theta^q \frac{\int_{\theta_r}^{\theta} \frac{d\theta}{\psi^2(\theta)}}{\int_{\theta_r}^{\theta_s} \frac{d\theta}{\psi^2(\theta)}} \quad [4-91]$$

where q is a correction factor equal to 2.

Chapter 4: Relationship between geotechnical and hydrogeological properties

Mualem (1976) investigated a conceptual model similar to that of Childs and Collis-George (1950) and derived the following equation for predicting relative hydraulic conductivity:

$$K_r(\theta) = \Theta^q \left(\frac{\int_{\theta_r}^{\theta} \frac{d\theta}{\psi(\theta)}}{\int_{\theta_r}^{\theta_s} \frac{d\theta}{\psi(\theta)}} \right)^2 \quad [4-92]$$

where q is a correction factor equal to 0.5 but depends on specific soil-fluid properties and may vary considerably for different soils.

Campbell (1974) proposes that unsaturated hydraulic conductivity can be expressed in terms of the Campbell equation (Equation 4-32). The unsaturated hydraulic conductivity can therefore be expressed as:

$$K(\theta) = \frac{\sigma^2 \theta_s^2}{2\rho_w \eta \psi_a^2 (2b+1)(2b+2)} \left(\frac{\theta}{\theta_s} \right)^{2b+2} \quad [4-93]$$

where b is the Campbell parameter as defined in Equation 4-32.

Van Genuchten (1980) applied Equation 4-33, describing the soil-water characteristic curve, to both the Burdine (1953) and Mualem (1976) equations. Since the Van Genuchten equation is not a closed-form equation, Van Genuchten applied a restriction to the equation, thereby causing the soil-water characteristic curve equation to lose some flexibility, but enabling the calculation of unsaturated hydraulic conductivity to take place. The modified Equation 4-33 can be expressed as:

$$\Theta = \left[\frac{1}{1 + (\alpha\psi)^n} \right]^m \quad \text{where } m = 1 - 1/n \quad \text{or} \quad m = 1 - 2/n \quad [4-94]$$

where restriction is represented by $m = 1 - 1/n$, unsaturated hydraulic conductivity as a function of soil suction can be expressed in terms of the Mualem (1976) model to yield:

$$K(\psi) = \frac{K_s \left\{ 1 - (\alpha\psi)^{mn} \left[1 + (\alpha\psi)^n \right]^{-m} \right\}^2}{\left[1 + (\alpha\psi)^n \right]^{m\ell}} \quad \text{and} \quad m = 1 - 1/n \quad [4-95]$$

where ℓ corresponds to the correction factor, q , introduced in the Mualem (1976) model with a value of 0.5.

The function can also be expressed in terms of the Burdine (1953) model with the restriction, $m = 1 - 2/n$, to yield:

$$K(\psi) = \frac{1 - (\alpha\psi)^{n-2} \left[1 + (\alpha\psi)^n \right]^{-m}}{\left[1 + (\alpha\psi)^n \right]^{m\ell}} \quad \text{where } m = 1 - 2/n \quad [4-96]$$

where ℓ corresponds to the correction factor, q , introduced in the Burdine (1953) model with a value of 2.

Van Genuchten (1980) suggests that the Brooks and Corey pore-size parameter, λ , is related to the van Genuchten parameters m and n through $\lambda = mn$. Rawls *et al.* (1992) maintain that the Van Genuchten

parameter, α , is related to the air-entry value, ψ_a through $\alpha = 1/\psi_a$. Equations 4-95 and 4-96 can therefore be expressed as a function of air-entry value, the Brooks and Corey pore-size index and saturated hydraulic conductivity. Rawls *et al.* (1992) indicate that the air-entry value, Brooks and Corey pore-size index parameter and saturated hydraulic conductivity can be estimated from soil fraction data by means of Equations 4-30, 4-42, 4-43 and 4-44 respectively. The unsaturated hydraulic conductivity as a function of volumetric water content can then be estimated, provided that porosity and soil fraction data are available.

Fredlund *et al.* (1994) suggest that a statistical function should be based on the Childs and Collis-George (1950) and Nielson *et al.* (1960) model (Equation 4-90), because of the fact that both the Burdine (1953) and Mualem (1976) models contain a correction factor dependent on the properties of the soil. Fredlund *et al.* (1994) applied the Fredlund and Xing (1994) function (Equation 4-39 and 4-40) to the Nielson *et al.* (1960) model. The Fredlund and Xing function that describing the soil-water characteristic curve, can be expressed as:

$$\theta = C(\psi) \frac{\theta_s}{\left\{ \ln \left[e + \left(\frac{\psi}{a} \right)^n \right] \right\}^m} \quad [4-97]$$

where $C(\psi)$ is a correction factor that can be expressed as:

$$C(\psi) = \frac{\ln \left(1 + \frac{\psi}{C_r} \right)}{\ln \left(1 + \frac{1\,000\,000}{C_r} \right)} \quad [4-98]$$

The Fredlund *et al.* (1994) model that describes the unsaturated hydraulic conductivity as a function by applying the Nielson *et al.* (1960) model (Equation 4-90), can be expressed as:

$$K_r(\psi) = \Theta^q(\psi) \frac{\int_{\ln(\psi)}^b \frac{\theta(e^y) - \theta(\psi)}{e^y} \theta'(e^y) dy}{\int_{\ln(\psi_a)}^b \frac{\theta(e^y) - \theta_s}{e^y} \theta'(e^y) dy} \quad [4-99]$$

where $b = \ln(1\,000\,000)$, e is the natural number 2.7182..., y is a dummy variable of integration representing the logarithm of soil suction and Θ^q is a correction factor, Θ is the normalised volumetric water content or relative degree of saturation that has been defined as:

$$\Theta = \frac{\theta - \theta_r}{\theta_s - \theta_r} \quad [4-100]$$

4.4.3 The effect of hysteresis

It has already been shown that the soil-water characteristic curve is hysteretic by nature, i.e. the shape of the curve differs with regard to wetting and drying cycles. Since the unsaturated hydraulic conductivity is derived from the soil-water characteristic curve, it is clear that the unsaturated hydraulic conductivity will

also exhibit hysteretic behaviour (**Figure 4.2**). Investigations by Nielson and Biggar (1961), Topp (1969) and others showed that the hysteretic behaviour of unsaturated hydraulic conductivity as a function of soil suction, $K(\psi)$, is much more pronounced than in the case of unsaturated hydraulic conductivity as a function of volumetric water content, $K(\theta)$. Topp and Miller (1966) observed, that for the soils tested, the maximum difference in values for the soil-water characteristic curve due to hysteresis was 300 per cent while in the case of $K(\psi)$ hysteretic loops, the difference was 20 000 per cent. In contrast, the difference in values for the $K(\theta)$ hysteretic loops was only 40 per cent. Various other authors have likewise observed this phenomenon (Nielson & Biggar, 1961; Topp, 1969; Talsma, 1970 and others). The hysteretic behaviour of unsaturated hydraulic conductivity is inconsistent with the statistical models that assume each soil to have single pore-size distribution function. From a practical point of view, it is acceptable to use only one branch of the hysteretic loops to compute unsaturated hydraulic conductivity. It is recommended that unsaturated hydraulic conductivity be expressed as a function of volumetric water content to minimise the effect of hysteresis.

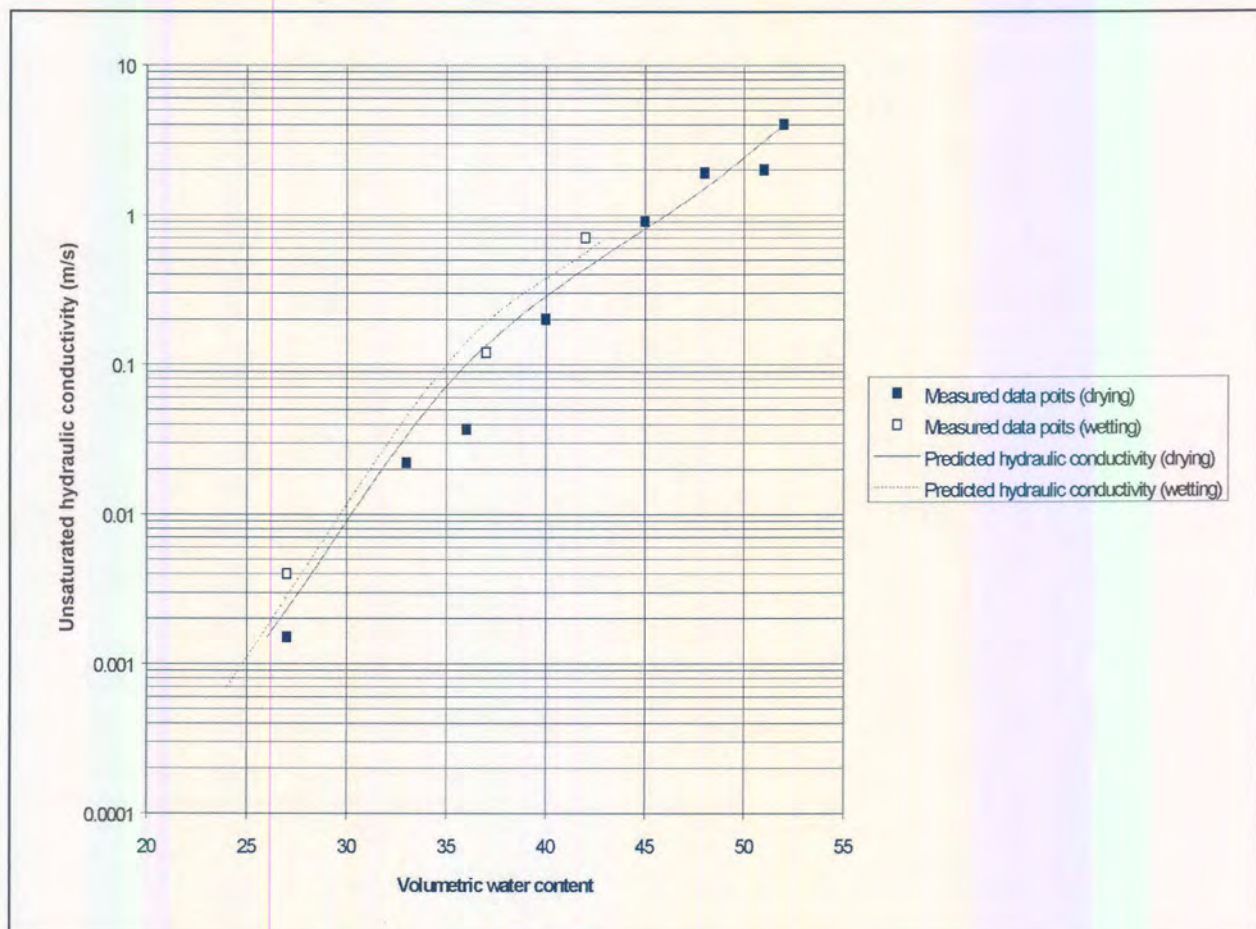


Figure 4.2: Experimentally determined and estimated unsaturated hydraulic conductivity as a function of volumetric water content (after Top and Miller, 1966)